

**SEDIMENTARY EVOLUTION AND SOURCE ROCK POTENTIAL OF LOWER JURASSIC
SUCCESSIONS IN THE MANDAWA BASIN, SE TANZANIA**

Dissertation

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Declaration

I certify that the conceptualization and execution of this research titled sedimentary evolution and source rock potential of Lower Jurassic successions in the Mandawa Basin, SE Tanzania was carried out by me under the supervision of Prof. L. Schwark.

In addition, I affirm that no part of this thesis has been submitted in fulfillment of the award of Doctoral degree at any university, neither was it published nor submitted for publication. Furthermore, I wish to state that I adhered to the Rules of Good Scientific Practice of the German Research Foundation while embarking on this research and preparing this thesis. I so state that none of my previous academic degrees has been withdrawn at any time.

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Abstract

Lower Jurassic successions that form syn-rift sedimentary packages in the Mandawa Basin preserve a complete stratigraphy that records tectonic processes related to the break-up of Gondwana as well as evidence of an extension of the Neothethys Sea (a.k.a. the Somali Ocean) into the East African Margin prior to the opening of the Indian ocean, and thus is considered to be important in Tanzania's geological history. Previous exploration activities have indicated the presence of a working petroleum system in the subsurface, with key elements such as source rocks, reservoirs, and seals present in both syn-rift and post-rift sedimentary packages. Syn-rift sedimentary packages have received little attention compared with post-rift sequences that are outcropping and have good biostratigraphic control. An understanding of the sedimentary facies evolution, depositional conditions of syn-rift sedimentary packages is important in assessing the petroleum potential of Lower Jurassic successions in the Mandawa Basin.

In this thesis, an integrated geochemical approach, which includes bulk organic matter analysis (CNS, TOC), stable isotope analysis of organic matter ($\delta^{13}\text{C}_{\text{org}}$), and programmed pyrolysis analysis, was employed to determine the variation in organic matter types and quality with respect to lithofacies evolution in the Mandawa Basin during Early Jurassic times. In addition to geochemical data, Spectral Gamma Ray logs (Th, U, K) and sediment lithotype data (carbonates, salts, and clastics) were integrated in the study to improve an interpretation and for reconstructing the basin-fill of the Lower Jurassic successions in the Mandawa Basin.

Variations in depositional settings for the eastern, western and northern, southern central parts of the basin during the Early Jurassic have been revealed by this study. These variations may be attributed to changes in subsidence, eustatic rise in sea level and local climate. Southern central parts of the basin, which subsidized at relatively higher rate, exhibited more reduced conditions as opposed to shallow, more oxic depositional settings prevailing in the eastern and western flanks. These conditions switched to less oxic depositional settings during Toarcian times, when restricted marine conditions were established. In addition to these findings, the relationship between sediment input (carbonates, clastics and organic matter) during the Early Jurassic was

also illustrated, this providing insights into factors regulating the distribution of organic matter within the Mandawa Basin.

The data obtained from the investigation of outcrops and well data in Lower Jurassic sediments show variability in TOC and source rock potential. TOC, kerogen yield (S₂) and hydrogen index (HI) are generally low, and organic matter is composed mainly of type III (gas prone) to IV (inert) kerogens with lesser admixture of type II-III kerogen (oil-gas prone), which have been deeply buried (>2000m) and undergone a wider range of thermal diagenesis, from early to over mature stage. This suggest that source intervals found within the sediments of Lower Jurassic successions have poor to fair source rock potential, and only few intervals have been locally enriched due to periodic marine ingression that persisted since the Pliensbachian period.

Furthermore, this study revealed that the source rock potential (S₂, HI) varies spatially in the basin, highest source potential has been observed in the southwestern and northwestern parts, suggesting that these areas were more favorable for the deposition and preservation of organic matter. Low source rock potential observed in the eastern and western flanks, would indicate degraded organic matter and significant terrestrial influx from the emerging lands in these parts. Changes in depositional conditions in an evolving rift basin, driven by changes in relative subsidence, eustatic sea level change and sediment influx are the proposed mechanisms for the organic matter distribution and source rock enrichment during Early Jurassic times in the Mandawa Basin. Organic matter was poorly preserved when high sediment flux exceeds subsidence rate, generally in the eastern and western but also southern part. Whenever there was a high subsidence rate coupled with eustatic rise in sea level and lower sediment supply, the productivity and preservation of organic matter in deep lake to restricted marginal marine conditions were enhanced.

Zusammenfassung (German Abstract)

Unterjurassische Sukzessionen, welche die syn-Rift Ablagerungen im Mandawa Becken formten, überliefern eine vollständige stratigraphische Abfolge, welche die tektonischen Prozesse aufzeichnet, die zum Zerfall von Gondwana und zur Ausdehnung des Neotethys Ozean (auch bekannt als Somali Ozean) in den Ostafrikanischen Kontinentalrand vor der Öffnung des Indischen Ozeans führten und die demzufolge als essentiell für die Geologische Geschichte Tansanias gelten müssen.

Bisherige Explorationstätigkeiten deuten die Präsenz eines intakten „Petroleum Systems“ an, welches die Schlüsselemente Muttergestein, Migrationspfade, Reservoirgestein, und Deckgestein in den syn-Rift und post-Rift Sedimentpakete beinhaltet. Syn-Rift Abfolgen haben bisher wenig Aufmerksamkeit erhalten gegenüber Post-Rift Abfolgen, die über Aufschlüsse zugänglich sind und für die eine gute biostratigraphische Kontrolle gegeben ist. Ein verbessertes Verständnis der sedimentfaziellen Entwicklung sowie der Ablagerungsbedingungen der Syn-Rift-Sedimente ist von entscheidender Bedeutung für eine Bewertung des Petroleumpotentials der Unterjurassischen Abfolgen im Mandawa Becken.

In dieser Dissertation wurde ein integrierter Ansatz verfolgt, der die Pauschalanalytik des Organischen Materials, (CNS, TOC), dessen stabile Kohlenstoffisotopie ($\delta^{13}\text{C}_{\text{org}}$), sowie die Programmierte Pyrolyse beinhaltet, um die Variation in der Komposition und Qualität des Organischen Materials und der Entwicklung der Lithofaziestypen im Mandawa Becken während des Unterjura abzubilden. In Ausweitung der geochemischen Daten wurde Bohrlochgeophysik unter Einsatz Spektraler Gammastrahlungsmessung (Th, K, U) und Sedimentlithotypanalyse (Karbonate, Evaporite, Klastika) integriert, um die Interpretation und Rekonstruktion der Beckenfüllung im Mandawa Becken zu verbessern.

Variation im Ablagerungsmilieu zwischen den östlichen, westlichen, nördlichen und südzentralen Kompartimenten des Mandawa Beckens im frühen Jura wurden in dieser Studie aufgedeckt. Diese Veränderungen werden assoziiert mit Variation in der Beckensubsidenz, eustatischen Meeresspiegel-schwankungen, und lokaler Klimaentwicklung. Im südzentralen Teil des Beckens,

wo hohe Subsidenzraten herrschten, entwickelten sich stärker reduzierende Milieubedingungen, während in den flacheren östlichen und westlichen Kompartimenten stärker oxidierende Bedingungen dominierten. Während des Toarcium veränderte sich das Ablagerungsmilieu zu stärker reduzierenden Bedingungen. Das Verhältnis zwischen Sedimenteintrag (Karbonate, Evaporite, Klastika und organisches Material) im frühen Jura wurde berücksichtigt, um die Faktoren zur Regulierung der Anreicherung organischen Materials im Mandawa Becken zu deuten.

Die Bohrungs- und Aufschlussdaten dokumentieren die hohe Variabilität im TOC und Muttergesteins-potential der Unterjurassischen Sedimente. TOC, Kerogenausbeute (S₂) und Wasserstoffindex (HI) sind generell niedrig und das Organische Material ist hauptsächlich vom Typ III (gashöfzig) bis IV (inert) mit geringeren Beimengungen vom Typ II (öl- und gashöfzig). Kerogene wurden tief versenkt (>2000m) und erfuhren eine weite Spanne an thermischer Diagenese, vom frühreifen bis zu überreifen Stadium. Dies bedingt, dass potentielle Muttergesteinsintervalle in den Jurassischen Abfolgen bestenfalls ein moderates Quellenpotential besitzen und nur lokale Anreicherungen an Muttergesteinspotential vorliegen, die sich im Zuge mariner Ingressionen seit dem Pliensbachium aufgebaut haben.

Das Muttergesteinspotential (S₂, HI) zeigt hohe räumliche Variation im Mandawa Becken, wobei die höchsten Potentiale im südwestlichen und nordwestlichen Kompartiment verortet sind, was auf dortige Ablagerungsbedingungen für präferenzielle Erhaltung des Organischen Materials deutet. Niedrige Kohlenwasserstoffpotentiale in den östlichen und westlichen Beckenflanken gehen auf oxisch stärker degradiertes Organisches Material und einen höheren terrigenen Sedimenteintrag zurück. Veränderungen in den Ablagerungsbedingungen eines initialen Riftbeckens, angetrieben durch relative Unterschiede in der Subsidenz, eustatische Meeresspiegelschwankungen und Sedimentakkumulation kontrollierten die Verteilung und Qualität des organischen Materials im Mandawa Becken während des frühen Jura. Organisches Material wurde schlecht erhalten, wenn der Sedimenteintrag die Subsidenzrate übertraf, was vor allem in den östlichen und westlichen Kompartimenten auftrat. Wann immer eine hohe Subsidenzrate mit einem eustatischen Meeresspiegelanstieg und niedrigem Sedimenteintrag

koinzidierte, erhöhte sich die aquatische Produktivität sowie die Erhaltungsrate für abgelagertes organisches Material in tieferen Seen oder in abgeschnürten marinen Randbecken.

Table of Contents

Acknowledgments	iii
Declaration	vi
Abstract	viii
Zusammenfassung (German Abstract).....	x
List of Figures.....	xx
List of Tables.....	xxvii
Abbreviations	xxix
Introduction	1

Chapter One

A regional synthesis of Pliensbachian - Toarcian Geological evolution of East African Margin

Abstract	6
1. Introduction.....	7
1.1. Regional Geological overview	10
1.1.1 Permo - Carboniferous to Triassic (Karoo) geological evolution.....	11
1.1.2 Early Jurassic geological evolution in the East African Coastal Margin.....	13
1.2 Pliensbachian and Toarcian lithostratigraphy of the East African Coastal Margin.....	15
1.2.1 Luuq Manderia Basin (SW Somalia)	16
1.2.1.1 Waney (Uaneii) Formation (Lower Toarcian)	17
1.2.1.2 Baidoa Formation (Aalenian-Callovian)	17
1.2.2 Ahl-Mado Basin (NE Somalia).....	18
1.2.2.1 Adigrat Formation (? Triassic-Pre-Toarcian)	19
1.2.2.2 Dhagabo Formation (Lower Toarcian).....	19
1.2.2.3 Qaranye (Hamanlei) Formation (Aalenian-Callovian).....	20

1.2.3 Lamu Basin (SE-Kenya)	20
1.2.3.1 Mazeras Formation (Upper Triassic to Lower Jurassic)	21
1.2.4 Mandera Basin (NE Kenya)	22
1.2.4.1 Mansa Guda Formation (Triassic)	23
1.2.4.2 Didimtu Formation (Toarcian)	23
1.2.5 Majunga Basin (NW Madagascar)	24
1.2.5.1 Isalo Formation (? Triassic)	25
1.2.5.2 Beronono Formation (Toarcian-Aalenian to ? Lower Bajocian)	25
1.2.6 Morondava Basin (SW-Madagascar)	26
1.2.6.1 Andafia Formation (Toarcian-Aalenian)	27
1.2.7 Mandawa Basin (SE-Tanzania).....	27
1.2.7.1 The Mbuo Formation (Pliensbachian)	29
1.2.7.2 Nondwa Formation (Lower Toarcian to Middle Toarcian)	30
1.2.7.3 Mihambia Formation (Upper Toarcian-Aalenian)	31
1.3. Discussion	31
1.3.1 Implications on lithofacies, stratigraphy and tectonics	39
1.3.2 Implications for Petroleum Systems.....	41
1.4. Conclusions.....	41
References.....	42

Chapter Two

Lithofacies of early Jurassic successions derived from spectral gamma ray logging in the Mandawa Basin, SE Tanzania

Abstract	56
2.0 Introduction.....	57

2.1 Geological setting and stratigraphy of the study area.....	58
2.2 Materials and methods	62
2.2.1. Sampling of drill cuttings	62
2.2.2 Total carbon (TC) and total sulfur (TS) analysis	63
2.2.3 Total organic carbon (TOC)	63
2.2.4 Sediment composition (carbonate, anhydrite, clastics and crystalline rock)	63
2.2.5 Spectral Gamma Ray Log data acquisition	64
2.2.6 Borehole quality check	64
2.2.7 Establishing U,Th, and K background concentration.....	64
2.3. Results	65
2.3.1 U,Th, and K concentrations according to lithofacies.....	65
2.3.2 Stratigraphical variability of Th, U, and K abundances.....	66
2.3.3 Sediment composition and stratigraphic variability of U/K, Th/U, and Th/K ratios	68
2.4 Discussion	71
2.4.1 Thorium-Uranium as an indicator of paleo-redox conditions.....	71
2.4.2 Thorium-potassium as an indicator detrital mineral composition and sediment lithofacies.....	75
2.4.3 Th/K as Provenance indicator	79
2.5 Conclusions.....	80
References.....	81

Chapter Three

Total Organic Carbon variability of Lower Jurassic successions in the Mandawa Basin, SE Tanzania

Abstract	98
----------------	----

3.0 Introduction.....	99
3.1 Geological and stratigraphy setting	101
3.2 Materials and methods	104
3.2.1 Wireline logs data acquisition and processing	104
3.2.2 Passey's ΔLOGR techniques.....	104
3.2.3 Total carbon.....	105
3.2.4 Total organic carbon (TOC) _{raw} , TOC _{cbt-free} , and TOC _{cbt+salt-free}	105
3.2.5 Sediment composition (carbonates, anhydrites, clastics).....	105
3.2.6 Total organic carbon (TOC) derived from the empirical relationship between measured TOC and well log data.....	106
3.4. Results	107
3.4.1 Geophysical well-log responses to various lithologies.....	107
3.4.2 Sediment composition and total organic carbon (TOC %)	108
3.4.3 GR, RHOB, DT, and ILD logs responses in organic-rich intervals	110
3.4.4 ILD-DT separation (ΔLOGR) as an indicator of organic richness in shale intervals	111
3.5. Discussion	112
3.5.1 Total organic carbon (TOC) variability.....	112
3.5.1.1 Variability in TOC derived from log responses and cuttings data	112
3.5.1.2 Variability in ΔLogR derived TOC	115
3.5.2 Influence of depositional conditions on organic matter enrichment and variability	117
3.5.2.1 Th/U and U concentration as an indicator of depositional conditions	117
3.5.2.2 Interplay between depositional environment (facies), sediment composition, and organic matter variability.....	119
3.5.2.3 Relationship between changes in sediment input and organic matter	120

3.6 Conclusions.....	122
References.....	123

Chapter Four

Assessment of kerogen types and source rock potential of Lower Jurassic successions in the Mandawa Basin, SE Tanzania

Abstract	132
4.0 Introduction.....	133
4.1. Geological and stratigraphy setting	135
4.2 Materials and methods	138
4.2.1 TOC and Programmed pyrolysis analysis (Hawk®)	138
4.2.2 Carbon isotopes analysis of organic matter	139
4.3 Results	139
4.3.1 Downhole variation in organic matter quality and quantity of three wells based on cutting samples.....	139
4.3.2 Downhole variation in organic matter quality and quantity of core samples	144
4.3.3 Variation in organic matter quality and quantity of a Nondwa outcrop section	146
4.4. Discussion	147
4.4.1 Quality assurance of pyrolysis data	147
4.2.2 Organic matter types and quality for four different locations.....	148
4.4.2.1 Organic matter quality from cuttings, core and outcrop samples for four different location (3 borehole wells)	148
4.4.3. Comparison of organic matter maturity of four locations in the Mandawa Basin ...	151
4.4.3.1 Organic matter types versus Maturity.....	151
4.4.3.2 Quality of organic matter with respect to maturity	154

4.4.4 Comparison of organic matter quantity and quality for all wells drilled in the Mandawa Basin	156
4.4.5 Palaeogeographical control of source rock source rock potential.....	159
4.4.5.1 Palaeogeographical control of source rock source rock potential during Pliensbachian (Mbuo Formation)	160
4.4.5.2 Palaeogeographical controls of source rock source rock potential during the Toarcian (Mbuo Formation).....	162
4.5 Conclusions.....	164
References.....	165
Chapter Five	
5.0 Conclusions and outlook	173
5.1 Conclusions	173
5.2 Outlook	176
Appendices.....	178
Chapter One	178
Chapter Two	182
Chapter Three.....	192
Chapter Four.....	195

List of Figures

Chapter One

Fig. 1.1: Comparison of major global events occurring during Early Jurassic including the Toarcian. Adapted and redrawn from (Bergelin, 2009; Dera et al., 2011), geological time scale was drawn using Timescale Creator 2019 (https://timescalecreator.org/index/index.php) and global sea level curve is adapted from (Haq, 2018).....	8
Fig. 1.2: Global positions of the Early Jurassic continents and known localities of Lower Jurassic sediments, locations of East African coastal basins are shown by the red box (modified from Silva et al., 2021).	9
Fig. 1.3: Location map of East African Coastal Basins, after (Downes and Boote, 2009).....	11
Fig. 1.4: Schematic map showing the distribution of Karoo rifting-associated continental sediment graben and basins of South East and Central Africa, after Frizon de Lamotte et al. (2015).	13
Fig. 1.5: (A) Conglomeritic Adigrat sandstone with detrital quartz exposed in Berbera Somalia (adapted from Ali, 2009), B: Mazeras sandstone intercalated with shale outcropping near the Town of Mazera in Lamu Basin (adapted from EAPCE, 2019), C: Highly deformed Nondwa evaporitic sequences in the Mandawa Basin, D: Toarcian shale with minor sandstone intercalation in the Majunga Basin (adapted from Banks, 2008).....	14
Fig. 1.6: Pliensbachian to Aalenian stratigraphy of basins along the East African Margin, depicting important formations including Toarcian strata. *The bases of these intervals have been interpreted as late Triassic by Hudson (2011; Hudson and Nicholas 2014).....	16
Fig. 1.7: Geological map of Somalia, depicting pre-Jurassic, Jurassic including Toarcian and Post Jurassic strata. Redrawn and modified from Mbede and Dualeh (1997) and Ali and Watts (2016).	18
Fig. 1.8: Geological map of Kenya showing pre-Jurassic, Jurassic including Toarcian and post Jurassic strata. Redrawn from Mbede and Dualeh (1997) and Akech et al. (2013).....	22
Fig. 1.9: Geological map of Madagascar depicting Karoo, Jurassic including Toarcian and post Jurassic sediments. Redrawn and adapted from Schlüter and Trauth (2006).	26

Fig. 1.10 : Geological map of Tanzania, depicting Pre-Jurassic, Jurassic including Toarcian and post Jurassic sediment. Adapted from (Pinna et al., 2008; Hudson, 2010; Barth et al., 2016).....	29
Fig. 1.11: Schematic late Triassic-Pliensbachian paleogeography of the East African Margin. ...	32
Fig. 1.12: Schematic early Toarcian (~183ma) paleogeography of East African Margin.....	34
Fig. 1.13: Toarcian (~ 183.1 Ma) global palaeogeographical map adapted from Frizon de Lamotte et al. (2015) and Scotese (2014). Red box demarcates the location of the East African Margin discussed here.....	37
Fig. 1.14: Schematic (Late Toarcian-Early Aalenian) paleogeography of the East African Margin.	39
Fig. 1.15: Chronostratigraphy of East African and adjacent Arabian basins (1. Tanzanian coastal basins, 2. Mandera and Lamu, 3. Morondava and Majunga, 4. Mandera-Luuq, 5. Ogaden Basins and adjacent areas) depicting pre-Jurassic clastics, continental Jurassic graben fills, and marine Jurassic deposition in the course of the Toarcian transgression. Redrawn from Bosellini (1992).	40

Chapter Two

Fig. 2.1: Geological map of Tanzania, showing the location of the Mandawa Basin and position of the Mita-Gamma-1 and East-Lika-1 wells investigated (adapted from Pinna et al. 2008; Hudson 2010; Barth et al. 2016).	59
Fig. 2.2: Stratigraphy of the Pindirolu Group (?Late Triassic-Mid Jurassic), depicting two controversial age assignments according to references 1. Balducci et al., 1992; 2. Alconsult and TPDC, 1997; 3. Hudson, 2010; 4. Hudson and Nicholas, 2014; 5. Fossum, 2019 versus 6. Veeken and Titov, 1996; 7. Stratadata, 2006; 8. Quinton and Copestake, 2007; 9. Corelab and TPDC, 2009; 10. Robertson and CGG 2017; 11. Macgregor, 2018.	60
Fig. 2.3: (A and B): Outcropping evaporite-shale intercalations in the Nondwa Formation. (C, D): Drill core slabs showing very fine to fine grained arkosic Mbuo sandstone with abundant flasers of organic material and lithic fragments.	61
Fig. 2.4: A summary of the workflow carried out in this study.	62

Fig. 2.5: Stratigraphical variability of the Th, U, and K SGRL abundances averaged for cuttings sample equivalents and their relationship to the total natural gamma ray intensity (GR API) in the sediments from the Mita-Gamma-1 well.	67
Fig. 2.6: Stratigraphical variability of the Th, U, and K SGRL abundances averaged for cuttings sample equivalents and their relationship to the total natural gamma ray intensity (GR API) in the sediments from the East-Lika-1 well.	68
Fig. 2.7: Stratigraphic trends in SGRL-derived U/K , Th/U, and Th/K ratios averaged for cuttings sample equivalents and their relationship to the sediment composition in the Mita-Gamma-1 well. The lithological column on the right was established based on geochemical analysis of cuttings.....	70
Fig. 2.8: Stratigraphic trends in SGRL-derived U/K, Th/U, and Th/K ratios averaged for cuttings sample equivalents and their relationship to the sediment composition in the East-Lika well. The lithological column on the right was established based on cuttings interpretation and geochemical analysis.	71
Fig. 2.9: Binary Th/U discrimination diagrams depicting lithofacies fields bordered by Th/U-ratios of 1, 3, 4, 6, and 27 based on global and regional data compilation: A (global data set), B (Mandawa Basin, this study), C (Mandawa Basin, Hudson 2010) and D (Proterozoic Buanji Basin, SW-Tanzania, Kasanzu et al., 2017).	73
Fig. 2.10: Binary Th/K discrimination diagrams depicting lithofacies fields and associated minerals bordered by Th/K-ratios of 0.5, 1, 2, 6, and 11 based on global and regional data compilation: A (global data set), B (Mandawa Basin, Mita-Gamma-1 well, this study), C (Mandawa Basin, Hudson 2010) and D (Proterozoic Buanji Basin, SW-Tanzania, Kasanzu et al., 2017).	77
Fig. 2.11: Th/U and Th/K discrimination plots for Jurassic sedimentary rocks and basement strata from the East-Lika-1 well based on SGRL analysis of this study according to sediment provenance and clay mineral types.	79

Chapter Three

Fig. 3.1: Geological map of Tanzania, showing the location of the Mandawa Basin and position of the Mita-Gamma-1 well investigated, as well as the Mbuo-1 and Mandawa-7 wells (adapted from Pinna et al., 2008; Hudson, 2010; Barth et al., 2016). The Mbuo Formation has not been encountered outcropping at the surface.....	102
Fig. 3.2: Comparison of high- and low-resolution GR, DT, RHOB, ILD and PEF logs responses to various lithologies penetrated by the Mita Gamma-1 well (high- and low-resolution logs are shown by red curve and black dotted traces, respectively).	108
Fig. 3.3: Variability in TOC_{raw} , TOC_{cbtf} , $TOC_{cbtf+sf}$ and their relationship to the sediment composition of the cuttings samples.	110
Fig. 3.4: Cross-plot depicting relationship between TOC_{raw} measured from cuttings samples and well log responses GR, DT, RHOB, and ILD averaged for intervals from which cuttings were collected. Regression lines shown are for all samples (green) and and excluding the five TOC-richest samples (red).....	111
Fig. 3.5: Comparison of ILD-DT-overlay and GR to the TOC content measured on cuttings samples. ILD and DT log trace separation indicates presence of organic enriched intervals in shales. ...	112
Fig. 3.6: Comparison of TOC content predicted from GR, DT, ILD, RHOB log response versus raw and corrected TOC concentrations determined on cuttings (cbtf = carbonate-free, sf = salt-free).	114
Fig. 3.7: Variability in the predicted and measured TOC content and its relationship to $\Delta LOGR$ (obtained using the Passey et al. 1990 technique).....	116
Fig. 3.8: Bubble plots depicting relationship between PEF and natural Gamma-Ray responses (GR) resolved for: A (TOC), B (U), C (Th), and D (K).	118
Fig. 3.9: Cross plots showing correlation between TOC versus A (U) and B (Th/U).	119
Fig. 3.10: Relationship between changes in the depositional environment to the TOC content and sediment composition.	120
Fig. 3.11: Cross plots depicting relationship between TOC content versus A (carbonates wt%) and B (clastics wt%).	121

Chapter Four

Fig. 4.1: Geological map of Tanzania showing the location of the Mandawa Basin and position of the East Lika-1, Mbuo-1 and Mita-Gamma-1 wells investigated as well as Mandawa-7, Mihambia-1, Kizimbani-1 and Mbate wells-1 (adapted from (Pinna et al., 2008; Hudson, 2010; Barth et al., 2016).	138
Fig. 4.2: Downhole variation in Total organic carbon (TOC), Kerogen yield (S ₂), Hydrogen Index (HI), Tmax and Production Index (PI) of cuttings samples in the sediment from the Mita Gamma-1 well.	142
Fig. 4.3: Downhole variation in Total organic carbon (TOC), Kerogen yield (S ₂), Hydrogen index (HI), Tmax and Production Index (PI) of cuttings samples in sediment from the East Lika-1 well.	143
Fig. 4.4: Downhole variation in Total organic carbon (TOC), kerogen yield (S ₂), Hydrogen index (HI), Tmax and Production Index (PI) of cuttings samples in sediment from the Mbuo-1 well.	144
Fig. 4.5: Variation in TOC, kerogen S ₂ yield, HI, Tmax, and PI in sediments from: A (Mbuo core), B (outcropping Nondwa evaporitic sequences).....	146
Fig. 4.6: Comparison of organic matter quality in the sediments of Lower Jurassic successions from four different location in the Mandawa Basin: A (Mita Gamma-1), B (East Lika-1), C (Mbuo-1), D (core and outcrops samples).	150
Figure 4.7: Comparison of organic matter quality in sediment Lower Jurassic successions from four different locations in the Mandawa Basin: A (Mita Gamma-1), B (East Lika-1), C (Mbuo-1), D (core and outcrops samples).	151
Fig. 4.8: Comparison of organic matter quality vs. maturity in the sediments of Lower Jurassic successions from four different locations in the Mandawa Basin: A (Mita Gamma-1), B (East Lika-1), C (Mbuo-1), D (Core and outcrops samples).	153
Fig. 4.9: Comparison of organic matter quality of the sediments of Lower Jurassic successions with respect to maturity ($\delta^{13}\text{C}_{\text{org}}$ vs. HI) from four different locations in the Mandawa Basin: A (Mita Gamma-1), B (East Lika-1), C (Mbuo-1), D (core and outcrops samples).	156

Fig. 4.10: Comparison of organic matter quality of Lower Jurassic intervals from all wells in the Mandawa Basin using: A (Modified van Krevelen diagram) and B (TOC versus total hydrocarbon yield). Data from Kizimbani-1, Mihambia-1, Mandawa-7, Mbate-1 wells have been compiled from literature (Alconsult and TPDC, 1997; Corelab and TPDC, 2009; McCabe, 2021).	158
Fig. 4.11: Comparison of organic matter quality of all wells from the Mandawa Basin using: A (Tmax vs. HI) and B (Tmax vs. Present day depth). Data from Kizimbani-1, Mihambia-1, Mandawa-7, Mbate-1 wells have been compiled from literature (Alconsult and TPDC, 1997; Corelab and TPDC, 2009; McCabe, 2021).	159
Fig. 4.12: Comparison of source rock quantity (TOC) and quality (S2, HI) of sediments from 7 wells in Mandawa Basin with respect to paleogeography during the Pliensbachian (Mbuo Formation). Paleogeographical map is adapted from (Hudson and Nicholas 2014; Mccabe 2021; Gama and Schwark, 2023a). Data from Kizimbani-1, Mandawa-7, Mbate-1 wells have been compiled from the literature (Alconsult and TPDC, 1997; Corelab and TPDC, 2009; McCabe, 2021).....	162
Fig. 4.13: Comparison of source quantity (TOC) and quality (S2, HI) of sediments from six wells in the Mandawa Basin in relation to paleogeography during early the Toarcian (Nondwa Formation). Paleogeographical map is adapted from (Hudson and Nicholas 2014; Mccabe 2021, Gama and Schwark, 2023). Data from Kizimbani-1, Mandawa-7, Mihambia-1 wells have been compiled from the literature (Alconsult and TPDC, 1997; Corelab and TPDC, 2009; McCabe, 2021).....	164

List of Tables

Chapter One

Table 1.1: Compilation of early Toarcian Biostratigraphy data of the five East African coastal sedimentary basins. * <i>Spiriferina rostrata</i> var. <i>Madagascariensis</i> reclassified as <i>Calyptoria madagascariensis</i> after Baeza-Carratalá et al., 2018.....	16
--	----

Chapter Two

Table 2.1: Compilation of concentration ranges and mean values for the radiogenic elements K, Th, and U in sedimentary rocks and derivation of Th/U and Th/K ratios applicable to sedimentary rock typing. Data compiled from: 1. Moore, 1957; 2. Adams and Weaver, 1958; 3. Moore 1960; 4. Turekian and Wedepohl, 1961; 5. Pliler and Adams, 1962; 6. Horn and Adams, 1966; 7. Ronov and Migdisov, 1970; 8. Wedepohl, 1970; 9. Moore 1971; 10. Cosgrove, 1972; 11. Beeson, 1980; 12. McLennan and Taylor, 1980a; 13. Dypvik and Eriksen, 1983; 14. Dypvik and Bue, 1984; 15. Myers and Wignall, 1987; 16. Myers and Bristow, 1989; 17. Van Buchem et al., 1992; 18. Tribouvillard et al., 1994; 19. Hayashi et al., 1997; 20. McRoberts et al., 1997; 21. Luczaj, 1998; 22. Asiedu et al., 2000; 23. Svendsen and Hartley, 2001; 24. Martinius et al., 2002; 25. Lüning et al., 2004; 26. Fello et al., 2006; 27. Yamaguchi, 2006; 28. Kasanzu et al., 2008; 29. Mernagh and Miezeitis, 2008; 30. Correia et al., 2013; 31. Paredes et al., 2015; 32. Kasanzu et al., 2017; 33. Amedjoe et al., 2018; 34. Reolid et al., 2020; 35. Güngör Yeşilova and Yeşilova, 2021; 36. Marín et al., 2021; 37. Phujareanchaiwon et al., 2021; 38. Wei et al., 2021.....	65
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Abbreviations

$\Delta\log R$: Delta logR

Δt : Sonic Travel Time

μs : Microsecond

API: American Petroleum Institute

C: Carbon

$CaCO_3$: Calcium Carbonate

cbt: Carbonate

cbtn: Carbonate free

cc: Cubic centimeters

CH_4 : Methane

CIEs: Negative Carbon Isotope excursions

CO_2 : Carbon dioxide

DT: Sonic travel time

E-W: East-West

Ft: Feet

g: Gram

GR: Gamma-Ray

GRL: Natural Gamma Ray Log

HC: Hydrocarbon

HCl: Hydrochloric acid

HI: Hydrogen Index

ILD: Deep Induction resistivity log

K: Potassium

KCl: Potassium Chloride

Km: Kilometer

LAS: A file format for the interchange of 3-dimensional point cloud data

LOM: Level of organic maturation

M: Estimated slope

m:Meters
Ma:Million Years
mg: Milligram
N:Nitrogen
NE:North East
NE-SE:North East-South East
NPHI: Neutron porosity log
N-S: North-South
NW-SE: North West-South East
OI:Oxygen Index
OM:Organic matter
PEF:Photoelectric Factor
PI: Production Index
ppm: parts per million
R: Resistivity
 R^2 = The coefficient of determination
RHOB: Bulk Density
Ro: Vitrinite reflectance
Rr : Vitrinite reflectance scale
S:Sulphur
S2: Korogen Yield
Sf: Salt free
SGRL:Spectral Gamma ray logging
SSW-NNE; South South West-North North East
SWC: Side Wall Core
TC:Total Carbon
Th:Thorium
TIC:Total Inorganic Carbon
Tmax : Temperature of maximum pyrolyzate yield expressed in °C

TPDC: Tanzania Petroleum Development Corporation

U: Uranium

V-PDB: Vienna PeeDee Belemnite

wt% : Weight percentage

XRF: X-ray fluorescence

Introduction

The onshore Mandawa Basin is one amongst many East African Coastal Basins that evolved in response to the break-up of Gondwana and the drifting of Madagascar during the early Mesozoic era. The basin evolved in multiple stages of extension (rifting phases) including the initial Permo-Triassic (Karoo) rift that did not advanced into the oceanization stage but rather created NNE-SSW intracratonic rift basins that were filled by continental sediment, followed by continued rifting during the Toarcian (~180Ma) that resulted into further disintegration of Gondwana and break up. This event was followed by rift to drift transition in the Middle Jurassic (~170Ma) period, which initiated seafloor spreading and opening of the western Somali Basin resulting in a widely spread deposition of marine sediments in the region. Seafloor spreading between Madagascar and Africa continued until 120Ma, when strike-slip motion between Madagascar and India began. From there onwards, the basin developed into a passive continental margin characterized by marine, coastal, and partly terrestrial sediment accumulation, primarily influenced by eustatic rise sea level and tectonics, and it persisted into the Neogene period (epoch). This resulted in a stratigraphic sequence of five groups listed chronologically as Pindirolu, Mandawa, Mavuzi, Kilwa, and Songosongo.

This study focuses on the Pindirolu Group that represents the syn-rift sedimentary package and is entirely made up of evaporites, siliciclastic, and carbonate lithofacies and consists of three formations (Mbuzi, Nondwa, and Mihambia Formations). These sediments recorded rifting events associated with earliest stage of Gondwana Break up as well as evidence of marine connection between the Neotethys Sea and the East African Margin terrain prior to the main phase of opening of the Indian Ocean during Middle Jurassic times. During this phase important elements of petroleum systems such as lacustrine and restricted marine shale, serving as potential source rocks, basal sandstone and post Toarcian shallow marine carbonates possessing high quality reservoir to store hydrocarbons, as well as shale/evaporites that could act as effective seals, were deposited.

To date, there have been few studies undertaken on the sediments of the Pindirolu Group. These studies focused on stratigraphy and sedimentology and relatively few studies on the source rock

potential of the Mbuo and Nondwa Formations. Investigations were based on few cuttings/sidewall samples collected from scattered wells. As result, both stratigraphic and lateral variability in source qualities with respect to facies changes in the heterolith Pindiro Group have been poorly documented. The primary objectives of this study are to examine sedimentary facies evolution, its prevailing deposition conditions as well as its implication on source rock potential in sediments of Lower Jurassic successions. The findings of this study have improved an understanding of lithofacies, stratigraphy and paleogeography of Pliensbachian-Toarcian strata in a regional scale (seven sub- basins in the East African Margin). In addition, prevailing redox conditions, as well as stratigraphic and lateral variability in TOC and source rock potential of Pliensbachian – Toarcian strata in the Mandawa Basin have been documented. Furthermore, an integration of geophysical wireline logs (Th/U, Th/K), sediment lithotypes (carbonate %, clastic %, salt %), and paleogeographic data provides an insight into relationship between changes in sediment input and organic matter distribution in the basin during Pliensbachian-Toarcian times.

This study was guided by nine research questions:

- How did Pliensbachian-Toarcian strata in the East African Margin evolve?
- What kind of paleoenvironment prevailed in the East African Margin during Pliensbachian-Toarcian?
- Which lithofacies types were deposited in the Mandawa Basin during Pliensbachian - Toarcian?
- What are the conditions of deposition prevailing during the Early Jurassic (Mbuo, Nondwa, and Mihambia Formations) in the Mandawa Basin?
- How was total organic carbon (TOC) distributed and enriched across the sediments of the Lower Jurassic successions?
- How can TOC variability across the sediment of Lower Jurassic successions be linked with changes in sediment lithotypes and redox conditions prevailing?
- What is the type of organic matter deposited in the Mandawa Basin during Early Jurassic times?
- How do organic matter types and quality of source intervals in the Lower Jurassic sediment successions vary spatially and laterally?

- What are the geological factors that govern organic matter quality distribution in the Lower Jurassic sediment successions in the Mandawa?

This project achieved its objectives through the application of geological, geophysical and geochemical research methodologies, details of each method are provided below:

Geological work involves lithological logging and sampling at Dar es Salaam's TPDC core repository, as well as field work in Lindi Region. About 185 rock cuttings samples from three bore holes (Mbuo-1, Mita Gamma-1, and East Lika-1) that penetrated sediments of Lower Jurassic successions (Mbuo, Nondwa, and Mihambia Formations), 25 core samples from the sediments of Lower Mbuo Formation in the Mbuo-1 well, and ten samples from the sediment of Nondwa evaporitic sequences outcropping in the southern central part of the basin were collected and used in this study. In addition, drill cutting sampling from the bore hole was supplemented by geophysical wireline logs to allow quasi-continuous documentation of lithology at higher resolution (~15cm).

In this study the following geochemical research methodologies were applied for the assessing organic facies, determining depositional conditions and kerogen types, evaluating petroleum generative potential, as well as the basin's thermal history and its associated maturation of kerogen and expulsion of hydrocarbons. These are analysis are as follows:

- Carbon, Nitrogen and Sulfur (C, N, S) analysis (Elementar®)
- Total organic carbon (TOC) analysis (Eltra®)
- Sediment lithotypes (carbonates, salts, clastics)
- Stable isotope analysis of organic matter ($\delta^{13}\text{C}$)
- TOC and Programmed pyrolysis analysis (Hawk®)

In addition to geochemical data derived from these analyses, Spectral Gamma Ray logs (Th, U, K) and sediment lithotype data (carbonates, salts, and clastics) were also integrated in the study to improve an interpretation of depositional conditions prevailing as well as changes in detrital sediment inputs through time. This approach allowed for reconstruction of the basin-fill of the Lower Jurassic successions in the Mandawa Basin.

This study consists of five chapters, a brief summary of each chapter is provided below;

In chapter One, a regional synthesis of the Pliensbachian-Toarcian geological evolution of East African Coastal Margin is presented. This synthesis provides a regional context for the subject matter, highlighted gaps in the existing literature and discussed all controversies in the published literature to enhance the geological understanding of these strata in the region. This chapter is titled: *Review of earliest Toarcian geological evolution in the East African Coastal Margin: Paleogeography, stratigraphy, and facies implications.*

In chapter Two, sedimentary facies evolution and depositional conditions of sediments of Lower Jurassic successions in Mandawa basin was reconstructed by integrating Spectral Gamma Ray logs (Th, U, and K), lithological logging as well as sediment composition data (carbonates, salts, and clastics). The findings indicate that during Pliensbachian times, deposition of clastic sediments into shallow, well mixed oxic water depositional settings occurred primarily in the east and western flanks. In contrast, more finer grain materials were deposited under oxygen-deficient conditions prevailing in the central and deeper parts of the basin. A prolonged period of clastic sedimentation was interrupted by two major episodes of marine incursions during the deposition of the sediment of Nondwa and Mihambia Formations as deduced from a decline in Th/U and Th/K ratios. This chapter is titled: *Lithofacies of early Jurassic successions derived from spectral gamma ray logging in the Mandawa Basin, SE Tanzania.*

Chapter Three, as continuation of chapter two, discusses TOC variability with respect to basin-fill history and changes in depositional conditions during Early Jurassic period in the Tanzanian Mandawa Basin as examined by integrating geophysical wire line and geochemical data. Results reveal that sediments from the Lower Jurassic successions exhibit high variability in TOC along stratigraphy. This study revealed three organic matter enrichment intervals, i) at the transition of Upper Mbuo to basal Nondwa Formation, in the Nondwa Formation, and in the Mihambia Formation. Hereby, (i) indicates organic matter preservation in a limnic environment, whereas (ii and iii) indicate organic matter under suboxic conditions upon marine incursions. An inverse linear

relationship between Th/U and TOC signifies relatively shallow deposition in a well-mixed water column under oxic conditions. Inverse trends in TOC versus clastic components, a positive linear relationship between TOC and carbonates, and higher amount of clastics (>57 wt%) in the Mbuo and Mihambia Formations suggest a dominantly clastic sedimentary system with relatively stable carbonate and organic matter input in proximity to the source area. This chapter is titled: *Total organic carbon variability of lower Jurassic successions in the Mandawa Basin, SE Tanzania*.

Chapter Four present a comprehensive geochemical analysis of source rock intervals in Lower Jurassic successions (Mbuo, Nondwa and Mihambia Formations) in the Mandawa Basin, in order to constrain organic matter types and quality. Organic matter richness was measured by TOC analysis, while quality and thermal maturity of organic matter was assessed by programmed pyrolysis, complemented by organic carbon isotope analysis ($\delta^{13}\text{C}_{\text{org}}$) of sediment samples collected from outcrops, cores, and cuttings samples. Programmed pyrolysis data show that organic matter is composed mainly of type III (gas prone) to IV (inert) kerogens with lesser contribution by mixed type II-III kerogen (oil-gas prone). Sediments have been deeply buried (>2000m) and undergone a wide range of thermal diagenesis, from early to over mature stages. Isotopic composition ($\delta^{13}\text{C}_{\text{org}}$) data of organic matter are consistent with pyrolytic kerogen typing, indicating preferentially terrestrial organic matter sources that changed to marine organic matter upon marine incursions, accompanied by a shift from heavier (-19 to -22 ‰ V-PDB) to lighter (-24 to -27 ‰ V-PDB) isotopic composition. This chapter is titled: *Assessment of kerogen types and source rock potential of Lower Jurassic successions in the Mandawa Basin, SE Tanzania*.

Chapter Five, summarize the key findings of our study and provide recommendations for future research work.

Chapter One

A regional synthesis of Pliensbachian - Toarcian Geological evolution of East African Margin

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Abstract

In the past decades, lower Toarcian strata have gained attention on a global paleogeographical scale but these strata have been poorly studied along the East African Margin. This study presents a comprehensive review of the Toarcian geological evolution of the East African Coastal Margin focusing mainly on the Mandawa Basin in Tanzania, the Morondava and Majunga Basins in Madagascar, the Lamu and Mandera Basins in Kenya, and the Luuq Mandera and Ahl-Mado Basins in Somalia. We discuss controversies in published reports regarding facies, stratigraphy, and pathways of marine ingressions during the Toarcian transgressive episode that led to the extension of a Neotethys sea arm (the Somali Ocean) into the East African Margin prior to the main phase of Gondwana drifting and the opening of the Indian Ocean.

The earliest Toarcian transgressive phase was accompanied by rifting between East Africa and Madagascar, which most likely rejuvenated Karoo rifting, drove marine influxes and inundated existing depressional troughs in the Ahl-Mado, Luuq Mandera, SW Majunga, and Mandawa Basins. This resulted in the formation of the Arab-Indo-Madagascar embayment, a NNE-SSW striking epicontinental sea (Somali Ocean), which was linked to the Tethys Sea southward via the horn of Africa route, as well as in widespread deposition of Toarcian marine to coastal-lagoonal to continental sediments. The occurrence of major facies changes from predominantly continental to shallow/marginal marine facies and the widespread distribution of an Arab-Indo-Madagascan *Bouleiceras* ammonite association confirms the connection between the East African coastal sedimentary terrain and the Tethys Sea during the Toarcian. Influxes of marine waters from Tethys Sea were limited in extent to the rift graben regions and reactivated Karoo

faults system. Restricted marine deposition within a limited epicontinental sea likely ended during Late Aalenian-Bajocian times, when establishment of fully marine conditions began due to opening of the Indian Ocean.

1. Introduction

The Toarcian stage of the Lower Jurassic records the first major global environmental perturbation of the Mesozoic and is thus regarded as one of the most important periods in Earth's history. Among the environmental changes that characterize this time interval is an increase in atmospheric CO₂ and enhanced greenhouse conditions that lead to the first large-scale Mesozoic Toarcian Oceanic Anoxic Event (T-OAE) (Jenkyns 1988; Röhl et al., 2001; Jenkyns 2010; Silva et al., 2021). The T-OAE is thought to have been caused by global warming, as average temperature increased by ~ 6-9°C (Svensen et al., 2007; Dera et al., 2011; Swaby, 2018; Ruebsam et al., 2020). Similarly, a significant increase in greenhouse gases such as CO₂, and ¹³C-depleted CH₄ was linked to the global warming phenomenon (Hesselbo et al., 2000, 2007; Sabatino et al., 2009; Holz, 2015) and carbon dioxide concentration in the atmosphere was substantially higher than it is today (McElwain et al., 2005; Svensen et al., 2007; Holz, 2015; Ruebsam, et al., 2020). Carbon dioxide may have been released as a result of degassing following tectonic events such as sea floor spreading, continental rifting, or the emplacement of large igneous provinces (LIP) on southern Gondwana, especially the Karoo Ferrar flood basalts in southern Africa and Antarctica (McElwain et al., 2005; Svensen et al., 2007), or greenhouse gases released from melting ice caps in high to mid latitude regions (**Fig. 1.1**). The latter, however, is still in contention (Littler et al., 2010; Rubino et al., 2015; Ruebsam et al., 2019; Ruebsam, et al., 2020). Temperature increases and high carbon dioxide concentrations impacted many terrestrial and marine habitats, a phenomenon termed as the “Jenkyns Event” (T-OAE) (Jenkyns, 1988, 2010; Hesselbo et al., 2000) (**Fig. 1.1**).

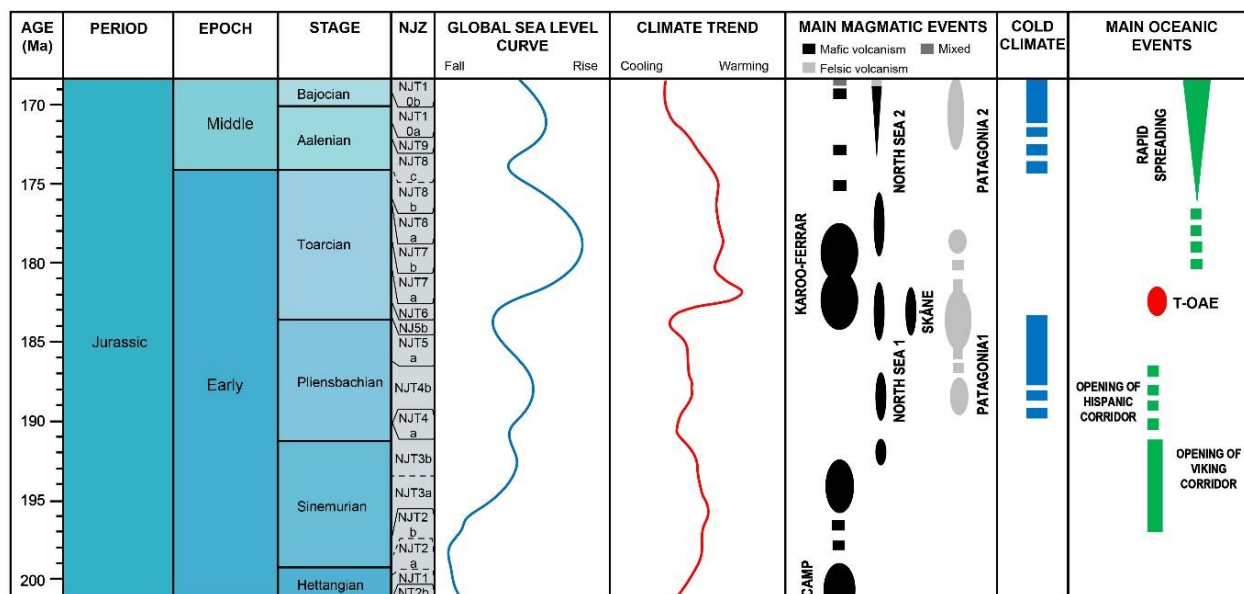


Fig. 1.1: Comparison of major global events occurring during Early Jurassic including the Toarcian. Adapted and redrawn from (Bergelin, 2009; Dera et al., 2011), geological time scale was drawn using Timescale Creator 2019 (<https://timescalecreator.org/index/index.php>) and global sea level curve is adapted from (Haq, 2018).

The T-OAE appears to have been short-lived for (< 1 Myr) (Svensen et al., 2007; Suan et al., 2008; Dera et al., 2011; Boulila et al., 2014; Ruebsam et al., 2019) and is well documented by recurrent negative carbon isotope excursions (CIEs) in both marine and terrestrial archives (Hesselbo et al., 2000, 2007; Kemp et al., 2005; Jenkyns, 2010; Littler et al., 2010; Them et al., 2017; Ruebsam et al., 2019). The T-OAE was further associated with deposition of organic-rich sediments (Jenkyns, 1988, 2010), which frequently possess excellent oil generating potential (Baudin et al., 1990; Yi et al., 2013; Krishna, 2017; Bishop, 2020). Most of the previous studies focused on the European boreal realm (Northern Europe), the western Tethyan sea (Northern hemisphere) and very few on the southern hemisphere, specifically on the South American continent (**Fig. 1.2**). Some studies have been undertaken in Northern Africa and very limited studies in East African Margin (**Fig. 1.2**).

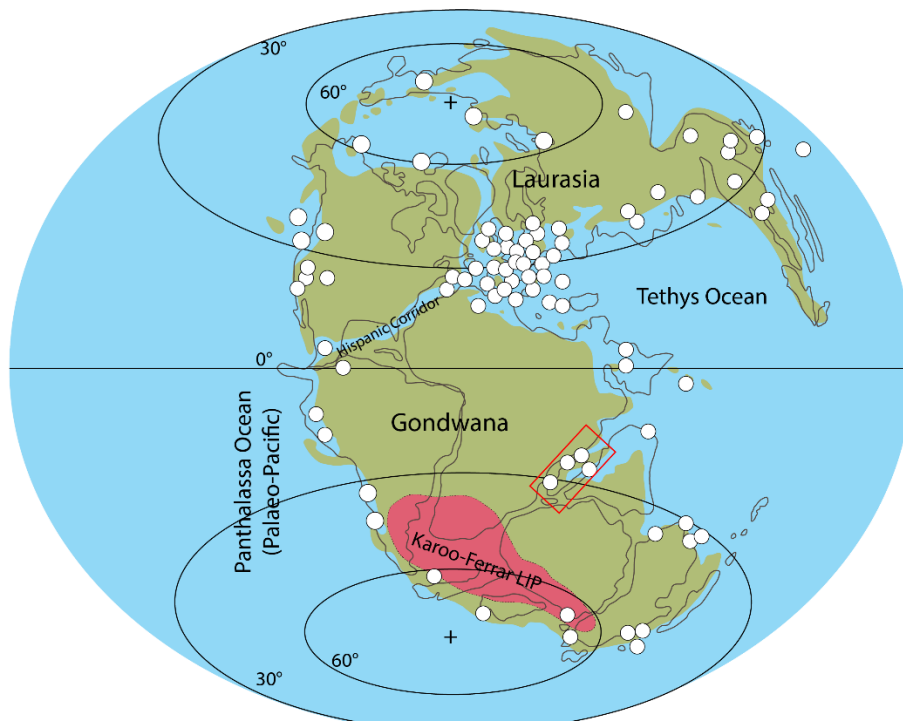


Fig. 1.2: Global positions of the Early Jurassic continents and known localities of Lower Jurassic sediments, locations of East African coastal basins are shown by the red box (modified from Silva et al., 2021).

Generally, for East African coastal basins, there has been controversy on the inception of the Toarcian marine transgression, the spatiotemporal configuration of paleo-seaways and basins as well as their concomitant sedimentary facies due to poor biostratigraphical data, lack of high quality outcrops, limited well numbers and complex tectono-sedimentary history. It is often assumed that all sediments beneath the widespread Bajocian-Bathonian transgressive phase are interpreted as Karoo strata (Kent et al., 1971; Kamen-Kaye and Barnes, 1979; Balduzz et al., 1992; Worku and Astin, 1992). In turn, continental to marine Toarcian facies tend to be overlooked or underdiscussed in comparison to the Karoo and Bajocian-Bathonian sequences. To date, there are few studies devoted to Toarcian strata in the East African region. Hence, this study aims to re-examine the existing knowledge and views on lower Toarcian lithostratigraphy and paleogeography of the East African Coastal Margin in order to address existing controversy regarding the geological evolution of these strata. In addition, an enhanced understanding of lithostratigraphy and paleogeography in the region will allow an assessment of possible petroleum systems including the Toarcian strata that globally are renowned for their oil potential.

1.1. Regional Geological overview

The East African Coastal Margin covers all sub-basins that extend from the coast of Somalia in the North to those in Mozambique in the South and includes the coastal plains of Tanzania, Kenya and Somalia (Kajato, 1982; Coffin and Rabinowitz, 1988; Coffin, 1992; Mpanda, 1997; Zhou et al., 2013; András et al., 2014). These basins were formed by the breakup of Gondwana and the drifting of Madagascar (Bosellini, 1986, 1992; Mascle et al., 1987; Salman and Abdula, 1995; Zhou et al., 2013; Frizon de Lamotte et al., 2015; Wen et al., 2015). Therefore, fragmentation of Gondwana and the drift of Madagascar are significant aspects of this review.

Any attempt to detail the geological evolution and paleogeography of these basins should consider the paleoposition of Madagascar. Its interpretation based on paleomagnetism has been debated for many years due to lack of high quality geological and geophysical data (Embleton and McElhinny, 1975; Bosellini, 1986; Geiger, 2004). Apparently, four major hypotheses on the paleoposition of Madagascar have been postulated; (i) it was placed adjacent to South Africa and Mozambique, (ii) it maintained its position at least since the Permian (iii) it was located west of India, and (iv) it was positioned adjacent to the East African coast and opposite to Tanzania, Kenya and Somalia (Kent et al., 1971; Förster, 1975; Kamen-Kaye and Meyerhoff, 1979; Bosellini, 1986; Coffin, 1992; Salman and Abdula, 1995; Mpanda, 1997). Recent improvement of paleomagnetic data confirmed that Madagascar was situated to the East of Kenya and Tanzania (Embleton and McElhinny, 1975; Cannon et al., 1980; Boast and Nairn, 1982; Coffin and Rabinowitz, 1987, 1988; Rais-Assa, 1988; Walsh and Scotese, 1992; Geiger, 2004; Schettino and Scotese, 2005; Eagles and König, 2008; Papini and Benvenuti, 2008; Krause et al., 2019). An improved understanding of the paleoposition of Madagascar and the role played by the Karoo and young tectonic events sheds light on refinement of paleogeography and evolution of the East African Margin, including depocentres bearing Toarcian strata.

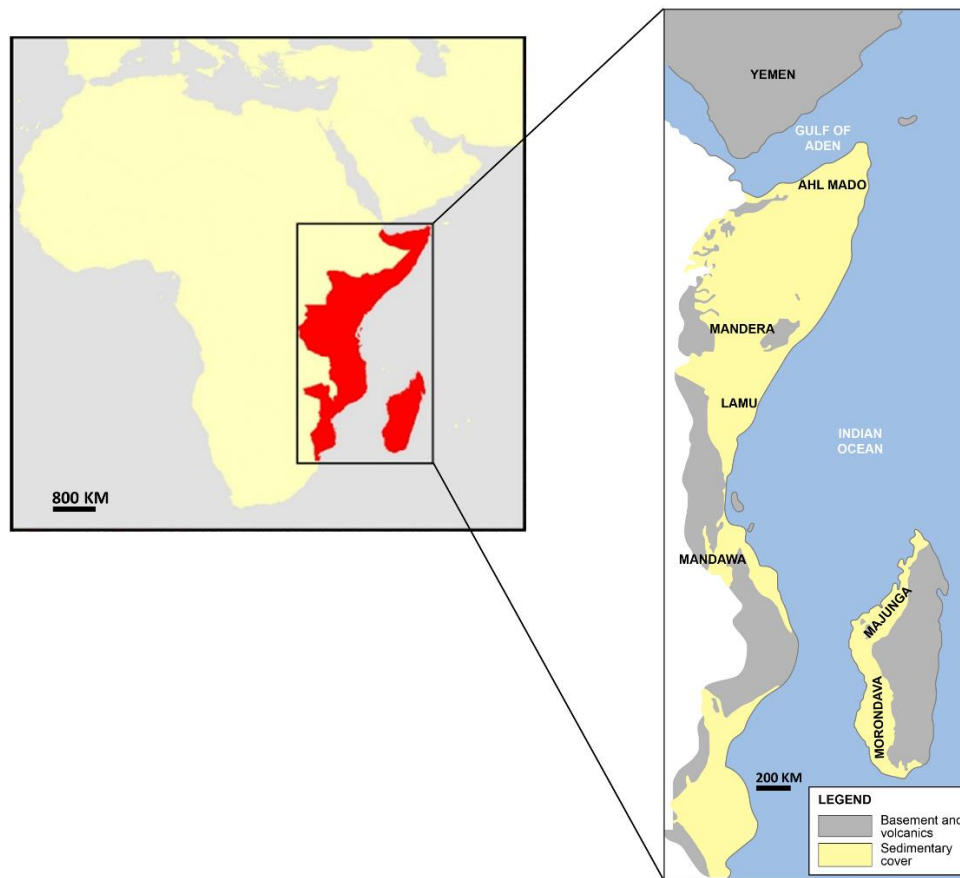


Fig. 1.3: Location map of East African Coastal Basins, after (Downes and Boote, 2009).

1.1.1 Permo - Carboniferous to Triassic (Karoo) geological evolution

Most of Gondwana is thought to have remained stable until Permo-Carboniferous times, when disintegration took place and continued until the late Triassic era, a period commonly referred to as the Karoo stage in Africa. Karoo strictly applies to continental rifting that affected the African continent (**Fig. 1.4**) (Cannon et al., 1980; Kreuser et al., 1990; Worku and Astin, 1992; Wopfner, 1994, 2002; Kreuser, 1995; Mbede and Dualeh, 1997; Mpanda, 1997; Delvauxi, 2001; Reeves, 2018). Karoo rifting was characterized by the subsidence, crustal extension and thinning of the low relief and extensive Precambrian peneplain (Coffin, 1992; Mbede and Dualeh, 1997; Delvauxi, 2001). There was no common mechanisms controlling the individual rifts and graben formations (Macgregor, 2015). During the Permian, the stress regime was purely strike-slip resulting in

formation of pull-apart basins (Coffin, 1992; Delvaux, 2001; Schandelmeier et al., 2004; Macgregor, 2015, 2018), a scenario that persisted up into the Triassic when the stress regime changed into trans-tensional that enforced extensional depositional systems (Schandelmeier et al., 2004). Triassic rifting events were succeeded by purely tensional tectonic regimes accompanied by widespread formation of half graben systems (Schandelmeier et al., 2004; Macgregor, 2015, 2018). The different tectonic regime resulted in a wide distribution of fluvial lacustrine to deltaic continental deposits that developed in half graben structures generally trending NE-SE, NNW-SSE and locally E-W. In exceptional cases, the Karoo basins contained marine deposits, implying short lived marine incursion into these half grabens (Kreuser et al., 1990; Wopfner, 1994, 2002; Macgregor, 2018). Along the East African Coastal Margin, Permo-Triassic marine successions have been reported from the Selous Basin in Tanzania, the Morondava Basin in Madagascar, and the Lamu Basin in Kenya (**Fig. 1.4**) (Boast and Nairn, 1982; Hankel, 1987, 1992; Coffin and Rabinowitz, 1988; Coffin, 1992; Wopfner, 1994, 2002; Geiger, 2004). The thickness attained by Karoo successions varies largely from place to place and regionally can reach maxima up to 12 kms (Kreuser et al., 1990; Wopfner, 1994). The majority of lithological units of the Karoo Group successions are genetically related (Wopfner, 2002), however, due to different levels of subsidence and depositional environment, substantial lithofacies variation do exist (Mbede, 1991; Mbede and Dualeh, 1997; Mpanda, 1997; Kapilima, 2003b).

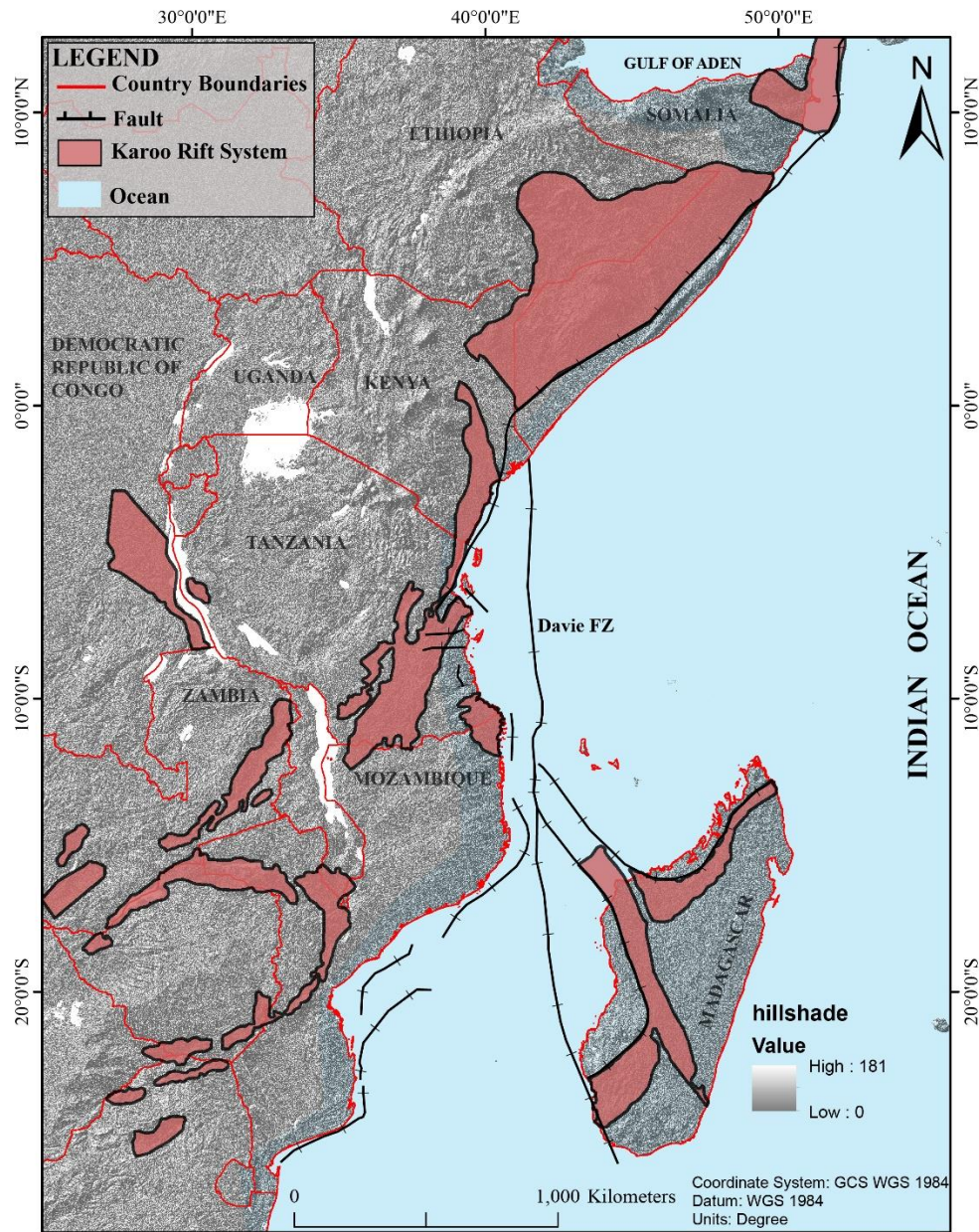


Fig. 1.4: Schematic map showing the distribution of Karoo rifting-associated continental sediment graben and basins of South East and Central Africa, after Frizon de Lamotte et al. (2015).

1.1.2 Early Jurassic geological evolution in the East African Coastal Margin

A major mantle plume affected SE-Gondwana prior to its breakup, leading to regional uplift during the Hettangian to Sinemurian, which was followed by rifting accompanied by volcanisms during the early to late Lias (Macgregor, 2018; Reeves, 2018). This rifting event probably attained its

peak during the Toarcian (Macgregor, 2018). Situated in the vicinity of East Gondwana, the East African Coastal Margin was affected by this rifting, resulting in extensive Karoo or Stromberg basalt sheets and extreme large dyke swarms in Southern Africa (**Fig. 1.13**). There was regionally localized volcanic activity in the East African Margin, including a portion of southern Tanzania and Somalia (Kamen-Kaye and Barnes, 1979; Hankel, 1992; Worku and Astin, 1992; Buscaglione et al., 1993; Kreuser, 1995). Additionally, this event reactivated the Karoo rifting and hugely modified the topography of the area covered by the present day East African Margin. As a result of relatively high plate extension rates and eustatic sea level rise of app. 50 to 80 m (Haq, 2018), influxes of marine waters from the Tethys Sea invaded the lowland margin and inundated subsidized graben structures (Cannon et al., 1980; Mpanda, 1997; Hudson, 2010; Reeves, 2018), where they deposited Toarcian marine strata. The trans- or ingressions into the revitalized Karoo troughs/grabens occurred sporadically due to differences in regional subsidence and local uplift that prevailed in all parts of the East African Coastal Margin (Kent, 1972, 1974; Buscaglione et al., 1993; Mbede and Dualeh, 1997; Mpanda, 1997).

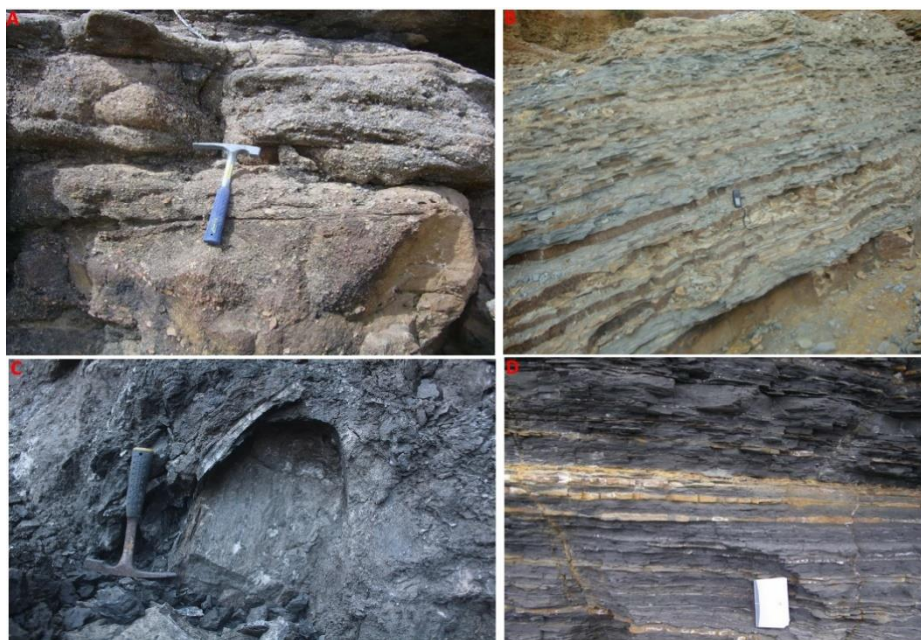


Fig. 1.5: (A) Conglomeritic Adigrat sandstone with detrital quartz exposed in Berbera Somalia (adapted from Ali, 2009), B: Mazeras sandstone intercalated with shale outcropping near the Town of Mazera in Lamu Basin (adapted from EAPCE, 2019), C: Highly deformed Nondwa evaporitic sequences in the Mandawa Basin, D: Toarcian shale with minor sandstone intercalation in the Majunga Basin (adapted from Banks, 2008).

1.2 Pliensbachian and Toarcian lithostratigraphy of the East African Coastal Margin

This section reviews Toarcian lithostratigraphic units of seven Mesozoic basins in the East African Coastal Margin (**Figs. 1.3 and 1.5**) from North to South. While reviewing, we emphasize on rocks types constituting different formations in various isolated basin, which are stratigraphically correlated based on important Toarcian macro- and nanno-index fossils summarized in **Table 1.1**. An outline of Toarcian geology and lithostratigraphy of individual sub-basins will be discussed in detail below.

Region	Basin	Formation	Main lithology	Important (Index) fossil	Stratigraphic position	Main references	General references
North Somalia	Ahl Mado	Daghabo	Marl/ Limestone	Ammonites: <i>Bouleiceras</i> spp. (<i>nitescens</i>), <i>Hildaite</i> sp. and <i>Protogrammoceras</i> spp. (var. <i>Madagascariense</i>); Brachiopods: <i>Spiriferina</i> <i>rostrata</i> var. <i>Madagascariensis</i> * = <i>Calyptoria</i> <i>madagascariensis</i> ; Bivalves: <i>Pelecypods</i> spp. (<i>Pholadomya</i> , <i>Lima</i> , <i>Camptonectes</i> , <i>Weyla</i> <i>ambongoensis</i>)	Semicelatum and elegantulum ammonite subzones, Lower Toarcian	Arkell, 1956; Kamen-Kaye 1978; Mette 1993	Luger et al.,1994; Fantozzi and Ali Kassim 2002; Krishna ,2017; Martinez and Joral, 2020
South Somalia	Lugh Mander	Waney (Uanei)	Marl/ Limestone	Ammonites: <i>Bouleiceras</i> spp. (<i>arabicum</i> , <i>elegans</i> , <i>nitescens</i> , <i>rectum</i>) <i>Protogrammoceras</i> spp. (var. <i>Madagascariense</i>) and <i>Hildaite</i> sp.; Brachiopods: <i>Spiriferina</i> <i>rostrata</i> var. <i>Madagascariensis</i> *= <i>Calyptoria</i> <i>madagascariensis</i>	Semicelatum and elegantulum ammonite subzones, Lower Toarcian	Arkell, 1956; Kamen-Kaye, 1978; Canuti et al., 1983; Buscaglione et al., 1993; Chiocchini et al.,2002	Angelucci et al.,1981; Riccardi, 1991; Ali Kassim et al., 2001; Krishna, 2017; Carratala et al., 2018; Martinez and Joral, 2020
North Madagascar	Majunga /Diego	Beronono equivalent	Marl/ Limestone	Ammonites: <i>Bouleiceras</i> spp. (<i>nitescens</i> , <i>elegans</i> , <i>rectum</i> , <i>tumidum</i> , <i>ibericum</i>); Brachiopods: <i>Spiriferina</i> <i>rostrata</i> var. <i>Madagascariensis</i> *= <i>Calyptoria</i> <i>madagascariensis</i> ; Bivalves: <i>Pecten</i> spp. (<i>Ambongoensis</i>)	Semicelatum and elegantulum ammonite subzones, Lower Toarcian	Arkell,1956; Colligon et al., 1959; Kamen-Kaye, 1978, Buffetaut et al., 1981; Geiger, 2004; Geiger and Schweigert, 2006	Riccardi, 1991; Luger et al.,1994; Pappini 1995, Pappini and Benvenuti, 2008; Krishna, 2017; Martinez and Joral, 2020
North Kenya	Mandera	Didimtu	Marl/ Limestone	Ammonites: <i>Bouleiceras</i> spp. (<i>arabicum</i> , <i>nitescens</i>); Brachiopods: <i>Spiriferina</i> <i>rostrata</i> var. <i>Madagascariensis</i> *= <i>Calyptoria</i> <i>madagascariensis</i> ; Cephalopods: <i>Nautilids</i> spp. (<i>Paracenoceras</i> <i>anomphalum</i>)	Semicelatum and elegantulum ammonite subzones, Lower Toarcian	Ayers, 1952; Arkell, 1956; Kamen-Kaye 1978	Thompson and Dodson, 1960; Joubert 1963; Martinez and Joral 2020
South Tanzania	Mandawa	Nondwa	Shale, anhydrite	Calcareous nannofossil (<i>Lotharingius</i> <i>hauffii</i> , <i>Carinolithus</i> <i>superbus</i>)	NJ5-NJ6 nannofossils subzones, Lower Toarcian	Karega, 2001; Stratadata, 2006; Corelab and TPDC, 2009; Robertson and CGG, 2017	Veeken and Titov 1996; Mbede and Dualeh, 1997; Quinton and Copestake, 2007; Hudson, 2010; Brownfield, 2016; Davison and steel, 2018; Macgregor, 2018; McCabe, 2021

Table 1.1: Compilation of early Toarcian Biostratigraphy data of the five East African coastal sedimentary basins.
 **Spiriferina rostrata* var. *Madagascariensis* reclassified as *Calyptoria madagascariensis* after Baeza-Carratalá et al., 2018.

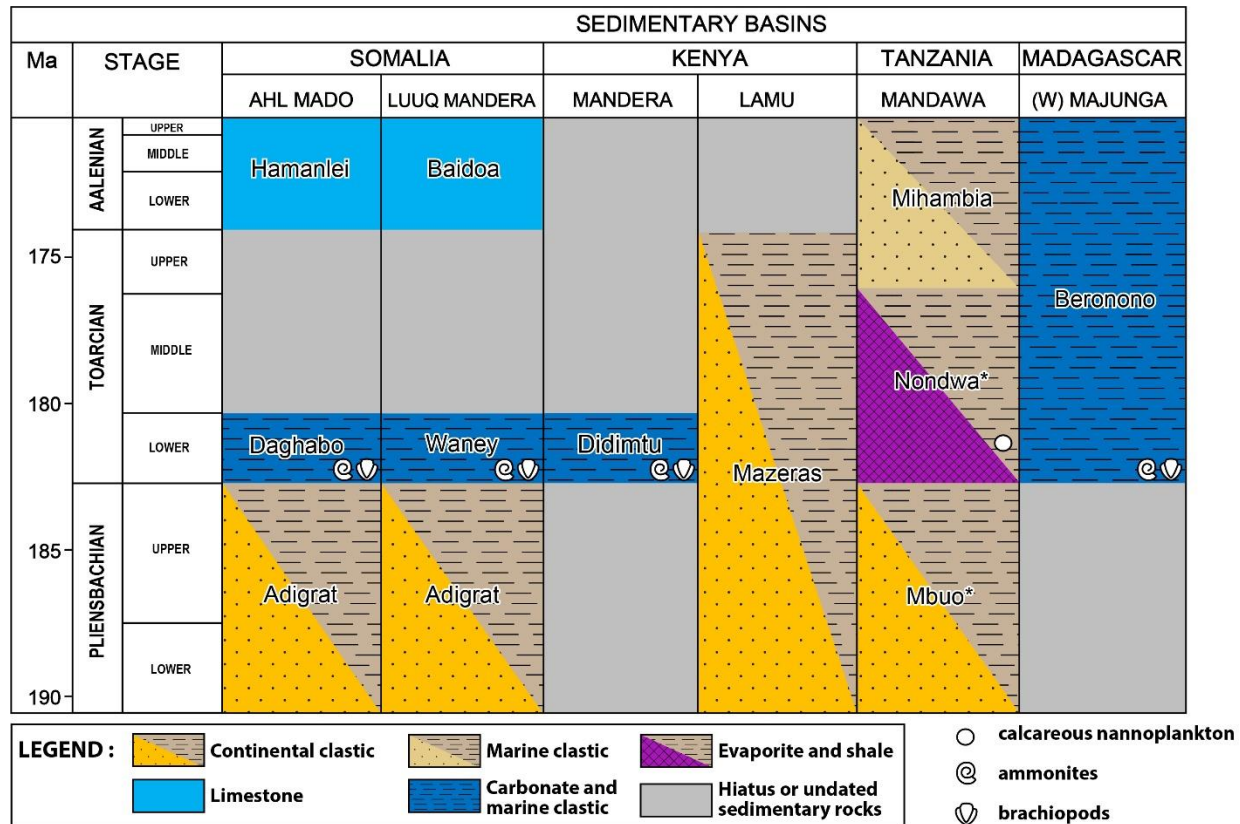


Fig. 1.6: Pliensbachian to Aalenian stratigraphy of basins along the East African Margin, depicting important formations including Toarcian strata. *The bases of these intervals have been interpreted as late Triassic by Hudson (2011; Hudson and Nicholas 2014).

1.2.1 Luuq Mander Basin (SW Somalia)

The Luuq Mander Basin is located in SW Somalia (**Figs. 1.3 and 1.7**) and is demarcated to the East by Proterozoic to early Paleozoic basement of the Bur region and to the West by crystalline basement rocks of Northeast Kenya (**Fig. 1.7**). This basin is part of the large East African Karoo rifting system filled with Mesozoic sediment up to 4km in thickness (Ali Kassim et al. 1987; Ali Kassim et al. 2001), and generally trends SSW-NNE (Ali Kassim et al. 1987; Mbede and Dualeh 1997; Ali Kassim et al. 2001). Outcropping Jurassic sediments dip shallowly to the Northwest in the southern part and to the Southeast in the northern part, indicating a gently dipping syncline

(Fig. 1.7) (Canuti et al. 1983; Ali Kassim et al. 1987; Ali Kassim et al. 2001). Liassic sediment including Toarcian strata comprise the oldest deposits in the basin **(Fig. 1.7)** and are subdivided into the Deleb (Adigrat) Formation (Pliensbachian) and the Waney (Uanei) Formation (Lower Toarcian) **(Fig. 1.6)**.

1.2.1.1 Waney (Uanei) Formation (Lower Toarcian)

The Waney (Uanei) Formation on average is 70m thick and overlies the Deleb (Adigrat) Formation unconformably **(Fig. 1.6)**. It is subdivided into three sections: lower, intermediate and upper members (Canuti et al. 1983; Ali Kassim et al. 1987; Buscaglione et al. 1993; Mbede and Dualeh 1997; Ali Kassim et al. 2001). The basal member is conglomeratic and highly bioturbated, with recrystallized nodules in place and is composed of marl bearing quartz clasts and limestones. Interbeds of thick marls to limestones, bioturbated calcilutites and calcarenite make up the intermediate member. The top member is made up of marl, red fossiliferous shale, calcilutites and calcarenites, which commonly exhibit cross bedding and breccia beds, or graded bedding (Ali Kassim et al., 2001). Facies and key sedimentary features suggest that this formation was deposited in a lagoonal to open shelf environment receiving intermittent influx of clastic material (Buscaglione et al., 1993; Ali Kassim et al., 2001). This formation contains important fossils **(Table 1.1)** such as gastropods, brachiopods, and ammonites, which date to the early Toarcian (Buscaglione et al., 1993; Mbede and Dualeh, 1997; Ali Kassim et al., 2001).

1.2.1.2 Baidoa Formation (Aalenian-Callovian)

The Baidoa Formation conformably overlies the Waney Formation **(Fig. 1.6)**, whereby the entire succession is dominated by oolitic limestones with shale interbeds and abundant shell fragments, which are further subdivided into the Baidoa and Goloda members (Buscaglione et al., 1993; Mbede and Dualeh, 1997; Ali Kassim et al., 2001). The Baidoa member is made up by interbedded layered calcarenites and bioclastic calcilutites of approximately. 100 m thickness that have oncoidal, peloidal, oolitic, packstone to bioclastic textures with cross bedding. The Goloda member is made up of calcilutites, limestones and dolomites that reach up to 600m thickness (Ali

Kassim et al., 2001). Previously, the entire Baidoa succession was indifferently placed in the Pliensbachian to Bathonian period (Mbede and Dualeh, 1997). This review adapts an Aalenian to Callovian age assignment on the basis of textures and fossils (Buscaglione et al., 1993; Ali Kassim et al., 2001). Lamellibranchiates, gastropods, echinoderms, foraminifers, ostracods, radiolarians, sponge spicules, bivalves, calcareous algae and corals are among the abundant fossils found in the succession. Sediments were deposited in a wider range of depositional environments, including open to narrow continental shelf, and the transition from narrow to open platform and ramp (Ali Kassim et al., 2001).

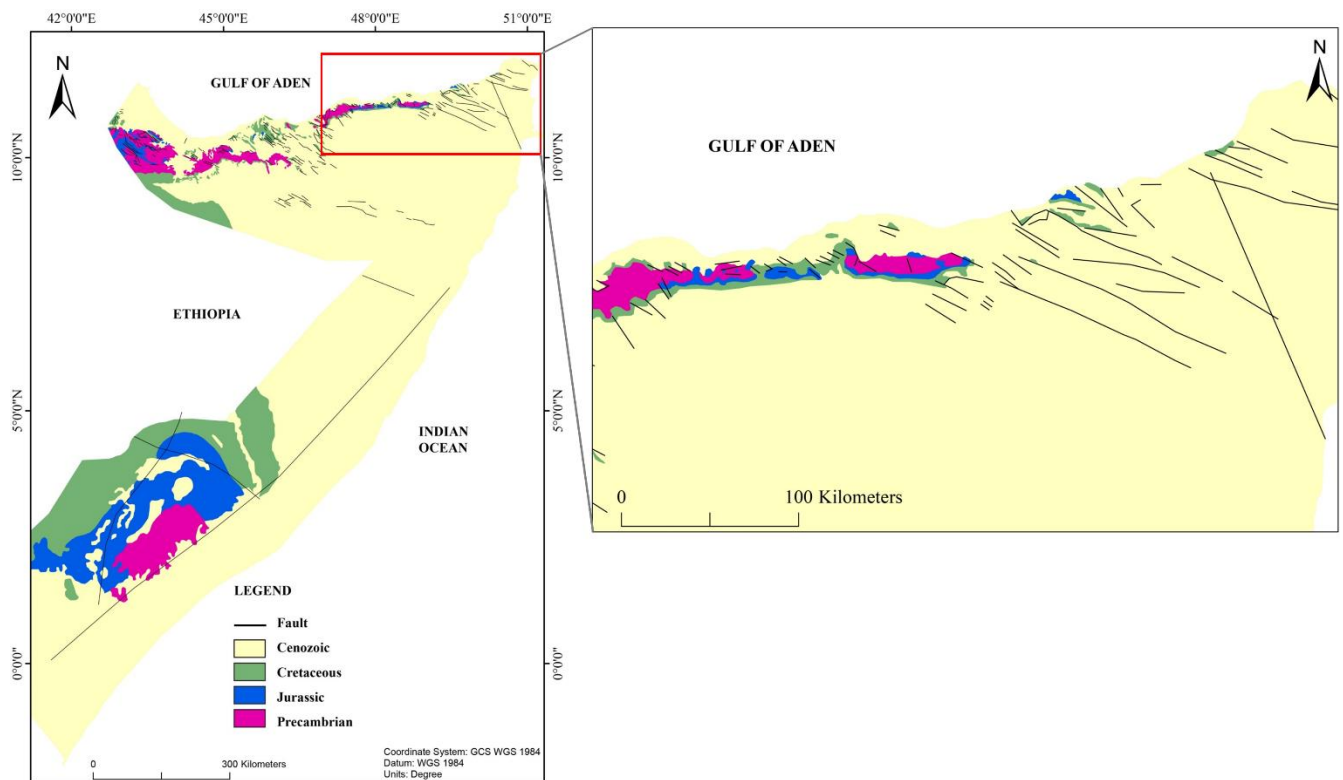


Fig. 1.7: Geological map of Somalia, depicting pre-Jurassic, Jurassic including Toarcian and Post Jurassic strata. Redrawn and modified from Mbede and Dualeh (1997) and Ali and Watts (2016).

1.2.2 Ahl-Mado Basin (NE Somalia)

The Ahl-Mado Basin is located in the northern horn of Africa (**Fig. 1.3**) and bounded on the North by the Gulf of Aden, on the South by the Nogal valley, on the West by the Somalia plateau, and on East by the Bosaso Basin (Ali, 2009). Evolution of this basin was linked with the development

of the Great Somalia Basin and concurrent opening of the Indian Ocean, which commenced during Triassic times (Mette, 1993; Luger et al., 1994; Fantozzi and Ali Kassim 2002). Except for the Erigavo area, where Cretaceous sediment directly overlies on Pan African basement, Jurassic sediments, including the Toarcian succession, outcrop widely in the Great Somali Basin (**Fig. 1.7**) (Luger et al., 1994). The thickness of Mesozoic sediments in the basin highly varies due to young faulting and hence is difficult to establish. Lower Toarcian successions correspond to the Dhagabo Formation (**Fig. 1.6**).

1.2.2.1 Adigrat Formation (? Triassic-Pre-Toarcian)

The Adigrat Formation (**Fig. 1.5a**) lies unconformably on Pan African basement (Mette, 1993; Luger et al., 1994; Fantozzi and Ali Kassim, 2002). This formation is mostly made up of sandstone and conglomerate (**Fig. 1.5a**). The lowermost strata consist of several meters of quartzose and schist clasts, rich in occasional detrital quartz and mica conglomerate, and the upper part is composed of interbeds of quartzose-micaceous sandstone and cross-bedded quartzose conglomerate (Fantozzi and Ali Kassim, 2002). Because both the upper and lower members are devoid of fossils, the inferred age of the formation is based on its relative stratigraphic position. The thickness of this formation varies greatly, possibly due to paleomorphological features (Mette, 1993; Fantozzi and Ali Kassim, 2002; Ali and Watts, 2016). A maximum thickness of 80-100m has been reported for the Ahl-Mado area (Fantozzi and Ali Kassim, 2002). This formation is widespread in East Africa and in the southern Arabian Peninsula (Kohlan Formation) (**Fig. 1.15**). Numerous authors proposed fluvial deposition with localized marine incursions as the depositional environment of this formation (Barnes, 1976; Bosellini, 1992; Papini, 1995). However, based on lithological characteristics, this review favors braided river to low sinuosity meandering depositional environment (Ali Kassim et al. 1987; Ali Kassim et al. 2001; Wolela 2008).

1.2.2.2 Dhagabo Formation (Lower Toarcian)

The Dhagabo Formation rests conformably on the Adigrat Formation (**Fig. 1.6**) (Mette, 1993; Luger et al., 1994; Ali and Watts, 2016). The transitional contact between the Dhagabo and the

Adigrat Formation is marked by the first appearance of marine strata (Ali Kassim et al., 2001; Fantozzi and Ali Kassim, 2002). The strata are mostly fossiliferous marls and limestones with ammonite (*Bouleiceras* sp. and *Hildaite* sp.), brachiopods, dasycladacean algae, and some fragments of cephalopods (Mette, 1993; Fantozzi and Ali Kassim, 2002). This formation is dated as of early Toarcian age based on these fossils. Due to faulting, the thickness of Dhagabo Formation varies greatly in this basin, the maximum thickness of the succession reaches 600m (Luger et al., 1994; Fantozzi and Ali Kassim, 2002). Biofacies and carbonate microfacies interpretation indicates normal marines (shelf lagoon) to marginal marine (protected lagoon) settings (Mette, 1993; Luger et al., 1994).

1.2.2.3 Qaranye (Hamanlei) Formation (Aalenian-Callovian)

The Hamanlei Formation lies unconformably on the Dhagabo Formation (**Fig. 1.6**) and is primarily composed of sandstones, conglomerates, fossiliferous dolomites, and gypsum. Numerous fossils including gastropods, algae and pelecypods have been discovered in the dolomite (Mette, 1993; Luger et al., 1994). Although the exact thickness is unknown due to faulting, drill wells have penetrated approximately 1775-2175m of this formation (Coffin and Rabinowitz, 1988; Mbede and Dualeh, 1997). Facies interpretation indicates an inter-supratidal depositional environment (Mette, 1993).

1.2.3 Lamu Basin (SE-Kenya)

The Lamu Basin is located along the present-day coast of SE Kenya (**Figs. 1.3 and 1.8**) and it is Kenya's largest basin, covering an area of approximately 170,000km² (Mbede, 1987; Nyagah, 1995; Nyaberi and Rop, 2014). This basin evolved as a failed arm of a tri-radial rift system during the early stages of continental breakup in the late Paleozoic. Rifting was successful in the Mesozoic, resulting in the breakup of Madagascar, drifting, and the development of a continental passive margin filled with sediment until today (Mbede, 1987; Nyagah, 1995; Nyaberi and Rop, 2014). In this basin, the Mazeras Formation (**Fig. 1.5b**) (Late Triassic-Early Jurassic) represents

Toarcian strata (Caswell, 1956; Cannon et al., 1980; Coffin and Rabinowitz, 1988; Rais-Assa, 1988; Hankel, 1994; Mbede and Dualeh, 1997; Chiocchini et al., 2005; Akech et al., 2013).

1.2.3.1 Mazeras Formation (Upper Triassic to Lower Jurassic)

The Mazeras Formation lies unconformably on the Mariakani Formation (**Figs. 1.5a and 1.6**) (Cannon et al., 1980; Rais-Assa, 1988; Mbede and Dualeh, 1997). The strata are primarily composed of medium to coarser sandstone of 450m thickness that is divided into three distinctive units. The lower unit is made up of banded arkosic yellow sandstone and a thin horizon of silicified wood, the middle unit consists of coarse arkosic sandstones with occasional conglomerates, and the upper unit comprises coarser sandstones with pebbly horizons (Coffin and Rabinowitz, 1988; Schlüter and Trauth, 2006). The lower to middle units represent a coarsening upward sequence. Whereas the upper unit displays cross bed stratification (Mbede 1997), the lower half of the Mazeras Formation comprises a thin horizon of late Triassic silicified wood (Cannon et al., 1980; Rais-Assa, 1988; Mbede and Dualeh, 1997). This implies that deposition of the Mazeras Formation probably commenced in the Late Triassic and continued into the Early Jurassic (Caswell, 1956; Cannon et al., 1980; Rais-Assa, 1988; Mbede and Dualeh, 1997). On the basis of facies and sedimentary structure associations, the Mazeras Formation was deposited in a deltaic to aeolian environment (Mbede and Dualeh, 1997). Strata are devoid of fossils, so that age assignment is inferred from relative position to underlying and overlying strata, which are abundant in fossils (Rais-Assa, 1988; Mbede and Dualeh, 1997).

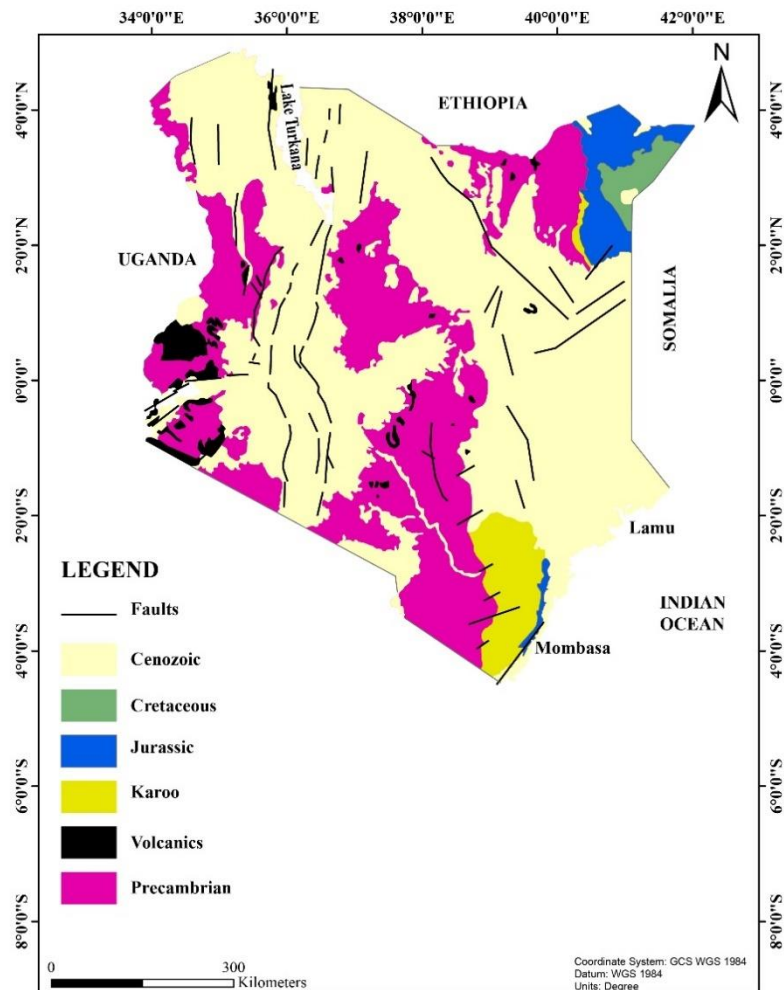


Fig. 1.8: Geological map of Kenya showing pre-Jurassic, Jurassic including Toarcian and post Jurassic strata. Redrawn from Mbede and Dualeh (1997) and Akech et al. (2013).

1.2.4 Mandera Basin (NE Kenya)

The Mandera Basin is located on the northeastern corner of Kenya (**Figs. 1.3** and **1.8**) and covers an area of roughly 44,000km². It is considered to be a regional counterpart of the larger Mandela Basin, which stretches from Somalia to Ethiopia (Ayers, 1952; Joubert, 1963; Worku and Astin, 1992; Nyagah et al., 1997; Wolela, 2008). This is a rift basin that formed as a result of the breakup of Gondwana, the drifting of Madagascar, and the subsequent opening of the Indian Ocean from the late Paleozoic to Mesozoic periods. The opening of the Indian Ocean was followed by the deposition of marine material, which lasted until the Lower Cretaceous, when the Neotethys sea began to withdraw northward (Mbede, 1987). The Mandera Basin contains continental and

marine sediments of Paleozoic and Mesozoic age, with a maximum thickness of 10 km. In this basin, the Didimtu Formation represents well-known Toarcian strata that outcrop in several localities in NE Kenya (**Fig. 1.8**). (Ayers, 1952; Arkell, 1956; Thompson and Dodson, 1960; Joubert, 1963; Swarzenski and Mundorff, 1977; Schlüter and Trauth, 2006; Galácz, 2017; Sproule, 2017; Jain, 2019).

1.2.4.1 Mansa Guda Formation (Triassic)

The Mansa Guda Formation, which is overlain by the Didimtu Formation, rests unconformably on Precambrian basement and is composed of a fine to cross bedded, occasionally lenticular medium-grained sandstone, quartzite, and conglomerate successions (Ayers, 1952; Thompson and Dodson, 1960; Joubert, 1963; Swarzenski and Mundorff, 1977; Papini, 1995; Schlüter and Trauth, 2006; Akech et al., 2013; Kuria, 2013). It outcrops discontinuously near the Didimtu area and extends further south to Tarbaj (Thompson and Dodson, 1960). Thickness varies on the surface from the North to South, from approximately 6m to 610m (Thompson and Dodson, 1960). Facies association and distribution indicate a deltaic depositional environment linked to a hinterland with high-energetic erosion and most probably extremely arid conditions (Thompson and Dodson, 1960). This formation is barren of fossils and a Triassic age blanket assignment is based on its relative stratigraphic position and its resemblance to similar rocks in Somalia and Ethiopia (Thompson and Dodson, 1960; Swarzenski and Mundorff, 1977).

1.2.4.2 Didimtu Formation (Toarcian)

The Didimtu Formation lies unconformably on the Triassic Mansa Guda Formation (Ayers, 1952; Thompson and Dodson, 1960; Joubert, 1963; Swarzenski and Mundorff, 1977; Kuria, 2013). Strata consist mainly of thick-bedded fossiliferous limestone with occasionally thin marly interbeds, which are subdivided into four distinctive members (Ayers, 1952). The lower member consists of a 6m thick dark grey and barren limestone 3 to 19m thick. Fine to coarser brown-grey spotted limestone overlies the lower member, followed by a 3 to 11m thick coquinoïdal limestone constituting an intermediate member. The upper member consists of 10 to 24m highly

fossiliferous limestone and calcitic mudstones and marly interbeds, which commonly display parting (Thompson and Dodson, 1960). *Bouleiceras Arabicum* ammonites were encountered in this bed by Ayers (1952). This formation contains diverse fossils such as brachiopods, gastropods, corals, cephalopods (ammonites and belemnites) and lamellibranchiates (Ayers, 1952; Arkell, 1956; Thompson and Dodson, 1960; Joubert, 1963). The presence of brachiopods (*Spiriferina rostrata* var. *Madagascariensis*) and Ammonites (*Bouleiceras arabicum*) demonstrate an Toarcian age (Ayers, 1952; Arkell, 1956; Thompson and Dodson, 1960; Joubert, 1963; Galácz, 2017; Jain, 2019). On the basis of lithology, ichnofacies and fossils association, the sediments were deposited in shallow water environments (Thompson and Dodson, 1960).

1.2.5 Majunga Basin (NW Madagascar)

The Majunga Basin is located on Madagascar's western coast (**Figs. 1.3 and 1.9**) and stretches for approximately 50 kilometers on the southern side of the Ampasindava Peninsula to 200 kilometers on the Cap St. Andre (Boast and Nairn, 1982; Razafindrazaka et al., 1999). It is one of three Madagascan sedimentary basins formed as result of the breakup of Gondwanaland and the drifting of the Madagascar microplate along the Davie fracture zone and the proto Mozambique Channel, which subsequently separated the Madagascar-India block from the African continent (Geiger, 2004). The distribution of sediments is controlled by the Sahondralava-Ihopy horst, which formed different depression separated by highs and divided this basin into the eastern and northern Loza-Mahavays and the Mitsinjo sub-basins, respectively (Boast and Nairn, 1982). The Majunga Basin has been found to host a 10km thick successions of Late Paleozoic to Cenozoic sediments (Papini, 1995; Papini and Benvenuti, 1998). Toarcian strata are represented by the lowermost part of the Beronono Formation, which outcrops mostly in the south-western part of the basin (Geiger et al., 2004; Geiger and Schweigert, 2006). It is worth noting that prior to the denomination by Geiger et al. (2004), the Beronono Formation had previously been termed as Isalo-III or Upper Isalo Formation (Boast and Nairn, 1982). Furthermore, the Isalo terminology has been used interchangeably referring to sandstones from various depositional sequences ranging from Late Triassic to the Middle Jurassic (Boast and Nairn, 1982; Razafindrazaka et al., 1999;

Geiger and Schweigert, 2006), which can be misleading and confusing. To avoid this inconsistency, the Isalo, the Beronono, and its lateral equivalent the Andafia Formation are considered lithostratigraphically distinct in this review.

1.2.5.1 Isalo Formation (? Triassic)

The Isalo Formation unconformably lies on the Sakamena Group and is overlain by the Beronono Formation (Toarcian-Aalenian) (**Fig. 1.6**). It consists primarily of approximately 600m of uniform fine to coarse sandstone with minor conglomerate beds (Boast and Nairn, 1982; Razafindrazaka et al., 1999; Geiger et al., 2004). As previously stated, the Isalo Formation appears to be part of larger classification that includes the Isalo I up to IIIC units, spanning from the Late Triassic to Middle Jurassic times (Boast and Nairn, 1982; Hankel, 1992; Luger et al., 1994; Razafindrazaka et al., 1999). This review, however, adapts a Triassic age proposition (Geiger, 2004; Geiger et al., 2004), determined from fossilized trees in the Andafia (Beronono Equivalent) Formation. Therefore, the Isalo I-II units are continental Triassic in age and represent the true “Isalo Formation”. This formation commonly displays cross bedding stratification and was deposited in fluvial channel to braided stream complex (Geiger et al., 2004).

1.2.5.2 Beronono Formation (Toarcian-Aalenian to ? Lower Bajocian)

The Beronono Formation unconformably lies on top of the Isalo and is overlain by the Bemaraha Formation (**Fig. 1.6**). It is a lateral stratigraphic equivalent of the Andafia Formation in the Morondava Basin and it is subdivided into three members that reach a thickness of about 100m (Geiger et al., 2004; Krishna, 2017). The lower part is 35m thick and made up of shales and marls interbedded with thinly bedded sandy limestone. The middle section is made up by a 47m thick unit of sandstones and shales-marls with thin interbeds of limestones. The upper part consists of a 20m thick fossiliferous shale with bioclastic limestone interbeds (Geiger and Schweigert, 2006). In general, limestones are fossiliferous and show coarsening upward in all sections, whereas shales and marls are almost barren of fossils. Ammonites (*Bouleiceras nitescens*), calcareous

foraminifera, ostracods, bivalves, gastropods, echinoderms, and algae are among the important Early Toarcian-Aalenian age fossils found in this formation (Collignon et al., 1959; Buffetaut et al., 1981; Geiger and Schweigert, 2006). Facies analysis indicates a diverse depositional environment ranging from fluvial deltaic, near shore shelf to shore face, shallow to open marine with periodical storm deposits (Geiger, 2004; Geiger et al., 2004).

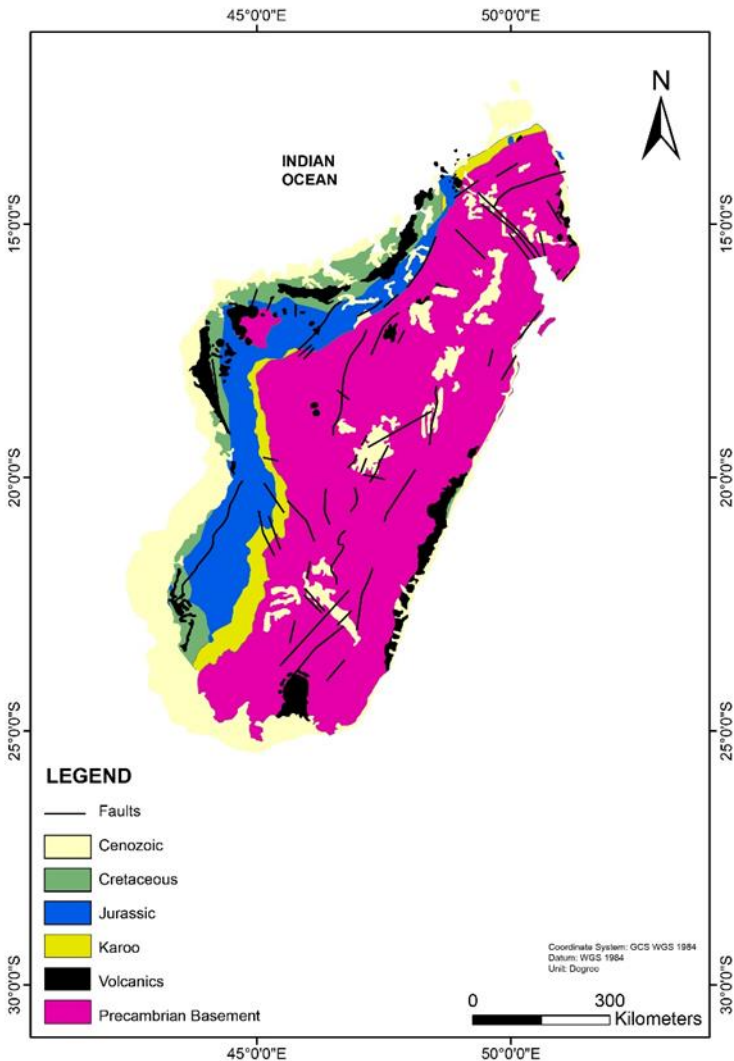


Fig. 1.9: Geological map of Madagascar depicting Karoo, Jurassic including Toarcian and post Jurassic sediments. Redrawn and adapted from Schlüter and Trauth (2006).

1.2.6 Morondava Basin (SW-Madagascar)

The Morondava Basin is located in the eastern part of Mozambique (**Figs. 1.3 and 1.9**). It extends approximately 1000 kilometers along the western coast, to the Cap St. Andre on the northern coast and is bound by Precambrian Basement on the East (**Fig. 1.9**). Furthermore, it covers the coastal plain and continental shelf areas, making it Madagascar's largest basin (**Fig. 1.9**) (Boast and Nairn, 1982; Coffin and Rabinowitz, 1988; Geiger et al., 2004; Harris et al., 2006). The basin evolution is linked to Madagascar's separation from Africa along the Davie transform fault and ridge zone, which initiated the formation of a transform margin filled with an approximately 11km thick sediment package of Late Paleozoic to Cenozoic age (Harris et al., 2006; Giese et al., 2012). Facies association and basin geometry indicates that deposition was fault-controlled, which allowed for marine influence in down-warp regions (Harris et al., 2006). Toarcian strata is corresponding to the Andafia Formation forming a lateral equivalent to the Beronono Formation of the Majunga Basin. The Andafia Formation shows poor outcropping, and thus there is no complete Toarcian sedimentary section to be encountered in Morondava Basin (Boast and Nairn, 1982; Coffin and Rabinowitz, 1988; Geiger et al., 2004).

1.2.6.1 Andafia Formation (Toarcian-Aalenian)

The Andafia Formation lies unconformably on the Isalo Formation and is overlain by the Bemaraha Formation. It was previously termed as the upper part of Isalo Group. Strata mainly consist of shale and marls, with minor interbeds of limestones and thin siltstones/fine sandstones (Geiger and Schweigert, 2006). The thickness of the formation varies greatly in both the surface and the subsurface. Up to 600m thick sedimentary successions have been found in the subsurface via drilling. In the central and northern parts, only the lower successions consisting of marl, shale, and limestone are exposed (Geiger et al., 2004). The Geology of the Andafia Formation in the central and northern parts is comparable to that of Beronono Formation in the Majunga Basin described above (**Section 1.2.5**).

1.2.7 Mandawa Basin (SE-Tanzania)

The 15,000km² large onshore part of the Mandawa Basin is located along Tanzania's Southeast coast (**Figs. 1.3 and 1.10**). It is bound to the South by the Ruvuma basement saddle, to the West by the Masasi basement spur, and to the north by the Rufiji trough (**Fig. 1.10**) (Alconsult and TPDC, 1997; TPDC, 2003; Stratadata, 2006; Hudson and Nicholas, 2014). It contains the only drilled evaporite basin in East Africa, with the most complete Jurassic succession (Kent et al., 1971; Alconsult and TPDC, 1997; TPDC, 2003). The Mandawa Basin is one of the failed rift systems in East Africa that was initiated during early stages of the breakup of Gondwana (Balduzz et al., 1992; Veeken and Titov, 1996; Einvik-Heitmann, 2016; Fossum et al., 2018). Approximately 4-8km of Mesozoic and Cenozoic sediments have been deposited in this basin (Veeken and Titov, 1996). Toarcian strata are included within the Pindirolu Group (Pliensbachian-Aalenian), which contains the Mbuo Formation (?Pliensbachian), the Nondwa Formation (? Early Toarcian to Middle Toarcian), and Mihambia Formation (? Late Toarcian to Aalenian) (**Fig. 1.6**). An older age assignment for the Mbuo and base of the Nondwa Formation has been postulated previously (Kent et al., 1971; Hudson, 2010; Hudson and Nicholas, 2014).

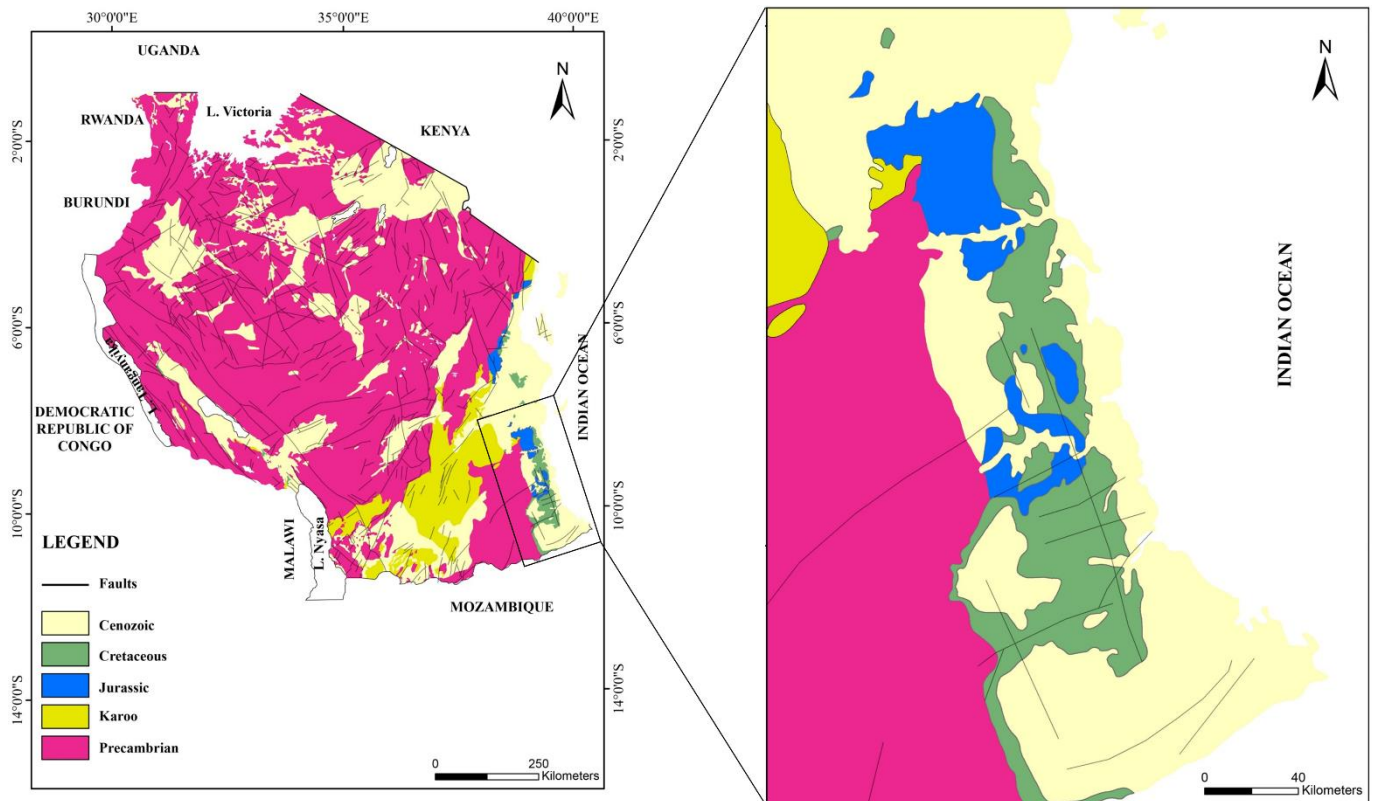


Fig. 1.10 : Geological map of Tanzania, depicting Pre-Jurassic, Jurassic including Toarcian and post Jurassic sediment. Adapted from (Pinna et al., 2008; Hudson, 2010; Barth et al., 2016).

1.2.7.1 The Mbuo Formation (Pliensbachian)

The Mbuo Formation rests unconformably on Precambrian basement and is overlain by the Nondwa Formation (**Fig. 1.6**) (Alconsult and TPDC, 1997; Stratadata, 2006; Quinton and Copestake, 2007; Corelab and TPDC, 2009; Hudson and Nicholas, 2014). The formation has only been encountered in the subsurface, with thicknesses ranging from 140m to 1770m. The entire succession is mostly made up of thinly interbedded shales, siltstones and coarser to conglomeritic sandstones, with occasional thin limestones beds (Corelab and TPDC, 2009; Hudson and Nicholas, 2014; Robertson and CGG, 2017; McCabe, 2021). It is subdivided into a lower and an upper member, whereby the lower member consists of sandstones with rare silt-claystones while the upper member is mainly made up by thick claystones with alternating thin dolomite beds (Hudson and Nicholas, 2014). To date, there still is a major disagreement on the age of this formation with Hudson and Nicholas (2014) or Barth et al. (2016) suggesting a Late Triassic age based on

palynology. This contrasts with other age assignments (Balduzz et al., 1992; Veeken and Titov, 1996; Stratadata, 2006; Quinton and Copestake, 2007; Corelab and TPDC, 2009; Robertson and CGG, 2017; Macgregor, 2018) proposing a Pliensbachian age based on both palynological and nannofossils data. This review concurs with the latter age assignment and further argues that an older age assignment might have been drawn based on reworked Triassic sediment. Lithofacies and biofacies associations indicate a wide range of depositional environments from alluvial, to fluvial and lacustrine (Hudson, 2010; Hudson and Nicholas, 2014).

1.2.7.2 Nondwa Formation (Lower Toarcian to Middle Toarcian)

The Nondwa Formation lies unconformably on the Mbuo Formation (**Figs. 1.5c and 1.6**) and is overlain by the Mihambia Formation. It comprises one of the oldest outcropping sedimentary units in the Mandawa Basin (**Fig. 1.10**). Strata are primarily made up by thick blocky units of halite interbedded with thinly interbedded anhydrite, shales and carbonates (Stratadata, 2006; Quinton and Copestake, 2007; Corelab and TPDC, 2009). The true thickness of this formation is difficult to establish at the surface outcrops due to salt diapirism (**Fig. 1.5c**), and varies greatly in the subsurface from 58m (MitaGamma-1 well) to 2802m (Mandawa-7 well) (Quinton and Copestake, 2007; Corelab and TPDC, 2009, 2009; Hudson, 2010; Hudson and Nicholas, 2014). Important Early Toarcian to Middle Toarcian calcareous nannofossils (**Table 1.1**) have been reported from drilling (Mita-Gamma, Mbuo, Mbate, East-Lika wells) (Karega, 2001; Quinton and Copestake, 2007; Corelab and TPDC, 2009; Robertson and CGG, 2017). However, a Late Triassic age was assigned to the base of the Nondwa Formation (Kent et al., 1971; Kent, 1972, 1974; Coffin and Rabinowitz, 1988; Wopfner, 1994; Kagya, 1996; Mpanda, 1997; Kapilima, 2003b; Hudson, 2010; Hudson and Nicholas, 2014; Riding, 2019). Nevertheless, this review argues that the unconformity at the base of the Nondwa Formation, which is found across all drilled wells (**Supplement-1.1**) is likely linked to an early Toarcian rifting event associated with a transgressive phase. The Nondwa facies varies from a shallow lacustrine or coastal lagoonal settings to restricted marine depositional environments formed in an arid climate (Hudson, 2010; Hudson and Nicholas, 2014).

1.2.7.3 Mihambia Formation (Upper Toarcian-Aalenian)

The Mihambia Formation lies on the Nondwa Formation and is overlain by the Mtumbei Formation (Stratadata, 2006; Quinton and Copestake, 2007; Corelab and TPDC, 2009; Hudson, 2010; Hudson and Nicholas, 2014). The entire succession consists of thinly interbedded sandstones and shales with minor carbonates (Alconsult and TPDC, 1997; Quinton and Copestake, 2007; Corelab and TPDC, 2009). Successions of the Mihambia Formation occur in the subsurface with variable thicknesses ranging from 247m to 598m (Hudson and Nicholas, 2014). The Mihambia Formation is rich in diverse microfauna and palynomorphs of Late Toarcian to Aalenian age (Corelab and TPDC, 2009; Hudson, 2010; Hudson and Nicholas, 2014). Litho- and biofacies indicate a shallow and marginal marine depositional environment (Corelab and TPDC, 2009; Hudson, 2010; Hudson and Nicholas, 2014).

1.3. Discussion

Lower Toarcian marine strata of the east African Margin were deposited following a prolonged period of rifting associated with deposition of continental sediments with only moderate marine contributions in graben or depression structures (Cannon et al., 1980; Coffin and Rabinowitz, 1988; Coffin, 1992; Luger et al., 1994; Kreuser, 1995; Mbede and Dualeh, 1997; Mpanda, 1997; Geiger, 2004; Schlüter and Trauth, 2006; Papini and Benvenuti, 2008; Macgregor, 2018). This Toarcian rifting event controlled sedimentation patterns and was accompanied by multiple phases of reactivation (Mpanda, 1997; Schandelmeier et al., 2004; Catuneanu et al., 2005; Papini and Benvenuti, 2008; Corelab and TPDC, 2009; Macgregor, 2015, 2018). Continental sedimentation in the rift grabens (**Figs. 1.4 and 1.11**) most likely continued until 183Ma, when rifting coincided with a major rise of the sea level, and the depositional regime shifted towards shallow/marginal marine conditions in eastern Africa, while widespread volcanism occurred in southern Africa (Svensen et al., 2007) (**Fig. 1.13**). These changes are linked to extensional stress regimes along the Tethyan margin of the Gondwana supercontinent (Boast and Nairn, 1982; Luger et al., 1994; Kreuser, 1995; Geiger, 2004; Catuneanu et al., 2005; Papini and Benvenuti, 2008).

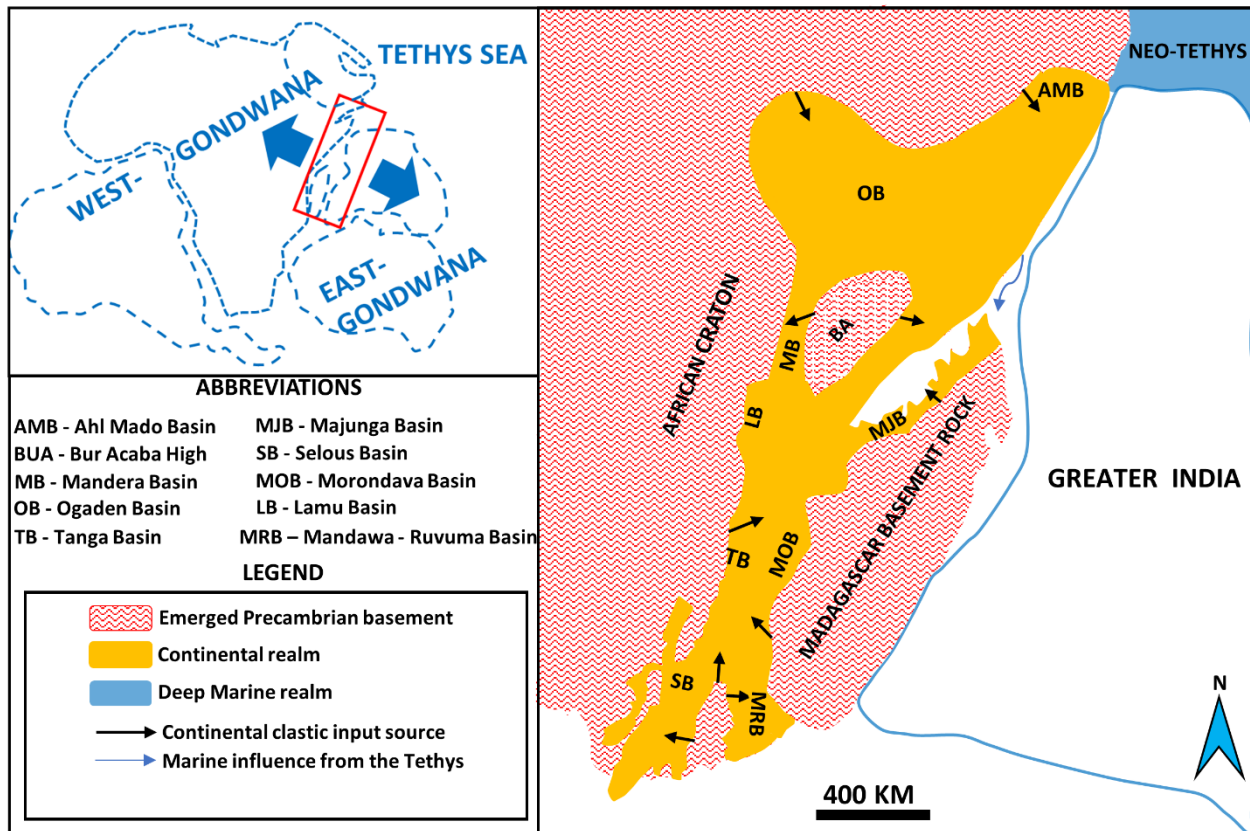


Fig. 1.11: Schematic late Triassic-Pliensbachian paleogeography of the East African Margin.

Rifting during the earliest Toarcian coincided with a global rise of sea level (Haq, 2018), and as a result, a narrow and shallow seaway that had been created since the Karoo rifting (Wopfner, 1994) continued to be enlarged and to form an embayment or gulf that extended southward from the Arabian Peninsula (Yemen), along northern to southern Somalia, into other parts of NE Kenya, Coastal Tanzania and North Madagascar, the so-called “Somali Ocean”. The southernmost extension of this arm of the Neotethys ended in the Mandawa-Ruvuma Basins in Tanzania, and its counterpart basins in Madagascar (**Figs. 1.12 and 1.13**). It is assumed that there was no further southward marine invasion presumably due to plume related uplift south of the present-day Davie Ridge fracture zone that could have acted as a morphological barrier (Geiger et al., 2004, 2006). Influxes of marine ingressions were sporadic due to differential rates of subsidence and uplift in the various rift grabens and depressions and hence, few areas were reached by this extension of the Tethyan sea (Arkell, 1956; Kent, 1972, 1974; Cannon et al., 1980; Riccardi, 1991; Buscaglione et al., 1993; Mette, 1993; Luger et al., 1994; Geiger, 2004; Schandelmeier et al., 2004;

Papini and Benvenuti, 2008). As result, the existing depressions or troughs along the East African Margin were inundated by marine incursions from the Tethyan Sea and this marked the first deposition of marine Toarcian strata. Paleo-current data indicate that influxes of marine waters occurred along NW-SE, NE-SW and N-S directions, which complies with the drainage pattern of the Karoo depression (Cannon et al., 1980; Buscaglione et al., 1993; Mette, 1993; Geiger et al., 2004; Papini and Benvenuti, 2008).

Marine invasion into shallow regions along the East African Margin via the horn of Africa utilized either the Somali-Madagascan (northwest of the Madagascar microplate) or Indo-Madagascans (southeast of the Madagascar microplate) seaways (**Fig. 1.12**) (Angelucci et al., 1981; Boccaletti et al., 1988; Coffin and Rabinowitz, 1988; Petters, 1991; Buscaglione et al., 1993; Mette, 1993; Schlüter and Trauth, 2006; Papini and Benvenuti, 2008; Wolela, 2008; John, 2016). Both migrations routes along the Trans Somalia-Madagascar and Indo-Madagascan seaways (NE-SE to SW) have been proposed in paleogeographical reconstructions (Dercourt et al., 2000; Scotese, 2014).

The existing depressions in northeastern Somalia (the Ahl-Mado and Ogaden Basins) were flooded via the Somali-Madagascan seaway (**Fig. 1.12**). Further to the South, an area covering the present-day Somali embayment and Somalia coastal basins was indented and this sea-arm continued further southwest into the present-day Mandera Basin (also termed Luuq Basin) in SW Somalia (**Fig. 1.12**). Alternative ingress pathways of the Trans Somali-Madagascar seaway via the Majunga Basin, separated from the Mandera Basin by the Bur Acaba paleohigh are more plausible (**Fig. 1.12**) because the northwesterly ingress path via the Mandera Basin would have to overcome the southwestern edge of the paleohigh, which is not considered to be feasible. Influxes of marine ingress in Somalia were limited by the Nogel uplift in the north and the uplifted Precambrian Bur Acaba basement high in the south (**Fig. 1.12**). The distribution of shallow and deeper facies (**Figs. 1.2 and 1.12**) was linked to bathymetric variation (Boccaletti et al., 1988; Buscaglione et al., 1993) in the intensively structured seaways.

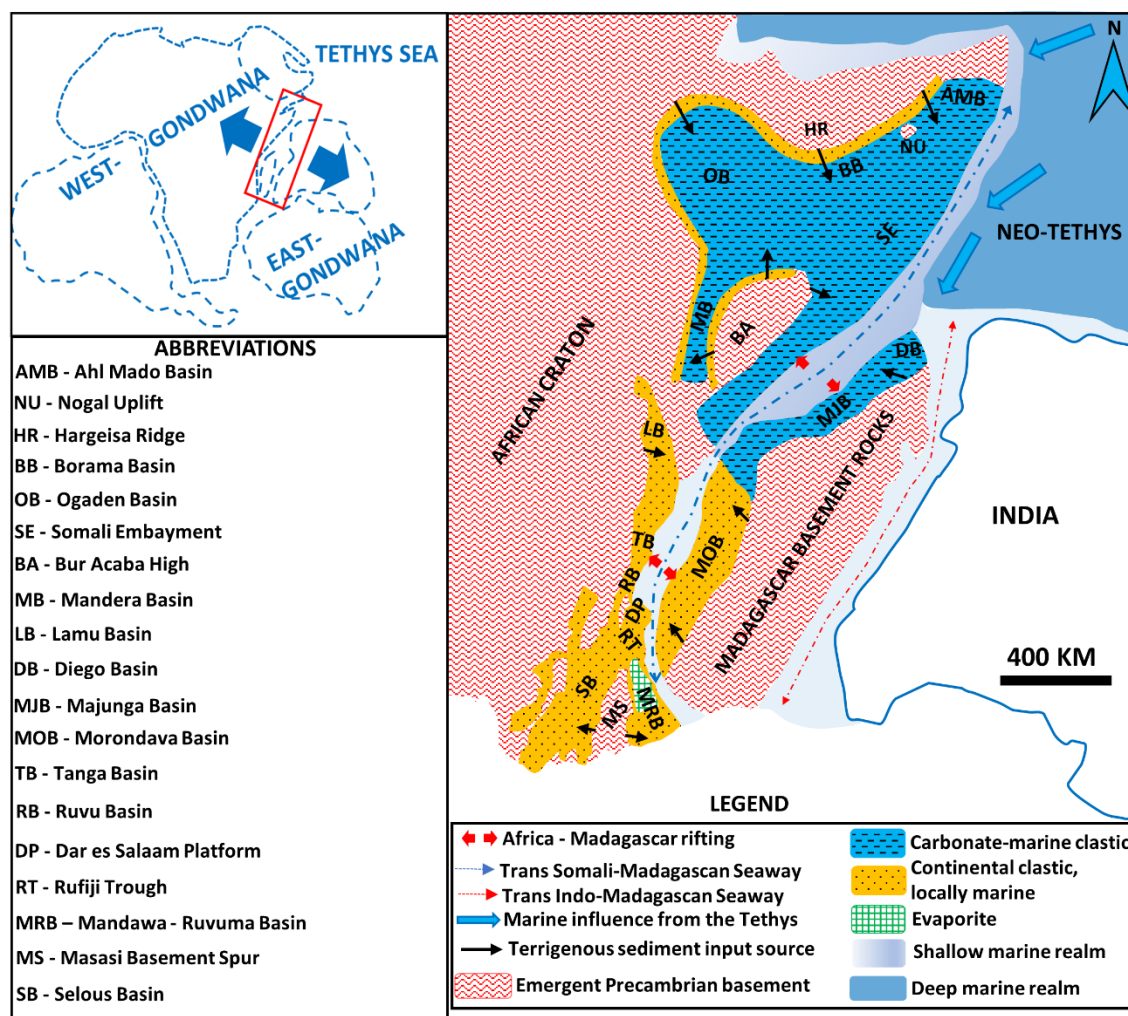


Fig. 1.12: Schematic early Toarcian (~183ma) paleogeography of East African Margin.

In NE Kenya, the Toarcian sea invaded the southern tip of the Manderia Basin via its adjacent part in Somalia and the Manderia Basin extended further south to the present day areas of the Kwanza rift (Ayers, 1952; Arkell, 1956; Thompson and Dodson, 1960; Joubert, 1963; Geiger, 2004; Sproule, 2017). Both southward and eastward pathways into the Manderia Basin are plausible, and are compatible with proposed paleogeographical reconstructions (Scotese, 2014). The sea arm of the Manderia Basin was not connected to the present day Lamu Basin, most likely due to intensive faulting that had taken place since late Triassic and succeeded during the Toarcian. Faulting created local barriers to marine ingression and thus deposition of the continental Mazeras sandstone continued (Cannon et al., 1980; Coffin and Rabinowitz, 1988; Rais-Assa, 1988; Coffin,

1992) (**Fig. 1.4**). This proposition concurs with the Early Jurassic structural time series maps of (Macgregor, 2018) and furthermore, apatite fission track data (Foster and Gleadow, 1993). It is presumed that the faulted and uplifted area acted as source of clastic materials into the Lamu Basin and hence deposition of the Mazeras sandstone continued during the Toarcian (Ayers, 1952; Caswell, 1956; Joubert, 1963; Kamen-Kaye, 1978; Cannon et al., 1980; Rais-Assa, 1988; Hankel, 1994; Papini, 1995; Mbede and Dualeh, 1997; Papini and Benvenuti, 2008).

Marine ingressions from the North invaded into the Majunga Basin via the northern Diego Basin during the early Toarcian and extended further south into the North of the Morondava Basin, whereas the central and southern part of the Morondava Basin was not invaded until the late Toarcian (Kamen-Kaye, 1978; Boast and Nairn, 1982; Luger et al., 1994; Geiger, 2004; Geiger et al., 2004; Geiger and Schweigert, 2006; Schlüter and Trauth, 2006; Papini and Benvenuti, 2008). Facies distribution indicates a relatively higher subsidence rate in the southern part of the Majunga Basin that allowed marine ingressions from the southern direction via the Indo-Madagascan seaway causing deposition of carbonatic facies (**Supplement 1.1; Fig. 1.12**). In contrast, the northern part of the Morondava Basin was partly emerged and hence dominated by continental-marine mixed facies (**Supplement 1.1; Fig. 1.12**). Both the southward and the northward marine ingressions pathways into the Morondava Basin are plausible and in agreement with regional paleogeographical reconstructions (Dercourt et al., 2000; Scotese, 2014).

In Tanzanian coastal basins, there still is a controversy regarding the timing of Toarcian marine invasion and pathways and no consensus has been reached until today. Based on ammonoid distribution and stratigraphy, many scholars (Arkell, 1956; Kent, 1972; Kamen-Kaye, 1978; Cannon et al., 1980; Boast and Nairn, 1982; Kapilima, 1984, 2003a, 2003b; Petters, 1991; Riccardi, 1991; Luger et al., 1994; Mbede and Dualeh, 1997; Chiocchini et al., 2005) proposed marine ingressions did not invade the present-day Tanzanian coastal basins until the Late Aalenian. This proposition, however, fails to account for the thick marine evaporites of the Nondwa Formation in the Mandawa Basin. Deposition of the thick Nondwa evaporites requires both periodical marine connectivity and relatively high subsidence (Mpanda, 1997; Kapilima, 2003b, 2003a; TPDC, 2003). Recent exploration drilling has enabled more biostratigraphy work to be undertaken from deeper wells (**Supplement table 1.1, Supplement-1.2**) and refinement of both

paleogeography and proposition of marine pathways in the Tanzanian coastal sedimentary basins. Important biostratigraphy information used in this study includes (Sissingh et al., 1989; Karega, 2001; Stratadata, 2006; Corelab and TPDC, 2009; Robertson and CGG, 2017). Therefore, in this review we propose that the earliest Toarcian marine incursions invaded the Tanzanian coastal sedimentary basins from the North via the Tanga-Ruvu Basins, the Dar es Salaam platform, and Rufiji Trough and continued southward into the Mandawa-Ruvuma Basin (**Fig. 1.12**). This proposition is supported by paleogeographical reconstructions of Scotese (2014) who favors the same southward pathways. Extent of marine invasion in Mandawa Basin was limited by the Tanga fault in the North, the E-W-striking Utete-Tagalala fault in the Rufiji Trough and the Lindi fault (Kent et al., 1971; Mbede, 1991; Mbede and Dualeh, 1997; Mpanda, 1997; Kapilima, 2003b, 2003a; TPDC, 2003). An alternative access of marine incursion via the Indo-Madagascan sea pathway as proposed by Scotese (2014) would require a continuous southward extension of an Neotethys arm along the eastern margin of the Madagascan microplate into the present-day areas of North Mozambique and Southern Africa, a scenario for which sedimentological evidence currently is unavailable.

By considering published information (Riccardi, 1991; Dercourt et al., 2000; Scotese, 2014), and (Ayers, 1952; Arkell, 1956; Caswell, 1956; Thompson and Dodson, 1960; Joubert, 1963; Kent, 1972, 1974; Kamen-Kaye, 1978; Cannon et al., 1980; Angelucci et al., 1981; Canuti et al., 1983; Bosellini, 1992; Buscaglione et al., 1993; Hankel, 1994; Luger et al., 1994; Papini, 1995; Kapilima, 2003a, 2003b; Geiger, 2004; Schandelmeier et al., 2004; Chiocchini et al., 2005; Papini and Benvenuti, 2008; Sproule, 2017), the paleogeography of earliest Toarcian strata was revisited and refined. Our paleogeographical proposition depicts an extension of Neotethys arms into the present-day East African Margin via the Ahl-Mado and Mandera Basins in Somalia, the Mandera and Lamu Basins in Kenya, the Majunga and Morondava Basins in Madagascar, and the Mandawa Basin in Tanzania. This complex and shallow seaway controlled the distribution of Toarcian shallow, marginal, restricted marine to mixed facies successions that overly the Karoo continental strata of the East African Margin. It is well supported by the affinities and distribution of the various species of *Bouleiceras* genus (*Arabicum*, *elegans*, *rectum*, *nitescens*, *tumidum*, *ibericum*) ammonoid fauna and Early Jurassic calcareous nannofossils that are diagnostic features of

Toarcian strata (**Table. 1**). During Early Jurassic times, the entire East African Margin experienced differential crustal thinning initiated by the breakup of East and West Gondwana (**Fig. 1.13**). This lead to a differential uplift and subsidence along the East African Margin that in the early Toarcian was accompanied by a global rise in sea level of about 60 to 80m (Haq, 2018). This sea level rise allowed for ingression of Tethyan waters and marine sediment was deposited in the most subsidized part of continental rift grabens and depressions. Emerged landmasses sourced clastic materials into these basins. It is worth noting that in Tanzania, Madagascar, and SE Kenya this stratum by many scholars previously was considered to be part of the Karoo sediment cycle. In this study we propose that the Lower Toarcian strata is not part of Karoo system but rather represents rejuvenated Karoo grabens and depressions filled by marine sediment upon the Toarcian transgression and sea level highstand.

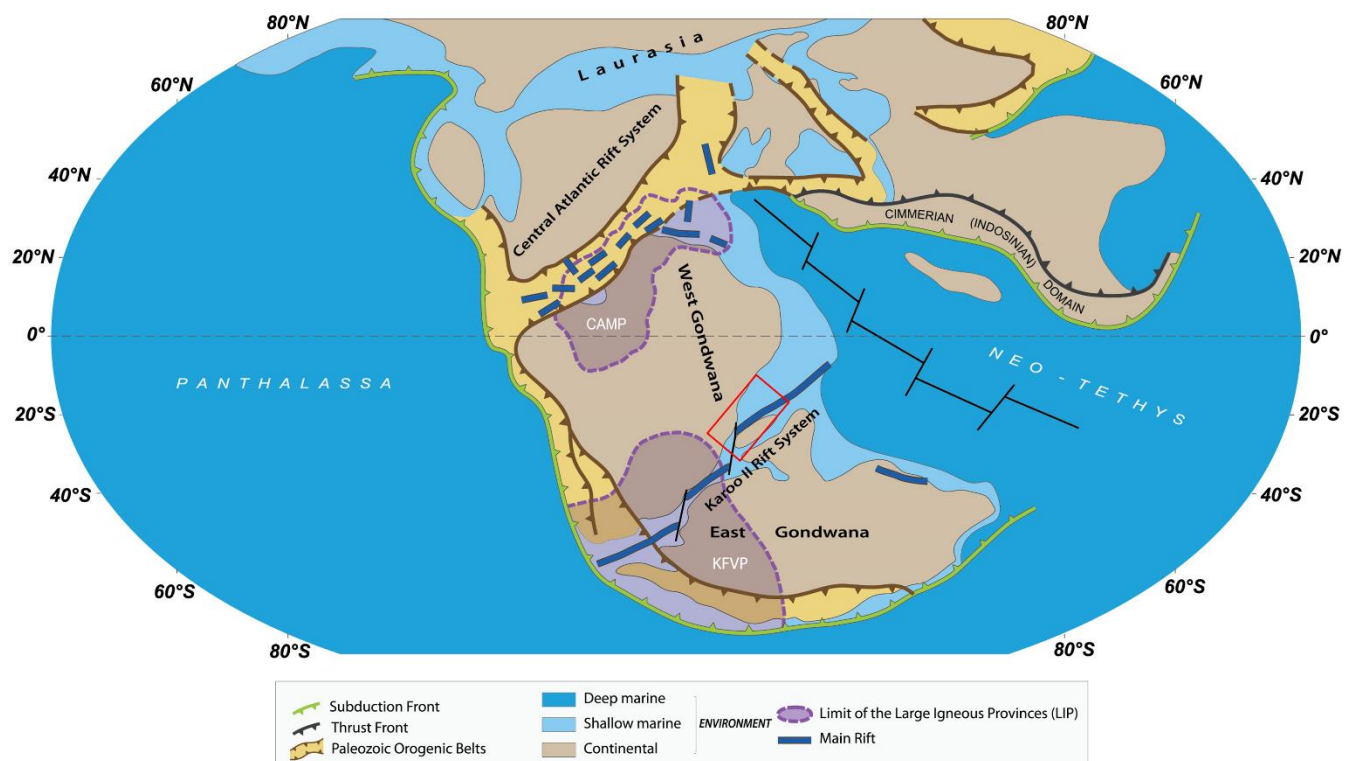


Fig. 1.13: Toarcian (~ 183.1 Ma) global palaeogeographical map adapted from Frizon de Lamotte et al. (2015) and Scotese (2014). Red box demarcates the location of the East African Margin discussed here.

early Toarcian flooding of depressions/grabens (marine ingression) in the East Africa margin due to eustatic rise in sea level was followed by a long-term fall in sea level and tectonic subsidence in the late Toarcian-earliest Aalenian (Kamen-Kaye, 1978; Boast and Nairn, 1982; Coffin and Rabinowitz, 1988; Luger et al., 1994; Papini and Benvenuti, 2008; Haq, 2018). These processes resulted in a drastic change in depositional conditions and paleogeography, some areas transitioned from shallow-open marine to more inner shelf-/platform carbonate environments, while in others, lagoonal/restricted marine conditions evolved into shallow marine conditions (**Fig. 1.14**). Variations in depositional environments along the margin probably resulted from differential subsidence and sedimentation rates (Bosellini, 1992; Luger et al., 1994; Mpanda, 1997; Ali Kassim et al., 2001; Kapilima, 2003b; Geiger et al., 2004; Geiger and Schweigert, 2006). A combination of shallow waters and high energy sedimentation resulted in general carbonate shelf deposition conditions in the Ahl-Mado Basins of Somalia, and the Mandera Basin of Kenya Somalia, and the Majunga Basin of Madagascar (Ali Kassim et al., 1987, 2001; Buscaglione et al., 1993; Fantozzi and Ali Kassim, 2002; Geiger et al., 2004). Due to an uplift of hinterland caused by rifting, predominantly continental depositional conditions persisted in Lamu and in the Central and South of the Morondava Basins. (**Fig. 1.13**). The change in tectonic subsidence, local climate, sea level, and sedimentation rate in the Mandawa Basin, and possibly the Ruvuma Basin, is likely to have led to the transition from restricted marine deposits (Nondwa Formation) to clastic shallow marine depositional conditions (Mihambia Formation) (Kajato, 1982; Kapilima, 2003b; Hudson and Nicholas, 2014). A continental - marine transitional depositional condition (Ngerengere beds) was sustained in the Central and Northern parts of the Tanzanian coastal margin owing to slow subsidence rates (Kajato, 1982; Kreuser, 1995; Mpanda, 1997; Kapilima, 2003b). Deposition in the continental to shallow/marginal marine continued across various parts of the East African Margin until the Bajocian-Bathonian when the Somali Basin opened up and open marine depositional conditions were established.

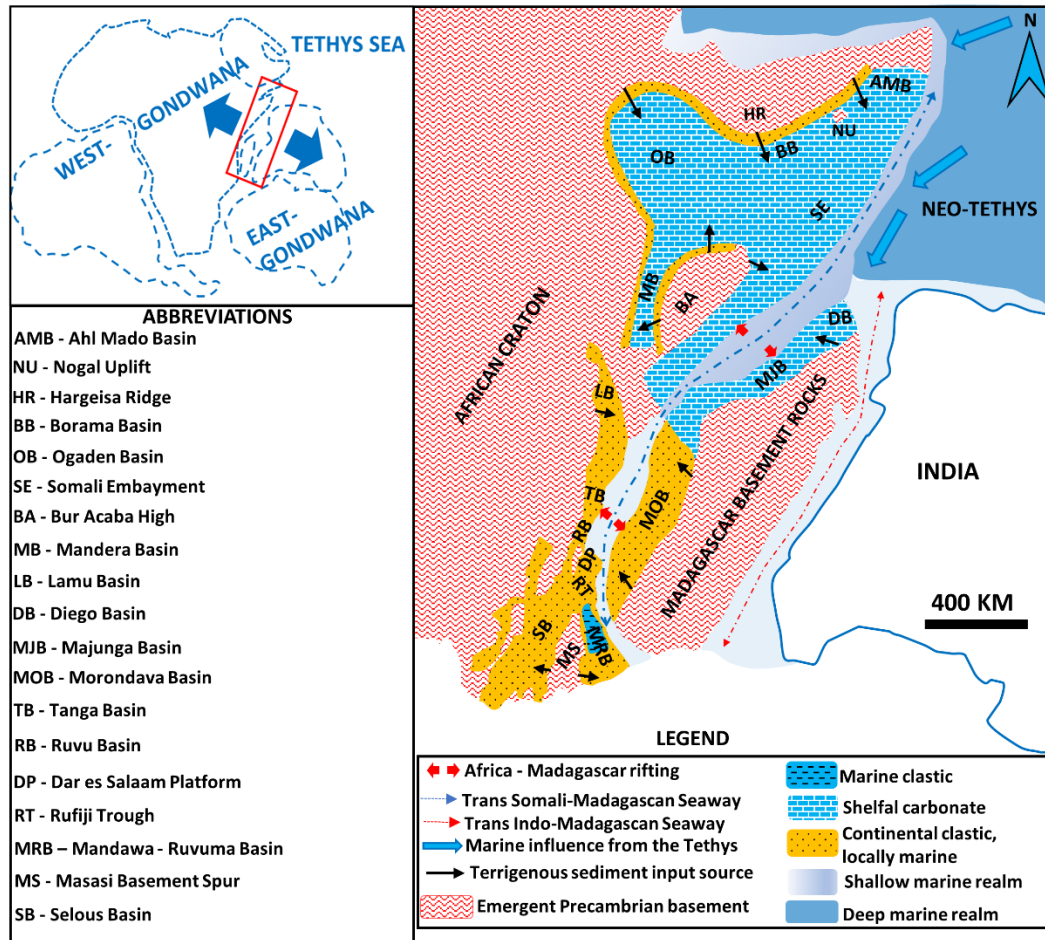


Fig. 1.14: Schematic (Late Toarcian-Early Aalenian) paleogeography of the East African Margin.

1.3.1 Implications on lithofacies, stratigraphy and tectonics

All seven sub-basins investigated in this study generally show major gradual facies change from continental to marine conditions during early Toarcian times (**Figs. 1.6 and 1.13**). Similarities in lithofacies and stratigraphy probably reflect related regional rifting events linked to the separation of East and West Gondwana (**Figs. 1.6 and 1.13**). During the Late Triassic to Early Liassic, parts of the East African Margin were subject to regional uplift and denudation (Petters, 1991; Foster and Gleadow, 1993; Hankel, 1994; Delvaux, 2001; Wolela, 2008). These events were accompanied by down-warping (Boccaletti et al., 1988; Petters, 1991) and the creation of subsiding zones separated by structural highs such as Bur Acaba and Nogal high in Somalia, the

Pande and Masasi basement high in southern Tanzania, the Sahondralava-Ihopy ridge, Cap St. Andre ridge, and Bekodoka massifs in Madagascar. It is envisaged that up to 2000m of overburden were removed upon uplift (Foster and Gleadow, 1993; Delvaux, 2001). The beginning of clastic accumulation was probably diachronous due to differential rifting and relief of the various regions along the East African Margin (Bosellini, 1992). These continental deposits commonly overly basement and generally display fining upward megacycles characterized by mixtures of alluvial fan, meandering and lacustrine depositional settings. Changes in depositional styles are either due to variation in paleoclimate and decrease in source area relief linked to the regional tectonic regime (Kapilima, 1984; Hankel, 1987, 1994; Mbede, 1991; Petters, 1991; Kreuser, 1995; Wolela, 2008). Deposition of the Early Liassic Adigrat Formation and its laterally equivalent strata in East African and Arabian sub-basins (**Fig. 1.15**) probably marked the end of this regional peneplanation and the beginning of new depositional phases (Kent et al., 1971; Petters, 1991; Hankel, 1994; Wolela, 2008).

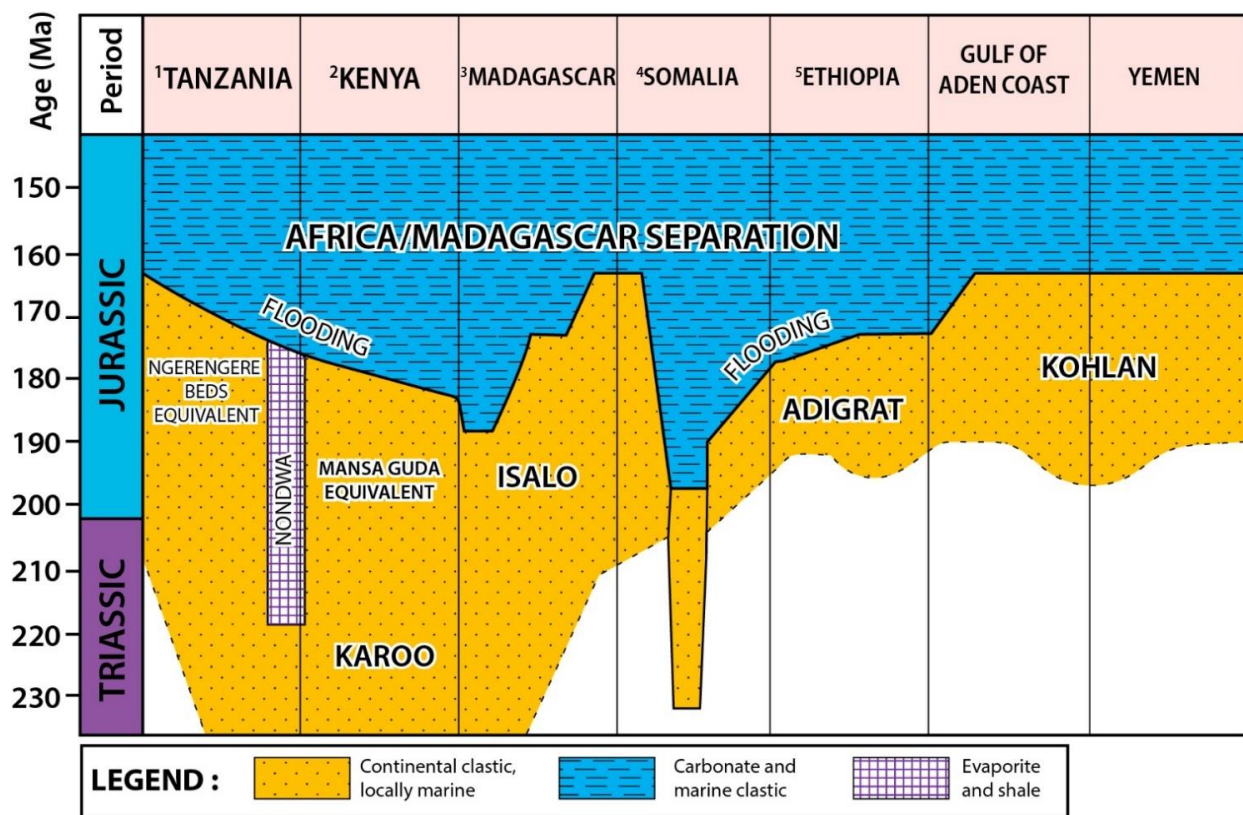


Fig. 1.15: Chronostratigraphy of East African and adjacent Arabian basins (1. Tanzanian coastal basins, 2. Mendera and Lamu, 3. Morondava and Majunga, 4. Mendera-Luuq, 5. Ogaden Basins and adjacent areas) depicting pre-

Jurassic clastics, continental Jurassic graben fills, and marine Jurassic deposition in the course of the Toarcian transgression. Redrawn from Bosellini (1992).

A rifting phase accompanied by an eustatic rise in sea level continued progressively into the early Toarcian leading to the down-warping and flooding of the pre-existing grabens (Kent et al., 1971; Cannon et al., 1980; Rais-Assa, 1988; Buscaglione et al., 1993; Papini and Benvenuti, 2008) (**Fig. 1.15**). Deposition of Toarcian marine strata commenced at the same time, whereby variation of Toarcian lithofacies (**Fig. 1.6**) indicates differences in environmental of deposition and subsidence (Mbede, 1991; Buscaglione et al., 1993; Mpanda, 1997; Kapilima, 2003b; Hudson and Nicholas, 2014). Examination of Toarcian lithofacies in the region indicates maximum subsidence was attained in basins of southern coastal Somalia and southern Tanzania (Mandawa) as well as in the South of the Majunga Basin (Angelucci et al., 1981; Boccaletti et al., 1988; Coffin and Rabinowitz, 1988; Mbede and Dualeh, 1997; Hudson and Nicholas, 2014).

1.3.2 Implications for Petroleum Systems

The Toarcian transgressive phase was coupled to a global oceanic anoxic event and widely distributed deposits of thick organic rich sediments potentially serving as hydrocarbon source rocks (**Supplement table 1.2**). Both basal sandstone and post Toarcian carbonates and sandstones possesses high quality reservoir properties to store hydrocarbons generated from Toarcian source rocks (**Supplement table 1.2**). The overlying widely distributed Bajocian-Bathonian marls/shales or evaporites could act as effective seal or cap rock. Both older and younger fault structures promote feasible migration pathways and tectonic trapping. Whereas the petroleum potential in off-shore regions of Tanzania is already exploited, the on-shore reserves along the East African Margin are still untapped and may warrant further exploration.

1.4. Conclusions

Evolution of Toarcian strata in the East African coastal margin is linked to the rifting process that lead to the breakup of East and West Gondwana and a coeval global rise in sea level. Breakup of

Gondwana initiated during Karoo times did not advance into oceanization stages but rather produced graben-like depression that were filled by continental Karoo sediment. Further rifting during the early Toarcian was accompanied by transgression, reactivated older Karoo grabens, and permitted marine ingressions from the North via the Horn Africa. Thus, a narrow seaway, strait or gulf formed between Africa and Madagascar, extending Neotethys sea-arms southwards and depositing Toarcian marine strata along the East African Margin. Due to differential uplift, subsidence, erosion rate, and bathymetry, the Toarcian lithofacies vary greatly from shallow marine carbonates and clastics to shallow marine evaporites in the East African Margin to volcanics in Mozambique and South Africa. Shallow marine facies were dominant in areas where there were no barriers restricting marine ingression such as in the Mandera-Luuq and Ahl-Mado Basins in Somalia or the Mandera Basin in NE Kenya. Continental clastic deposition was common in uplifted parts that acted as barriers to marine ingression such as the present-day Lamu embayment. Mixed facies were predominantly deposited in slow subsiding basins with periodical marine incursion, as for example in the North Majunga and Morondava Basins in Madagascar, the Tanga-Ruvu Basin, the Rufiji Trough, and Dar es Salaam platform in Tanzania. Evaporites developed in a restricted marine embayment and under relatively high subsidence rates, as in the Mandawa Basin, and probably offshore in the Somalia and Morondava Basins. Flood basalts and volcanoclastic facies were dominant in the southernmost part of Africa, where a mantle plume was located. Two main marine pathways that were limited by older faults systems and emerged basement high structures have been proposed in this study.

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Chapter Two

Lithofacies of early Jurassic successions derived from spectral gamma ray logging in the Mandawa Basin, SE Tanzania

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Abstract

Spectral Gamma Ray log (SGRL) data has been widely employed to derive subsurface lithological information and to perform well correlation when core is not available. Previous studies mainly targeted homogenous facies whereas few attempts addressed heterolithic successions. The SGRL investigation here aimed at deriving lithofacies and further constrain conditions of deposition prevailing during the Early Jurassic (Mbuo, Nondwa and Mihambia Formations) in the Mandawa Basin, onshore Tanzania, where core and outcrop data is rare. By use of binary discrimination diagrams of U, Th, K, composition fields were defined for various lithofacies types (black shale, red shale, siltstone, sandstone, chert, anhydrite, halite, and carbonate) compiled from literature data. Sediment composition for two wells in the Mandawa Basin was then deduced from cuttings and SGRL-data and compared with literature-derived geochemical provenance data from adjacent regions and strata. Binary discrimination diagrams revealed the presence of multi-mineral components in the heterolithic facies type, which in concert with diagenetic and detrital mineral dilution effects, led to low compositional contrasts between provenance fields and to overlap of different lithofacies types, complicating interpretation and limiting effectiveness of the SGRL-based facies differentiation. For the Early Jurassic strata, an upward increase in the Th/K ratio suggests an enhanced terrigenous influx into the basin due to intensified weathering. High Th/U and Th/K in the study region indicate a relatively shallow, well mixed and oxic depositional

setting in proximity to the source area opposing the more distal, deeper and reduced depositional settings that prevailed in the southern central part of the basin. Th/U and Th/K excursions in the Nondwa and Mihambia Formations suggest a shift towards less oxic depositional conditions upon two major marine incursions, corroborated by a decline in the clastic fraction and concomitant increase in carbonate.

2.0 Introduction

Over many decades, conventional natural gamma ray (GRL) geophysical logs have been widely available from exploration drilling and extensively used as primary lithology and clay type indicators (Fertl et al. 1982; Fertl and Frost 1982; Dypvik and Bue 1984; Nielsen et al. 1987; Hurst 1990; Van Buchem et al. 1992; Rider 2000; Catuneanu 2006; Simíček et al. 2012; Gould et al. 2014; Paredes et al. 2015; Alqubalee et al. 2019; Phujareanchaiwon et al. 2021; Bataller et al. 2022). This geophysical technique, however, cannot differentiate distinct signals coming from either Th, U, or K that can provide more detailed lithological information. The increasing use of Spectral Gamma Ray logging (SGRL) and the invention of portable gamma ray- or XRF-tools have enabled measurement of the total radiation and quantification of individual Th, K, and U elements to be very easy and fast both in borehole and in outcrop (Myers and Wignall 1987; Wignall 1994; Parkinson 1996; Bessa and Hesselbo 1997; McRoberts et al. 1997; Ruffell et al. 2003; De Vleeschouwer et al. 2017; Dill et al. 2017; Bábek et al. 2018; Sêco et al. 2018; Sêco et al. 2021). Relative abundances of U, Th, and K in sedimentary rocks vary in accordance with the mineral assemblage (McLennan et al. 1980; Myers and Bristow 1989; Rider 2000; Ehrenberg and Sva 2001; De Vleeschouwer et al. 2017), therefore these elements are indicative of lithology. In addition, SGRL data can provide continuous lithological information and thus facies or provenance change for an entire succession penetrated by a borehole. Supplementary to continuous logging information, drill cuttings collected over variable intervals of a succession have been used as a complementary indicator of lithology and for calibration of geophysical logs. Drill cuttings always are taken as composites over a collection interval (usually 3 to 10 m) and hence do not represent the full range of individual lithologies or alternations thereof that may exist over a cuttings composite interval. Geophysical logging allows separation of smaller depth intervals or lithology

alternations by using SGRL data. However, interpretation of SGRL data is complicated by its sensitivity to: borehole conditions, drilling additives, penetration depth, vertical tool resolution (~30cm) and lithology/minerals mixtures that may be affected by post depositional processes (i.e. weathering and diagenetic effect that alter element composition).

To date, lithofacies and basin evolution information of the Lower Jurassic successions in the Mandawa Basin has been hampered by the limited number of wells drilled, a high degree of weathering and incompleteness of exposed outcrop sections, and thus, most interpretations of lithofacies evolution rely on drill cuttings only. In this study, we attempt to elucidate the applicability of geophysical SGRL for deriving lithological information and sedimentary facies evolution over time in early Jurassic successions of the Mandawa Basin.

Methodology has been developed based on a global database compiled from the literature to define baseline Th, U, and K background concentrations from which various Th/K and Th/U lithofacies field have been mapped. These binary discrimination diagrams were then compared with SGRL derived Th/U, Th/K and sediment composition data of Lower Jurassic strata (Mbuo, Nondwa and Mihambia Formations) from the Mandawa Basin including basement crystalline rock, to derive the lithofacies and paleo-redox conditions. When complementary geochemical analysis of outcrop samples was available via literature, this information was synthesized with the SGRL data to account for the regional provenance.

2.1 Geological setting and stratigraphy of the study area

The on-shore Mandawa Basin lies along the SE coast of Tanzania (**Fig. 2.1**). The basin is bounded to the S by the E-W trending basement high of the Ruvuma saddle, to the N by the Rufiji trough, to the W by the border fault of the Masasi basement spur and covers an area of approximately 15,000km² (Alconsult and TPDC 1997; TPDC 2003; Stratadata 2006).

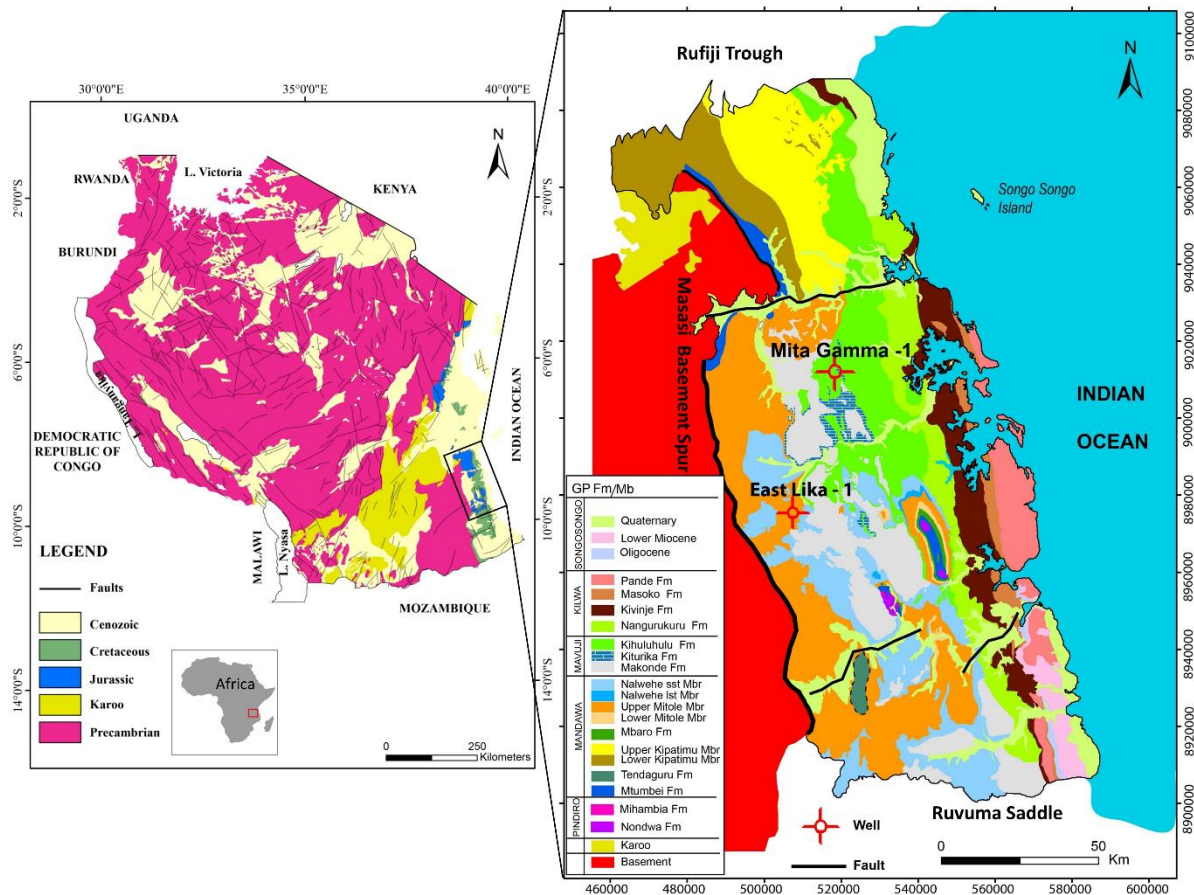


Fig. 2.1: Geological map of Tanzania, showing the location of the Mandawa Basin and position of the Mita-Gamma-1 and East-Lika-1 wells investigated (adapted from Pinna et al. 2008; Hudson 2010; Barth et al. 2016).

Basin evolution and sedimentation were controlled by major rifting events associated with the breakup of Gondwana and the drifting of Madagascar (Balducci et al. 1992; Veeken and Titov 1996; Einvik-Heitmann 2016; Fossum 2019). The Permo-Triassic Karoo rifting event, creating NNE-SSW intracratonic basins filled by continental Karoo clastic sediments, was succeeded by crustal extension during the Toarcian (~180 Ma) that further fragmented Gondwana. Crustal extension was accompanied by basin subsidence that allowed marine in-/transgression, which led to deposition of regional source and cap rocks such as the Nondwa Shale. A transition from rifting to drifting began in the ~170 Ma period (Middle Jurassic), leading to the initiation of seafloor spreading and opening of the western Somali and Mozambique Basins, and a southward drift of Madagascar along the N-S trending transform fault of the Davie Fracture Zone (Eagles and König 2008; Gaina et al. 2013). Seafloor spreading between Madagascar and Africa terminated around 120 Ma when strike-slip motion between Madagascar and India began (Eagles and König 2008;

Reeves 2018; Fossum 2019; Fossum et al. 2019). Following the Gondwana breakup, sediment accumulation was predominantly marine to partly terrestrial and continued into the Neogene resulting in a stratigraphic succession of five groups listed here in chronological order: Pindiro, Mandawa, Mavuji, Kilwa and Songosongo (Hudson 2010; Fossum et al. 2019). This study focuses on the Pindiro Group that is primarily siliciclastic with evaporites and carbonates (**Fig. 2.2**) and constituted by the Mbuo, Nondwa and Mihambia Formations.

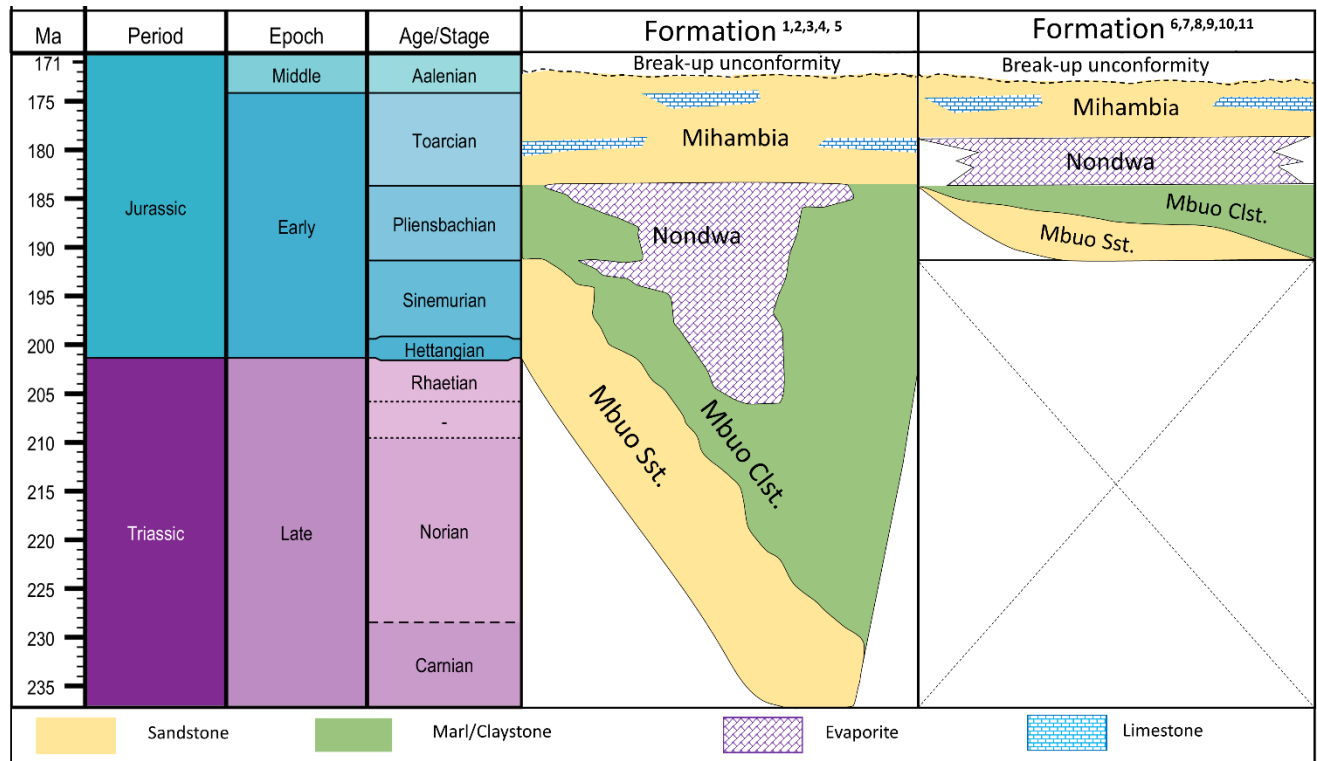


Fig. 2.2: Stratigraphy of the Pindiro Group (?Late Triassic-Mid Jurassic), depicting two controversial age assignments according to references 1. Balduzz et al., 1992; 2. Alconsult and TPDC, 1997; 3. Hudson, 2010; 4. Hudson and Nicholas, 2014; 5. Fossum, 2019 versus 6. Veeken and Titov, 1996; 7. Stratadata, 2006; 8. Quinton and Copestake, 2007; 9. Corelab and TPDC, 2009; 10. Robertson and CGG 2017; 11. Macgregor, 2018.

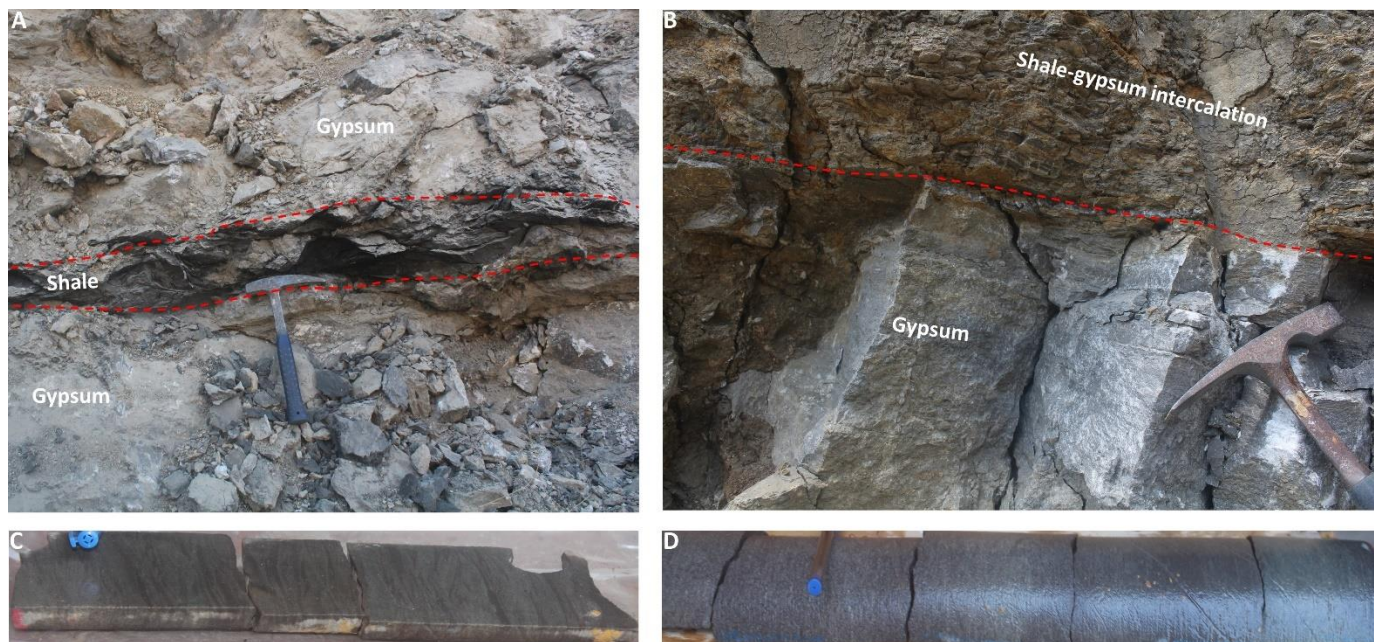


Fig. 2.3: (A and B): Outcropping evaporite-shale intercalations in the Nondwa Formation. (C, D): Drill core slabs showing very fine to fine grained arkosic Mbuo sandstone with abundant flasers of organic material and lithic fragments.

The Mbuo Formation lies unconformably upon the basement and predominantly consists of fine to coarser sandstones, intercalated with siltstones and claystones with minor dolomites (**Fig. 2.3**). Strata were deposited in alluvial, fluvial and lacustrine environments (Stratadata 2006; Corelab and TPDC 2009; Hudson 2010; Hudson and Nicholas 2014). The Nondwa Formation overlies the Mbuo Formation and comprises evaporites and shales (**Fig. 2.3**) deposited in a shallow restricted marine embayment, and is overlain by the Mihambia Formation. The latter is constituted dominantly by clastic sediments with subordinate limestones deposited in a shallow marine setting (Stratadata 2006; Corelab and TPDC 2009; Hudson 2010; Hudson and Nicholas 2014). Due to limited biostratigraphic information, the chronology of these formations is discussed controversially and deviating age assignments do exist for the base of Mbuo and Nondwa Formations as summarized in **Fig. 2.2** (Kent et al. 1971; Balduzz et al. 1992; Wopfner 1994; Kagya 1996; Alconsult and TPDC 1997; Schlüter 1997; Quinton and Copestake 2007; Corelab and TPDC 2009; Hudson 2010; Robertson and CGG 2017; Davison and Steel 2018; Macgregor 2018; Godfray and Seetharamaiah 2019; Abay et al. 2019).

2.2 Materials and methods

Geological and geochemical methodologies were applied to achieve the objectives of this study (**Fig. 2.4**). Details of individual research methods are described below.

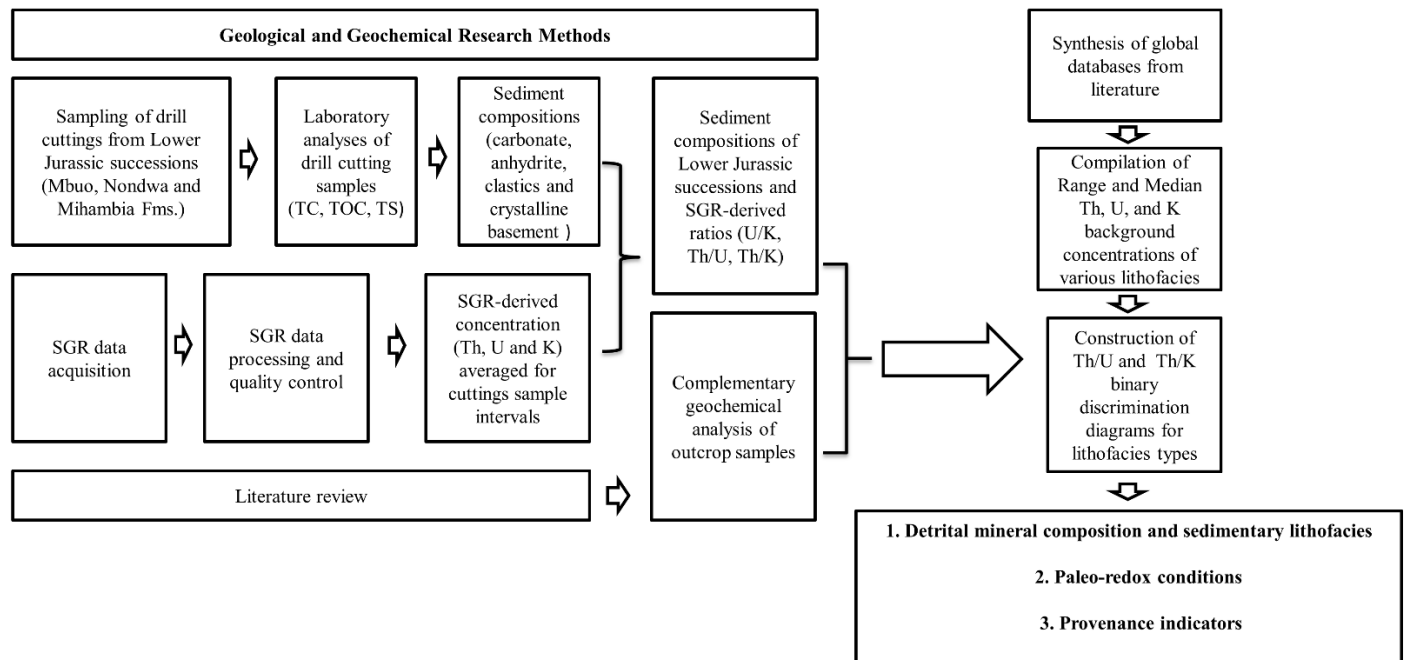


Fig. 2.4: A summary of the workflow carried out in this study.

2.2.1. Sampling of drill cuttings

A total of 105 samples, each constituting 3m of composite drill cuttings that cover the Lower Jurassic succession was collected at the Tanzanian Oil Company core repository. Lithotypes of cuttings were documented during sampling and in combination with existing well reports (Halliburton 1996; Halliburton 1997; Stratadata 2006; Quinton and Copestake 2007; Corelab and TPDC 2009; Robertson and CGG 2017) were used to construct the lithological columns shown in **Figs. 2.4 to 2.7**. From each composite cuttings sample a representative aliquot of about 16-20 g was washed, dried and crushed to powder for geochemical analysis.

2.2.2 Total carbon (TC) and total sulfur (TS) analysis

Total carbon (TC) and total sulfur (TS) contents were determined by use of a VARIO-III elemental analyser (Elementar®) using 10mg of sample weighted into a tin capsule followed by combustion at 1150°C. Results were reproducible by $\pm$ (0.05 %).

2.2.3 Total organic carbon (TOC)

Total organic carbon was obtained by decarbonatization of ~ 50mg of powdered sample weighted into a ceramic crucible via successive treatment with 10% and 25% HCl to remove calcite and dolomite if present. After washing to neutral and drying, samples were analyzed via combustion in an Eltra Helios CS Analyzer (Eltra®). Results were reproducible by $\pm$ (0.05%).

2.2.4 Sediment composition (carbonate, anhydrite, clastics and crystalline rock)

Determinations of cumulative carbonate and evaporite mineral contents were achieved via dissolution and removal by 10% and 25% HCl treatment of a ~800-1200 mg sample aliquot followed by washing with deionized water for neutralization and drying at 40° C for 24 hrs. Cumulative amounts of evaporites and carbonate in the sample were determined from the weight loss upon treatment.

Carbonate contents were calculated from proportions of total inorganic carbon (TIC) calculated by difference of TC – TOC. TIC was converted into weight % carbonate by multiplying with a stoichiometry factor of 8.33, assuming calcite as main carbonate phase.

Proportions of evaporite minerals (halite, gypsum, anhydrite) were determined from the difference between cumulative carbonate + salt fractions versus carbonate. The difference between the cumulative carbonate + salt fractions and the total sample weight was taken to represent the clastic fraction.

2.2.5 Spectral Gamma Ray Log data acquisition

Log measurements were acquired by rising a SGRL-calibrated tool in an uncased well, with a data spacing of 0.15 ft following Halliburton® standardized procedures. All logs were processed and available in composite format in both analogue and digital form (LAS). LAS format data was imported into an Excel® spreadsheet and averaged over 10ft intervals to match with cuttings sample intervals. SGRL data from the Mita-Gamma-1 well covers the entire Lower Jurassic succession, whereas SGRL data from the East-Lika-1 well was available for the Mbuo Formation only.

2.2.6 Borehole quality check

Gamma ray log measurement is influenced by drilling induced effects, particularly caving and additives from the drilling mud. Caving will reduce detector responses and drilling fluid such KCl or bentonite may increase K, and/or Th concentrations measured by a Spectral Gamma Ray tool. Gamma ray and calliper data, as well as technical drilling reports were examined to identify the potential presence of cavings and types of mud used during drilling. Caving was recognized in the Nondwa evaporitic sequences. Drilling reports showed that both of the two wells were drilled by the same contractor using a mixture of gel/polymer salt saturated water, whereby SGRL responses were not affected by drilling additives.

2.2.7 Establishing U,Th, and K background concentration

Definition of U, Th, K background concentrations representative of selected lithofacies types was achieved via a literature compilation. Element concentration ranges as well as average and median concentrations (**Table 2.1**) were determined and binary discrimination diagrams (Figs. 2.9a and 2.10a) were constructed to map the lithofacies types.

2.3. Results

2.3.1 U,Th, and K concentrations according to lithofacies

Data compilation revealed a high variability in relative abundance of the radiogenic elements in different sedimentary rocks (**Table 2.1, Supplementary Table 2.1**). Substantial K content is present in shale, siltstone and sandstone, and diminished K-contents are present in carbonates and evaporites. Whereas U is commonly enriched in black shales, carbonates and other clastic rocks do not exhibit high variability in U concentrations, as opposed to very low concentrations of U in evaporites.

Lithology	K (%)		Th (ppm)		U (ppm)		Th/U		Th/K	
	Range	Median	Range	Median	Range	Median	Range	Median	Range	Median
Black shale ^{2,5,12,13,14,15,18,25,26,29,30,36,37}	1-4	2.3	2-24	12	2.7-53	7.3	1-3	1.4	2-6	4.7
Grey/green shale ^{2,7,8,10,13,17,19,23,29,32,34}	1.6-6.8	3.8	5.9-19.2	11.8	1.5-5.1	3	3-4	3.8	1.6-5.4	3.7
Red/yellow shale ^{2,27,28,32}	4.1-4.9	4	14.6-19	17.6	1.8-2.8	2.5	7-8	7.1	3.6-4.8	4.8
Siltstone ^{11,16,20,22,28,31,33}	1.3-3.8	2.2	8-15.8	11.6	1.7-3.7	2.6	4-6	3.9	3-10.8	4.9
Sandstone ^{1,2,3,13,14,16,23,24,29,31,37}	0.7-4.9	1.6	2-12	8.4	0.8-3.5	2.4	2.6-5.8	3.5	2.5-11	5.6
Carbonates ^{2,3,4,6,7,10,21,25,29,30,38}	0.4-1.3	1	0.2-3.1	2.2	0.2-7	2.2	0.2-1.3	1	0.5-2.4	2.2
Anhydrite/Gypsum ^{2,5,6,7,8,29,35}	0.2-0.25	0.2	0.15-0.2	0.2	0.1-0.2	0.2	1-1.5	1	0-1.2	1.2
Halite ^{2,5,6,7,8,29}	0.2-0.3	0.3	0-0.2	0.2	0-0.1	0.1	0-2	2	0-1.2	1.2

Table 2.1: Compilation of concentration ranges and mean values for the radiogenic elements K, Th, and U in sedimentary rocks and derivation of Th/U and Th/K ratios applicable to sedimentary rock typing. Data compiled from: 1. Moore, 1957; 2. Adams and Weaver, 1958; 3. Moore 1960; 4. Turekian and Wedepohl, 1961; 5. Pliler and Adams, 1962; 6. Horn and Adams, 1966; 7. Ronov and Migdisov, 1970; 8. Wedepohl, 1970; 9. Moore 1971; 10. Cosgrove, 1972; 11. Beeson, 1980; 12. McLennan and Taylor, 1980a; 13. Dypvik and Eriksen, 1983; 14. Dypvik and Bue, 1984; 15. Myers and Wignall, 1987; 16. Myers and Bristow, 1989; 17. Van Buchem et al., 1992; 18. Tribouillard et al., 1994; 19. Hayashi et al., 1997; 20. McRoberts et al., 1997; 21. Luczaj, 1998; 22. Asiedu et al., 2000; 23. Svendsen and Hartley, 2001; 24. Martinus et al., 2002; 25. Luning et al., 2004; 26. Fello et al., 2006; 27. Yamaguchi, 2006; 28. Kasanzu et al., 2008; 29. Mernagh and Mieizitis, 2008; 30. Correia et al., 2013; 31. Paredes et al., 2015; 32. Kasanzu et al., 2017; 33. Amedjoe et al., 2018; 34. Reolid et al., 2020; 35. Güngör Yeşilova and Yeşilova, 2021; 36. Marín et al., 2021; 37. Phujareanchaiwon et al., 2021; 38. Wei et al., 2021.

Likewise, sandstone, siltstone, and red shale contain high Th concentrations relative to other rock types. The results reveal that in general, sedimentary rock can be subdivided by geochemical classification into various fields based on Th/U and Th/K ratios (**Table 2.1**) that reflect minerogenic composition and depositional conditions, respectively.

2.3.2 Stratigraphical variability of Th, U, and K abundances

SGRL-concentration data reveal a high variability in Th, U, and K along the sediment succession of wells Mita-Gamma-1 and East-Lika-1 (**Figs. 2.4 and 2.5; Supplementary Table 2.1**) that follows the trend in natural gamma ray (GR API) response. The overall GR trend appears to be mainly controlled by Th and K concentrations ($R^2=0.90, 0.87$, respectively) and least by U contributions ($R^2=0.35$). All anomalous GR values at depths of 2060-2070, 2140-2150, and 2230-2240 m in the Mita-Gamma-1 and 1640-1650, 1860-1890m in the East-Lika-1 wells coincide with highest concentration in Th.

K, Th, and U abundances systematically vary with lithofacies, with enhanced Th proportions in sandstone/siltstone intervals whereas K and U contributions are elevated in shales. Minimum concentrations of all three radiogenic elements are associated with salt, carbonate, and basement intervals. A tendency of Th, U, and K concentrations to co-vary with carbonate abundance is noted in sediments from both wells, except for a section in the upper part of the Mihambia Fm., where carbonate influx is accompanied by an increase of U and decrease of Th, K. All major lithological boundaries including the boundary between basement and the Mbuo Fm. or interspersed salt layers can be recognized on the basis of individual Th, U, and K concentrations (**Figs. 2.4 and 2.5**). Crystalline basement rock is characterized by its fairly constant concentration of the three radiogenic elements, whose abundances sharply increase at the base of the Mbuo Formation and maintain at high levels in both wells. The Nondwa and Mihambia Fm. sediments in the Mita-Gamma-1 well not only show lower concentrations of Th, K, and U than the Mbuo Fm. but also substantial variation in their respective radiogenic element concentrations (**Fig. 2.5**). Two main trends can be differentiated: a trend to comparable increase/decrease of all elements at 2280-2200m and 2200-2150m depth, respectively, and an opposite U covariation with Th and K at 2050 to 2010m in the Mihambia Formation. In contrast, sediments from the East-Lika-1 well display a trend of systematic increases in Th, U, and K from the crystalline bedrock to the base of the Mbuo Fm.; from there on, a relatively invariant trend maintains throughout the entire formation (**Fig. 2.5**).

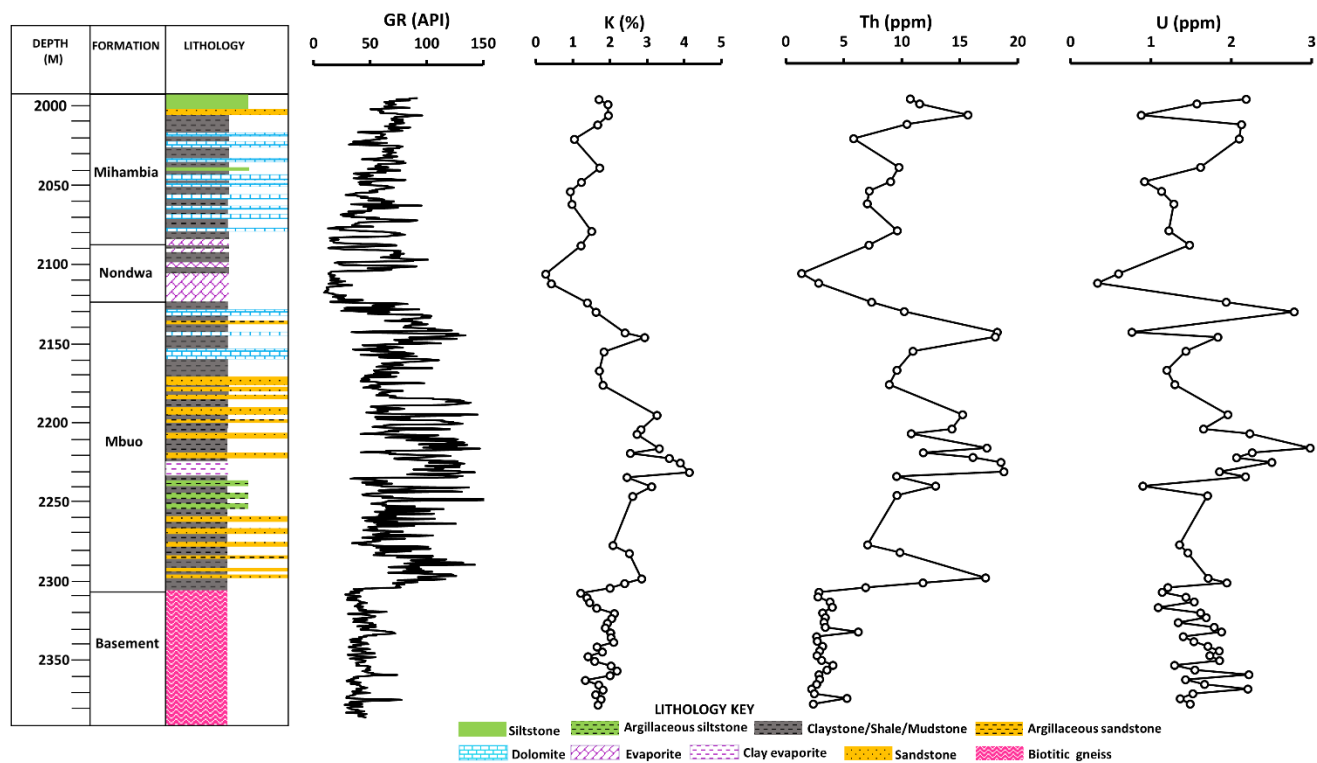


Fig. 2.5: Stratigraphical variability of the Th, U, and K SGRL abundances averaged for cuttings sample equivalents and their relationship to the total natural gamma ray intensity (GR API) in the sediments from the Mita-Gamma-1 well.

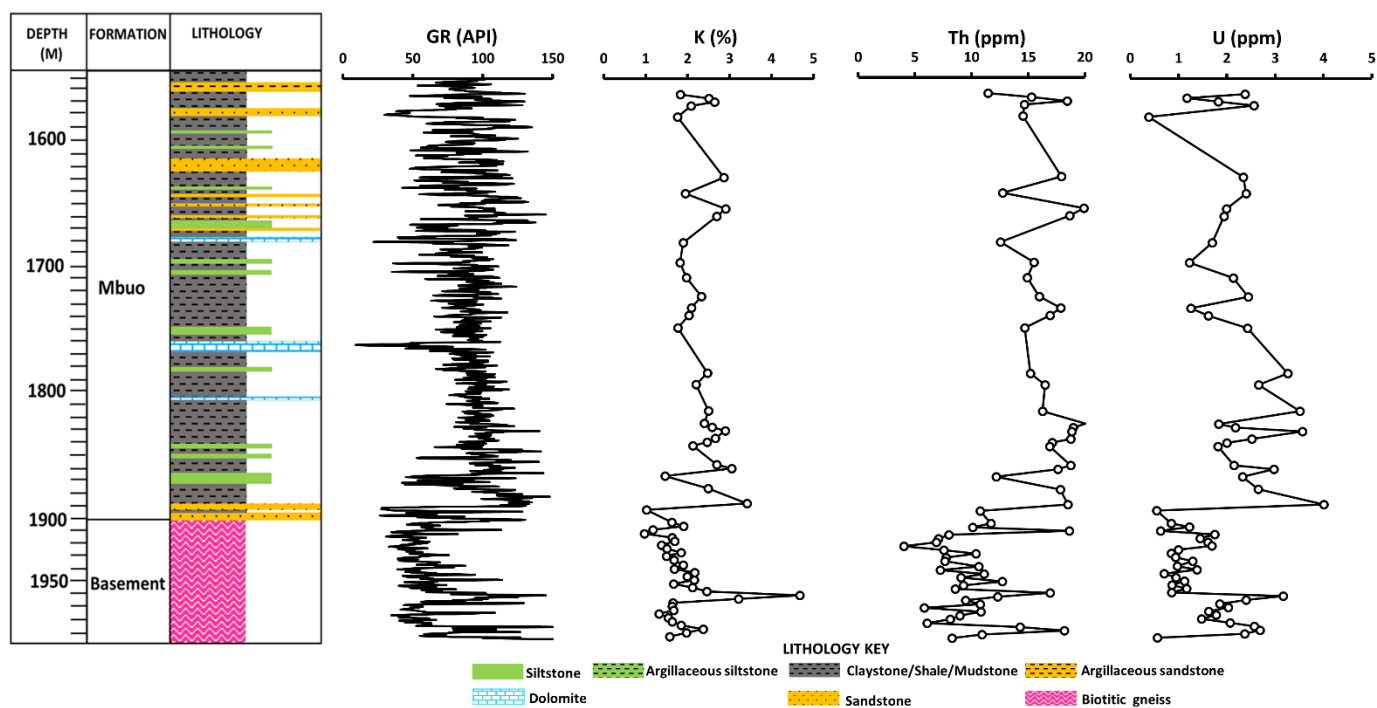


Fig. 2.6: Stratigraphical variability of the Th, U, and K SGRL abundances averaged for cuttings sample equivalents and their relationship to the total natural gamma ray intensity (GR API) in the sediments from the East-Lika-1 well.

2.3.3 Sediment composition and stratigraphic variability of U/K, Th/U, and Th/K ratios

In addition to absolute concentrations of the three radiogenic elements used for lithofacies typing, their relative proportions can guide in deriving further information on sediment facies, in particular on the redox regime prevailing, the intensity of weathering and the proportion of terrigenous influx into marine/limnic basins.

In both wells, the U/K ratios of the rocks range between 0.3 to 2.3 and do not show much variation between the basement and the Mbuo Formation. A substantial increase in average U/K ratios from about 0.7 in the Mbuo Formation to average values of 1.3 in the Nondwa and Mihambia Formations is noted for the Mita-Gamma-1 site (**Fig. 2.7**). This is ascribed to a higher degree of weathering for the hanging wall strata penetrated by this well. Within basement and Mbuo strata of the East-Lika well, the U/K ratio remains rather constant at values around 0.9, recording only episodic shifts in weathering stage.

Rather invariant Th/U ratios ranging from 2 to 4 are recorded for the basement strata of the Mita-Gamma-1 well (**Fig. 2.7**). As this ratio is mainly controlled by U uptake of sediments at low redox potential, a process not relevant in crystalline bedrock, the invariance is as expected. For the East-Lika well though, very high Th/U ratios ranging between values of 5 to 20 occur in an interval in the uppermost basement rocks (**Fig. 2.8**). Below 1950 m depth, basement rocks exhibit Th/U ratios ranging from 2 to 6 and thus quite similar to those noted for the Mita-Gamma-1 well. The U depletion in the basement strata between 1990 to 1950 m is reflected by the basal Mbuo Formation showing Th/U ratios of 5 to 20 and thus identical to the underlying basement.

The ratio of Th/K exhibits an overall upward increasing trend in the sediments from the Mita-Gamma-1 well (**Fig. 2.7**) with values constantly below 5 in the basement rocks. Except for a single value of about 16 determined in the basement, Th/K ratios range between 1 to 8 in the East-Lika strata, with slightly higher but rather invariant values of 4 to 8 in the Mbuo Formation (**Fig. 2.8**).

Radiogenic elements ratios correlate with lithofacies in both wells. The rock record of the East-Lika well shows low variability in sediment composition in the dominantly clastic Mbuo Formation. A higher variability in sediment composition occurs in the Mita-Gamma-1 well, where evaporites are present in two intervals and various carbonate intercalations occur, both of which show elevated U/K ratios.

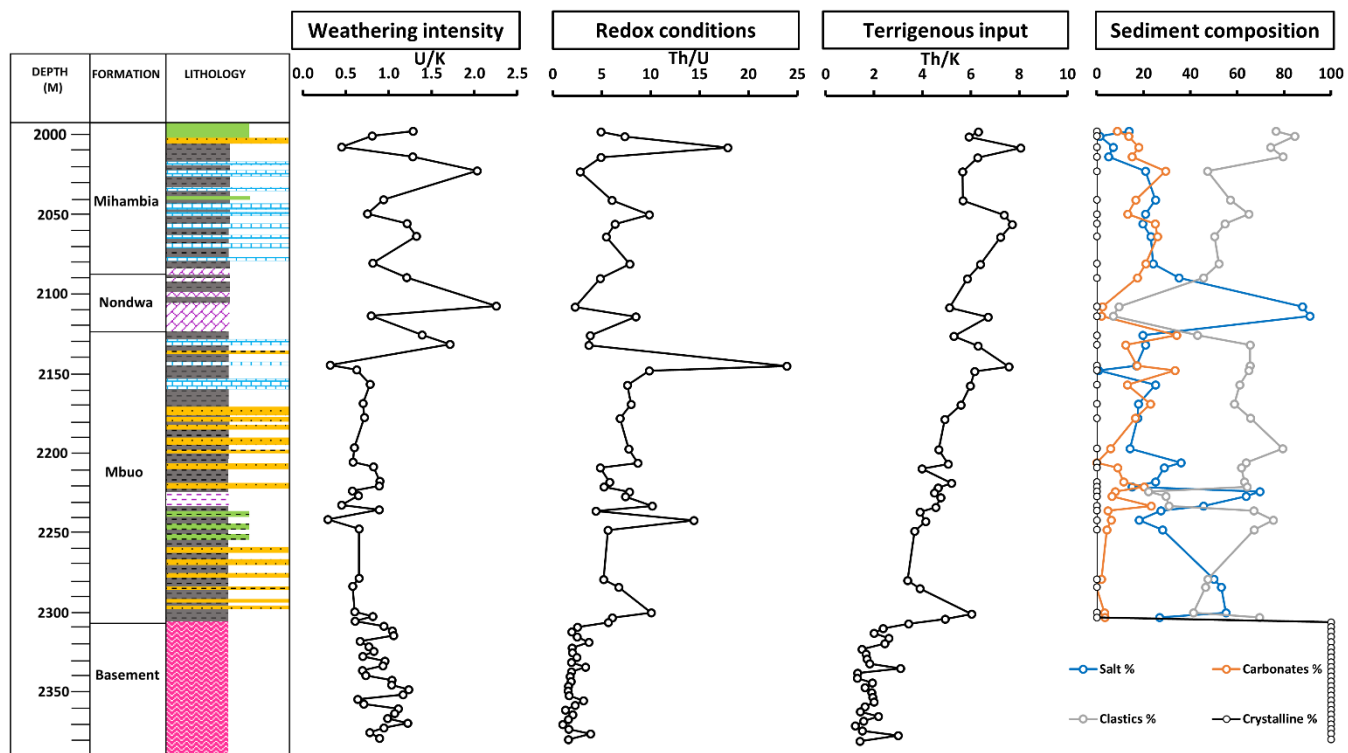


Fig. 2.7: Stratigraphic trends in SGRL-derived U/K , Th/U, and Th/K ratios averaged for cuttings sample equivalents and their relationship to the sediment composition in the Mita-Gamma-1 well. The lithological column on the right was established based on geochemical analysis of cuttings.

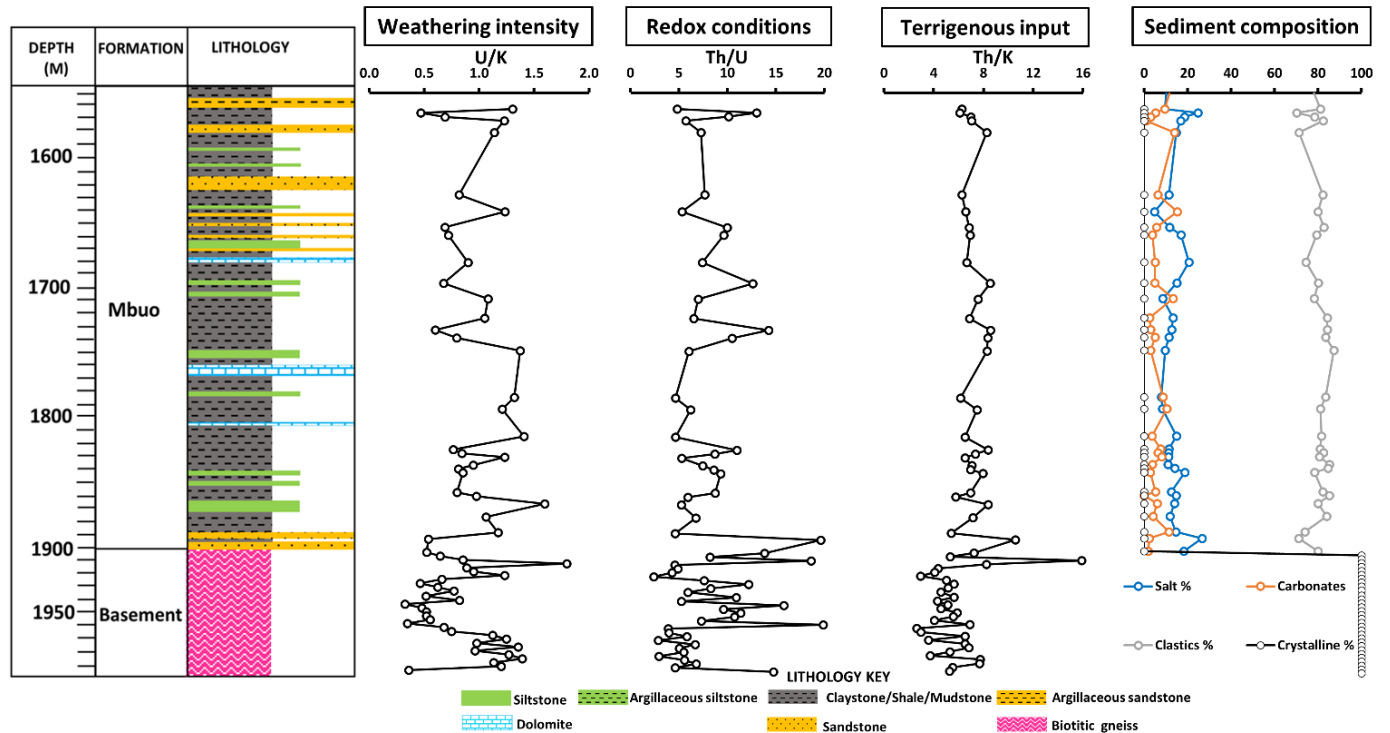


Fig. 2.8: Stratigraphic trends in SGRL-derived U/K, Th/U, and Th/K ratios averaged for cuttings sample equivalents and their relationship to the sediment composition in the East-Lika well. The lithological column on the right was established based on cuttings interpretation and geochemical analysis.

2.4 Discussion

2.4.1 Thorium-Uranium as an indicator of paleo-redox conditions

Sedimentary rocks exhibit Th/U ratios ranging from <0.02 to >20 , thus spanning two orders of magnitude. Th/U ratios <2 are typical for marine deposits and stand in contrast to Th/U ratios >7 common for a continental origin (Adams and Weaver 1958; Fertl et al. 1982; Isaksen and Bohacs 1995; Wignall and Twitchett 1996; Rider 2000; Lüning and Kolonic 2003; Marenco et al. 2016). Variation in Th/U ratios may reflect a series of geological processes that may ultimately enrich or deplete one of the elements either during weathering, transportation or deposition, or as a function of redox conditions prevailing in pore waters (Adams and Weaver 1958; Kelepertsis 1981; McLennan and Taylor 1991). In reducing environments U tends to be precipitated as U^{4+} and becomes enriched relatively to Th, whereas in oxidizing environments U^{4+} is oxidized to more

water-soluble U^{6+} , which partitions back into the water column (Osmond and Ivanovich 1992). Therefore, the Th/U ratio serves as an important paleoenvironment indicator for distinguishing lithofacies, leaching/weathering and paleoredox conditions. In order to assess lithofacies from SGRL-based Th/U ratios, we compiled concentration data of the two elements in various lithotypes from the literature (**Table 2.1, Supplementary Table 2.1**) and constructed a general lithofacies discrimination diagram (**Fig. 2.9a**). In order to account for regional provenance, we used geochemical data determined on outcrop samples in the wider study region for Jurassic (Hudson, 2010) and Proterozoic strata (Kasanzu et al., 2017) as depicted in **Fig. 2.9c** and **Fig. 2.9d**, respectively. The data set in this study generated from SGRL is shown in the discrimination diagram of **Fig. 2.9b**.

Most of the data analyzed here cluster in a field delineated by U/Th ratios of 6 to 27 (**Fig. 2.9b**) and are ascribed to an oxic-continental source, with a few data points placed in a field delineated by U/Th ratios between 4 to 6 and ascribed to a typical siltstone-sandstone provenance. The oxic-continental field is assumed to be associated with the Mozambique Belt provenance and reveals a differentiation within that field according to Th concentration being highest in shales, followed by siltstones, intermediate in sandstones and lowest in basement rocks (**Fig. 2.9b**). The lowest concentrations of both Th and U are found in evaporite samples, whereas carbonates show variable but generally low Th concentrations, exceeding those in basement rocks. This indicates that either the provenance of the crystalline basement of the Mozambique Belt has a unique signature, which is incompatible with the global Th/U systematics or that almost all sediments studied derived their Th/U signature from oxic-continental clastics subjected to intensive weathering and leaching of U. High Th/U-ratios of 5 to 20 determined for crystalline basement in the East-Lika well argues for a source rock signal but intensive leaching upon weathering and transportation cannot be excluded.

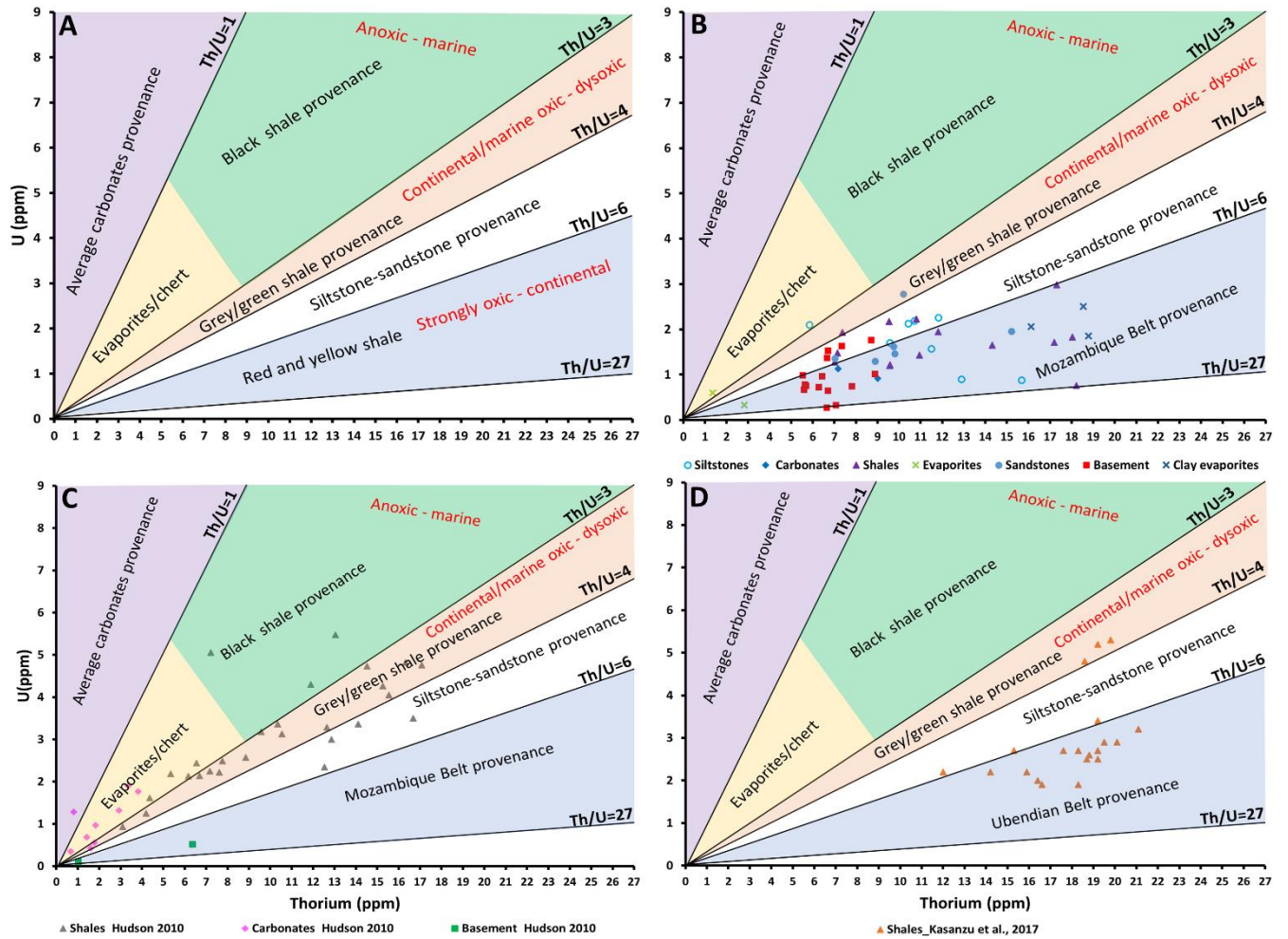


Fig. 2.9: Binary Th/U discrimination diagrams depicting lithofacies fields bordered by Th/U-ratios of 1, 3, 4, 6, and 27 based on global and regional data compilation: A (global data set), B (Mandawa Basin, this study), C (Mandawa Basin, Hudson 2010) and D (Proterozoic Buanji Basin, SW-Tanzania, Kasanzu et al., 2017).

A comparison of the data generated here with those reported for the Mandawa Basin by Hudson (2010) reveals that outcrop samples from the latter study show higher U concentrations, with most shales samples clustering in the “grey/green-shale” provenance field bordered by Th/U ratios of 3 to 4 (**Fig. 2.9c**). Several shale samples are located in the marine-anoxic provenance field, which is considered to reflect the distal locations of the outcrops sampled. The study sites investigated in this work are located on the basin margin, where shallower water depth caused more intensive mixing and oxygenation of the water column, conditions not favorable for reducing pore waters and associated U-precipitation. The offset in Th/U ratios between the

outcrop derived deep basin and the well-derived basin margin samples thus reflects the redox regime prevailing.

Carbonates from the outcrop study plot in the evaporite field close to the carbonate segment potentially due to U removal by leaching. The two basement rocks from outcrop that were studied by Hudson (2010) plot in the Mozambique Belt provenance field but in comparison with the data generated in this study are highly depleted in U and Th, thus pointing to intensive weathering and leaching.

Adjacent to the Mandawa Basin and its Jurassic-Cretaceous sediment fill, the Buanji Basin of SW-Tanzania is located. It is thought to have received its Proterozoic sediment infill from the Ubendian Belt (Kasanzu et al. 2017), which is compatible with the Mozambique Belt. Shales from the Buanji Group populate the Ubendian Belt field (**Fig. 2.9d**) and are highly compatible with the younger strata of the Mandawa Basin, which were sourced from the same basement complex. Three samples from the Buanji Group are positioned in the dysoxic marine shale field, which may indicate deposition under more restricted conditions and precipitation of U at lower E_h -values.

Deviation of Mandawa and Buanji Basin samples from the standard lithological /depositional environment classification scheme can be due to a variety of factors including specific source rock chemical composition, dilution effects, hydrodynamic sorting, and diagenetic alteration or redox regimes prevailing that in turn elevate or diminish the Th/U ratio. In carbonates and evaporites, the Th/U-ratios are mostly controlled by the presence of detrital minerals and organic matter (Adams and Weaver 1958; Adams et al. 1959; Graf 1962; Moore 1971). These impurities may raise Th/U-ratios of carbonates to approach those of siliciclastic rocks (Adams and Weaver 1958; Adams et al. 1959). In siliciclastic rocks the Th/U-ratios are sensitive to depositional conditions (Adams and Weaver 1958; McLennan and Taylor 1980b) and will vary in accordance with the parent rock types, distance to source and diagenetic alteration (Adams and Weaver 1958; Nielsen et al. 1987; Rider 2000). Presence of organic matter may promote uranium fixation in the water column or surface sediment (McLennan and Taylor 1980b; Myers and Wignall 1987; Rider 2000) and thereby reduce the Th/U-ratio. Rapid sedimentation and hydrodynamic sorting seem to play significant roles in the distribution of Th-U in sandstones and resistate minerals (Adams and Weaver 1958). Due to fast deposition, the Th/U-ratio in greywackes will be virtually identical to

that of crystalline rocks and vary significantly from pure orthoquartzites (Adams and Weaver 1958; Murray and Adams 1958; Galbraith and Saunders 1983; Van Buchem et al. 1992).

We conclude that almost all samples in this study were deposited in oxic to strongly oxic environments whereby a significant fraction of the primary U was leached and no U added via reducing conditions. In contrast, the sediments studied by Hudson (2010) and Kasanzu et al. (2017) originate from deeper basin settings and were deposited in dysoxic to oxic-dysoxic water columns, respectively. Carbonate and evaporites may deviate from the global classification scheme due to presence of detrital minerals and diagenetic leaching whilst strongly oxic depositional would account for deviation in clastic rocks. The very few of the siltstones that plotted in the grey/green dysoxic shale provenance field could be associated with episodes of marine incursions into the Mandawa Basin upon sea-level highstands. These incursions were accompanied by enhanced organic matter deposition and preservation under oxygen depletion yielding higher amounts of total organic carbon (TOC) noted in Mandawa sediments (Emanuel et al. 2020; McCabe 2021). Low U content and average concentrations of TOC < 3% (Emanuel et al. 2020; McCabe 2021) could imply burn-down/remobilization after brief marine incursions (Lüning and Kolonic, 2003). The wide spread of data points in the oxic continental clastic field can be ascribed to hydraulic sorting for components of different grain size and presence of heavy minerals.

2.4.2 Thorium-potassium as an indicator detrital mineral composition and sediment lithofacies

Thorium is a chemically-physically stable trace element that occurs in resistate heavy minerals or alternatively, is adsorbed to clay minerals (Langmuir and Herman 1980; Nielsen et al. 1987; Myers 1989; Davies and McLean 1996; Parkinson 1996; Rider 2000; Gould et al. 2014) as opposed to K that is commonly abundant but chemically unstable in many rock forming minerals including feldspars, clays, and micas (Fertl et al. 1982). Fractionation of Th and K in sedimentary environments is driven by their physical and chemical properties. Th is transported as particulate solid matter and ultimately retains its original composition while K is soluble and continuously altered due to its chemical reactivity (Rider 1990; Ruffell et al. 2003). Absolute and

relative abundance of Th and K may provide clues on mineralogical composition changes that may assist in clay mineral or rock typing. It has been generally observed that variation of Th/K ratios is primarily a function of detrital minerals but may also depend on neo-formed glauconite minerals. Sedimentary rocks exhibit wide variation in Th/K ratios, whereby higher ratios ($\text{Th/K} > 6$) may be related to heavy mineral enriched or highly matured lithology and predominance of kaolinite clay minerals. In contrast, lower ratios ($\text{Th/K} < 4$) can specifically be associated with enrichment of feldspars, mica, glauconite or illite minerals (Rider 2000; Basu et al. 2009; Inwood et al. 2013). Moreover, variation in Th/K ratios has been linked to changes in transport mechanisms, source area proximity, or topography, post depositional processes and other local factors involving tectonics and climate (Doveton 1991; Hesselbo et al. 1997; Schnyder et al. 2006; Basu et al. 2009; Inwood et al. 2013). The Th/K-ratio despite its multifaceted influences provides an essential tool that can offer lithofacies information linked to the evolution of a sedimentary basin through time.

Globally derived data on Th and K concentrations (**Table 2.1; Supplementary Table 2.1**) depending on sediment type were employed to create binary discrimination diagrams defining different lithofacies types. Element ratios indicative of clay mineralogy were adapted from Quiren et al. (1982). Following the same procedure as applied in chapter 2.4.1 for Th/U ratios, the binary discrimination diagrams were generated for the global data compilation to define fields that are delineated by Th/K ratios of 0.5, 1, 2, 4, and 11 (**Fig. 2.10 A**). These discrimination diagrams were then populated with data determined from Jurassic succession located in the marginal (**Fig. 2.10b**) and distal (**Fig. 2.10c**) parts of the Mandawa Basin (Hudson 2010) and Proterozoic successions from the Buanji Basin (Kasanzu et al., 2017).

The heterolithic nature of cuttings samples analysed in this study prevents data to cluster exclusively in one of the end-member fields, e.g. either pure evaporites or carbonates (**Fig. 2.10**) or kaolinites. Most data points populate the clastic regime of shales, siltstones, and sandstones, whereby dispersal along the Th-concentration axis is due to dilution by quartz, carbonate or feldspars. The Mandawa Basin outcrop sediments studied by Hudson (2010) comprise mainly pelitic shales and less coarser psammitic to psephitic clastics and accordingly cluster in the shales field, indicative of mixed layer clay minerals. Again, dilution with preferentially carbonate controls variation along the Th- and K-concentration axis (**Fig. 2.10c**). Data points from Proterozoic shales

of the Buanji Basin investigated by Kasanzu et al. (2017) all cluster in the shales field as well and due to a lack of carbonate and quartz dilution display a narrow range in Th-concentration (**Fig. 2.10d**).

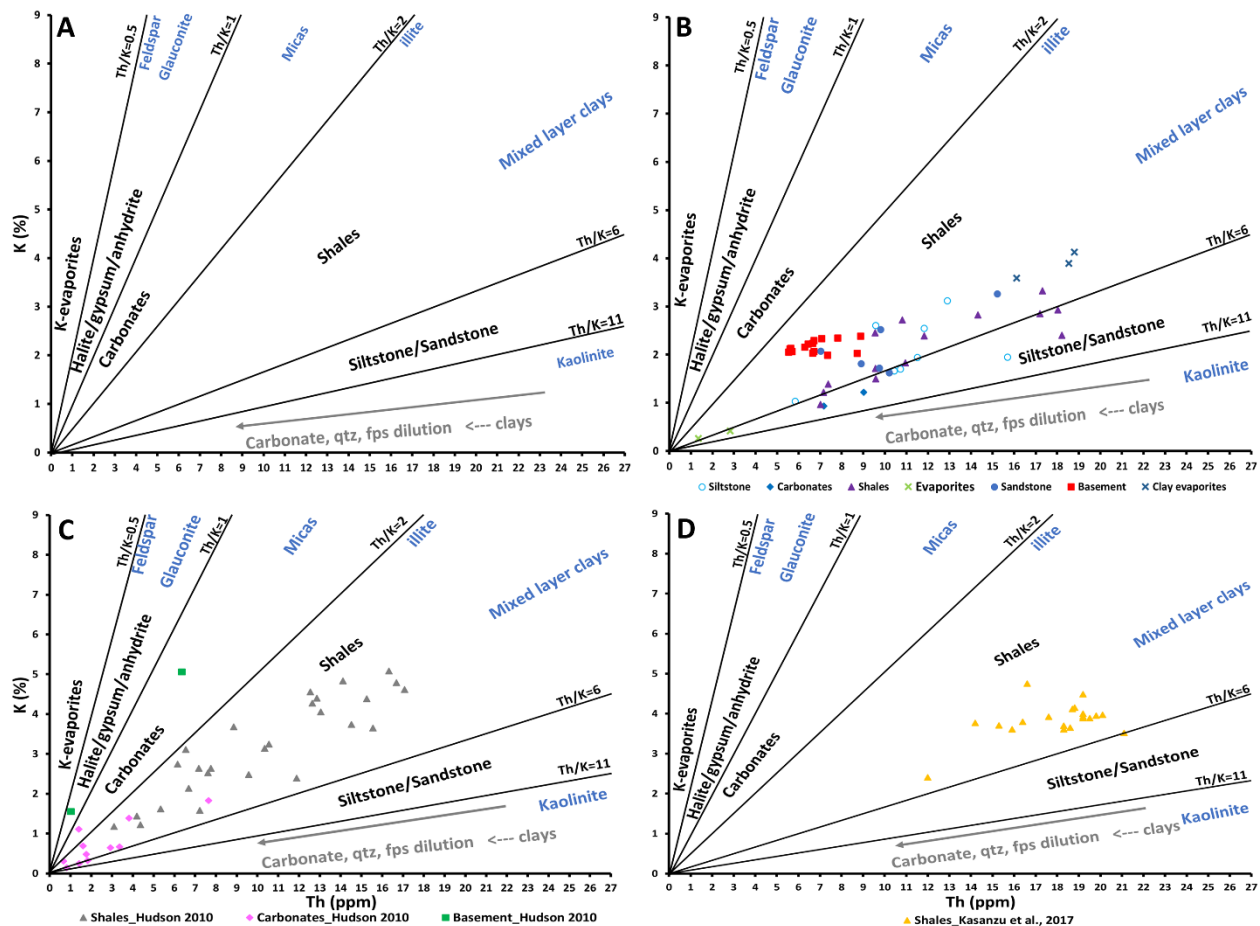


Fig. 2.10: Binary Th/K discrimination diagrams depicting lithofacies fields and associated minerals bordered by Th/K-ratios of 0.5, 1, 2, 6, and 11 based on global and regional data compilation: A (global data set), B (Mandawa Basin, Mita-Gamma-1 well, this study), C (Mandawa Basin, Hudson 2010) and D (Proterozoic Buanji Basin, SW-Tanzania, Kasanzu et al., 2017).

A rather narrow range in Th/K ratio of about 2 to 6, implies the presence of similar clay mineral assemblages dominated by mixed layer clays for the Mandawa distal basin outcrop samples and the Proterozoic Buanji Basin shales (**Figs. 2.10c and 2.10d**). Higher Th/K ratios in the proximal settings of the Mandawa Basin indicate an elevated proportion of kaolinite clays (**Fig. 2.10b**). Several studies pointed out that the Th/K-ratio is sensitive to grain size or mineral content of the

sediment (Rider 1990; Rider 2000; Schnyder et al. 2006; Inwood et al. 2013; Simíček and Bábek 2015), whereby a distribution of Th and K according to grain size is affected by diagenetic and hydraulic sorting during transport and deposition (Van Buchem et al. 1992; Simíček et al. 2012). Such grain size and mineralogy effects are more pronounced in heterolithic than in homogeneous facies (Rider 1990; Simíček and Bábek 2015) due to the presence of multiple mineral components. Occurrence of detrital components, organic matter, and lithic fragments/clasts in the Mbuo core samples, and in shales in the Nondwa evaporitic outcrop section (**Fig. 2.3**) attest to a heterolithic nature of our samples. We further propose that a slight variation in Th/K-ratios between the samples from the proximal basin section studied here and the distal basin samples studied by Hudson (2010) is controlled by source proximity. Proximal facies generally exhibit high Th/K ratios as opposed to low Th/K typical of distal facies due to differential hydraulic settings and clay flocculation (Parkinson 1996; Bessa and Hesselbo 1997; Hesselbo et al. 1997; Schnyder et al. 2006; Inwood et al. 2013). Th/K contrast in carbonates between the proximal and distal Mandawa Basin sediments may imply reduction in terrigenous influxes due to a distant source area. The low variance in Th-K ratios determined for the Proterozoic Buanji Basin sediments shown by Kasanzu et al. (2017) may indicate chemically more mature sediments than those of the Jurassic Mandawa basin fill.

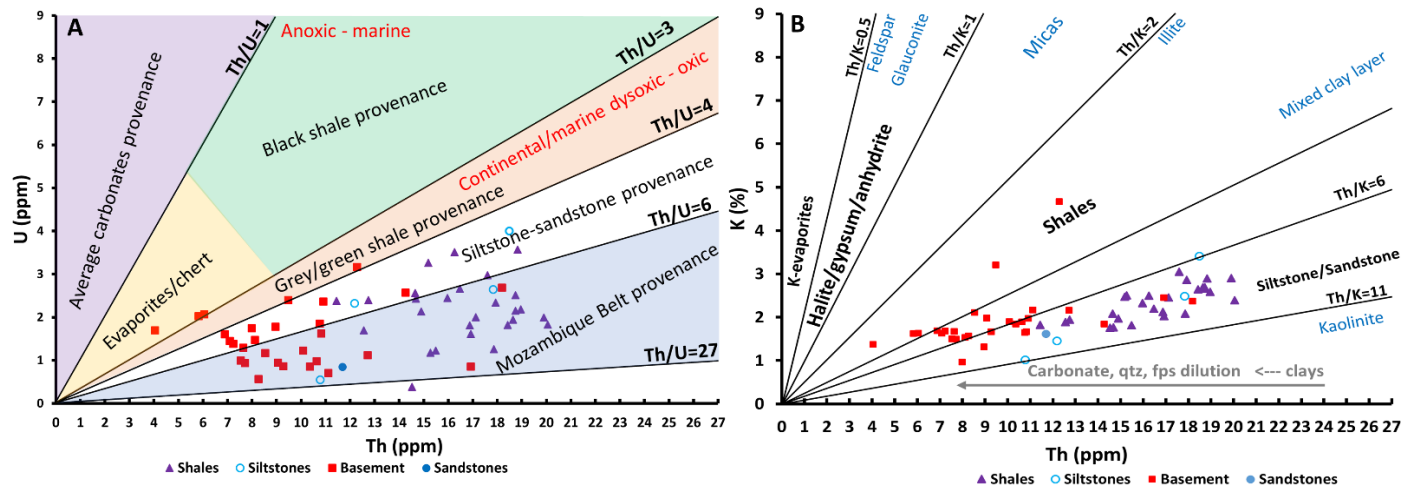


Fig. 2.11: Th/U and Th/K discrimination plots for Jurassic sedimentary rocks and basement strata from the East-Lika-1 well based on SGRL analysis of this study according to sediment provenance and clay mineral types.

The sediments from East-Lika exhibit exceptionally high Th/K ratios and in the discrimination diagram (**Fig. 2.11b**) almost exclusively cluster in the siltstone/sandstone rather than the shale field. This can be interpreted as either a higher degree of weathering and longer transport distances of pedogenic derived kaolinite or in contrast the opposite, a rather short transport, low degree of weathering and thus an inheritance of the composition of basement source rocks. The latter option is preferred here due to a clastic dominance of the Mbuo Formation in the East-Lika section (**Fig. 2.8**) and a low degree of carbonate dilution.

2.4.3 Th/K as Provenance indicator

Th/K ratios of sedimentary rocks can provide important information on source rock composition when climate, hydraulic sorting and post-depositional chemical changes are taken into consideration. Provenance shift may be accompanied by changes in mineral composition that can be reflected in Th-K data (Parkinson 1996; Bessa and Hesselbo 1997; Leeder et al. 1998; Ehrenberg and Sva 2001; Deconinck et al. 2003; Schnyder et al. 2006; Inwood et al. 2013; Simíček and Bábek 2015; Armstrong-Altrin et al. 2015; Chavez et al. 2016; Akinlotan 2018; Bábek et al. 2018; Nkongho et al. 2022; Ramezani et al. 2022; Akono et al. 2022). Data for the proximal Mandawa Basin sediments as depicted in lithological columns (**Figs. 2.7 and 2.9**) and binary diagrams (**Figs.**

2.10 and 2.11) do not show sufficient variation to constrain shifts in provenance, possibly due a homogeneous sediment source area or sediments were well mixed prior to deposition (Fralick and Kronberg 1997; Garzanti et al. 2008; Amorosi and Sammartino 2018; Ekoa Bessa et al. 2021; Mounteney et al. 2021; Lučić et al. 2021). Steady increases in Th/K ratios during deposition of the Mbuo Formation suggest that more pedogenic/detrital material was supplied from the hinterland and deposited quickly under oxic environments probably due to rejuvenated rifting associated with the breakup of Gondwana. Abrupt increases of Th/K toward the base of the Mbuo Fm. in both wells may imply either an uplift, erosion and fast deposition that could be linked with initiation of rifting events in the region (Veeken and Titov 1996; Fossum 2019). Excursions in Th/K- and Th/U-trends in the Nondwa and Mihambia Formations may indicate short-lived marine incursions into a restricted brackish/limnic Mandawa Basin that complementary caused minor increases in U abundance due to less oxic conditions, a process unrelated to changes in sediment source areas. In contrary, the early Jurassic paleogeography suggests much more variable sediment sources that might include recycled older Karoo sedimentary rocks (Hudson 2010; Hudson and Nicholas 2014), a model not supported by the sediment composition of proximal Mandawa Basin sediments studied here. We propose that the Mozambique Belt basement constituted the main provenance region and sedimentation patterns were mainly controlled by rifting, periodical marine incursions, and local climatic conditions. A series of paleo-highs associated with the Pan-African basement in the study region is assumed to have generated locally variable accommodation space and restricted water circulation and communication between sub-basins that ultimately controlled sediment supply and deposition.

2.5 Conclusions

In this study we demonstrate the application of spectral gamma-ray logging (SGRL) tools to derive lithofacies information on the Lower Jurassic successions in the Mandawa Basin, Tanzania, which are not sufficiently accessible via core and outcrop.

In order to apply radiogenic U/K/Th-concentrations and ratios as lithofacies and clay mineralogy indicators, we constructed discrimination diagrams compiled from a global set of literature data.

Our study reveals that heterolithic facies, amalgamation of different lithologies in cuttings-composite intervals, hydrodynamics sorting and post depositional processes, particularly diagenesis and salt diapirism can complicate facies interpretation. Multiple mineral components in heterolithic facies and those introduced by either detrital mineral components or diagenetic effects may both lead to a lower contrast in Th, U, and K proportions thus complicating facies assignment in discrimination diagrams. Despite these limitations, the SGRL-method has proven to be effective for highlighting the changes in detrital sediment input, hydrodynamics effect, weathering intensity, and diagenetic effects through time.

Although no highly reducing regime leading to deposition of organic-rich black shales was present in the Mandawa Basin, paleoredox conditions prevailing during deposition and a source-sink relationship can be established from SGRL data. Availability of more SGRL-data from the region may permit execution of well to well correlation that will unravel lateral variability in terms of redox conditions and sediment input between distal versus proximal parts of the basin. We propose that during the Early Jurassic oxic to strongly oxic depositional condition prevailed in the western (East-Lika) and eastern flanks (Mita-Gamma-1) of the Mandawa Basin, whereas more oxygen deficient conditions prevailed in southern central and deeper part of the basin. Freshly eroded sediment was supplied and deposited fast into the proximal basin part due to rifting activity during deposition of the Mbuo Formation. Sedimentation was interrupted by two major episodes of marine incursions during deposition of the Nondwa and Mihambia Formations as deduced from a decline in Th/U- and Th/K-ratios.

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Chapter Three

Total Organic Carbon variability of Lower Jurassic successions in the Mandawa Basin, SE Tanzania

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Abstract

Over many decades, an understanding of total organic carbon (TOC) variability in various sedimentary basins has been constrained by the availability of suitable samples for geochemical analysis. Many attempts rely on selected sidewall core or cuttings samples, which do not provide a continuous lithological record of the succession. As a result, TOC variability as a response to facies evolution is often underestimated. An advancement in TOC evaluation from well logging techniques often has permitted a continuous analysis of source rock intervals, while eliminating issues associated with obtaining discrete physical samples. Geochemical and wireline geophysical logs investigation here aimed at constraining TOC variability and depositional conditions prevailing during the early Jurassic in the Tanzanian Mandawa Basin (Mbuo, Nondwa and Mihambia Formations), which to date are poorly documented. TOC was deduced from both well cuttings samples and wire log data. Depositional conditions were defined via the use of spectral Gamma-Ray (Th/U, U) logging and lithofacies data. Sediment composition and TOC from cuttings were compared to illustrate how organic matter distribution is influenced by changes in sediment type. Complex mineral composition of heterolithic facies dominated log-response, complicated log interpretation, and thus limited the effectiveness of well logging techniques to predict TOC. Cuttings data reveal three organic-rich intervals, i) at the transition of Upper Mbuo to basal Nondwa Formation with the highest TOC of 3.5 (wt%), ii) in the Nondwa Formation with a TOC of 2 (wt%), and iii) in the Mihambia Formation with a TOC of 2.8 (wt%). An inverse linear relationship between Th/U and TOC signifies relatively shallow deposition in a well-mixed water column under oxic conditions. Intervals with higher TOC imply organic matter preservation under suboxic

conditions upon two marine incursions in the Mihambia and Nondwa Formations, whereas a limnic environment prevailed at the Upper Mbuo to Nondwa transition. Inverse trends in TOC versus clastic components, a positive linear relationship between TOC and carbonates, and higher amount of clastics (>57wt%) in the Mbuo and Mihambia Formations suggest a dominantly clastic sedimentary system with relatively stable carbonate and organic matter input in proximity to the source area. We, therefore, argue that the organic rich intervals observed in the Upper Mbuo, Nondwa and Mihambia Formations were deposited during periods of higher organic productivity driven by a rise in eustatic sea level, and coincide with a decline in terrigenous input from the hinterland.

3.0 Introduction

Evaluation of prospective source rocks in a sedimentary basin is the most critical step in any petroleum exploration stage (Jarvie, 1991; Magoon and Dow, 1994; Peters and Cassa, 1994; Bolandi et al., 2015, 2017; Sêco et al., 2019). It will permit the identification of favorable areas that contain potential source beds capable of generating hydrocarbons and determine the stratigraphic position of these strata with respect to carrier beds (Hunt and Jamieson, 1956; Jarvie, 1991; Harris, 2005; Bolandi et al., 2015). Assessment of source rock is commonly undertaken via determination of the amount of organic matter (TOC), quality (Hydrogen Index), and thermal maturity (R_o , T_{max}) (Tissot and Welte, 1984; Magoon and Dow, 1994; Bolandi et al., 2015). Total organic carbon (TOC) and Hydrogen Index (HI) of source rock are the key parameters that determine the quantity of the generated petroleum and the type of kerogen, which controls the composition of hydrocarbons generated (Tissot and Welte, 1984; Magoon and Dow, 1994; Peters and Cassa, 1994). Quantity of organic carbon in source rock is expressed in weight percentage (wt%) of a sediment sample (Espitalié et al., 1984; Tissot and Welte, 1984; Jarvie, 1991; Magoon and Dow, 1994; Bolandi et al., 2015; Kenomore et al., 2017) and is determined geochemically using either outcrop or core samples. However, outcrop samples often are unavailable for the entire source section, may be weathered and therefore unsuitable for geochemical analysis. Coring, on the other hand, is expensive and an operation rarely performed on source intervals. Hence, cuttings and sidewall core samples are often used as a complementary

indicator of lithology and organic content. Drill cuttings are taken as composites over drilling intervals (3-10m) and hence do not provide accurate lithological information at higher resolution. Sidewall samples, on the other hand, suffer from a smaller sample size and thus measurement may be inaccurate or not representative (Serra, 2003). Consequently, geochemical analyses taken on cuttings and side core samples provide discrete points of data at unsatisfying resolution. This limitation eventually will render TOC measurement from cuttings/sidewall core to be stratigraphically non-representative of total organic carbon (TOC) variability with respect to lithofacies (Brosse et al., 1988; Bessereau et al., 1990; Serra, 2003).

To fill sampling gaps and eliminate depth uncertainties introduced by cuttings samples, borehole geophysical logs have been effectively used to evaluate source rock (Passey et al., 1990; Bessereau et al., 1990; Serra, 2003). Wireline geophysical logging techniques offer a nearly continuous analysis of the formation (often with high vertical resolution ~15cm) (Rider, 2000; Serra, 2003) and thus they are more representative of mean sediment properties even in heterogeneous rocks than cuttings/core data (Serra, 2003). The difference in physical properties between organic matter and mineral components of its host rock has enabled quick source rock identification from wireline logs (Meyer and Nederlof, 1984; Passey et al., 1990; Herron, 1991; Rider, 2000; Kamali and Allah Mirshady, 2004; Bolandi et al., 2015, 2017). Source rock intervals can be recognized by their lower density, lower sonic transit time, higher resistivity, and higher Gamma-Ray signal (Schmoker and Hester, 1983; Meyer, and Nederlof, 1984; Fertl et al., 1986; Herron, 1987; Fertl and Chilingar, 1988; Herron, 1991; Kenomore et al., 2017). To date, several studies have demonstrated the application of wireline logs for source rock evaluation including; natural Gamma-Ray (GR) log (Supernaw et al., 1978; Schmoker, 1981), density (RHOB) (Schmoker, 1979; Schmoker and Hester, 1983), sonic (DT) (Dellenbach et al., 1983), a combination of resistivity (ILD), density, and sonic logs (Meyer and Nederlof, 1984), a resistivity (ILD)-porosity (NPHI) logs overlay (Passey et al., 1990), a sonic and resistivity combination (Carpentier et al., 1991), and multivariate analysis of log data (Mendelson, 1985).

An understanding of total organic carbon (TOC) variability in sediments from the Pindi Group of Lower Jurassic successions of the Mbuo, Nondwa and Mihambia Formations in the Mandawa Basin, Tanzania is hampered by a limited number of wells drilled, high degree of weathering,

halokinesis, and incompleteness of outcrop sections. Thus only a few attempts (Kagya, 1996; Hudson, 2010; Emanuel et al., 2020) relying on drill cuttings samples, which are rather discontinuous in extremely heterogeneous formations of the Pindiro Group have been attempted. As a consequence, TOC distribution with respect to lithofacies to date has been poorly documented. This study attempts to provide comprehensive TOC distributions in sediment successions of the Mita Gamma-1 well in the Mandawa Basin by integrating wireline log derived and cuttings sample data. We apply univariate regression and $\Delta\log R$ techniques to predict semi-continuous evolution in TOC, whereby the reliability of each technique was validated to obtain the best approach when estimating TOC in the Mandawa Basin. To improve interpretation, Spectral Gamma-Ray logs (Th, U, K) and sediment lithotype (organic matter, clastics, evaporites and carbonates) data were used to assess the redox conditions prevailing and reconstruct the basin-fill history for the Jurassic successions.

3.1 Geological and stratigraphy setting

The onshore Mandawa Basin at the southeastern coast of Tanzania (**Fig. 3.1**), is one amongst of many coastal sedimentary basins on the East African continental margin that evolved in response to the breakup of Gondwana during the early Mesozoic era (Kent et al., 1971; Geiger, 2004; Nicholas et al., 2006; Hudson, 2010; Hudson and Nicholas, 2014; Caracciolo et al., 2020). The basin is bound to the South by the East-West trending basement high of the Ruvuma saddle, to the North by the Rufiji trough, to the West by the border fault of the Neoproterozoic Masasi basement spur and covers an area of approximately 15,000km² (Alconsult and TPDC, 1997; TPDC, 2003; Geiger, 2004; Stratadata, 2006).

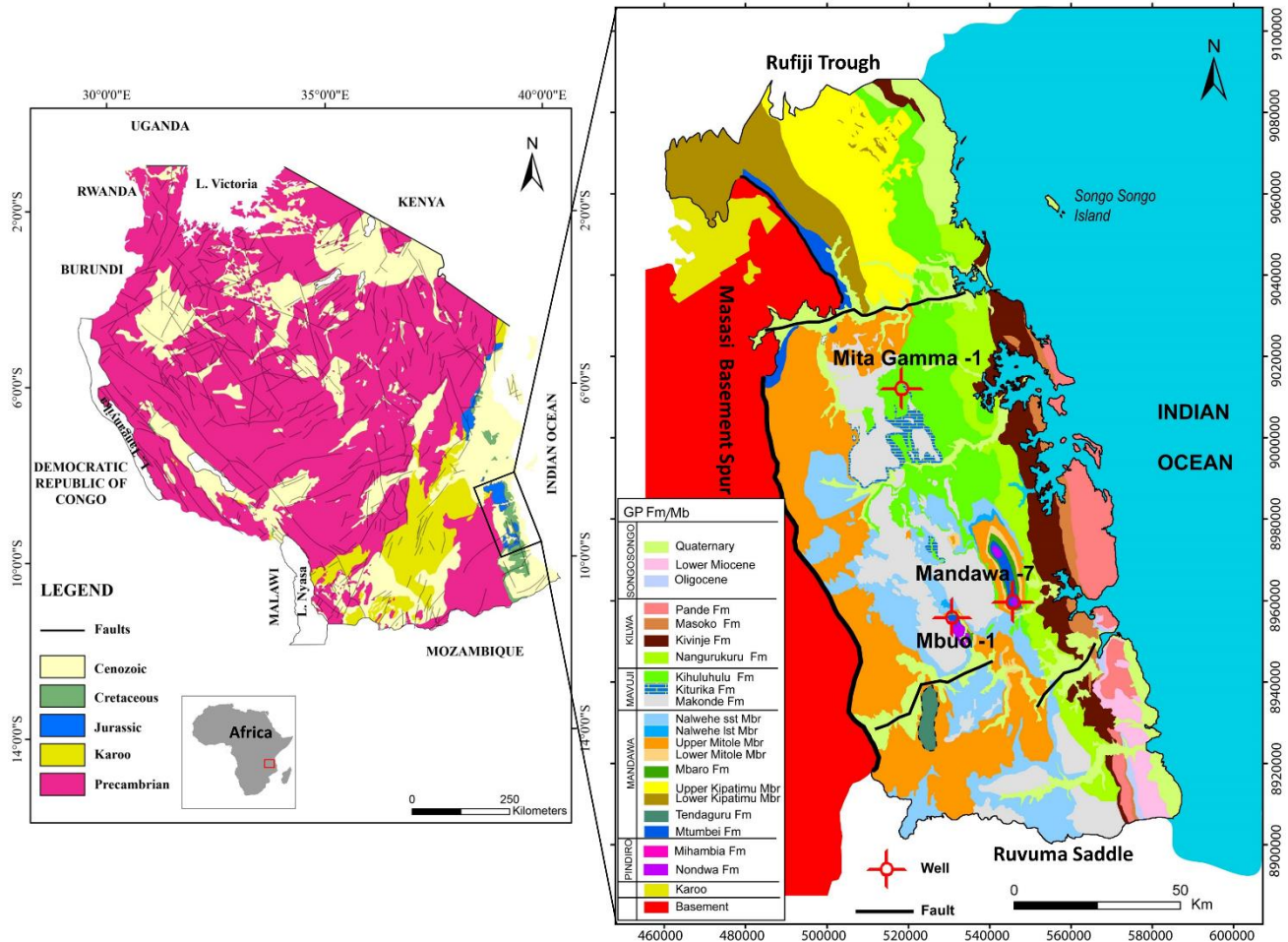


Fig. 3.1: Geological map of Tanzania, showing the location of the Mandawa Basin and position of the Mita-Gamma-1 well investigated, as well as the Mbuo-1 and Mandawa-7 wells (adapted from Pinna et al., 2008; Hudson, 2010; Barth et al., 2016). The Mbuo Formation has not been encountered outcropping at the surface.

Early Mesozoic sedimentation was mainly controlled by tectonics, fluctuation in sea level, and climate. During the basin initiation phase, sedimentation was predominantly clastic and filled graben-like structures, with occasional marine incursions (Mpanda, 1997; Kapilima, 2003). Further subsidence coupled with marine ingression probably during the Early Jurassic resulted in restricted marine conditions and widespread deposition of evaporites (Kent et al., 1971; Kagya, 1996; Caracciolo et al., 2020). Continued rifting during the late Early Jurassic led to an open marginal marine setting, whereby influx of terrigenous supply from the hinterland hindered the development of fully open marine conditions (Hudson and Nicholas, 2014). These were established in Bajocian times manifested by marine carbonates in the Middle Jurassic Mtumbei

Formation (Kent et al., 1971; Balduzz et al., 1992; Kapilima, 2003; Hudson, 2010) (**Fig. 3.1**). Marine to partly terrestrial sediment accumulation continued throughout the Neogene. The five main stratigraphic groups in chronological order are: Pindirolu, Mandawa, Mavuzi, Kilwa, and Songosongo (Hudson and Nicholas, 2014; Fossum et al., 2019). This study here focuses on the late Triassic-Early Jurassic successions (Pindirolu Group) that constitute the Mbuo, Nondwa, and Mihambia Formations (**Fig. 3.2**). The Mbuo Formation lies unconformably upon the basement. It is the oldest sedimentary deposit and has only been encountered in subsurface, with thicknesses ranging from 140m to 1770m (Hudson, 2010). The entire succession is mostly composed of thinly interbedded shales, siltstones and coarser to conglomeritic sandstones, with occasional thin limestone beds (Corelab and TPDC, 2009; Hudson and Nicholas, 2014; Robertson and CGG, 2017; McCabe, 2021). The formation is subdivided into two members (**Fig. 3.2**), the lower member consists of sandstones with silt-claystones and occasional salt-clays while the upper member is mainly made up by thick claystones with alternating thin dolomite beds and occasional thin sandstone beds (Stratadata, 2006; Corelab and TPDC, 2009; Hudson and Nicholas, 2014). A facies interpretation of sediments from the Mita Gamma-1 well (**Figs. 3.2 and 3.10**) reveals that the lower Mbuo member was deposited in a fluvial-deltaic environment, whereas the Upper Mbuo member was deposited in a lacustrine/limnic environment that episodically may have experienced some marine incursions. Such occasional and minor marine incursions have been recorded in the Mandawa-7 and Mbuo-1 wells of the deeper basin (**Fig. 3.1**) (Kagya, 1996; Hudson, 2010), where these features were encountered in both the lower and upper members of the Mbuo Formation and typify a basin initiation stage (Lambiase, 1990). The Nondwa Formation overlies the Mbuo Formation and comprises thick blocky units of halite interbedded with thinly interbedded anhydrites, shales and carbonates (Stratadata, 2006; Quinton and Copestake, 2007; Corelab and TPDC, 2009). It was deposited in a shallow restricted marine embayment, and is overlain by the Mihambia Formation. This is one of the oldest outcropping sedimentary units in the Mandawa Basin (**Fig. 3.2**). The true thickness of this formation is difficult to establish in surface outcrops due to salt diapirism and varies greatly in the subsurface from 58m to 2802m (Quinton and Copestake, 2007; Corelab and TPDC, 2009; Hudson, 2010; Hudson and Nicholas, 2014). The Mihambia Formation lies on top of the Nondwa Formation (**Fig. 3.2**) and is constituted mainly by clastic sediments with minor carbonates (**Fig. 3.2**) deposited in a shallow

marine setting (Hudson, 2010; Hudson and Nicholas, 2014). The formation has different thicknesses in the subsurface, ranging from 247 meters to 598 meters (Corelab and TPDC, 2009; Hudson and Nicholas, 2014; McCabe, 2021).

3.2 Materials and methods

3.2.1 Wireline logs data acquisition and processing

Gamma-Ray (GR), bulk density (RHOB), sonic (DT), photoelectric factor (PEF), and deep resistivity (ILD) log measurements were acquired in an uncased wellbore by Halliburton®, with a data spacing of 0.15 ft. All logs were processed and available in composite format in both analog and digital form (LAS) following Halliburton® standardized procedures. LAS format data was converted into an Excel® spreadsheet and averaged over 10ft intervals to match with cuttings sample intervals. Both averaged logs (low resolution) and high resolution (0.15ft) data, were utilized in the wire logs interpretation and source rock evaluation.

3.2.2 Passey's ΔLOGR techniques

Sonic (DT) – deep resistivity (ILD) logs overlay was employed to identify and quantify TOC following procedures established by Passey et al. (1990). Organic-lean intervals were identified where the two signal traces showed overlap; then the resistivity and sonic values chosen in this interval were used to define the baseline. Similarly, organic-rich intervals were identified by the maximum separation between the two signal traces and used to define ΔLogR , which depends on the level of organic maturation and has a linear relationship with TOC (Passey et al., 1990). Following definition of the baseline, ΔlogR values for the entire Lower Jurassic successions (Mbua, Nondwa, and Mihambia Formations) were computed using **equation (1)**. LOM in **Equation (2)** was derived from the correlation with R_r % using the vitrinite reflection scale (Hood et al., 1975), with R_r % values of the studied well adapted from the literature (Corelab and TPDC, 2009; McCabe,

2021). No appropriate baselines were achieved when overlying the bulk density and Neutron (Porosities) – (ILD) deep resistivity logs. Therefore, TOC estimates were computed based on empirical sonic and resistivity values using **equation (3)**.

$$\Delta \log R_{\text{sonic}} = \log(R / R_{\text{Baseline}}) + 0.02 (\Delta t - \Delta t_{\text{Baseline}}) \quad (1)$$

$$\text{LOM} = 69.925Vr^3 - 153.29Vr^2 + 115.66Vr - 20.453 \quad (2)$$

$$\text{TOC}_{\text{sonic}} = \Delta \log R_{\text{sonic}} * 10^{2.297 - 0.1688 * \text{LOM}} \quad (3)$$

3.2.3 Total carbon

Total carbon (TC) was determined by the use of a VARIO-III elemental analyzer (Elementar®) employing standard procedures described in Ruebsam et al. (2018) using 10mg weighted into a tin capsule. Results were reproducible by $\pm (0.01 \%)$.

3.2.4 Total organic carbon (TOC)_{raw}, TOC_{cbt-free}, and TOC_{cbt+salt-free}

Total organic carbon was determined by decarbonatization of ~50mg of powdered sample weighted into a ceramic crucible and successively treated with 10% and 25% HCl to remove calcite and dolomite, if present. After washing to neutral and drying, samples were analyzed in an Eltra Helios CS analyzer (Eltra®) following standard protocols described in Edegbaei et al. (2020). Results were reproducible by $\pm (0.04\%)$.

TOC_{cbt-free} was calculated from the ratio of TOC_{raw} and the proportion of weight loss upon decarbonatization expressed in % as $((\text{TOC}_{\text{raw}} / (100 - \% \text{ loss})) * 100)$. Similarly, TOC calculated for a carbonate- and salt-free (TOC_{cbt+Sf}) fraction was calculated as $(\text{TOC}_{\text{raw}} / (100 - \% \text{ carbonate plus salt})) * 100$.

3.2.5 Sediment composition (carbonates, anhydrites, clastics)

Determination of cumulative carbonate and evaporite mineral contents was achieved via dissolution and removal by 10% and 25% HCl treatment of a ~800-1200 mg sample aliquot followed by washing with deionized water for neutralization and drying at 40° C for 24.

The combined proportions of salt and carbonate in the sample were determined from the weight loss upon treatment. Carbonate contents were calculated from total inorganic carbon (TIC) obtained by the difference of TC – TOC. TIC was converted into weight % carbonate by multiplying with a stoichiometry factor of 8.33, assuming calcite as the main carbonate phase. Proportions of evaporite minerals (halite, gypsum, anhydrite) were determined from the difference between cumulative carbonate + salt fractions versus carbonates. The difference between the cumulative carbonate + salt fractions and the total sample weight was taken to represent the clastic fraction, ignoring the minor contributions of organic matter and sulfides.

3.2.6 Total organic carbon (TOC) derived from the empirical relationship between measured TOC and well log data

TOC based on the log-derived model was determined via comparison of the measured TOC of cuttings samples with their respective density, DT, GR, and ILD log data in binary diagrams following procedures described previously (Supernaw et al., 1978; Meyer and Nederlof, 1984; Mendelson, 1985; Fertl et al., 1986; Khoshnoodkia et al., 2011; Alshakhs and Rezaee, 2017). The aim was to derive separate linear regression equations used to compute TOC values for sediments from the Lower Jurassic sections. Four linear regression equations (4-7) versus TOC were obtained: GR-TOC (4), DT-TOC (5), RHOB-TOC (6), ILD-TOC (7). Log response to the organic matter concentration was determined directly by the magnitude of the regression slope coefficient (M) and correlation coefficient (R^2) of individual well log and cuttings data.

$$\text{TOC} = -0.0081 \text{ GR} + 1.5628 \quad (4)$$

$$\text{TOC} = -0.0133 \text{ DT} + 1.8899 \quad (5)$$

$$\text{TOC} = -0.2092 \text{ RHOB} + 1.5154 \quad (6)$$

$$\text{TOC} = 0.1058 \ln (\text{ILD}) + 0.6031 \quad (7)$$

3.4. Results

3.4.1 Geophysical well-log responses to various lithologies

Lithologies from well logs have been identified based on responses/characteristics patterns of logs suites (GR, DT, RHOB, ILD and PEF) within a particular unit. Results of the examined logs suite are generally comparable with the gross lithology derived from cutting samples (**Fig. 3.2 and Supplementary Table 3.1**). High-resolution logs signatures provide more detailed lithological information than low resolution logs signals averaged over (3m) intervals of the succession corresponding to drill cuttings samples. Variability in lithologies, bed boundaries, presence of thin beds, and apparent thickness of beds are better recognized from high-resolution wells logs than from the averaged well logs data. DT or GR logs appear to be more sensitive to small variations in lithology as indicated by the high degree of serrated curves compared to other logs (**Fig. 3.2**). The most of the irregular logs suite patterns correspond to shale, sandstones, siltstone or carbonate alternations in either the Mbua or Nondwa Formations, whereas evaporites and carbonates register more regular and often monotonous log patterns. Diagnostic features shown by the latter two lithologic units are lower GR, RHOB, and DT, very high ILD, higher PEF for evaporites and lower GR, lower DT, higher RHOB, ILD, and PEF for carbonate, respectively. Other lithologies such as shale, siltstone, and sandstone exhibit much variability and overlap in their (RHOB, ILD, PEF) logs signals, likely related to complex mineral composition. Maxima in GR or DT occur in shale intervals whereas siltstone, sandstone, and minor shale intervals generally show low-moderate response in their GR or DT.

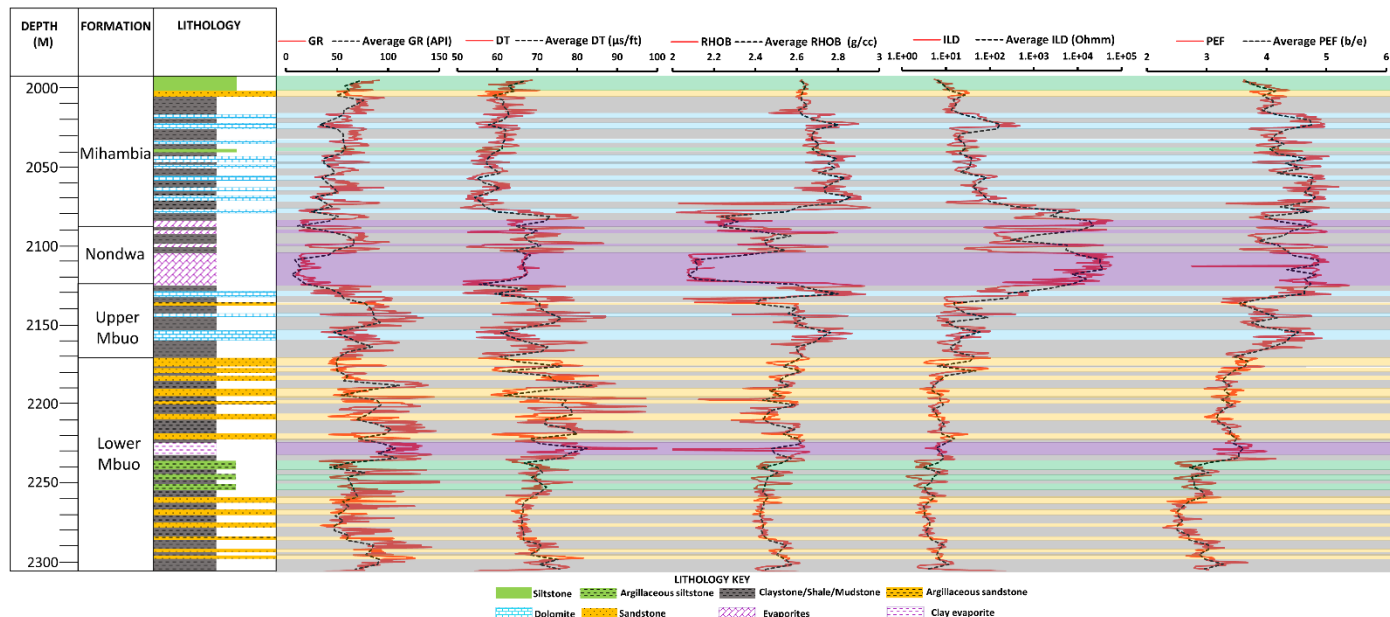


Fig. 3.2: Comparison of high- and low-resolution GR, DT, RHOB, ILD and PEF logs responses to various lithologies penetrated by the Mita Gamma-1 well (high- and low-resolution logs are shown by red curve and black dotted traces, respectively).

3.4.2 Sediment composition and total organic carbon (TOC %)

TOC analysis of well-cuttings samples reveals that sediments from the Pindiro Group (Mbuo, Nondwa and Mihambia Formations) exhibit high variability in TOC due to alternations of organic-rich and organic poor beds. TOC values range between 0.2 to 3.5 (wt%) with an average of 0.98 (wt%) (**Supplementary Table 3.2**). TOC values in sediments of the Lower Mbuo Formation range from 0.37 to 1.97 (wt%) with an average of 0.65 (wt%), whereas in Upper Mbuo Formation they range from 0.54 to 3.45 (wt%) with an average 1.36 (wt %). In sediments of the Nondwa Formation, TOC values range from 0.2 to 2.2 (wt%) with an average of 0.9 (wt%) and in sediments of the Mihambia Formation TOC values range from 0.50 to 2.82 (wt%) with an average 1.31 (wt%). Good to very good hydrocarbon source potential based on TOC occurs mainly in shaly intervals and poor to fair source potential occurs in evaporites and sandstones (**Fig. 3.3**). The overall TOC

pattern reveals three major maxima at 2113.5-2144.5m at the Upper Mbuo-basal Nondwa Formation boundary, at 2063.5-2107.5 m in the Nondwa-Mihambia Formation boundary and 2012.5-2063.5 m in the Mihambia Formation. The TOC content in sediments of the Lower Mbuo Formation is generally low (<1wt%) as opposed to the Upper Mbuo Formation where TOC is higher (>1wt%). Sediment composition derived from logs shows high stratigraphic variability (**Fig. 3.2**) and in general is in agreement with lithology derived from cuttings. The amount of clastics is higher in sediments of the Mbuo and Mihambia Formations (>57.5 %) and generally lower in evaporitic intervals. CaCO₃ ranges from 0-34 % with an average of 13 %, clastics range from 7 to 85 %, averaging 57 %, and salt/evaporites range from 1-91 %, with an average of 30 % (**Fig. 3.3 and Supplementary Table 3.2**). TOC concentrations correlate positively with carbonate% (**Figs. 3.3 and 3.11**) and covary negatively with the amount of clastics % (**Figs. 3.3 and 3.11**). Maxima in TOC in sediments of the Nondwa and Mbuo Formations are bound to thin shale layers embedded in evaporites or carbonates (**Fig. 3.3**). If TOC is calculated on a carbonate- and evaporite-free basis to eliminate dilution by high sedimentation rates, sediments from the Upper Mbuo, Nondwa and Mihambia Formations show overall higher TOC values than those of the clastic dominated Lower Mbuo Formation (**Fig. 3.3 and Supplementary Table 3.2**).

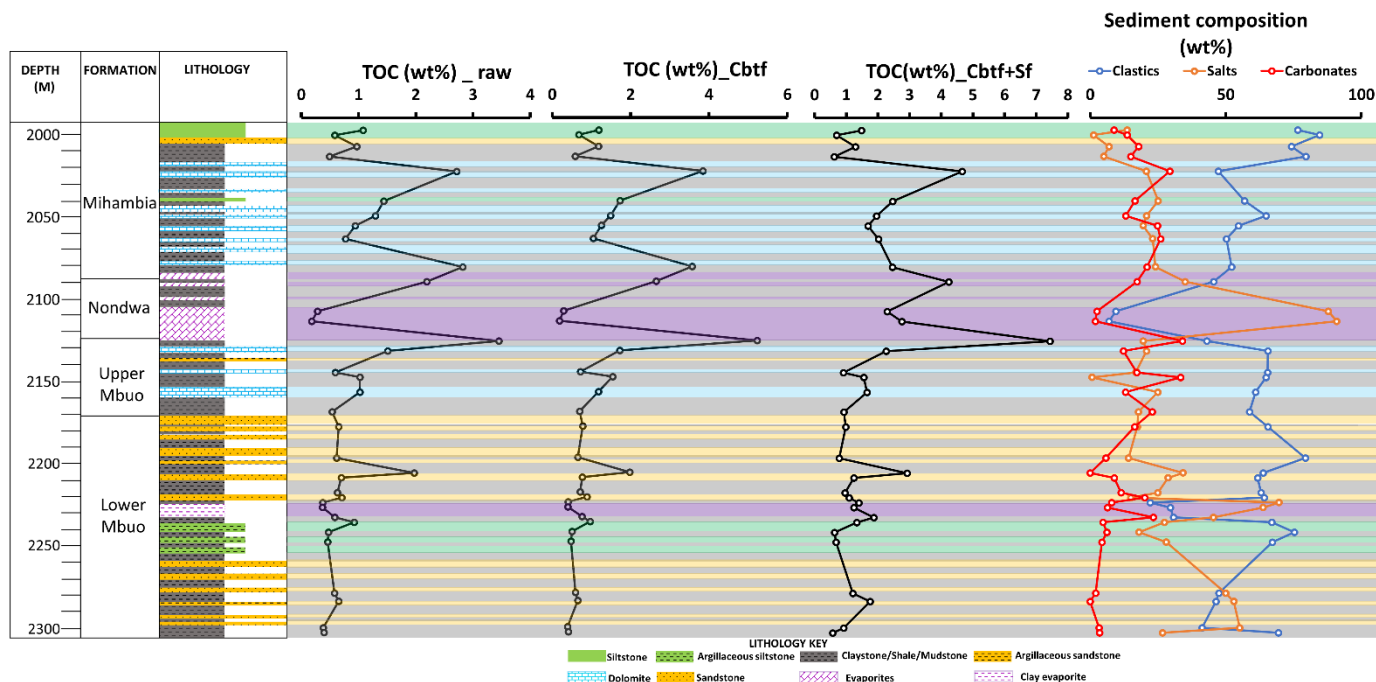


Fig. 3.3: Variability in TOC_{raw} , TOC_{cbtf} , $TOC_{cbtf+sf}$ and their relationship to the sediment composition of the cuttings samples.

3.4.3 GR, RHOB, DT, and ILD logs responses in organic-rich intervals

Direct log responses in organic-rich intervals were determined from the correlation between averaged GR, DT, RHOB, and ILD log suites and TOC from cuttings samples as shown by binary diagrams (**Fig. 4**). Log-derived TOC values predicted via **equations 4-7** generally revealed a poor correlation, if based on GR ($R^2=0.07$), RHOB ($R^2=0.002$), DT ($R^2=0.02$) versus TOC determined on cuttings samples. Log responses versus TOC measured on cuttings yielded negative coefficients of regression for GR ($M=-0.008$), RHOB ($M=-0.209$) and DT ($M=-0.013$), if all samples were included in the regression analysis (**Fig. 3.4**) and positive coefficients for ILD. A more differentiated distribution was obtained, if the five samples richest in TOC were excluded from regression analysis (**Fig. 3.4**). GR values averaged for cuttings intervals range from 16 to 120 API with an average of 72 API (**Supplementary Table 2**), whereby the highest GR signal corresponds to a TOC-lean salt-clay (**Fig. 3.4a**). DT values range from 55 to 82 ($\mu s/ft$) lacking covariation with TOC, whereby the highest travelling times again occur in TOC-poor salt-clays (**Fig. 3.4b**). Whereas the majority of the samples in the RHOB-TOC cross plot shows RHOB increasing with TOC, the five

samples with highest TOC deviate from this trend by showing lower density values (**Fig. 3.4c**). The relationship between ILD and TOC shows a reasonable positive correlation ($R^2=0.5$), if the two pure evaporite samples are excluded.

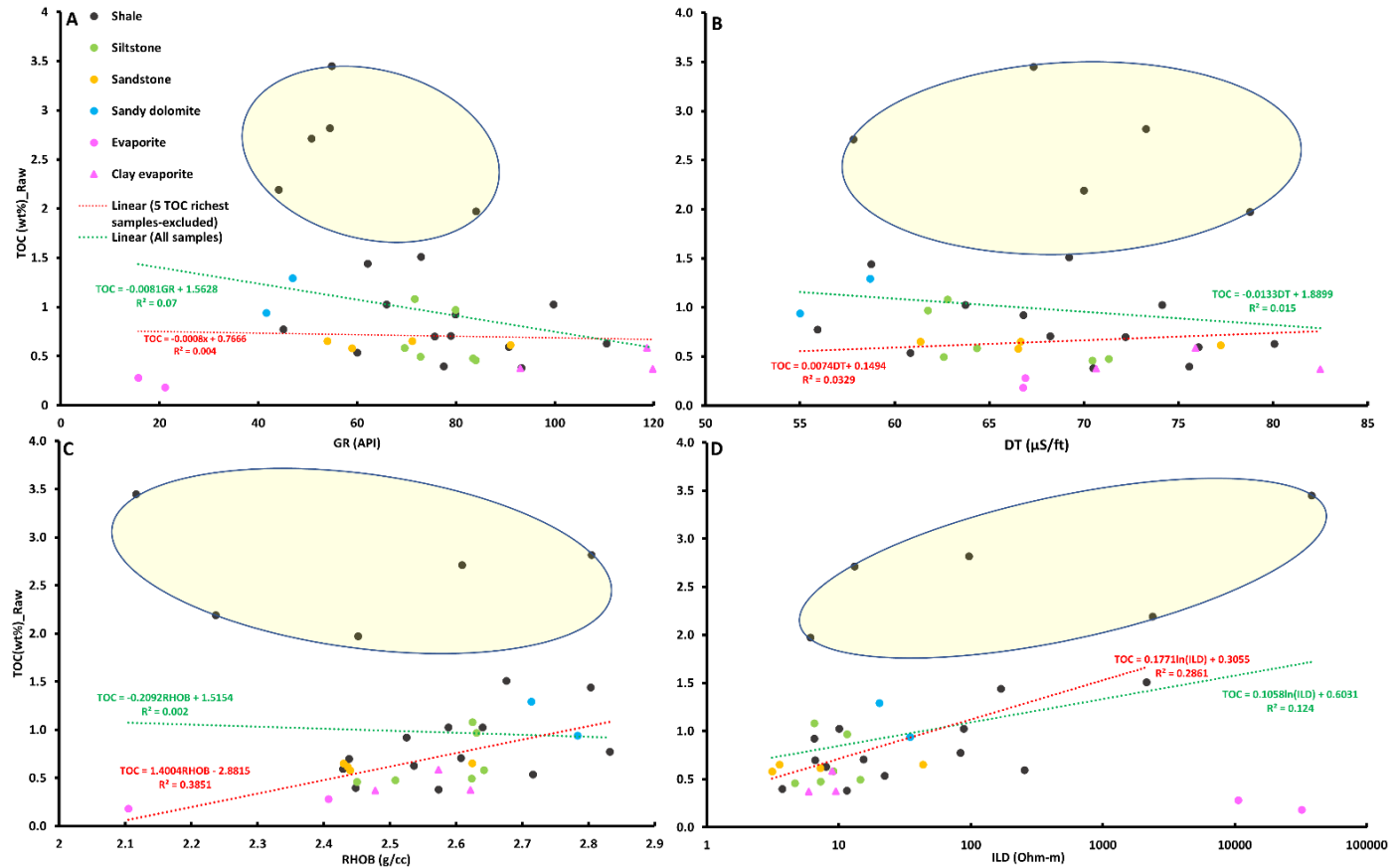


Fig. 3.4: Cross-plot depicting relationship between TOC_{raw} measured from cuttings samples and well log responses GR, DT, RHOB, and ILD averaged for intervals from which cuttings were collected. Regression lines shown are for all samples (green) and and excluding the five TOC-richest samples (red).

3.4.4 ILD-DT separation (ΔLOGR) as an indicator of organic richness in shale intervals

In addition to porosity (Neutron-Density, Sonic), resistivity log responses are used for identification of organic-rich intervals in various lithological units. Signal trace overlays at proper scale can further guide in the identification and depth profiling of organic-rich intervals (**Fig. 3.5**). Deep induction resistivity (ILD) and sonic (DT) traces overlapped at 2262m in the Lower Mbua Formation, giving baseline values of approximately 3 (ohm-m) for ILD and 67.5 (μS/ft) for DT

(Fig.3. 5). In contrast, no ILD-DT log-overlap (baseline) occurred in the Upper Mbua, Mihambia and Nondwa Formations. Separation of ILD-DT log-traces in shale intervals reveal the presence of organic matter. In contrast, an ILD-DT logs overlay points to an organic lean formation. Both former and latter observations are in agreement with GR and TOC measured from cuttings. Separation in sandstone/carbonate intervals may indicate the presence of reservoir lithology, whereas siltstone intervals commonly do not show any ILD-DT separation (see depth intervals of 2237-2240, 2245-2249, 2250-2255m). Differential ILD-DT separation within shale intervals along depth is directly linked with TOC variability. ILD-DT log-separation in a few intervals, however, appears to be correlated with corrected TOC. For example, in the salt-clay interval at 2225-2230m, ILD-DT log-separation is higher correlated with $(TOC_{cbtf+sf})$ than with TOC_{raw} .

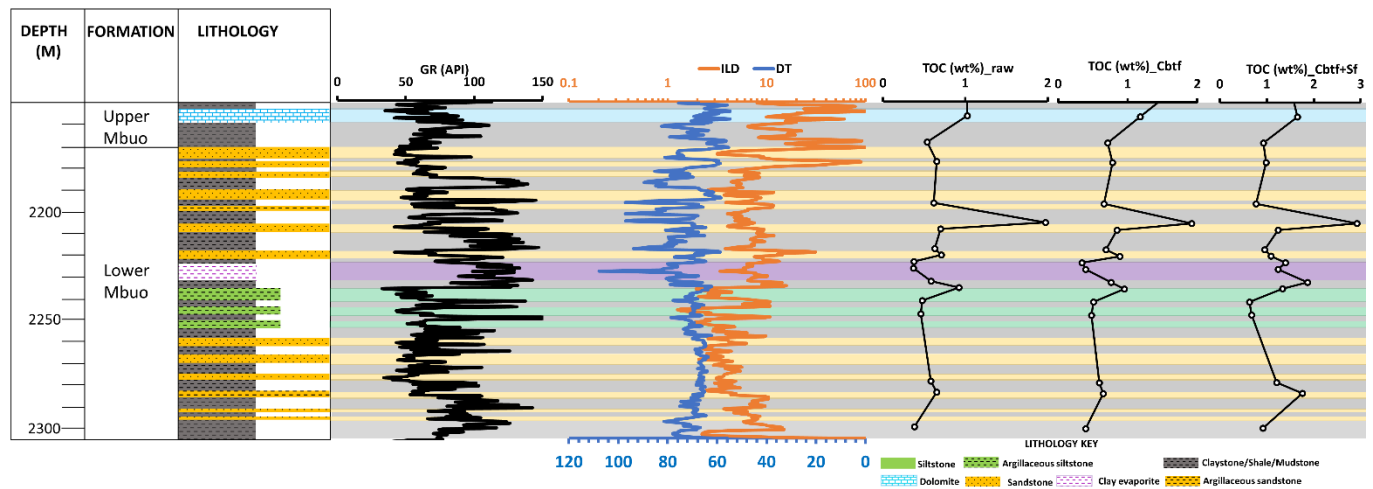


Fig. 3.5: Comparison of ILD-DT-overlay and GR to the TOC content measured on cuttings samples. ILD and DT log trace separation indicates presence of organic enriched intervals in shales.

3.5. Discussion

3.5.1 Total organic carbon (TOC) variability

3.5.1.1 Variability in TOC derived from log responses and cuttings data

Most data analyzed show that TOC computed by the models applied here are not in agreement with the TOC values obtained from geochemical analyses of cuttings samples. The GR, RHOB, and DT log responses are less influenced by the presence of organic matter, if compared to other dependencies such as variable mineral composition of the rock matrix, fluid types and pore space volume (Mendelson, 1985; Passey et al., 1990; Jacobi et al., 2008; Kenomore et al., 2017). Variability and discrepancies of log-derived TOC to the measured TOC along the sediment succession reflect changes of these mainly lithology-dependent variables (**Fig. 3.6**). TOC predicted from ILD in contrast, shows a reasonable correlation with TOC measured on cuttings. Log responses in the evaporitic Nondwa and for the carbonate-rich intervals in the Mbuo and Mihambia Formations appear to be highly influenced by lithology and thus affect predicted TOC (**Fig. 3.6**).

Comparison of the correlation coefficient (R^2) of all four logs indicates that in the sediments studied here, the deep-induction resistivity log (ILD) is more sensitive to the organic matter as compared to the GR, DT, and RHOB logs.

The inconsistencies between log-predicted and measured TOC to a large extent may be due to spatial resolution and depth matching between logging depth and driller's depth. In order to allow comparison with cuttings, log signals (GR, DT, RHOB) were averaged over 3m intervals resulting in loss of depth resolution. Depth assignments for wireline-logging and for cuttings is assumed to vary by about one meter as seen by inconsistencies between cuttings lithology and logging signal (**Fig. 3.2**). Hence, correlation between cuttings collection intervals and averaged log-data will be affected by offsets in depth assignment. Geophysical logs differ by spatial resolution or penetration depth (Rider, 2000) and here ILD tends to cumulate the whole bed effects (thin beds) and therefore is not much affected even when averaged (Rider, 2000). Thin bed effects are not registered in the high-resolution ILD logs indicating poor bed resolution as a typical feature in ILD logs and therefore average ILD logs will be always the least affected by resolution issues compared to other logs. This is expressed in the best correlation of ILD with TOC in the data set generated here.

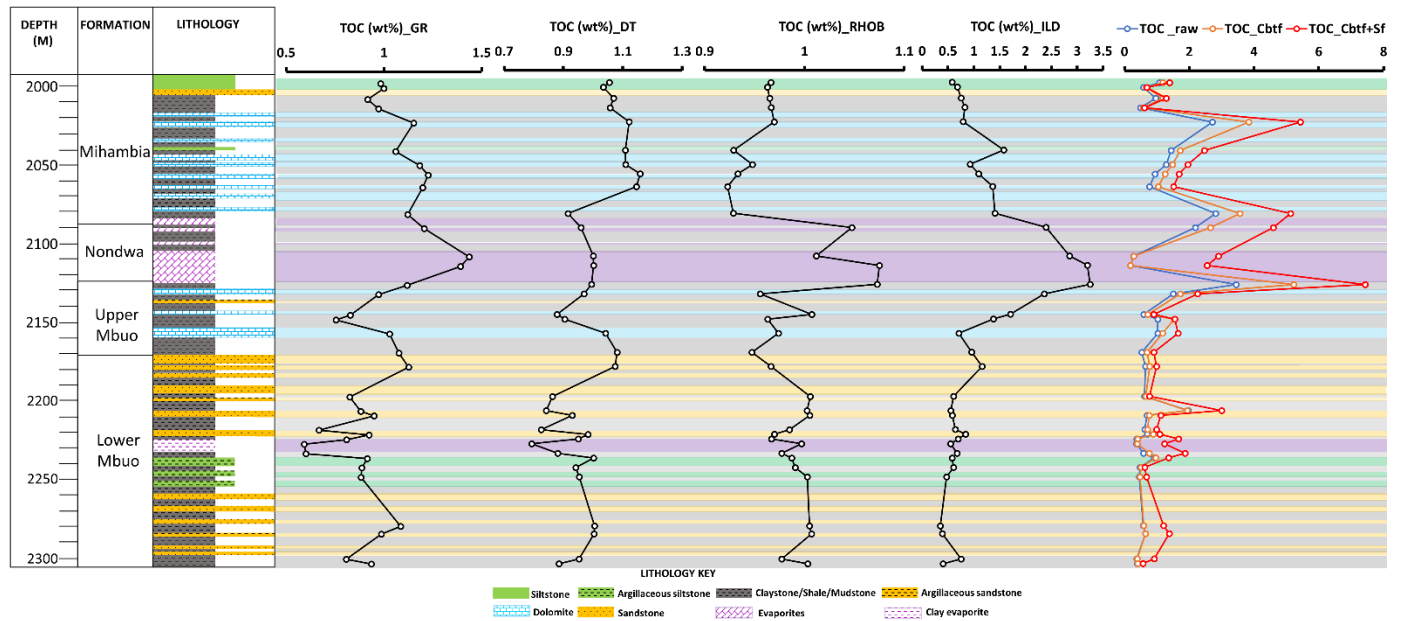


Fig. 3.6: Comparison of TOC content predicted from GR, DT, ILD, RHOB log response versus raw and corrected TOC concentrations determined on cuttings (cbtf = carbonate-free, sf = salt-free).

In addition to the loss of depth resolution caused by averaging well logs signals across 3m intervals of sedimentary successions, the observed discrepancy between measured and predicted TOC could be explained by preferential responses to varying and complex mineral composition over organic matter concentration in a heterolithic facies (Meyer, and Nederlof, 1984; Mendelson, 1985; Fertl et al., 1986; Rider, 2000; Jacobi et al., 2008). During the acquisition of well logs, the variable mineral composition of the rock matrix, fluid types, pores space volume, and amount of organic matter contained in a sediment are all taken into account (Rider, 2000). Therefore, a substantial amount of organic matter must be present in a sediment for source rock intervals to be evaluated by well logging techniques (Schmoker, 1979; Fertl et al., 1986; Khoshnoodkia et al., 2011). Typically, organic rich intervals have higher U concentrations ranging from 15-60 (ppm), higher DT ranges ranging from 150-1985 ($\mu\text{s}/\text{ft}$), very high ILD, and relatively lower RHOB of 1.2 (g/cc), all of which are often positively correlated with TOC concentration (Meyer and Nederlof, 1984; Rider, 2000; Alizadeh et al., 2018). In contrast, the log data examined here generally correlate negatively with the TOC, indicating that the mineral composition of heterolithic facies predominates over the organic matter response.

3.5.1.2 Variability in ΔLogR derived TOC

The TOC calculated via the model has been validated against the TOC measured on cuttings samples and the corrected TOC (TOC_{cbtf} , $\text{TOC}_{\text{cbtf}+\text{sf}}$) that eliminates dilution due to elevated carbonate and evaporite concentration, as shown in **Fig. 7**. Most of the results presented here show that TOC derived by the model is highly variable along the stratigraphic succession due to an alternation of organic rich and poor beds of different lithologies. The ΔLogR derived TOC is in good agreement with measured TOC of cuttings samples taken at 2131.5, 2147.5, 2208.5, 2217.5, and 2220.5m in the Mbuo and at 2000.5, 2007.5, 2013.5, 2040.5, 2049.5m in the Mihambia Formations. In contrast, TOC values are higher than those of cuttings samples in the Nondwa Formation and in the Mbuo Formation at 2132, 2155, 2205, 2215, 2235m. In the Mihambia Formation at 2038, 2060, and 2065m, the predicted TOC seems to correlate more closely with TOC corrected than with TOC raw. The ΔlogR model data revealed no TOC in sandstone or siltstone intervals, and further implies that organic lean shale is present in the sandstone-shale sequence at 2260-2280m in the Lower Mbuo Formation (**Fig. 3.7**).

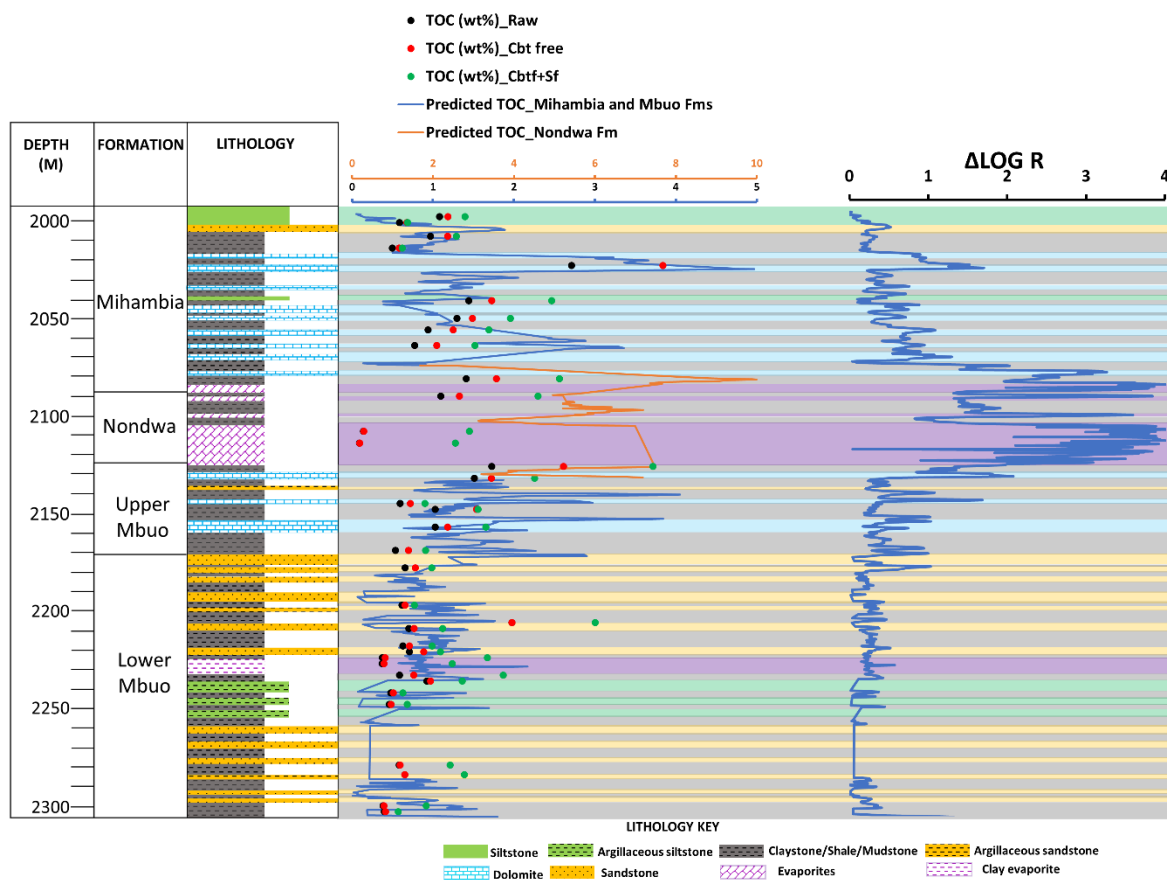


Fig. 3.7: Variability in the predicted and measured TOC content and its relationship to $\Delta \text{LOG R}$ (obtained using the Passey et al. 1990 technique).

A comparison of the TOC derived from ΔlogR and the TOC from cuttings (**Fig. 3.7**) showed that TOC is highly dependent on changes in lithology, as shown by the anomalous ΔLogR in the evaporites and carbonates (**Fig. 3.7**). These anomalies resulted from an increase in both resistivity and sonic travel time in non-porous and caved evaporites and not from organic matter content in the sediment. Therefore, TOC predictions in these intervals would be unreliable.

Apart from the change in lithologies, TOC inconsistencies are due to differences in the vertical resolution of sonic/resistivity tools versus cuttings samples (Passey et al., 1990; Sêco et al., 2019; Tenaglia et al., 2020). TOC values calculated from well logs represent 15cm of sedimentary successions whereas the cuttings samples represent 3m composite intervals. Thus, TOC concentration from cuttings in the organic-rich and lean interbeds similar to the Mbuo and Mihambia Formations represents an average value (Huang et al., 1994, 1996) and thus is incomparable to the TOC derived from logs. In addition to deviations in TOC calculations due to

differences in the vertical resolution of cuttings and well log data, uncertainty can also arise from determining baseline values. TOC calculations according to the Passey et al. (1990) technique are based upon a single baseline regardless of the depth or position within a sediment succession. This can be unrealistic in heterolithic facies within and similar to sediments of the Pindiro Group (Mbuo, Nondwa and Mihambia Formations) (Passey et al., 1990; Huang et al., 1994; Kenomore et al., 2017).

3.5.2 Influence of depositional conditions on organic matter enrichment and variability

3.5.2.1 Th/U and U concentration as an indicator of depositional conditions

Using the Photoelectric Factor (PEF) and Spectral Gamma-Ray (SGRL) logging data and TOC measured on cuttings samples, we examined changes in depositional conditions that may relate to the accumulation and preservation of organic matter. In general, all of the data confirms that SGRL values and TOC content correlate negatively (**Fig. 3.8a**), (**Fig. 3.9**). The variation in SGRL values of various lithologies is most dramatically influenced by variation in Th and K, whereas variation in U is not as important to the variation in the SGRL values (**Fig. 3.8b-d**). The TOC generally decreases with increases in the Th/U ratio (**Fig. 3.9a**). Evaporites, clay evaporites, carbonates, sandstones, and siltstones exhibit lower TOC ranges from 0.2 to 1 (wt%) and relatively higher Th/U ratios (**Fig. 3.9b**). Uranium concentration is generally lower with a broader spread in the TOC-U binary plot (**Fig. 3.9b**).

Basin may have been disrupted by highly oxygenated water influx in the course of high terrigenous clastic sediment input, as a result of changes in tectonics and climate of the evolving graben/depression (Lambiase, 1990; Lüning and Kolonic, 2003; Hudson and Nicholas, 2014; Marín et al., 2021).

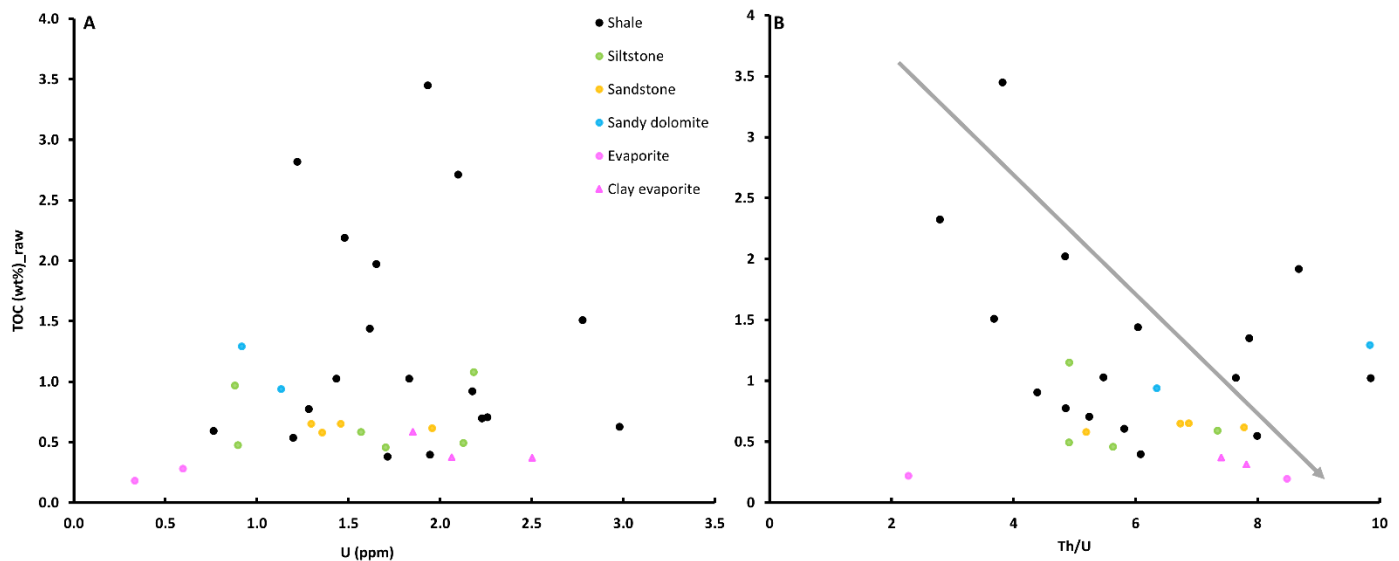


Fig. 3.9: Cross plots showing correlation between TOC versus **A** (U) and **B** (Th/U).

3.5.2.2 Interplay between depositional environment (facies), sediment composition, and organic matter variability

We examined the lithology (extracted from drill cuttings and well logs) and sediment composition data to determine depositional conditions related to TOC content in Lower Jurassic successions. It appears from the variations in lithology and facies that sediments from the Pindiro Group (Mbue, Nondwa, and Mihambia Formations) were deposited in a wide range of depositional conditions ranging from fluvio-deltaic, limnic, and restricted to shallow marine conditions and correlated with TOC concentrations (**Fig. 3.10**). During the early Pliensbachian (Lower Mbue Formation), braided rivers predominated that during the late Pliensbachian evolved into shallow lakes and lagoons. Restricted marine conditions prevailed during the early-middle Toarcian period, switching to sabkha and tidal channels in Late Toarcian-Aalenian times.

An assessment of organic carbon versus depth trends shows that environmental differences exerted control over organic matter because of different productivity, preservation, and destruction, which are directly related to basin evolution and eustatic level changes (Tyson, 1996; Harris, 2005; Holz, 2015) (**Fig. 3.10**). TOC peaks in successions of the Upper Mbuo, Nondwa and Mihambia Formations coincide with decreased clastic input and a rise in eustatic sea levels as opposed to generally lower TOC <1 (wt%) accompanied by higher terrigenous input in successions of the Lower Mbuo Formation (**Fig. 3.10**).

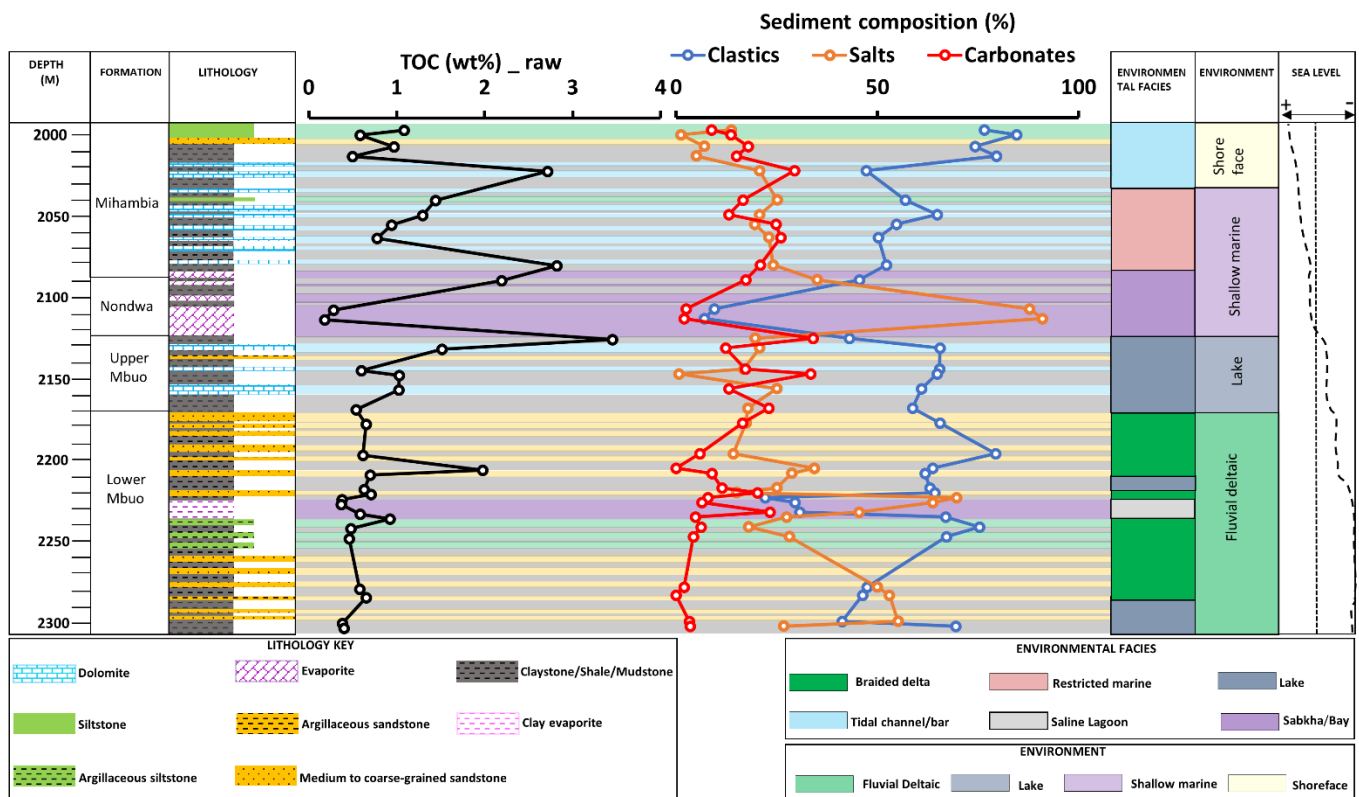


Fig. 3.10: Relationship between changes in the depositional environment to the TOC content and sediment composition.

3.5.2.3 Relationship between changes in sediment input and organic matter

To better understand the controls on the distribution of total organic carbon in the Lower Jurassic sediment, the three primary components of the sedimentary system (carbonate, organic matter, and clastics concentrations) were analyzed following the procedure established by Ricken (1993). Data examined here reveal a linear relationship between TOC and carbonate (**Fig. 3.11a**) and an

inverse relationship between TOC versus clastics (**Fig. 3.11b**). The former observation implies coupled carbonate and organic matter input via aquatic primary productivity in the system at two different productivity levels, represented by the full and the stippled arrow in **Fig. 3.11a**. The latter observation indicates a sedimentary system that is mainly controlled by varying degrees of siliciclastic deposition with constant organic matter and carbonate supply in background sediments (Ricken, 1993; Baudin et al., 2002; Buchem et al., 2005). Furthermore, the TOC-clastics binary plot (**Fig. 3.11b**) indicates that organic matter deposition during the Pliensbachian (Mbua Formation) and late Toarcian-Aalenian (Mihambia Formation) was controlled by clastic dilution, with higher TOC content occurring when there is minimal dilution in the system. The presence of variable TOC content at constant carbonate composition may also be explained by this association (**Figs. 3.10 and 3.11a**).

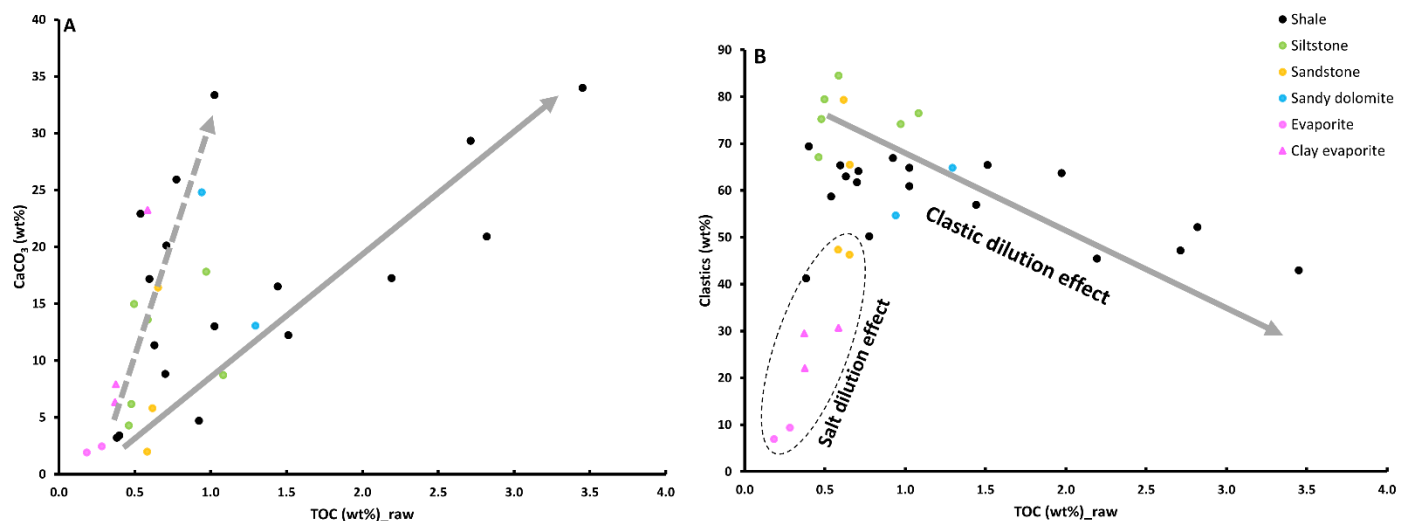


Fig. 3.11: Cross plots depicting relationship between TOC content versus A (carbonates wt%) and B (clastics wt%).

Therefore, we propose that higher terrigenous input and fluctuations in redox conditions in the water column played major roles in the stratigraphic distribution of the Lower Jurassic potential source rocks encountered in the Mita Gamma-1 well of the Mandawa Basin. Patterns of source rock heterogeneity differ in response to changing climate, sea level, and tectonic activity (Lambiase, 1990; Buchem et al., 2005; Hellenen et al., 2020; Marín et al., 2021). During the initial rifting phases (Lower Mbua Formation), dominantly fluvial-deltaic to shallow-lake-transition

conditions of sedimentation led to dilution and generally lower TOC content. Further rifting along with fluctuations in sea level created abrupt changes in the depositional environment, leading to episodically enhanced deposition and preservation of organic material during the Late Pliensbachian to Aalenian (Upper Mbuo, Nondwa and Mihambia Formations).

3.6 Conclusions

In this study, geochemical and wireline geophysical log data were used to constrain total organic carbon variability and depositional conditions in Lower Jurassic successions (Mbuo, Nondwa, and Mihambia Formations) of the Mandawa Basin, SE Tanzania, where organic matter variability is poorly documented. TOC was determined from drill cuttings samples (n=35) and well logs data for sediment covering the entire Lower Jurassic successions (Mbuo, Nondwa, and Mihambia Formations) penetrated by the Mita Gamma-1 well. Logs-derived TOC content revealed complex mineral composition of the heterolithic facies that complicated log interpretation, and thus limited the effectiveness of this technique in predicting TOC. Due to the heterogeneity in lithologies and the relatively low TOC contents in Lower Jurassic successions, TOC estimated via the Passey et al. (1990) methodology gives trustworthy data only for a limited number of intervals with homogeneous lithology within the Lower Jurassic succession.

Cuttings data reveal three organic-rich intervals at the transition of Upper Mbuo to basal Nondwa Formations with the highest TOC of 3.5 (wt%), in the Nondwa Formation with a TOC of 2 (wt%), and in the Mihambia with a TOC of 2.8 (wt%). Comparison of TOC, Th/U and sediment composition data for clastics (%), carbonates (%), evaporites (%) derived from cuttings samples suggests episodic deposition of organic matter under sub-oxic conditions. These were mainly controlled by fluctuation in sea level, tectonics and clastic dilution. Therefore, we hypothesize that during the initial rifting phase, fluvial-deltaic to shallow lake-transition depositional conditions, likely driven by rifting and climate, led to higher terrigenous input and clastic dilution, and generally lower TOC < (1wt%) in the Lower Mbuo Formation. Further rifting coupled with eustatic sea level rises led to abrupt changes in the depositional environment and preservation

of organic matter in the Upper Mbuo, Nondwa and Mihambia Formations corroborated by a decline in the Th/U ratio and the amount of clastic influx.

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Chapter Four

Assessment of kerogen types and source rock potential of Lower Jurassic successions in the Mandawa Basin, SE Tanzania

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Abstract

Variability in organic matter types and quality of source rock intervals across evolving rift-basins has been the focus of previous studies, which have shown that due to dynamic sedimentary regimes and water depths such source intervals occur sporadically and restricted to specific stratigraphic horizons. Such heterogeneous settings, require detailed geochemical analysis to document variations in source quality. We here present a comprehensive geochemical analysis of source rock intervals in Lower Jurassic successions (Mbuo, Nondwa and Mihambia Formations) in the Mandawa Basin, on-shore Tanzania, in order to constrain organic matter types and quality. Organic matter richness was measured by TOC analysis, while quality and thermal maturity of organic matter was assessed by programmed pyrolysis, complemented by organic carbon isotope analysis ($\delta^{13}\text{C}_{\text{Org}}$) of sediment samples collected from outcrops, cores, and cuttings samples. TOC and programmed pyrolysis data from existing literature was synthesized in this study to better understand organic matter quality and thermal maturity evolution on regional scale. These data were examined in relation to paleogeography to gain insight into the factors regulating the distribution of organic matter quality within the Mandawa Basin.

Average TOC in Lower Jurassic sediments fluctuate from 0.31 to 4.83wt% with an average of 1.4wt%, whereas, kerogen yield (S₂) and hydrogen index (HI) remain generally low, with values ranging from 0.01 to 8.6 (mg HC/g rock) with an average (0.92 mg HC/g rock) as well as from 2-233 (mg HC/g TOC) with an average of 46 (mg HC/g TOC), respectively. Organic matter is composed mainly of type III (gas prone) to IV (inert) kerogens with lesser admixture of type II-III kerogen (oil-gas prone), which have been deeply buried and undergone a wider range of thermal

diagenesis, from early to over mature stage. Isotopic composition ($\delta^{13}\text{C}_{\text{org}}$) of organic matter is consistent with kerogen typing, indicating exclusively terrestrial organic matter of heavier $\delta^{13}\text{C}_{\text{org}}$ signature under sea level lowstand conditions that changes to isotopically more depleted marine organic matter upon marine incursions. This implies that the investigated source intervals, have poor to fair source potential, and thus only few intervals of limited thickness in the Upper Mbuo and Nondwa, and Mihambia Formations of the Mita Gamma-1 and East Lika-1 wells, as well as in the Lower Mbuo Formation in the Mbuo-1 well that have been locally enriched due to periodic marine ingression that persisted since the Pliensbachian period possess oil generation potential. A comparison of TOC and programmed pyrolysis borehole data from this study to those compiled from literature in a paleogeographical context reveals high spatial variability. The southwestern and northwestern parts of the study area have the highest source potential, indicating high organic matter productivity in the dysoxic to anoxic water column, which promoted the preservation of organic matter. In contrast, the eastern and western parts generally show low source potential, indicating degraded organic matter and significant terrestrial influx from emergent land that were deposited in shallow, mixed oxygenated water with low to moderate organic matter productivity.

4.0 Introduction

Programmed pyrolysis is the standard and most widely used technique for screening petroleum generation and thermal maturity of source rock (Espitalie et al., 1977; 1984; Peters, 1986; Lafargue et al., 1998; Dembicki, 2009; Hart and Steen, 2015). Analysis is carried out on a small fraction of samples (~50mg) and provides a quick evaluation of organic matter (OM) richness via the total organic carbon content (TOC), OM-types (Hydrogen Index) and level of thermal maturity (T_{max}) of a source rock. These parameters are commonly reported in geochemical logs (parameter vs. depth), which are useful for mapping the stratigraphic extent of source beds as well as identifying areas where petroleum can accumulate (Peters, 1986; Bordenave et al., 1993; Peters and Cassa, 1994; Dembicki, 2009; Hart and Steen, 2015). This information is crucial for

evaluating the petroleum generative potential of a prospective area, and thus programmed pyrolysis is considered as an important guiding tool in petroleum exploration.

Programmed pyrolysis is routinely carried on rock samples collected either from outcrops or core, sidewall core or cuttings samples obtained during drilling to analyze source rock potential. Samples obtained from outcropping may be weathered or incomplete for the entire source section and thus problematic or even unsuitable for geochemical analysis (Littke et al., 1991; Bordenave et al., 1993; Ruebsam et al., 2018). Coring is expensive and rarely performed on source intervals, thus making cuttings samples or occasionally side wall core (SWC) more common to evaluate source rock intervals (Bordenave et al., 1993). SWCs collected from stratigraphically well defined geological successions may potentially suffer from small sample size and thus may not be representative of entire source intervals. Cuttings represent a composite samples of rock fragments taken over 3-5m intervals of sedimentary successions, and may be easily impregnated by drill fluid or affected by caving and hence will not provide accurate lithological information at high resolution (Busch et al. 1987; Bordenave et al., 1993; Serra, 2003). As result, limited geochemical analysis on cuttings or sidewall core samples can mislead in evaluation of source quality and quantity. Therefore, sufficient sampling is often required to assess the source potential in a particular area whereby cutting samples taken over a (5-10m) depth interval may still be suitable in non-source regions or in monotonous sedimentary sequences. However, in intervals where source rocks, particularly those of limited extent, are expected, geochemical sampling can be supplemented by geophysical wireline logs, to allow quasi-continuous analysis of source intervals at a higher resolution (~15cm) and determination of their thickness (Brosse et al., 1988; Passey et al., 1990; Bordenave et al., 1993; Serra, 2003; Godfray and Seetharamaiah, 2019; Gama and Schwark, 2023b).

Lower Jurassic successions that form syn-rift sedimentary packages in the Mandawa Basin consist of three distinct formations (Mbuo, Nondwa, and Mihambia Formations) that are entirely made up by siliclastic, evaporite, and carbonate facies. The stratigraphic and lateral distribution of organic matter types and quality of source rock intervals have been poorly investigated due to the limitation in wells drilled, degree of weathering and halokinesis, and incompleteness of outcropping sections (Alconsult and TPDC, 1997; TPDC, 2003; Corelab and TPDC, 2009; Hudson,

2010; McCabe, 2021). Most of the previous geochemical studies were based on a few cuttings/sidewall samples from the sediment of the Nondwa and Mbuo Formations collected from scattered wells (Kagya, 1996; Godfray and Seetharamaiah, 2019; McCabe, 2021). No attempt has been made to constrain the source rock potential in the Mihambia Formation sediment. As a result, both stratigraphic and lateral variability in source qualities with respect to facies changes in the heterolith Pindi Group to date have been poorly documented.

This study aimed at constraining organic matter types and quality of source intervals in the Lower Jurassic sediment successions encountered in the Mita Gamma-1, East Lika-1, and Mbuo-1 wells, as well as in the Nondwa evaporitic sequence outcropping in the Mandawa Basin. This was achieved through TOC, programmed pyrolysis, and carbon isotope analysis of cuttings, cores, and outcropping samples.

TOC and programmed pyrolysis data of source rock intervals compiled from the literature (Alconsult and TPDC, 1997; Corelab and TPDC, 2009; McCabe, 2021) were synthesized in this study to better understand how organic matter quality and thermal maturity evolution may be inferred from sediment of other wells (Mandawa-7, Mbate-1, Mihambia-1, Kizimbani-1) not sampled in this study. To further gain insight into factors that regulated organic matter quality distribution within the basin during Early Jurassic times in the Mandawa Basin, we examine TOC and programmed pyrolysis data in relation to paleogeography.

4.1. Geological and stratigraphy setting

The onshore Mandawa Basin (**Fig. 4.1**), is one amongst many East African Coastal Basins that evolved in response to the break-up of Gondwana and the drifting of Madagascar during the early Mesozoic era. It forms a restricted arm of a failed rift system that is bound to the South by the East-West trending basement high of the Ruvuma saddle, to the North by the Rufiji Trough, to the West by the border fault of the Neoproterozoic Masasi basement spur and covers an area of approximately 15,000 km² (Gama and Schwark, 2023a and references therein). Early Mesozoic, sedimentation was mainly influenced by tectonics, fluctuations in sea level, and climate (Kent,

1972; Kapilima, 2003; Abay et al., 2019; Caracciolo et al., 2020; Gama and Schwark, 2023b). During the Permo-Triassic era, clastic sedimentation with periodic marine incursions was predominant in the NNE-SSW intracratonic basins (Wopfner, 1994; Mpanda, 1997; Kapilima, 2003; Magohe, 2019; Kiswaka et al., 2023). This was followed by further subsidence and marine ingression during the Toarcian (~180 Ma), which resulted in restricted marine conditions and extensive deposition of evaporites (Geiger, 2004; Geiger and Schweigert, 2006; McCabe, 2021; Mshiu et al., 2022; Gama and Schwark, 2023a). A shift from rifting to drifting occurred around 170 Ma, resulting in the opening of the western Somali and Mozambique Basins. As a result, Madagascar started to drift southward along the N-S trending transform fault of the Davie Fracture Zone (Eagles and König, 2008; Gaina et al., 2013). Seafloor spreading between Madagascar and Africa is believed to have ceased around 120 Ma due to the onset of strike-slip motion between Madagascar and India (Eagles and König, 2008; Reeves, 2018; Fossum, 2019; Fossum et al., 2019).

As a result of the Gondwana breakup, sediment accumulation was mainly marine, coastal, and partly terrestrial, and it persisted into the Neogene period. This resulted in a stratigraphic sequence of five groups listed chronologically as Pindiuro, Mandawa, Mavuji, Kilwa, and Songosongo (Hudson, 2010; Fossum et al., 2019). This study focuses on the Lower Jurassic successions (Pindiuro Group) that constitute the Mbuo, Nondwa, and Mihambia Formations (Hudson, 2010; Gama and Schwark, 2023a).

The Mbuo Formation rests unconformably on the Precambrian basement and is overlain by the Nondwa Formation. The formation has only been encountered in the subsurface with thickness ranging from 140m to 1770m. The entire succession is mostly composed of thinly interbedded shales, siltstones and coarser to conglomeritic sandstones, with occasional thin limestone beds. The formation is subdivided into two members, the lower member consists of sandstones with silt-claystones and occasional salt-clays while the upper member is mainly made up by thick claystones with alternating thin dolomite beds and occasional thin sandstone beds (Stratadata, 2006; Corelab and TPDC, 2009; Hudson and Nicholas, 2014; Gama and Schwark, 2023a). Strata were deposited in a wide range of depositional environments from alluvial, to fluvial and lacustrine. The Nondwa Formation overlies the Mbuo Formation and is overlain by the Mihambia Formation. It is one of the oldest outcropping sedimentary units in the Mandawa Basin and that

comprises thick blocky units of halite interbedded with thinly interbedded anhydrites, shales and carbonates (Stratadata, 2006; Quinton and Copestake, 2007; Corelab and TPDC, 2009; Mshiu et al., 2022). It was deposited in a shallow restricted marine embayment. The true thickness of this formation is difficult to establish in surface outcrops due to salt diapirism and varies greatly in the subsurface from 58 m to 2802 m (Stratadata, 2006; Quinton and Copestake, 2007; Corelab and TPDC, 2009; Hudson and Nicholas, 2014; Mshiu et al., 2022).

The Mihambia Formation lies on the Nondwa Formation and is overlain by the Mtumbei Formation (Stratadata, 2006; Quinton and Copestake, 2007; Corelab and TPDC, 2009; Hudson and Nicholas, 2014). The entire succession consists mainly of clastic with minor carbonates deposited in a shallow marine setting (Alconsult and TPDC, 1997; Quinton and Copestake, 2007; Corelab and TPDC, 2009; Gama and Schwark, 2023a). Mihambia Formation occurs in the subsurface with variable thicknesses ranging from 247 m to 598 m.

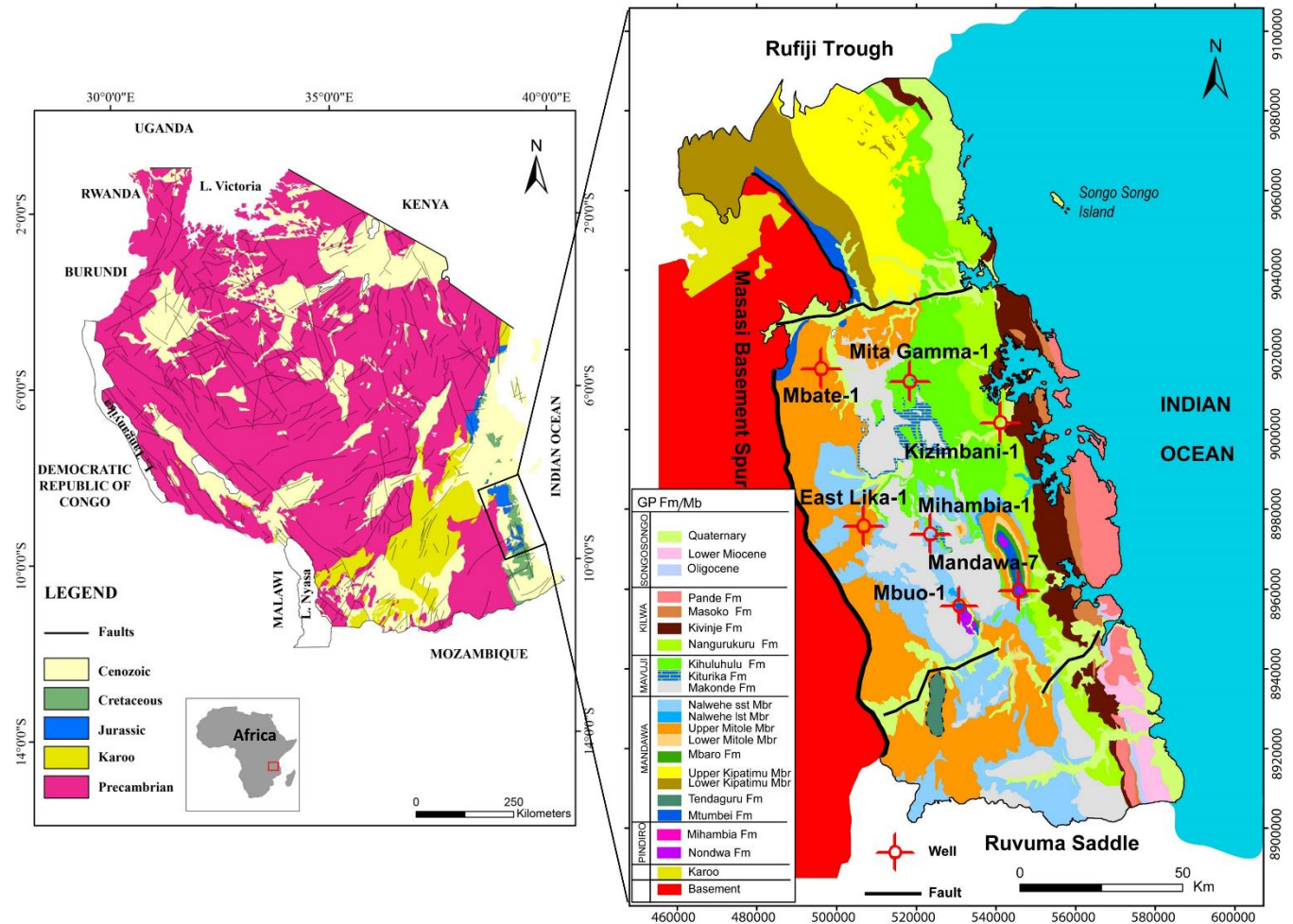


Fig. 4.1: Geological map of Tanzania showing the location of the Mandawa Basin and position of the East Lika-1, Mbuo-1 and Mita-Gamma-1 wells investigated as well as Mandawa-7, Mihambia-1, Kizimbani-1 and Mbate wells-1 (adapted from (Pinna et al., 2008; Hudson, 2010; Barth et al., 2016).

4.2 Materials and methods

4.2.1 TOC and Programmed pyrolysis analysis (Hawk®)

Organic matter was characterized by the use of programmed pyrolysis (Hawk®) following the standard protocol described in (Maende, 2017), which is compatible to Rock Eval methodology Lafargue et al. (1998) as demonstrated by Birdwell et al. (2017). Programmed pyrolysis delivers multiple parameters for organic matter characterization (e.g. Hart and Steen, 2015) including: S1

(quantity of free hydrocarbons expressed as mgHC/g of rock), S₂ (hydrocarbon generative potential expressed as mgHC/g of rock), hydrogen index (HI) (quantity of hydrocarbons generated from the kerogen relative to TOC, and expressed as mgHC/g TOC). Oxygen index (OI) (quantity of carbon dioxide generated from the kerogen relative to TOC, and expressed as mgCO₂/g TOC). Tmax (temperature of maximum pyrolyzate yield expressed in °C), TOC amount of organic matter expressed in wt% and carbonate carbon expressed in wt%.

4.2.2 Carbon isotopes analysis of organic matter

A total of 215 samples were subjected to stable carbon isotope analysis for bulk organic carbon, carried out on decalcified sample material. Prior to analysis, a well homogenous powdered samples were treated with hydrochloric acid (HCL, 10% and 25%) to remove inorganic (carbonate -and dolomite-bound) carbon, there after sample were washed and neutralized with deionized water and dried in an oven at 40c for 48h.

Stable carbon isotope analysis was performed using a Thermo Finnigan Delta V isotope ratio mass spectrometer coupled to a Flash EA via a Conflow III interface. Carbon isotope ratios are expressed in conventional delta notation: $\delta^{13}\text{C} (\text{‰}) = [(R_{\text{sample}} - R_{\text{standard}})/R_{\text{standard}} - 1] \times 1000$, where R is the ratio of ¹³C/¹²C of the sample and of the V-PDB standard for carbon. Reproducibility and accuracy were monitored by replicate standard and sample analysis and were better than 0.1‰.

4.3 Results

4.3.1 Downhole variation in organic matter quality and quantity of three wells based on cutting samples

Variation in OM content and quality of the sediments from three boreholes investigated are shown in the **Figs. 4.2, 4.3, and 4.4**. Most of the data show that organic matter content and quality

determined on the cutting samples varies substantially along stratigraphy (**Figs. 4.2, 4.3 and 4.4**). In general, organic matter contents in sediment samples from three wells range from 0.31 to 3.08 wt% with an average of 0.97wt%, whereas kerogen yield (S₂) and HI are generally low, varying from 0.01 to 6.07 (mg HC/g) with an average 0.64 for S₂ and 4 to 220 (mg HC/g TOC) with average 45 (mg HC/g TOC) for HI, respectively. There is a substantial heterogeneity in organic matter contents and quality of cuttings samples from the three wells.

TOC in the sediments of Mita Gamma-1 well fluctuate along stratigraphy, they range from 0.38 to 3.08 wt% with average of 1.1wt%, whereas S₂ and HI values are generally low; they vary from 0.13 to 6.07 (mg HC/g) with average 1.17 (mg HC/g) and 22 to 196 (mg HC/g TOC) with an average of 72 (mg HC/g TOC), respectively. Sediment samples from the Upper Nondwa and Mihambia Formations in the Mita Gamma-1 well have fair to good petroleum potential whereas, sediment samples from the Mbuo Formation generally exhibit poor source potential. TOC, S₂, HI analyzed in the sediments of the East Lika-1 well are generally low, TOC ranges from 0.31 to 1.67 wt% with average 0.75 wt%, whereas S₂ varies from 0.01 to 1.23(mg HC/g) with average 0.28 and HI ranges from 4 to 73 (mg HC/g TOC) with average of 27, respectively. Sediment samples of the Upper Nondwa in the East Lika-1 well exhibit fair source rock potential as opposed to poor source potential in the Mihambia and Lower Nondwa and Mbuo Formations. While TOC in the sediment of the Mbuo-1 well varies from 0.58 to 2.21 wt% with an average of 1.16 wt%, the S₂ fluctuates from 0.11 to 4.49 (mg HC/g) with an average of 0.65 (mg HC/g) and HI values range from 12 to 220 (mg HC/g TOC) with an average of 52 (mg HC/g TOC). Shale samples in the Lower Mbuo Formation show fair source rock potential as opposed to poor potential shown by Nondwa and Mihambia Formations.

Most of the shale samples that are enriched in organic content (TOC $\geq$ 1 wt %) and with sufficient S₂ yield and HI occur in a few localized interval. These have been encountered in the sediments of Upper Mbuo, Nondwa, and Mihambia Formations in the Mita Gamma-1 and East Lika-1 wells, as well as in the Lower Mbuo Formation in the Mbuo- 1 well. Other intervals containing shale, carbonates, evaporites, siltstone and sandstone samples are not enriched (TOC < 0.5wt %) and hence they do not show any enrichment in HI values due to lower S₂. Production indices (PI) in source facies (anything not sandstone) fluctuate along stratigraphy with values ranging from

(0.03-0.3), unless affected by drilling additives. OM-rich source facies give PI values of only <0.05 (in the East Lika-1 well), < 0.1 for Mbuo-1 well (note deeper burial and maturity), Mita Gamma-1 <0.15 (note moderately higher maturity). Consequently, thermal maturity is too low to allow for major petroleum generation and expulsion. Shale samples from sediment that are enriched in TOC, S2 and HI from all three wells generally show lower PI of < 0.25. It appears that both a higher PI (> 0.25) and a general trend increasing with depth occurs in sandstone, dolomite and few shale intervals. Whereby PI increases for well-suited samples from 0.04 to 0.20 in the Mbuo-1 well over a depth of 400 to 3000 m, in the Mita Gamma-1 well there no increase in PI for source facies, as demonstrated by PI values around 0.10 over a depth range of only 200 m (2000 -2200 m). In the East Lika-1 organic facies give PI values of 0.04 at 1200 m to approximately 0.25 at 1800 m depth. High porosity and permeability samples (siltstone, dolomite and sandstone) do not exhibit high PI due to low S2 (kerogen-lean). In contrast, high PI values observed over the depths of 2250, 2285, 2178m in the sediments from the Mita Gamma-1 well, and at the depths of 1080-1145, 1890m in the sediments from the East Lika-1 well, and at 350 and 2910m depth from the Mbuo-1 well sediments are attributed to drilling additives. Tmax values for source facies samples with sufficient S2 from all three wells range from 428 to 477 °C. In the East Lika well, they range from 429 - 477°C with an average 436°C, whereas in the Mita Gamma-1 well they range from 434 to 453°C with an average 442 and in the Mbuo-1 well they range from 428 to 435°C with an average 442°C. Generally, there is a faint downhole increase in the thermal maturity trend observed in all three wells. Sediment samples from the Mihambia Formation over a depth range 2008–2030 m in the Mita Gamma-1, 1220–1240 m in the East Lika-1, and in 300–550 m in the Mbuo-1 wells show an inverse maturity trend (higher Tmax at a shallower depth). Samples from the Upper Mihambia and Nondwa Formation show moderate generative potential in all three wells, whereas the Lower Mbuo Formation exhibits source rock potential in the Mbuo-1 well only.

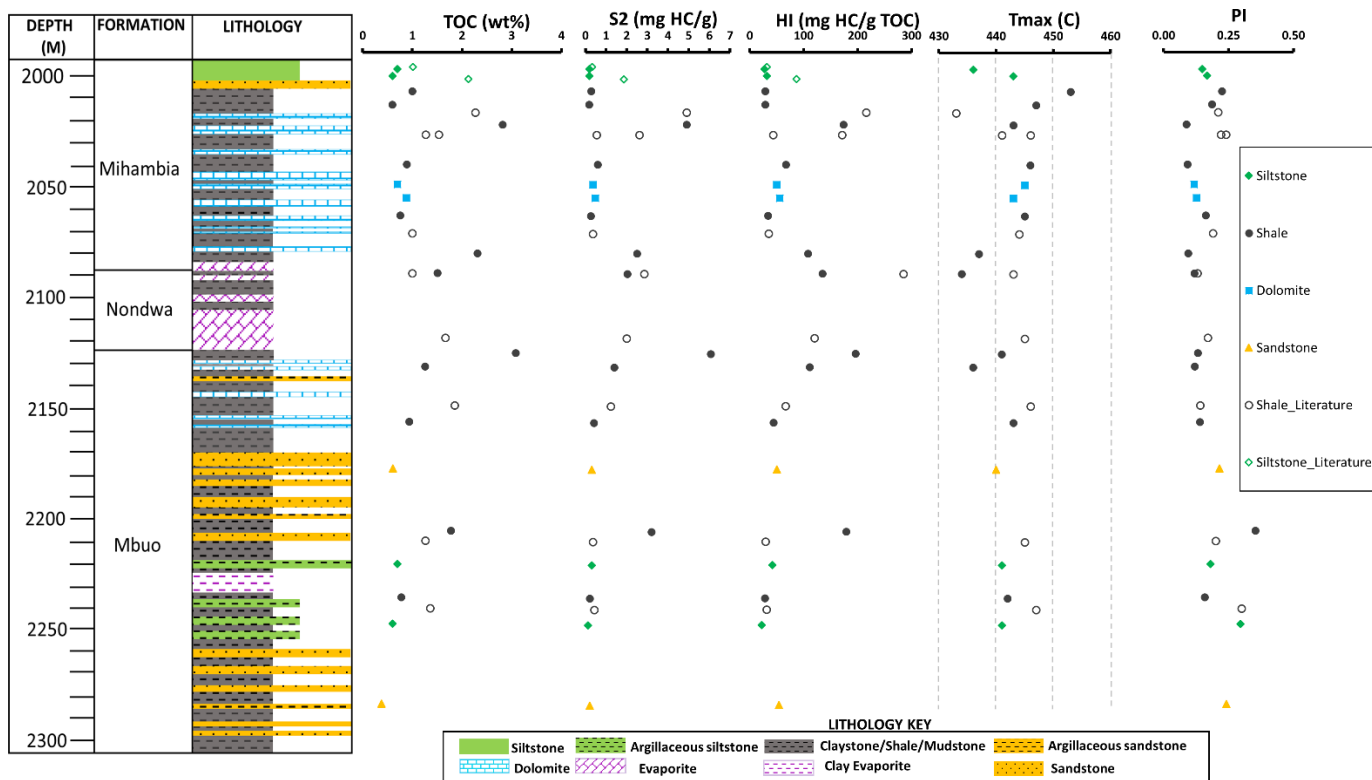


Fig. 4.2: Downhole variation in Total organic carbon (TOC), Kerogen yield (S2), Hydrogen Index (HI), Tmax and Production Index (PI) of cuttings samples in the sediment from the Mita Gamma-1 well.

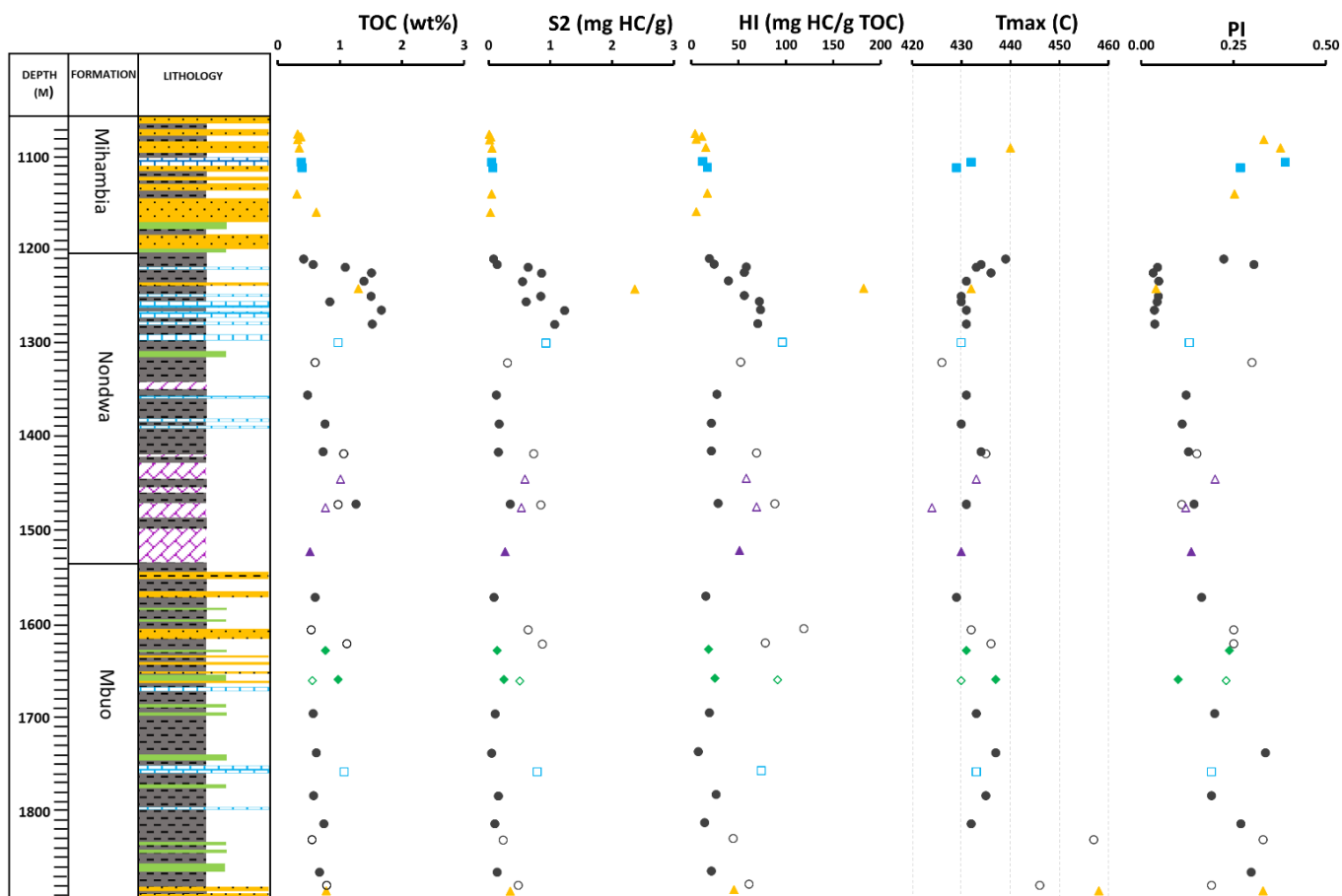


Fig. 4.3: Downhole variation in Total organic carbon (TOC), Kerogen yield (S2), Hydrogen index (HI), Tmax and Production Index (PI) of cuttings samples in sediment from the East Lika-1 well.

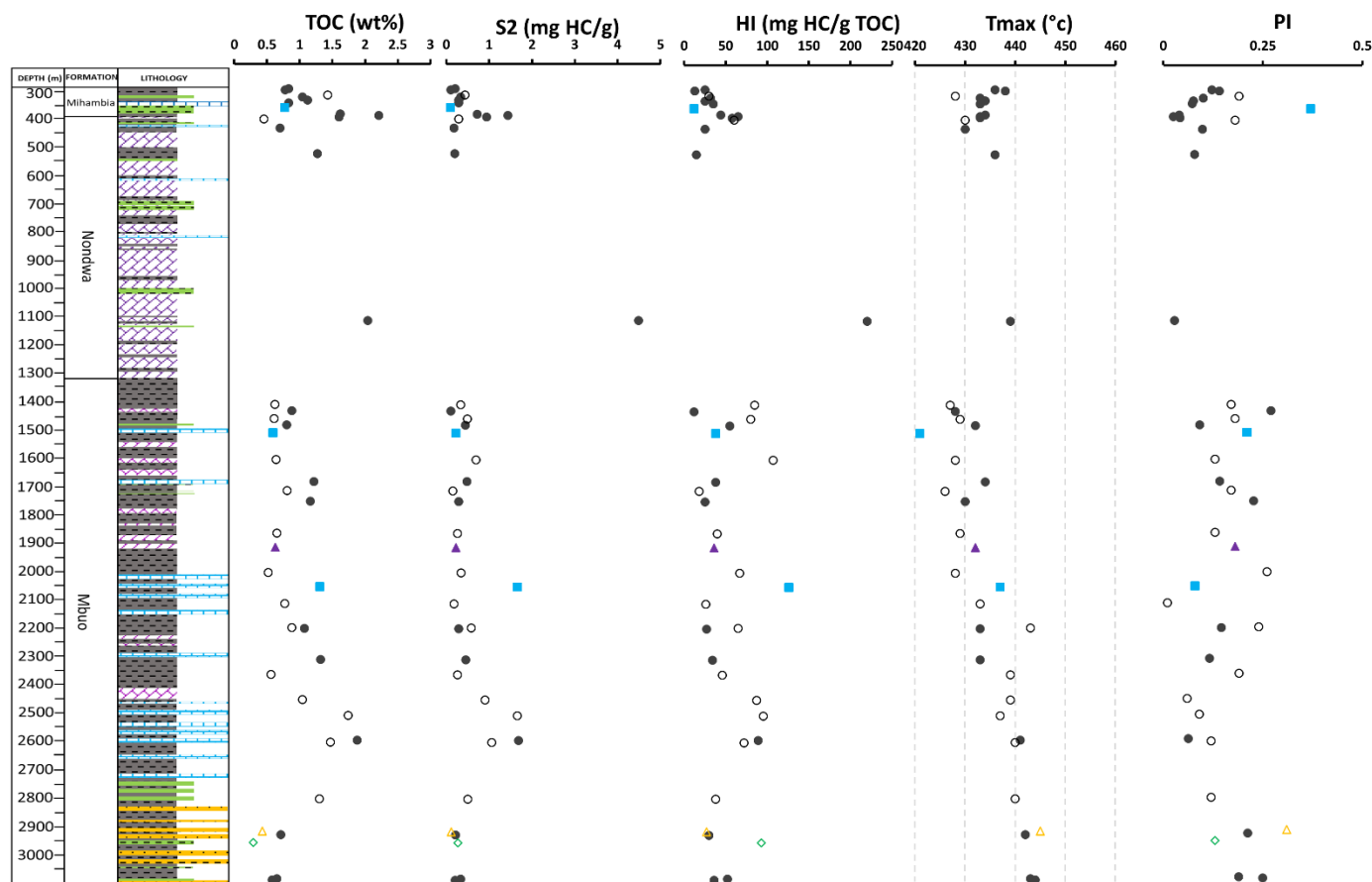


Fig. 4.4: Downhole variation in Total organic carbon (TOC), kerogen yield (S2), Hydrogen index (HI), Tmax and Production Index (PI) of cuttings samples in sediment from the Mbuo-1 well.

4.3.2 Downhole variation in organic matter quality and quantity of core samples

Variation in OM contents and quality of core samples investigated in the sediment of Mbuo-1 well covering 12m interval of sedimentary successions are shown in **Fig. 4.5**. Generally, both organic matter content and quality vary along the cored section (**Fig. 4.5**). TOC values range from 1.42 to 4.83 wt% with an average value of 3.1 wt%. S2 and HI values are generally low, only three samples over a depth range of 2808.8-2809.9 m of the cored section show enrichment with an average S2 and HI of 8.24 (mg HC/g) and 193 (mg HC/g TOC), respectively. Tmax in core samples from the Mbuo-1 well is generally higher compared to cuttings samples in the same well/intervals, values generally range from 445 to 450°C, suggesting that Tmax determined from the cuttings samples

may be suppressed (**Supplement table 4.1**). PI values of core samples vary dramatically along the sedimentary successions of the cored section with average values for source facies of about 0.06, whereas siltstone and sandstones may show higher PI (> 0.25) due to impregnation of drill additives. Shale and siltstone over a depth range 2807-2813 exhibit PI values ranging from 0.06-0.25 due to low S₂ content (organic lean facies). In contrast, three organically enriched shale samples over a depth range of 2808-2809 m possess a very low PI (< 0.05), indicative of true petroleum generation potential as shales are less affected by mud invasion. A comparison of TOC, S₂, and HI values in the Mbuo well-1 core samples, averaged over a 5-meter interval of the sedimentary successions, to those of cutting samples in the same well revealed compatible results, indicating that shale samples are only enriched to source quality within short sediment intervals in the Mbuo-1 well.

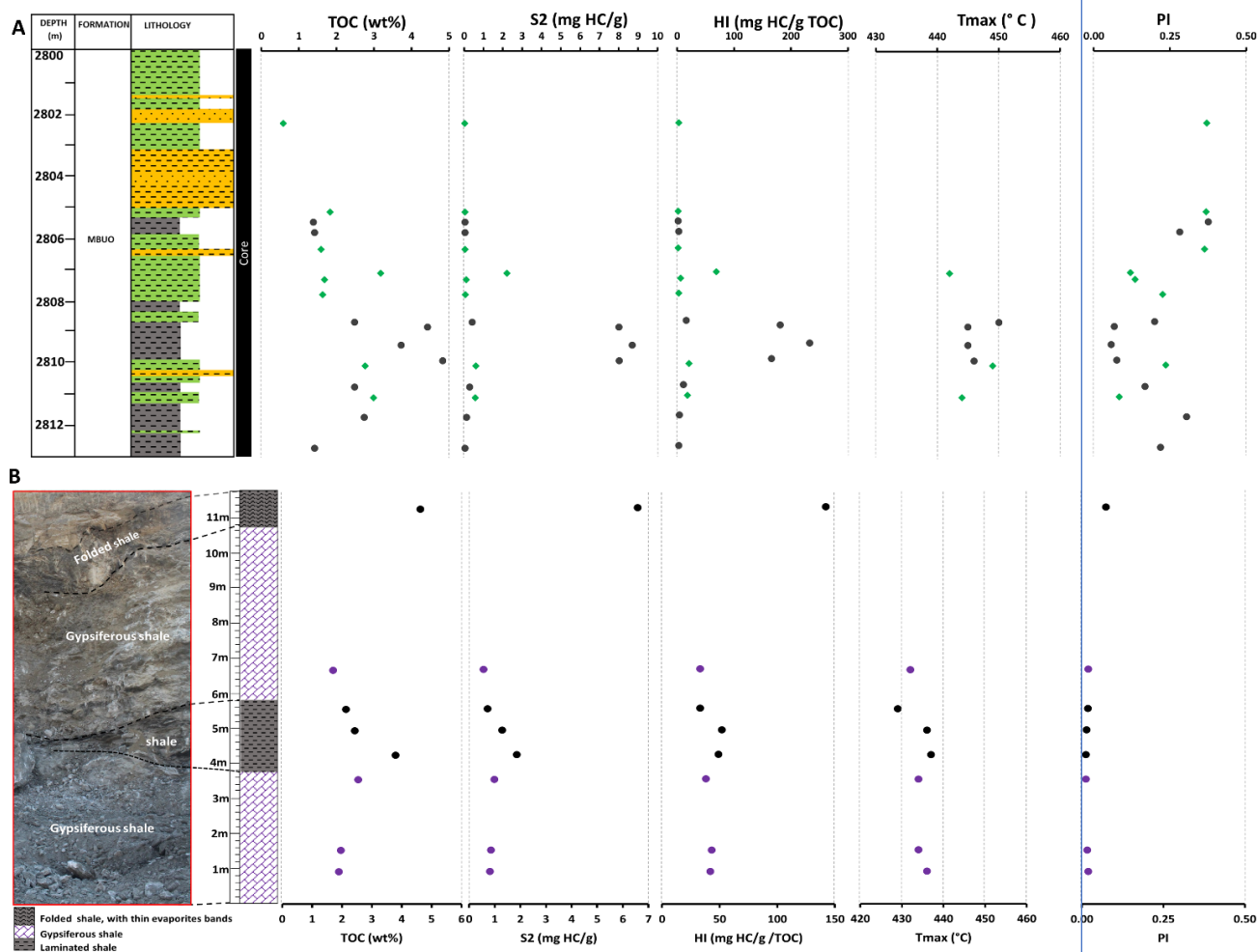


Fig. 4.5: Variation in TOC, kerogen S2 yield, HI, Tmax, and PI in sediments from: A (Mbuo core), B (outcropping Nondwa evaporitic sequences).

4.3.3 Variation in organic matter quality and quantity of a Nondwa outcrop section

Variation in OM contents and quality for Nondwa outcrop sediment samples investigated are shown in **Fig. 4.5**. It appears that both organic matter contents and quality vary with respect to lithofacies, with shale lithology being more organically enriched than gypsiferous shale (**Fig. 4.5**). Within shale the lithology, organic carbon content and quality increase, for example, over a depth interval of 3.8–5.8 m in an outcropping section conforming an internal variability of organic matter in heterolithic successions. Almost all shale samples are enriched in organic carbon (TOC ≥ 1 wt%) and poorly enriched in S2 yield and HI value. TOC in the Nondwa outcropping section

ranges from 1.70 to 4.62 wt% with an average of 2.64wt%, whereas S2 yield and HI values are generally low across the outcropping section. Only two samples located at 4.26-4.96 m and one sample located at 11.3m show enrichment with respect to other samples. S2 and HI values in these intervals are 1.58 (mg HC/g rock) and 51 (mg HC/g TOC) and 6.58 (mg HC/g rock) and 142 mg HC/g TOC, respectively.

Kerogen S2-content and HI of enriched shale samples (organofacies) generally show very low production indices ~ 0 ; this may be due to evaporative loss of volatiles during surface exposure over prolonged time. When TOC, S2, and HI values in Nondwa shale and gypsiferous shale samples taken at depths of 0-5m and 5-10m in outcropping section are averaged at intervals equivalent to cutting samples, the results are comparable to those obtained from the Nondwa cutting samples collected in the Mbuo-1 well. Values for such a comparison are 2.53 wt% for TOC, 1.16(mg HC/g rock) for S2, 45 (mg HC/g TOC) for HI versus and 2.82 wt% for TOC, 2.62(mg HC/g rock) for S2, and 69(mg HC/g TOC) for HI, respectively. Tmax values determined from outcrop samples generally range from 429 to 437°C with an average of 434°C, which indicates a maximum burial depth of around 2000- 2300m, followed by i.e. 2000-2300 m uplift due salt diapirism.

4.4. Discussion

4.4.1 Quality assurance of pyrolysis data

Discussion of organofacies and maturity is based on programmed pyrolysis measurements executed on a few core and outcrop samples but dominantly on cuttings samples, which are prone to alteration, i.e. impregnation with oil-based drilling mud. Hence, data quality assurance is of pivotal importance for interpretation and therefore, the sample suite is differentiated in a suite of best quality samples (termed “organofacies” samples) versus the remainder of the “standard” samples, considered to deliver data of lower suitability for organic geochemical interpretation. Sample quality assurance is based on TOC and S2 content, with samples showing the highest TOC and S2 being considered more reliable (Bordenave et al., 1993) these sample mostly constitute shale lithologies. The remainder of the “standard” samples are important to

identify trends along the succession studied and are critically screened for their utility for interpretation. Whereas TOC and HI values of standard samples may reflect primary sample composition, the S1 and PI values are affected and enhanced by mud contamination (Bordenave et al., 1993) to a variable degree, as will be discussed below. Degree of mud contamination is controlled by lithology, with sandstones and siltstones owing to their higher porosity and permeability being more susceptible for mud infiltration. Tmax values exhibit a tendency to be affected by oil-based mud infiltration, which may suppress Tmax values. A threshold of S2 abundance of 1 mg/g TOC here is taken as cutoff value for quality assurance of Tmax values (**Supplement table 4.1**), whereas other sources consider a threshold of 2 or even 4 mg/gTOC as appropriate (Bordenave et al., 1993; Peters and Cassa, 1994; Lafargue, Marquis and Pillot, 1998; Yang and Horsfield, 2020). Utilization of organofacies vs. standard sample data leads to substantial differences in interpretation between both data set, which is discussed in detail below. Above mentioned of averaging effect of source quality for cuttings intervals of 3 to 5 meters may explain why individual layers may be more enriched and source prolific as referred to in discussion below when addressing core samples.

4.2.2 Organic matter types and quality for four different locations

4.4.2.1 Organic matter quality from cuttings, core and outcrop samples for four different location (3 borehole wells)

The relationship between HI and OI (modified Van Krevelen diagram), S2 and TOC can give indication on the quality of organic matter or kerogen types (**Figs. 4.6 and 4.7**). Most of the cuttings samples from three boreholes analyzed in this study and along with data supplemented from literature have low HI values (< 200 mg HC/g TOC) and variable OI values. On a modified Van Krevelen diagram (**Fig. 4.6**), almost all samples plot within the kerogen types III to IV fields, with minor populations in mixed types II and III. Generally, for cuttings sediment samples from three wells it appears that HI do not decrease with increasing with thermal maturity (**Figs. 4.3, 4.4, 4.5**

and 4.6). Therefore, minor variation in HI and thus position of the kerogen samples observed in **Fig. 4.6** besides drilling mud alteration would represent primary compositional differences of the kerogen rather than differences in maturity (Bordenave et al., 1993). Core and outcrop samples plot towards the origin, indicating a higher and unaltered level of maturity and more refractory composition of organic matter. All sediment samples from the three boreholes show variation in OI whereas, core and outcrop samples plot toward the origin and do not show much variation in OI. The former indicates a differential degree of oxidation in water column, whereas the latter indicates high degree of maturity of the core and outcrop samples. Most of the samples analyzed in this study plot in kerogen type-III to type-IV fields, very few plots in the mixed kerogen type II/III region. Consequently, the best quality organofacies samples plot in the II/III fields, whereas the standard samples due to both poor preservation during deposition via aerobic degradation and alteration upon drilling show a refractory and oxidized composition. In addition, storage of cuttings due to their large surface area may have affected organic matter by secondary oxidation, which is prone to decrease HI and increase OI (Bordenave et al., 1993; Jarvie et al., 2004). Based on the presence of kerogen III and mixed types of kerogens in the samples, it is evident that most shale samples are primarily gas-prone, while other samples are unsuitable to generate hydrocarbons and very few samples are oil-prone. The TOC vs. S₂ binary plot (**Fig. 4.7**) reveals a very minor differences in OM, with almost all samples scattered in kerogen type-III/IV fields and a minority in the mixed kerogen type II-III. Almost 50% of the cuttings samples as well as, core and outcrop samples that has that have TOC concentration > 1wt %, exhibit low hydrocarbon source potential.

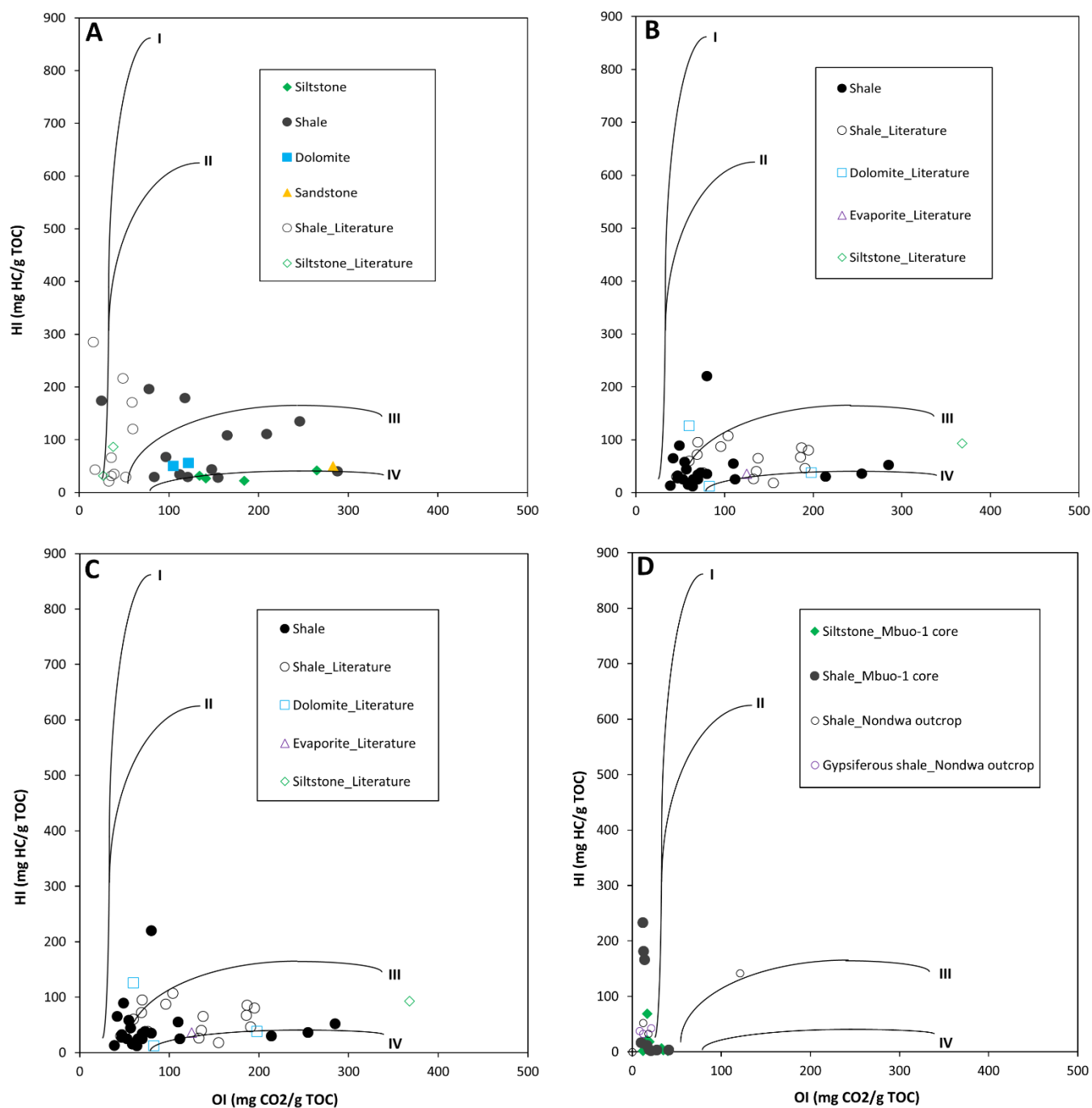


Fig. 4.6: Comparison of organic matter quality in the sediments of Lower Jurassic successions from four different location in the Mandawa Basin: A (Mita Gamma-1), B (East Lika-1), C (Mbuo-1), D (core and outcrops samples).

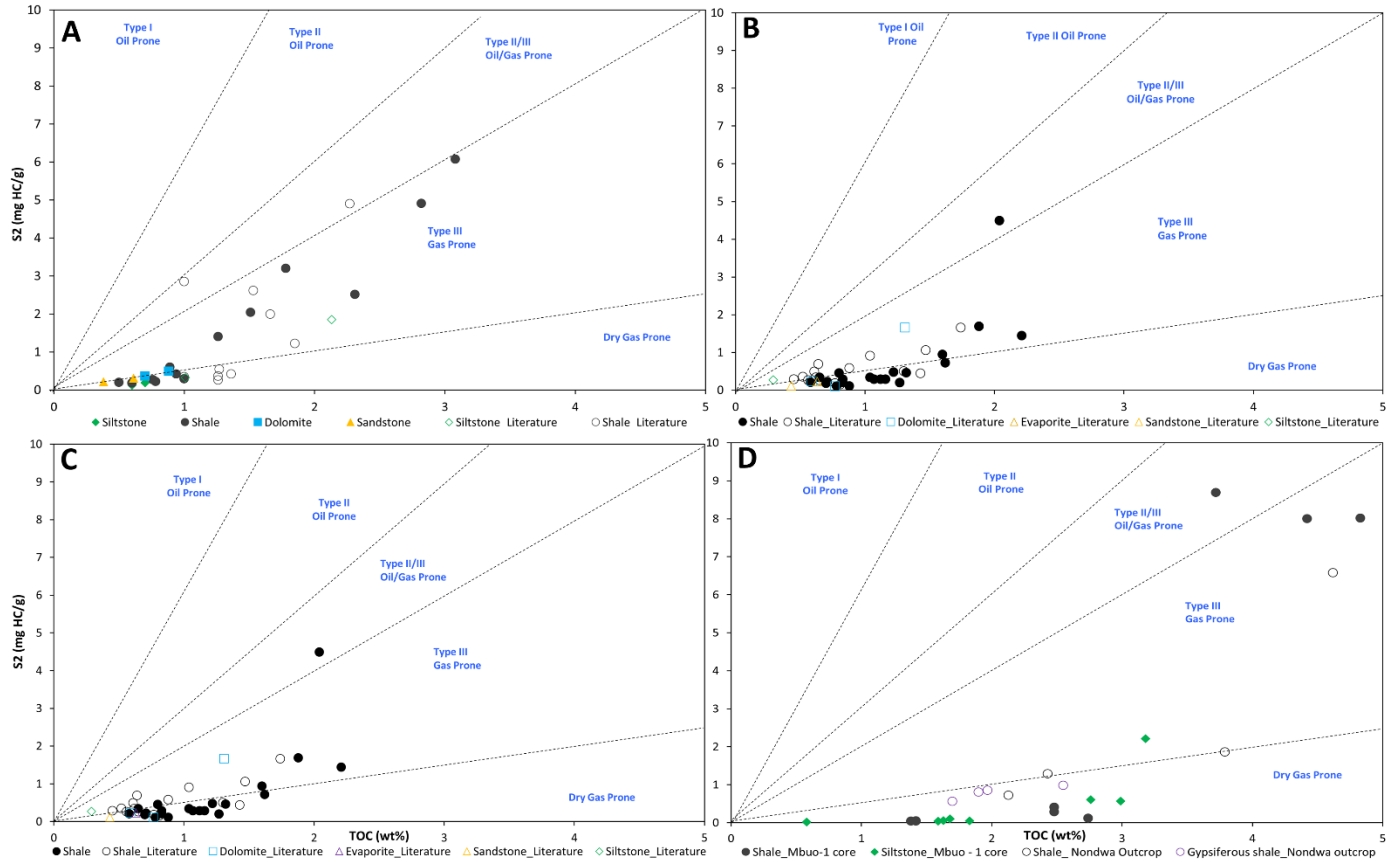


Figure 4.7: Comparison of organic matter quality in sediment Lower Jurassic successions from four different locations in the Mandawa Basin: A (Mita Gamma-1), B (East Lika-1), C (Mbua-1), D (core and outcrops samples).

4.4.3. Comparison of organic matter maturity of four locations in the Mandawa Basin

4.4.3.1 Organic matter types versus Maturity

Thermal maturity is another important factor for source rock evaluation in addition to organic richness and quality (Espitalie et al., 1977, 1984; Peters, 1986; Wopfner, 1994; Lafargue et al., 1998; Dembicki, 2009; Yang and Horsfield, 2020). Comparison of the downhole variation of Tmax of cuttings samples from three boreholes, core, and outcrop samples (**Figs. 4.2, 4.3, 4.4 and 4.5**) indicate a wide range of thermal maturity, ranging from onset of to within to the mature stage of the oil window. We employed a Tmax vs. HI binary plot (**Fig. 4.8**) to evaluate the effect of thermal maturity on organic matter for the sediment of Lower Jurassic successions in the Mandawa Basin.

Comparison of the Tmax vs. HI binary plot for four locations in the basin indicate that most of the organo-facies samples are mainly composed of Type III/IV kerogens and are within the onset to mature oil window zones with respect to the organic matter level of maturity. An average Tmax for samples analyzed from the sediment of Mita Gamma-1 is 440°C (oil window zone, whereas, in the East Lika-1 and Mbuo-1 wells is 435°C (onset of oil window). Differences in Tmax indicate differences in burial history between the three locations.

In general, sandstone and siltstone of the cuttings samples from three wells display a higher level of Tmax possibly due to drilling alteration. Average Tmax of the outcrop samples is 435 °C (onset of oil window), whereas core samples are more mature (450 °C) compared to other samples. Comparison of the Tmax values determined from core samples in the sediment of Mbuo Formation (2002-2014m) to the of cuttings intervals (2800m) in the same well (Mbuo well-1) suggests that the Tmax of the cuttings sample are suppressed due to oil impregnation. A comparison of the Tmax of Nondwa sediments from outcrop samples with those from drill holes indicates that outcrop samples have a higher Tmax because of deeper burial. Consequently, we suggest that the outcrop succession was buried approximately 2000-2300 m and then uplifted by salt diapirism.

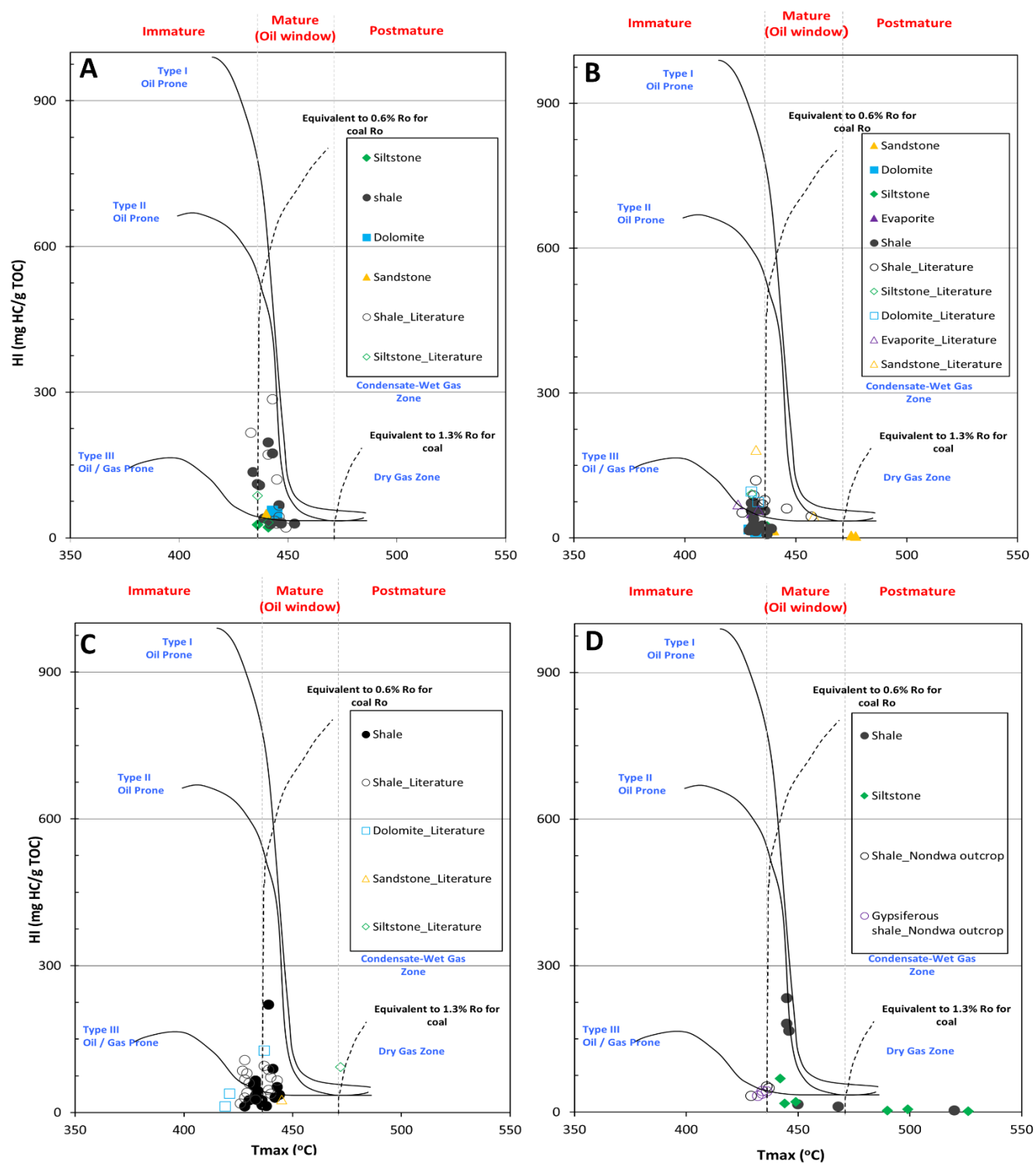


Fig. 4.8: Comparison of organic matter quality vs. maturity in the sediments of Lower Jurassic successions from four different locations in the Mandawa Basin: A (Mita Gamma-1), B (East Lika-1), C (Mbuo-1), D (Core and outcrops samples).

4.4.3.2 Quality of organic matter with respect to maturity

Stable carbon isotope ratios are commonly employed to trace the origin and fate of OM, to distinguish between marine and terrestrial plant sources and to identify OM that derived from land plants using the C3 or the C4 photosynthetic pathway (Buchardt et al., 1986; Hollander et al., 1992; Tyson, 1995; Meyers, 2003; Bauersachs et al., 2014). The HI corresponds to the quantity of pyrolyzable organic compounds relative to the total organic carbon in the samples, which is sensitive to maturity as opposed to the organic carbon isotope signature ($\delta^{13}\text{C}_{\text{org}}$), which is an indicator of the facies and types of organic matter that within the temperature range of the oil window is not affected by maturity (Bordenave et al., 1993). These two parameters were employed in the $\delta^{13}\text{C}_{\text{org}}$ vs. HI binary diagram (**Fig. 4.9**) in order to assess, if there is variation in the type of organic matter in sediments of Lower Jurassic successions in the Mandawa Basin. Most of the samples displayed in the $\delta^{13}\text{C}_{\text{org}}$ vs. HI binary diagram (**Fig. 4.9**) shows two main trends: Lower HI and heavier isotopic composition vs. higher HI and Lighter isotopic composition of organic matter (**Fig. 4.9**). The $\delta^{13}\text{C}_{\text{org}}$ vs. HI binary diagram (**Fig. 4.9a,c,d**) shows a broad isotopic composition range of -19.6 to -27.2 ‰ in samples from Mita Gamma-1, East Lika-1, Mbuo-1, and outcrop sediments, in contrast to the narrow range of -19.8 to -23.5‰ observed in East Lika-1 sediments (**Fig. 4.9b**). $\delta^{13}\text{C}_{\text{org}}$ variability may indicate changes in the proportion of marine (-24 to -28‰) vs. terrestrial organic matter (-19 to -22‰). For the Jurassic sediments, heavier isotopes values indicate terrestrial organic matter, whereas lighter isotopic composition corresponds to marine organic matter (Tyson, 1995). Based on the observed $\delta^{13}\text{C}_{\text{org}}$ vs. HI data trend (**Fig. 4.9**), most of studied sample are predominantly derived from terrigenous organic matter, with only smaller proportions of marine organic matter. Terrestrial-marine transitional samples show

mixed OM isotope and quality signatures. We therefore suggest that low HI values with corresponding heavier carbon (-22 to -24‰) isotopic composition point to depositional systems with high sediment influxes and probably transitioning from fluvial-lacustrine towards lacustrine to restricted marine conditions in an evolving rift basin. In contrast, changes in preservation conditions due to marine ingression resulted in an increase in HI and a shift to lighter isotopic composition. Low HI and intermediate isotopic composition of sediments indicate that organic matter is possibly sourced from a mixed terrestrial-marine environment. The occurrence of light isotopic compositions (-24 to -28‰) observed in the sediment samples of the Nondwa, Mihambia, and Mbuo Formations in the four locations, suggest that the Mandawa Basin experienced a short-lived marine incursion over a long time due to rifting, accompanied by an increase in sea level.

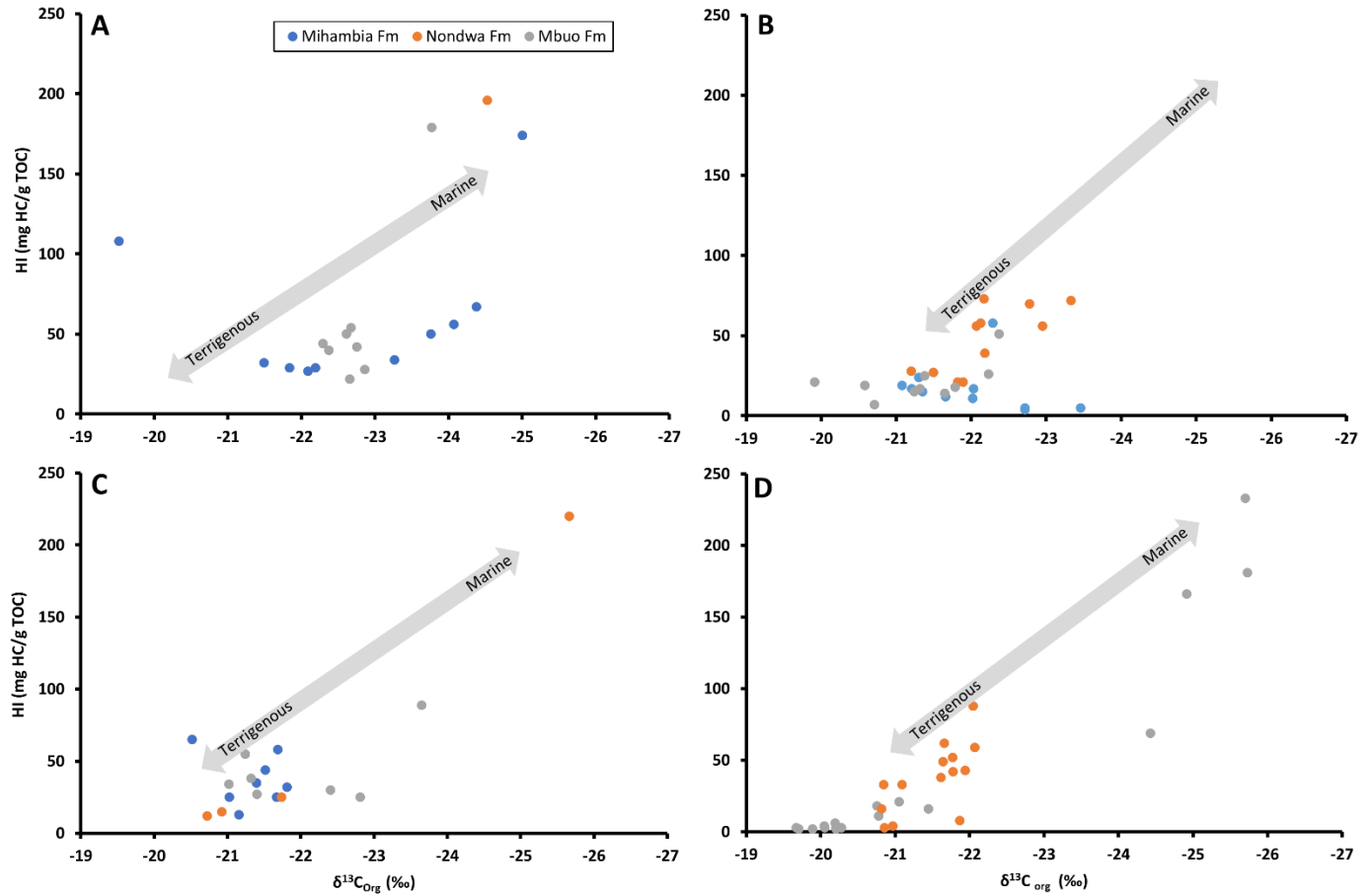


Fig. 4.9: Comparison of organic matter quality of the sediments of Lower Jurassic successions with respect to maturity ($\delta^{13}\text{C}_{\text{org}}$ vs. HI) from four different locations in the Mandawa Basin: A (Mita Gamma-1), B (East Lika-1), C (Mbuo-1), D (core and outcrops samples).

4.4.4 Comparison of organic matter quantity and quality for all wells drilled in the Mandawa Basin

To further understand how petroleum potential, thermal maturity evolution, and paleogeography in the Mandawa Basin evolved, we integrated data published from other wells drilled in the basin. The TOC and Rock Eval datasets of four wells (Mandawa-7, Mbate-1, Mihambia-1, and Kizimbani-1 wells) collected from either published or unpublished literature (Alconsult and TPDC, 1997; Corelab and TPDC, 2009; McCabe, 2021) were synthesized in this study.

We use a modified van Krevelen diagram (**Fig. 4.10a**) and TOC vs. total hydrocarbon yield binary plots (**Fig. 4.10b**) for source, facies and maturity assessment. Three trends of organic matter quality/types could be deduced (**Fig. 4.10a**): Recalcitrant (reworked organic matter with variable OI), terrigenous kerogen type-III and aquatic material of kerogen type-II/III in the sediments of Mandawa-7, Mbate-1 and occasionally in sediments from the Mbuo-1, East Lika-1 and Mita Gamma-1 wells. Almost 80% of all sediment samples either generated from this study or compiled from literature show poor generation potential. Samples from sediments of Mandawa-7 and Mbate-1 show good-excellent source potential and the few samples from the Mita Gamma-1, East Lika-1, and the Mbuo-1 core that show fair to good source potential imply better preservation and a preferentially aquatic marine organic matter deposited upon marine incursions. Core samples from sediment of the Mbuo-1 well have high TOC compared to other samples but they have poor generation potential due to the recalcitrant nature of the organic matter. Comparison of the Tmax vs. HI binary plot (**Fig. 4.11a**) from 7 locations indicate that most of the samples are mainly composed of Type II, mixed II/III to III/IV kerogens and with respect to the organic matter level of maturity are within the onset to mature oil window zones. This indicates that the Lower Jurassic sediments in the Mandawa Basin have been deeply buried and sediments have undergone thermal alteration, whereby slight differences in Tmax across wells would indicate different burial histories. Deeper burial is evident from the unaltered core samples of the Mbuo Formation in the Mbuo-1 well. There are obvious differences between HI values from sediments of the Mandawa-7 vs. Mbate wells, and other wells at low maturity that may be due to the sedimentary environment (different types of organic matter).

To further constrain maturity evolution of Lower Jurassic sediment in the Mandawa Basin, Tmax vs. present day depth plots were used (**Fig. 4.11b**). In general, the Tmax data from sediment samples from all wells plotted according to present-day depth show an increasing maturity trend with depth. However a few sediment samples from the East Lika-1, Mita Gamma-1, and Mandawa-7 wells exhibit inverse maturation trends at a given depth. The two former exhibit Tmax values seemingly too high, whereas the latter exhibits Tmax values too low (**Fig. 4.11b**). It is unlikely that an uplift of 1000m is responsible for the elevated Tmax in the East Lika-1 and Mita Gamma-1 wells, since other samples in the wells do not show this unusual trend. Consequently,

we propose that the anomalously elevated Tmax derive from different types of organic matter (recalcitrant /reworked organic matter or of previously coalified materials) that can be linked to the relative position of deposition sites to the hinterland. During Pliensbachian - Toarcian times high sediment influxes occurred in the East (MitaGamma-1) and West (East Lika-1) of the Mandawa Basin, which allowed deposition of reworked sediment at the basin margin as observed in these two wells. Moreover, Tmax may be suppressed due to the presence of thermally labile kerogen that may show elevated HI and high TOC (Snowdon, 1995), features that are similar to those found in the Nondwa sediment in Mandawa-7 well. Thus, the suppression of Tmax at a depth of 2800-3300m in the Mandawa-7 well sediment is proposed to be caused by oil-prone organic matter (algal type). There is no evidence for this Tmax suppression to be caused by evaporitic thermal retardation since it does not occur in other wells that contain this lithology.

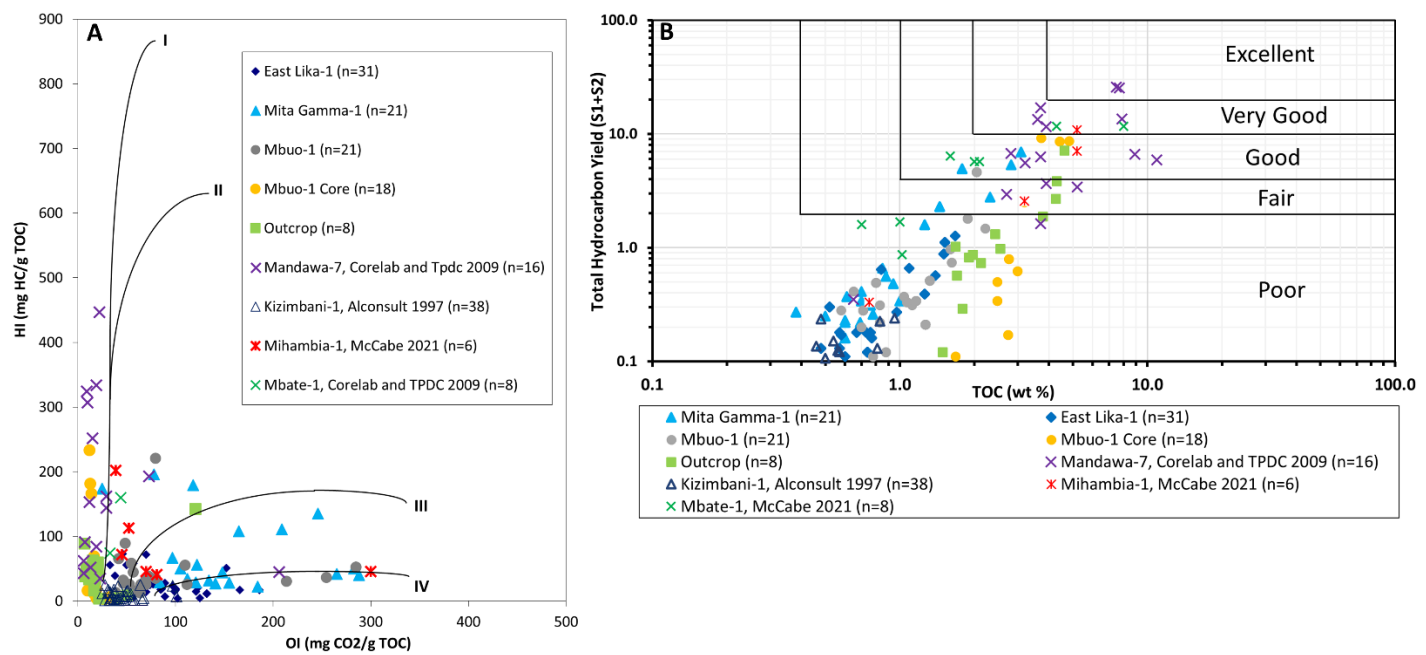


Fig. 4.10: Comparison of organic matter quality of Lower Jurassic intervals from all wells in the Mandawa Basin using: A (Modified van Krevelen diagram) and B (TOC versus total hydrocarbon yield). Data from Kizimbani-1, Mihambia-1, Mandawa-7, Mbate-1 wells have been compiled from literature (Alconsult and TPDC, 1997; Corelab and TPDC, 2009; McCabe, 2021).

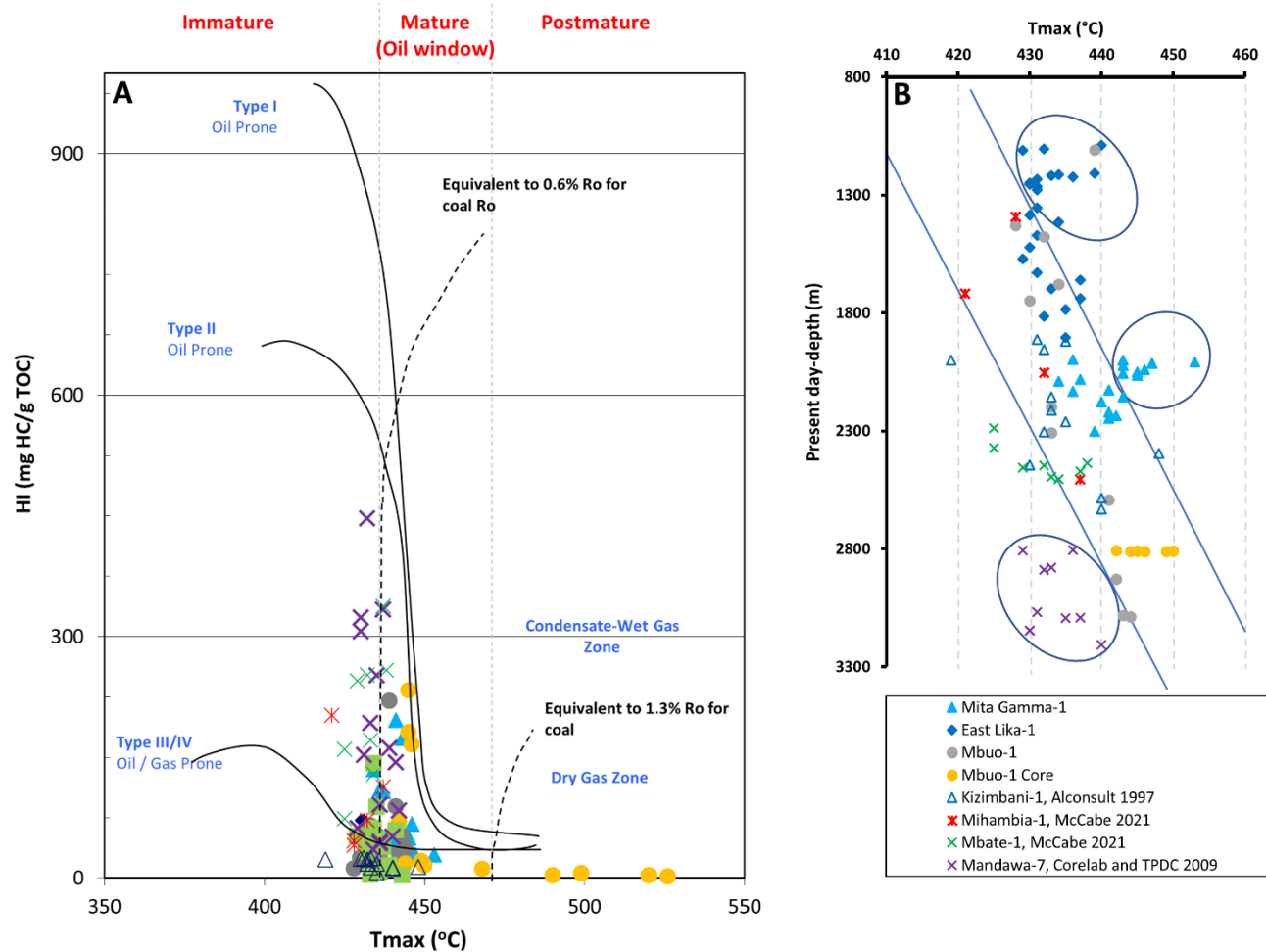


Fig. 4.11: Comparison of organic matter quality of all wells from the Mandawa Basin using: A (Tmax vs. HI) and B (Tmax vs. Present day depth). Data from Kizimbani-1, Mihambia-1, Mandawa-7, Mbate-1 wells have been compiled from literature (Alconsult and TPDC, 1997; Corelab and TPDC, 2009; McCabe, 2021).

4.4.5 Palaeogeographical control of source rock source rock potential

Paleogeography is controlled by interaction between tectonic processes, sea level fluctuations, and local climate and will influence sediment accommodation, types, and facies in both marine and continental depositional environments (Herbin et al., 1995; Leanza et al., 2000; Stasiuk and Fowler, 2004; Scotese, 2014; Cao et al., 2017; Gama and Schwark, 2023a). Therefore, spatial and temporal variations in organic matter recorded in sediment archives can be linked with the paleogeographical history of a particular area.

During the Late Triassic to Early Jurassic times the Mandawa Basin was a rift basin, which propagating northeast between the Matandu and Mbemkuru lineaments (**Figs. 4.12 and 4.13**), (Hudson, 2011; for a review see Gama and Schwark, 2023a). This rifting event is attributed to the regional extension between East Africa and Madagascar. The interplay between rifting events, sea level changes and local climate dominated the paleogeographic evolution of the Mandawa Basin, via the creation of accommodation space, formation of paleo-highs and regulation of sedimentation patterns in the basin. As result, the southern and central parts of the basin subsided at a much higher rate than the adjacent areas and formed the deepest part of the basin. Other parts of the basin such as at the western (Masasi Basement spur), eastern (Kiswere, Pande and Kizimbani), and southern margin (Ruvuma saddle) remained as structural highs and sourced most of the continental sediment into the basin. To understand the paleogeographical control of source rock potential of Lower Jurassic sediments in the Mandawa Basin, paleogeographical maps for two time slices that constitute the Pliensbachian and the Toarcian were reconstructed with respect to the position of the seven wells that penetrated Lower Jurassic successions in the Mandawa Basin (**Figs. 4.12 and 4.13**). TOC and programmed pyrolysis data from sediments of Lower Jurassic successions from four wells collected from the literature and those from three wells generated in this study were averaged and plotted on the Pliensbachian, respective Toarcian paleogeographic maps (**Figs. 4.12 and 4.13**).

4.4.5.1 Palaeogeographical control of source rock source rock potential during Pliensbachian (Mbuo Formation)

The Mandawa Basin during the Pliensbachian was considered to be a small endorheic basin, as the Masasi basement spurs and the Eastern rift high acted as barrier to rivers draining the hinterland from the West and East, thus any established rivers would probably be flowing through Malagasy trough to the North (McCabe, 2021; Gama and Schwark, 2023a). These two paleo-highs were positioned at the eastern and western basin margin, with two depocenters located in South of the Mbate-1 and Mbuo-1 wells constituting an important paleogeographical element during Pliensbachian times. Sedimentation was mainly continental to partly marine controlled by variations in sea-level and tectonics (**Fig. 4.12**). During sea level low stand, deposition of

continental sediments was limited to the basin margins, whereas coastal wetland to lacustrine/evaporitic conditions prevailed in the southern basin centre, which mostly likely was flooded during short-lived sea level highstand. The paleogeographical map indicates that sediment of the Mbuo Formation showed a wide range of depositional environment, ranging from continental, shallow, to deeper lake in dependence of distance to the sediment source area. These differences in depositional conditions would suggest variations in the nature and preservation conditions of organic matter across drill sites. TOC and programmed pyrolysis data in sediments of the Mbuo Formation on the paleogeographic map shows relatively elevated values in the Northern (Mbate-1 well) and in Southern (Mandawa -7 and Mbuo -1 wells) parts, and generally lower values (TOC <1wt%, S2 <1, and HI ≤50) in the Eastern (Mita Gamma-1, Kizimbani-1) and Western (East Lika-1) parts of the basin (**Figs. 4.12**). This observation points to variable deposition and preservation conditions of organic matter at these sites whereby large accommodation space located in the northern and southern parts was more favorable for organic matter preservation, as opposed to the western and eastern part that correspond to the basement high terraces of the basin margin, where organic matter was easily destroyed by oxidation and diluted by clastic input.

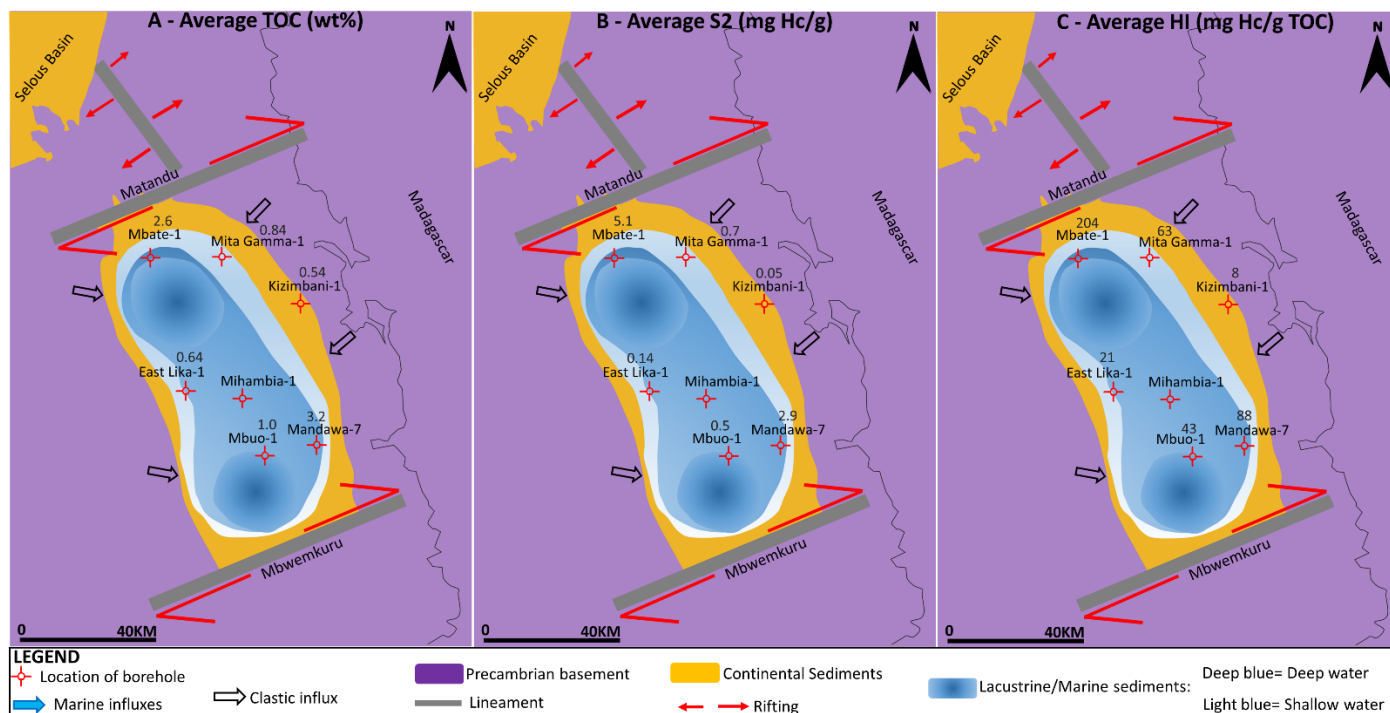


Fig. 4.12: Comparison of source rock quantity (TOC) and quality (S2, HI) of sediments from 7 wells in Mandawa Basin with respect to paleogeography during the Pliensbachian (Mbuo Formation). Paleogeographical map is adapted from (Hudson and Nicholas 2014; McCabe 2021; Gama and Schwark, 2023a). Data from Kizimbani-1, Mandawa-7, Mbate-1 wells have been compiled from the literature (Alconsult and TPDC, 1997; Corelab and TPDC, 2009; McCabe, 2021).

4.4.5.2 Palaeogeographical controls of source rock source rock potential during the Toarcian (Nondwa Formation)

Continued rifting during Toarcian coincided with transgression, resulting in an opening between the Mandawa Basin and Malagasy Trough that allowed marine in-/transgression, led to deposition of regional source and cap rocks such as the Nondwa Shale (Gama and Schwark, 2022) (**Fig. 4.13**). Existing paleo-highs in the western and eastern parts, rise in sea level and differential subsidence affected the paleogeographic evolution of the area. Sedimentation was mainly marine with partial terrigenous influx along the East-West margins. The paleogeographic map of the Toarcian period shows variations in TOC and programmed pyrolysis data for sediments of Nondwa Formation penetrated by 5 wells, with higher average TOC, S2 and HI values than in sediments from the Mbuo Formation, which is attributed to the Toarcian transgression and rifting that enhanced organic matter preservation conditions (**Fig. 4.12**). Palaeogeographical maps show that

the average TOC values in Nondwa sediment are all $\geq 1\text{wt}\%$, with the highest values found in the Mandawa-7 and Mihambia-1 wells. The source rock potential (S_2 , HI) varies spatially, with the highest values observed in the southwestern and northwestern parts, suggesting that these areas were more favorable for the deposition and preservation of organic matter. Low source rock potential observed in the Mihambia-1, Mbuo-1, East Lika-1 wells on the palaeogeographical maps (**Fig. 4.13**), would indicate degraded organic matter and significant terrestrial influx from the emerging lands in the western and eastern parts. These sediments were probably deposited in shallow, well-mixed and oxygenated waters with low to moderate productivity. The high source rock potential observed in the sediment of the Mandawa-7 well indicates that the southwestern area had persistently high productivity in the surface water accompanied by a dysoxic to anoxic water column, which promoted the preservation of organic matter.

We therefore argue that the observed distribution in TOC and source rock potential of the sediment in Mandawa Basin displayed on the paleogeographical maps may be related to changes in depositional conditions in an evolving rift basin, driven by changes in relative subsidence, eustatic sea level change and sediment influx. Organic matter was poorly preserved when high sediment flux exceeds subsidence rate, generally in the eastern and western but also southern part. Whenever high subsidence rates coupled to eustatic rise in sea level and lower sediment supply occurred, this enhanced productivity and preservation of organic matter in deep lake to restricted marginal marine conditions as shown in the Mandawa-7 well.

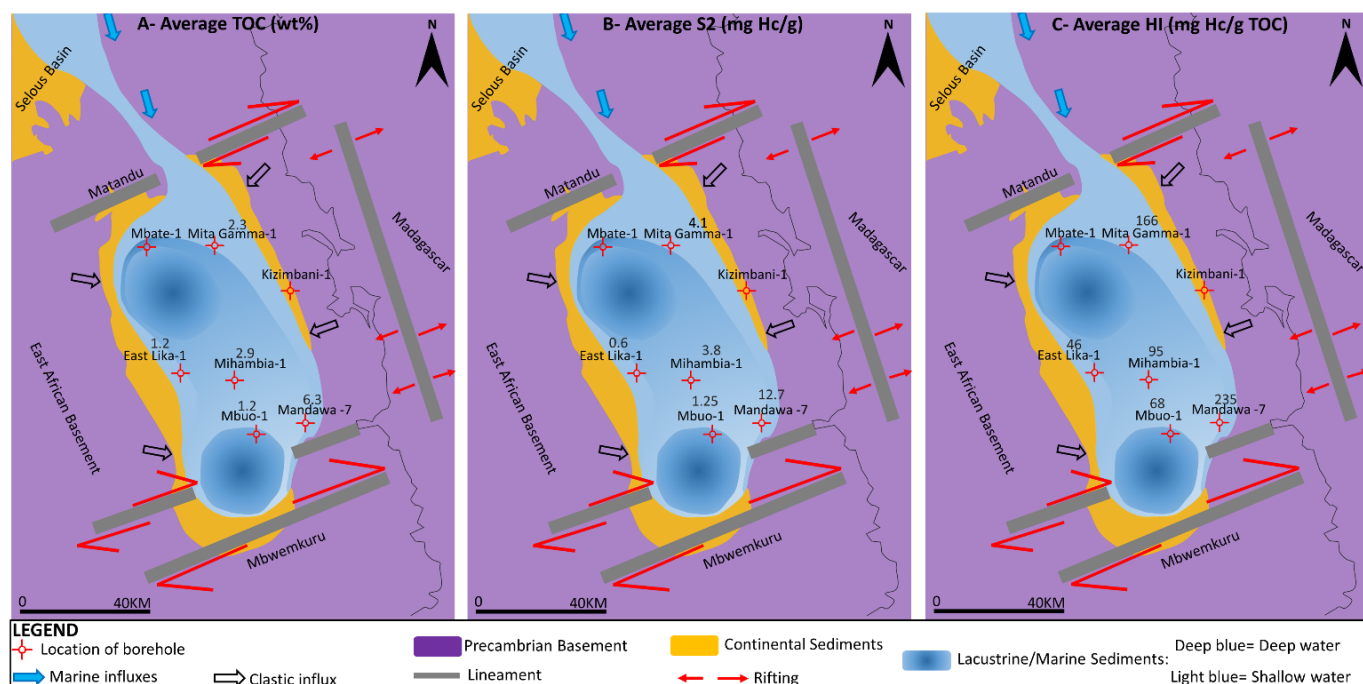


Fig. 4.13: Comparison of source quantity (TOC) and quality (S2, HI) of sediments from six wells in the Mandawa Basin in relation to paleogeography during early the Toarcian (Nondwa Formation). Paleogeographical map is adapted from (Hudson and Nicholas 2014; McCabe 2021, Gama and Schwark, 2023). Data from Kizimbani-1, Mandawa-7, Mihambia-1 wells have been compiled from the literature (Alconsult and TPDC, 1997; Corelab and TPDC, 2009; McCabe, 2021).

4.5 Conclusions

In this study, organic matter types and qualities in sediments of Lower Jurassic successions (Mbue, Nondwa and Mihambia Formations) in the Mandawa Basin, SE Tanzania have been characterized by programmed pyrolysis and isotope analysis. Programmed pyrolysis data show that organic matter is composed mainly of type III (gas prone) to IV (inert) kerogens with lesser contribution by mixed type II-III kerogen (oil-gas prone). Sediments have been deeply buried (>2000m) and undergone a wide range of thermal diagenesis, from early to over mature stages. Isotopic composition ($\delta^{13}\text{C}_{\text{org}}$) data of organic matter are consistent with pyrolytic kerogen typing, indicating preferentially terrestrial organic matter sources that changed to marine organic matter upon marine incursions, accompanied by a shift from heavier (-19 to -22 ‰ V-PDB) to lighter (-24 to -27 ‰ V-PDB) isotopic composition.

Our data indicate that source intervals found within the sediments of Lower Jurassic successions from the Mbuo-1, East Lika-1 and Mita Gamma-1 wells have poor to fair source rock potential. There are only few intervals in the Upper Mbuo and Nondwa, and Mihambia Formations in the Mita Gamma-1 and East Lika-1 wells, as well as, in the Lower Mbuo Formation in the Mbuo-1 well that have been locally enriched due to periodic marine ingression that persisted since the Pliensbachian period. This finding has implications on source rock screening in the Mandawa Basin, as most of the previous studies were based on a few samples per formation that only provided discrete points in heterogeneous and thick Pindiro Group. More detailed sampling and data quality assurance in this study allowed identification of source rock intervals and their stratigraphic variability. Synthesis of data generated in this study with literature data was used to reconstruct organic facies in a paleogeographical context revealing that the southwestern and northwestern parts of the study area appeared to have the highest source potential. This is attributed to enhanced organic matter surface productivity and preservation in the dysoxic to anoxic water column. In contrast, the eastern and western parts generally show low source potential, indicating degraded organic matter and significant terrestrial influx from emerging lands that were deposited in shallow, well-mixed and oxygenated water with low to moderate organic matter productivity.

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Chapter Five

5.0 Conclusions and outlook

5.1 Conclusions

This thesis examined the Pliensbachian - Toarcian geological evolution in the Mandawa Basin and its implication on source rock potential using multidisciplinary techniques. This was undertaken to constrain paleodepositional conditions, lithofacies, TOC variability, and source rock potential, as well as reconstructing basin-fill history for the Lower Jurassic strata in Mandawa Basin. To improve geological understanding at regional scale, all controversy on facies, stratigraphy and pathways of marine incursions prior/ or during Toarcian transgressive phase compiled from the published were discussed (**Chapter One**). This has enabled compilation of Pliensbachian - Toarcian lithostratigraphic sections as well as reconstruction of paleogeographic of seven sedimentary basins in the East African margin.

This study was based on 185 rock cuttings samples collected from three boreholes (Mbuo-1, Mita Gamma-1, and East Lika-1) that penetrated sediments of Lower Jurassic successions (Mbuo, Nondwa, and Mihambia Formations), 25 core samples from the sediments of Lower Mbuo

Formation in the Mbuo-1 well, and ten samples from the sediment of Nondwa evaporitic sequences outcropping in the southern central part of the basin. In addition, geophysical wireline logs data was supplemented in the study to allow quasi-continuous documentation of lithology at higher resolution (~15cm). The main findings of this study are outlined below.

➤ Lithofacies and TOC derived from wire line geophysical logs

This study has shown that the sediments of the Pindiro Group are made up of complex siliciclastic, carbonatic, and evaporitic heterolithic facies. The multi-mineral components in the heterolithic facies and as well as those introduced by either detrital minerals or diagenetic effects, may lower logs contrast in Spectral Gamma Ray (Th, U and K proportions). Moreover, these minerals appear to influence log measurements as compared to the presence organic matter and thus complicated both facies interpretation and organic matter prediction from wireline geophysical logs.

Despite these limitations, the SGRL method has proven to be effective for highlighting the changes in detrital sediment input, hydrodynamics effect, weathering intensity, and diagenetic effects through time

➤ TOC variability with respect to basin-fill history and changes in depositional environment

According to this study, the sediments of the Pindiro Group (Mbuo, Nondwa, and Mihambia Formations) were deposited in a wide range of depositional conditions ranging from fluvio-deltaic, limnic, and restricted to shallow marine conditions. Furthermore, it was revealed that changes in depositional conditions were correlated with TOC concentration. This imply that environmental differences exerted control over organic matter enrichment due to variations in productivity, preservation, and destruction, which are directly linked to basin evolution and eustatic level changes. TOC peaks in successions of the Upper Mbuo, Nondwa and Mihambia Formations coincide with decreased clastic input and a rise in eustatic sea levels as opposed to generally lower TOC <1 (wt%) accompanied by higher terrigenous input in successions of the Lower Mbuo Formation.

Inverse trends in TOC versus clastic components, a positive linear relationship between TOC and carbonates, and higher amount of clastics (>57 wt%) in the Mbua and Mihambia Formations suggest a dominantly clastic sedimentary system with relatively stable carbonate and organic matter input in proximity to the source area. This study proposes that organic matter enrichment in the Upper Mbua, Nondwa and Mihambia Formations were deposited during periods of higher organic productivity driven by a rise in eustatic sea level, and coincide with a decline in terrigenous input from the hinterland.

➤ Pliensbachian-Toarcian Paleodepositional conditions

This study reveals that paleodepositional settings during Pliensbachian-Toarcian fluctuated due to changes in subsidence and eustatic rise in sea level. During the Pliensbachian, the southern-central part of the basin which, experienced relatively high subsidence, exhibited a more reduced depositional setting compared to western and eastern flanks, which had relatively shallow, well-mixed, and oxic depositional settings in proximity to source area. The depositional setting persisted until Toarcian times, when further subsidence coincided with marine ingression/transgression that resulted in establishment of restricted marine conditions. This led to a shift towards less oxic depositional conditions in the east and western flanks, while more reduced conditions continued in the southern central part of the basin.

Continual rifting during late Early Jurassic led to an open marginal marine setting, reduced deposition persisted in the eastern flanks and as well in the southern central part of the basin, whereas high sediment influx in the western flank promoted oxic - strongly deposition condition.

➤ Organic matter types and quality of source rock intervals in the sediment of Lower Jurassic successions

This study has shown that organic matter for the studied source rock intervals in Jurassic sediments samples is composed mainly of type III (gas prone) to IV (inert) kerogens with lesser contribution by mixed type II-III kerogen (oil-gas prone). There are only few intervals in the Upper Mbua and Nondwa, and Mihambia Formations in the Mita Gamma-1 and East Lika-1 wells, as well as, in the Lower Mbua Formation in the Mbua-1 well that have been locally enriched due to

periodic marine ingression that persisted since the Pliensbachian period. These sediments have been deeply buried (>2000m) and undergone a wide range of thermal diagenesis, from early to over mature stages. This finding has implications on source rock screening in the Mandawa Basin, as most of the previous studies were based on a few samples per formation that only provided discrete points in the heterogeneous and thick Pindi Group. More detailed sampling and data quality assurance in this study allowed identification of source rock intervals and their stratigraphic variability. Synthesis of data generated in this study with literature data was used to reconstruct organic facies in a paleogeographical context revealing that the southwestern and northwestern parts of the study area appeared to have the highest source potential. This is attributed to enhanced organic matter surface productivity and preservation in the dysoxic to anoxic water column. In contrast, the eastern and western parts generally show low source potential, indicating degraded organic matter and significant terrestrial influx from emerging lands that were deposited in shallow, well-mixed and oxygenated water with low to moderate.

5.2 Outlook

The investigation conducted by this thesis was limited to the syn-rift sedimentary packages in the onshore part of the Mandawa Basin. The offshore part of the Tanzanian coastal basin, where these sediments are deeply buried, was not within the scope of this study. In the off-shore part, these sedimentary sequences have been inferred from seismic data but not yet penetrated by boreholes. The recent acquisition of high-quality subsurface data and drilling has led to the discovery of a huge volume of natural gas in Block 2, located approximately 100 km from the Mandawa Basin, as well as in the Rovuma Basin in Mozambique. Therefore, it will be worthwhile to undertake a study on the tectonic sedimentary evolution of the basin, which will provide insights on how these sedimentary packages are distributed. This will improve our understanding of the distribution of important elements of the petroleum system such as source rocks including (Permo-Triassic and Jurassic), reservoirs, seals, and migration pathways.

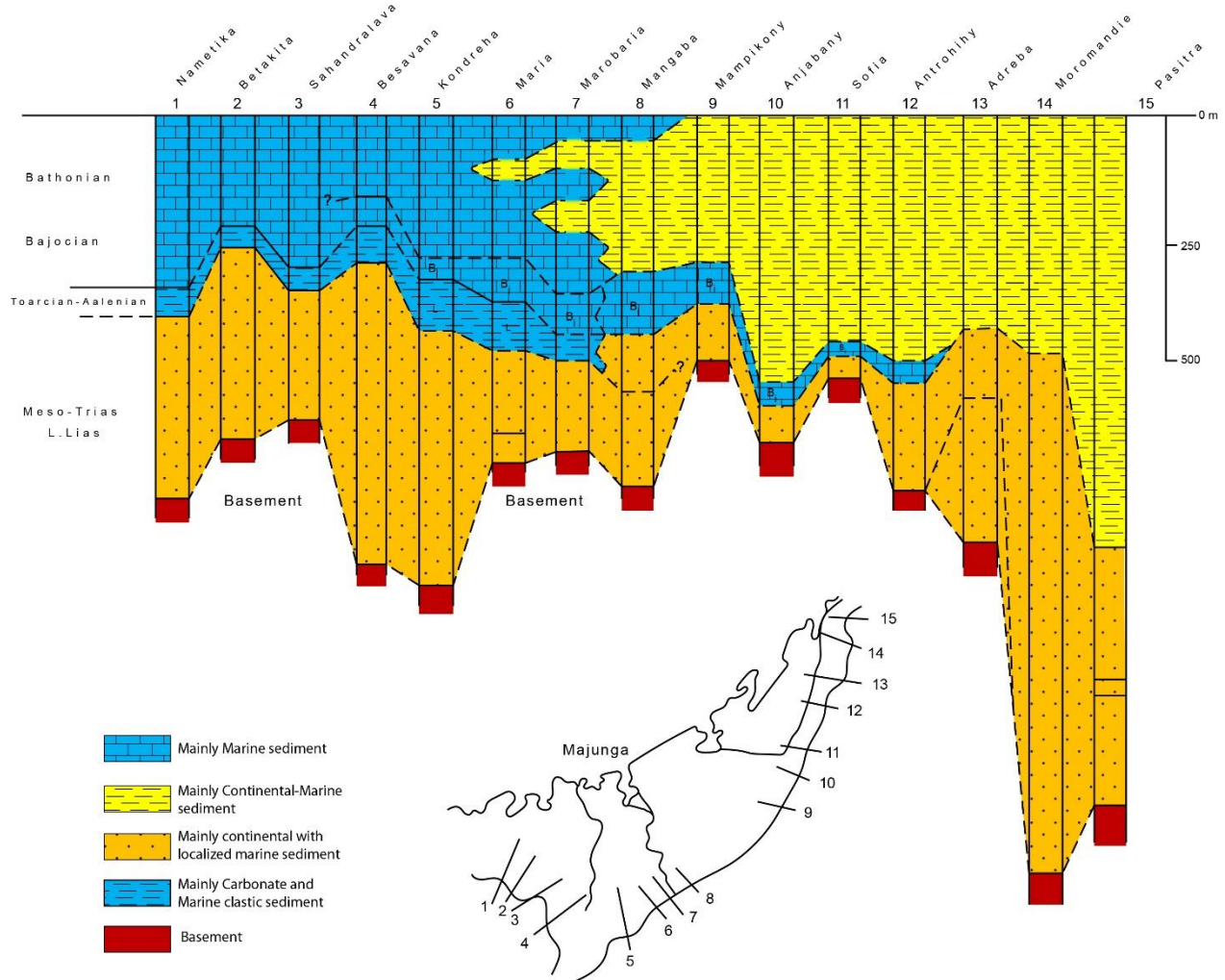
Due to the limited number of drilled wells, incomplete outcropping sections, and halokinesis, only a few biostratigraphy studies have been conducted on the sediments of the Mbuo and Nondwa

Formations in the Mandawa Basin. Despite efforts made by international oil companies to improve the biostratigraphy of Pindiro Group's sediments from six borehole that were drilled between 1990-2009, most of the findings from this works have not been published. Thus, a controversial age assignment for the bases of the Mbuo and Nondwa Formation exists as summarized in chapter 1. Therefore, there is a need to undertake detailed biostratigraphy work on the seven boreholes that have penetrated the sediment of Pindiro Group.

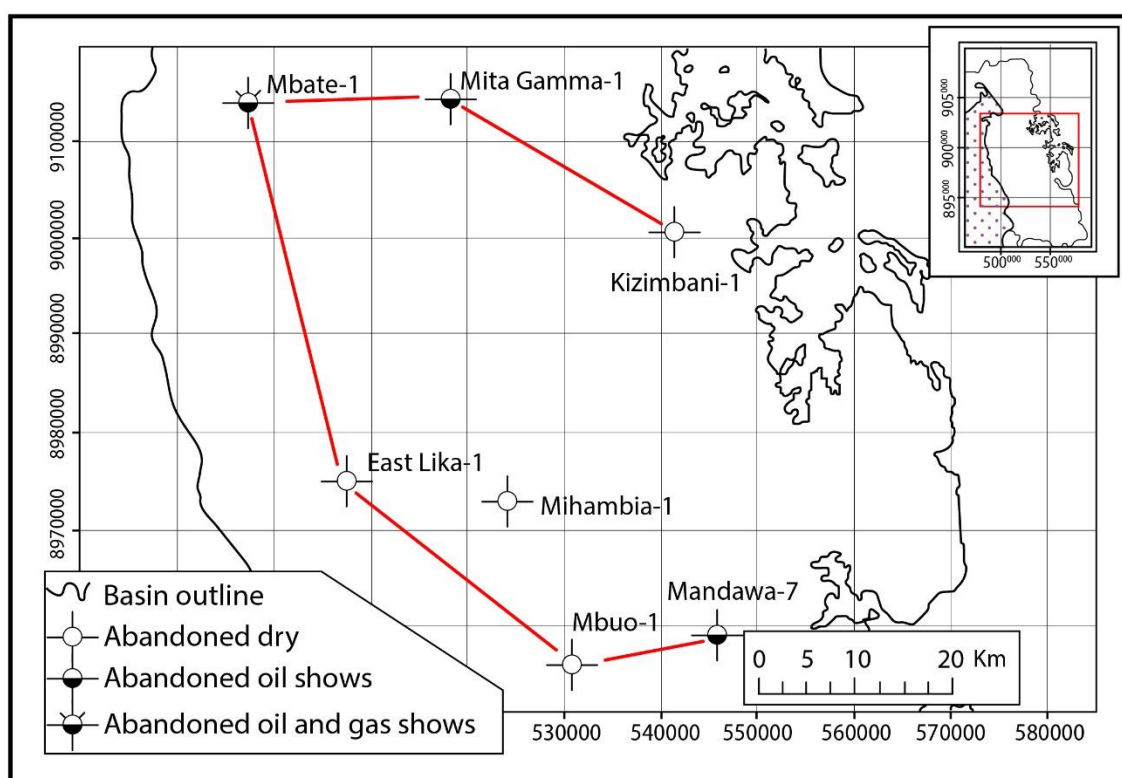
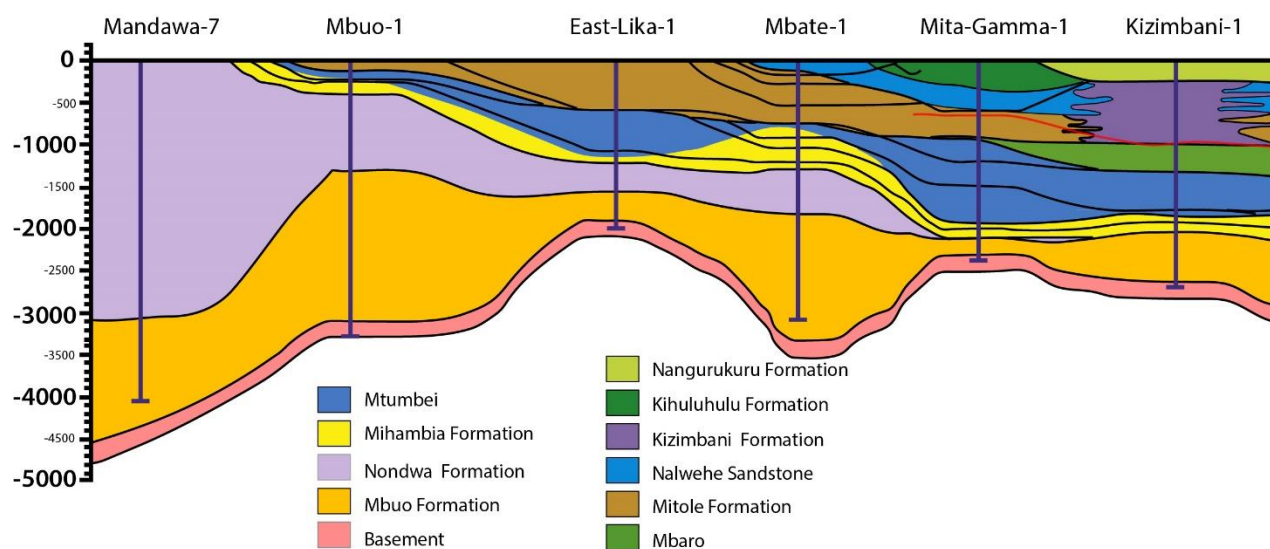
Reservoir properties of sandstone and carbonate units in the Mbuo and Mihambia sediments of the Pindiro Group, are generally poorly understood due to lack of core and outcrops. However, sidewall core (SWC) and geophysical wire line data can be used as supplementary sources of information. The interpretation of Spectral Gamma Ray data from two boreholes (East Lika-1 and Mita Gamma-1) indicates that the thickest packages of sandstone and limestone, located in the western and eastern flanks, have the potential to serve as reservoirs in the syn-rift package. Future studies should focus on these areas to gain a better understanding of their potential as reservoirs.

Appendices

Chapter One



Supplementary Fig. 1.1 : Pre-Toarcian and Post Toarcian Lateral Facies variation in Majunga Basin, adapted from (Boast and Nairn, 1982).



Supplementary Fig. 1.2: Lithostratigraphic correlation of pre-Toarcian and Toarcian strata penetrated by wells in Mandawa basin adapted from (Hudson, 2011; Hudson and Nicholas, 2014).

Country	Sedimentary basin	Well	Location		Total depth (m)	Oldest strata penetrated	References
			LATITUDE	LATITUDE			
Somalia	Ahl Mado	Sagaleh-1	9° 25' 00 " N	50° 32' 00"E	3212	Lower Jurassic	Barnes ,1976
		Cotton-1	9° 33' 00 " N	50° 32' 00"E	3310	Lower Jurassic	
	Borama Bihendula	Darrin	10° 39' 00 " N	49° 45' 00"E	2966	Basement	
		Las Anod 1	8° 28' 00 "N	47° 12' 00"E	1664	Basement	
		Faro Hills 1	9° 38' 00 "N	47° 47' 00"E	1637	Basement	
		Yaguri	8° 39' 00 "N	47° 02' 00"E	1441	Basement	
		Buran 1	10° 15' 00 "N	48° 54' 00"E	2437	Basement	
		Burhisso-1	8° 19' 00 "N	47° 54' 00"E	1551	Basement	
	Somali Embayment	El Cabobe-1	4° 14' 48 " N	47° 40 ' 42"E	4428	Upper Triassic	Coffin and Rabinowitz ,1988
		Coriole -1	1° 50' 39 " N	44° 33 ' 16"E	3518	Lower Jurassic	
		Brava-1	1° 04' 00" N	43° 31 ' 00"E	3810	Triassic	
		Gira	5° 30' 00 " N	48° 5 ' 00"E	3890	Lower Jurassic	Barnes, 1976
		Marai	4° 31' 00 " N	47° 25 ' 00"E	4115	Lower Jurassic	
		Obia-1	5° 56' 00 " N	48° 54' 00"E	4883	Lower Jurassic	
		Garad Mare-1			3250	Basement	
		Hat-1			1000	Basement	
		Merca-1	1° 52' 00 " N	44° 55' 00"E	3998	Lower Jurassic	
Kenya	Lamu basin	Ras Kaluj	3° 49' 00 " S	39° 15' 00"E	1537	Permian	Coffin and Rabinowitz , 1988
		Kencan	0° 18' 57 " S	39° 46' 16.6"E	3863	Permian	Sproule, 2012
	Mandera	Elgal-1	1° 22' 47 " S	39° 53' 09"E	1280	Permian	
		Elgal-1	1° 27' 32" S	39° 58' 06"E	1908	Triassic	
Madagascar	Majunga	Ambilobe	13° 03' 00 " S	48° 52' 40"E	1670	Triassic	Coffin and Rabinowitz , 1988
		Sophia 1	15° 29' 12 " S	47° 09' 00"E	3026	Lower Jurassic	
		Tuileri	15° 52' 00" S	46° 30' 00"E	2659	Lower Jurassic	
		C.St Andre-1	16° 11' 00" S	44° 28' 30"E	1668	Triassic	
		C.St Andre-2	16° 11' 00" S	44° 28' 30"E	2153	Permian	
		Chesterfield	16° 22' 00" S	43° 57' 30"E	4774	Triassic	
		Bemolanga-1	17° 41' 24" S	44° 48' 00"E	1501	Lower Jurassic	
	Morondava	Belinta 1	18° 21' 52" S	44° 52' 00"E	2529	Permian	
		Belinta 2	18° 18' 17" S	45° 03' 00"E	2513	Permian	
		Belinta 3	18° 20' 00" S	45° 18' 24"E	1201	Upper Triassic	
		Ankamotra 2	18° 28' 00" S	44° 18' 45"E	3506	Triassic	
		Maroaboaly 1	18° 33' 00" S	45° 09' 00"E	1986	Upper Triassic	
		Vaucluse 1	19° 28' 00" S	43° 59' 20"E	4027	Lower Jurassic	
		Serinan-1	19° 37' 00" S	44° 49' 00"E	3658	Upper Triassic	
		East Serinam 1	19° 36' 00" S	44° 43' 00"E	3048	Triassic	
		Ankasofotsy	20° 57' 00" S	44° 13' 00"E	2346	Upper Triassic	
		Mandabe 1	21° 06' 48" S	44° 57' 00"E	2749	Lower Jurassic	
		Skily 1	21° 51' 00" S	44° 07' 00"E	2832	Lower Jurassic	
		Ampandramitsetak	22° 07' 00" S	44° 39' 00"E	2701	Lower Jurassic	
		Amabalabe	21° 56' 51" S	45° 16' 00"E	2181	Permian	
		Mandevy 1	22° 22' 51" S	44° 12' 00"E	3911	Lower Jurassic	
		Lambosina 1	22° 38' 06" S	44° 34' 00"E	2589	Lower Jurassic	

		Manera 1	22° 51' 00" S	44° 18' 00"E	3911	Lower Jurassic	
		Vohindolo 1	22° 51' 00" S	44° 50' 00"E	3464	Triassic	
		Vohindolo Bis	22° 49' 33" S	43° 44' 00"E	3426	Upper Triassic	
		Bezaha Bis	23° 27' 16" S	44° 31'30.6"E	2114	Permian	
		Salonivo 1	23° 31' 00" S	44° 47' 00"E	2650	Permian	
		Antsokay 1	23° 15' 00" S	44° 12' 00"E	1139	Upper Triassic	
		Lac 1	24° 10' 00" S	43° 56' 00"E	2449	Triassic	
		Ambanasa 1	24° 10' 00" S	43° 56' 00"E	4670	Upper Permian-Lower Triassic	
		Vohibasias 1	22° 47' 55" S	45°40' 10"E	2878	Upper Permian-Lower Triassic	
Tanzania	Ruvu	Makarawe	05° 33' 10"	38°52'11"	3821	Triassic	Strata data 2006 ; Core lab and TPDC, 2009
		Kiwangwa	06° 21' 43"	38°32'56"	3514	Triassic	
	Dar es Salaam Platform	Kisangire	7°29' 09" S	38° 32' 42"E	3296	Lower Jurassic	Coffin and Rabinowitz , 1988
	Rufiji	Lukuliro	8°21' 31"S	38° 25' 00"E	2367	Lower Jurassic	Coffin and Rabinowitz 1988 ; Strata data 2006 ; Core lab and TPDC, 2009
		Kizimbani-1	9°02' 30"S	39° 22' 32"E	2697	Lower Jurassic	Balduzi et al., 1992 ; Coffin and Rabinowitz 1988
	Mandawa	Mandawa 7	9°25' 00" S	39° 25' 00"E	4065	Lower Jurassic	Kent 1971, Coffin and Rabinowitz 1988
		Mbuo-1	09°26'41" S	39°16'51"	3313	Basement	Shell, 1990; StrataData 2006 ; CoreLab and TPDC, 2009
		Mita Gamma-1	08°54'56" S	39°09'53"	2390	Basement	StrataData 2006 ; CoreLab and TPDC, 2009
		Eastlika	09°16'26" S	39°04'04"	2002	Basement	
		Mbate-1	08°55'08" S	38°58'29"	3317	Lower Jurassic	
		Mihambia	09°17'35" S	39°13'12"	2503	Lower Jurassic	
	Selous	Liwale-1	09°25'53" S	37°31'16"	1762	Permo-Triassic	StrataData 2006 ; CoreLab and TPDC, 2009
	Ruvuma	Lukuledi-1	10°09'56" S	39°39'58"	1941	Basement	

Supplementary Table 1.1: List of wells from 13 sedimentary basins in the East African Margin that have penetrated Permo-Triassic to Lower Jurassic sediments in the East African Margin.

Sedimentary Basin	Source bed	Lithology	Depositional environment	TOC (wt%)	Kerogen type	Level of maturation	References
Mandawa	Mbuo	Shale	Fluvial lacustrine to restricted marine	0.3-10	I,II and III	Immature to overmature	TPDC, 2003; Zhou et al., 2013; Davison and Steel 2018
	Nondwa	Shale		1.2-8.7			TPDC, 2003; Kagya, 1997; Zhou et al., 2013; Davison and Steel 2018
Majunga	Beronono	Shale	Restricted marine	1.3-23.6	II, III	Mature	Webster and Ensign 2007; Davison and

							Steel 2018; Davison, 2018, Boote, 2019
Morondava	Andafia	Shale and marl	Restricted marine	0.6-7.7	II	Mature	Davison and Steel 2018; Boote, 2019
Somali	Adigrat	Shale, coaly	Continental	6.59-29.6	III	Over mature	Ali and Lee 2019
	Hamanlei	Shale	Tidal flat and Lagoonal facies	3.4-6.4	II, III	Immature to mature	Zhou et al., 2013; Mohammedyasin et al., 2020

Supplementary Table 1.2: distribution of TOC in (Pliensbachian Toarcian) source rocks from various part of East African Margin basins complied from published literature.

Chapter Two

Lithology	K (%)	Th (ppm)	U(ppm)	Th/U	Th/K	References
Black shale						
Black shale (n=19)	n.d	13.8	11.2	1.2		Adams and Weaver, 1958
Mancos shale (n=131)	2	10.3	3.5	2.9	4.7	Pliier and Adams, 1962
Green river shale (n=3)	n.d	5.8	5.2	1.1		McLennan and Taylor, 1980
Black shale (n=35)	2	7.3	7.3	1.0	3.8	Dypvik and Eriksen, 1983
Janus Jellet black shale (n=2)	4	12.5	5.5	1.4	3.8	Dypvik and Bue, 1984
Jet rock black shale (n=50)	2.25	14.5	7.6	1.9	6.4	Myers and Wignall, 1987
Kimmeridge clay (n=12)	n.d	18.1	7.3	2.5		Tribovillard et al., 1994
Givetian black shale (n=12)	2	12.0	5.0	2.4	6.0	Lüning et al., 2004
Tanezzuft shale (n=10)	4	24.0	15.0	1.6	6.0	Fello et al., 2006
Carbonaceous shale (average abundance)	n.d	11.5	53.0	0.2		Mernagh and Mieztis, 2008
Vale das Fontes shale (n=20)	2.3	6.0	4.6	1.3	2.6	Correia et al., 2013
Hekking Black shale (n=10)	3.7	6.4	14.0	0.5	1.7	Marín et al., 2021
Huai Hin shale	1.2	2	2.7	1.6	2.6	Phujareanchaiwon et al., 2021
Range	1-4	6-26	4-53	1-3	1-6	
Median	2.3	12	7.3	1.4	4.7	
Grey/Green shale						
Grey/green shale (n=52)	n.d	13.1	3.2	4.1		Adams and Weaver, 1958
(Average Abundances)	2.3	12.3	3.2	3.9	5.4	Ronov and Migdisov, 1970
(Average Abundances)	3.0	12.0	3.7	3.2	4.0	Wedepohl, 1970
(Average Abundances)	3.2	12.0	3.0	4.0	3.8	Cosgrove, 1972

Cambro-silurian shale (n=16)	3.3	12.9	3.1	4.2	3.9	Dypvik and Eriksen, 1983
Lower Lias shale (n=30)	3.8	11.4	3.0	3.8	3.0	Van Buchem et al., 1992
Roswell Harbour shale (n=24)	4.8	7.8	2.0	3.9	1.6	Hayashi et al., 1997
Grey/green shale (n=11)	6.7	11.8	4.2	2.8	1.8	Svendsen and Hartley, 2001
(Average Abundance)	2.5	12.0	4.2	2.9	4.8	Mernagh and Mieztis, 2008
Buanji shale (n=3)	3.9	19.2	5.1	3.8	5.0	Kasanzu et al., 2016
Toarcian blue Marl	1.59	5.9	1.5	3.9	3.7	Reolid et al., 2020
Range	1.59-6.7	5.9-19	1.5-5	2.8-4	1.6-5	
Median	3.8	11.8	3.1	3.8	3.7	

Red/yellow shale						
Red/yellow shale (n=26)	n.d	14.625	1.76	8.3		Adams and Weaver, 1958
Red/yellow shale (n=8)	4.9	18	2.5	7.1	3.6	Yamaguchi, 2006
Ikorongo shale (n=30)	4.0	19	2.8	6.8	4.8	Kasanzu et al., 2008
Buanji shale (n=14)	4.1	18	2.3	8.1	4.4	Kasanzu et al., 2016
Range	4-5	14-19	1-3	7-8	3.6-4.8	
Median	4.0	17.6	2.5	7	4.8	
Siltstone						
Beaufort Siltstones (n=550)	3.8	15.8	3.7	4	4	Beeson, 1980
Namurian siltstones (n=40)	2.2	13.5	3.6	4	6	Myers and Bristow, 1989
Kössen Siltstones (n=9)	2.4	11.6	3.1	4	5	McRoberts et al 1997
Susiyama-Wakino Siltstones (n=3)	2.1	8.0	1.7	5	4	Asiedu et al 2000
Ikorongo Siltstones (n=13)	1.3	15.1	2.5	6	11	Kasanzu et al., 2008
Cerro Ballena Siltstones (n=179)	1.7	10.2	2.6	4	6	Paredes et al., 2015
Oti-Kwayu Siltstones (n=27)	2.7	8.1	2	4	3	Amedjoe et al., 2018
Range	1.2-3.8	8-15.8	1.7-3.7	4-6	3-11	
Median	2.2	11.6	2.6	3.9	4.9	
Sandstone						
Sandstone		5	1	5.0		Moore 1957
Sandstone (n=4)		4.7	0.8	5.8		Adams and Weaver, 1958
Sandstone		2.0	0.5	4.0		Moore 1960
Sandstone (n=9)	1.6	8.9	2.5	3.6	5.4	Dypvik and Eriksen, 1983
Sandstone (n=267)	0.7	8.0	2.4	3.3	10.8	Dypvik and Bue, 1984
Numurian Sandstone(n=28)	1.6	11.9	2.9	4.1	7.4	Myers and Bristow, 1989
Sandstone (n=26)	4.9	12.1	2.5	4.8	2.5	Svendsen and Hartley, 2001
Sandstone (n=90)	0.8	5.9	1.4	4.3	7.0	Martinius et al., 2002

Average abundance	1.8	5.0	1.5	3.3	2.8	Mernagh and Mieztis, 2008
Cerro Ballena Sandstone (n=230)	1.7	9.9	3.5	2.8	5.8	Paredes et al., 2015
Sandstone (n=33)	0.7	2.0	0.8	2.6	2.6	Phujareanchaiwon et al., 2021
Range	0.7-4.9	2-12	0.8-3.5	2.6-5.8	2.5-10.8	
Median	1.6	8.4	2.4	3.5	5.6	

Carbonates						
Average abundance (n=54)		1.7	2.2	0.8		Adams and Weaver, 1958
(Average Abundance)		2.0	2.0	1.0		Moore 1960
Carbonates		1.7	2.2	0.8		Turekian and Wedepohl, 1961
(Average Abundance)	0.43	0.2	0.2	1.0	0.5	Horn and Adams, 1966
Carbonate (Average Abundance)	1.0	2.0	1.7	1.2	2.1	Ronov and Migdisov, 1970
Average Carbonates		2.0	2.0	1.0		Cosgrove, 1972
Chase Carbonate	1.1	2.4	4.1	0.6	2.2	Luczaj 1998
Givetian Carbonate	1.0	1.6	7.0	0.2	1.6	Lüning et al., 2004
Average Abundance	0.6	1.2	1.9	0.6	1.9	Mernagh and Mieztis, 2008
Vale das Fontes Carbonates (n=5)	1.3	3.1	2.4	1.3	2.4	Correia et al., 2013
L. Cambrian Carbonate		0.4	1.3	0.3		Wei et al 2021
Range	0.3-1.3	0.2-3.1	0.2-7	0.2-1.3	0.5-2.4	
Median	1.0	2.2	2.2	1.0	2.2	
Evaporites						
Anhydrite	n.d	0.15	0.10	1.50		Adams and Weaver, 1958
Utah-Salado Anhydrite (n=10)	0.2	0.2	0.2	1.0	1.2	Stewart, 1963
Halite (n=9)		0.2	0.1	2.0		Stewart, 1963
Evaporites	0.4	0.2	0.2	1.0	0.5	Horn and Adams, 1966
(Average abundance) Halite	0.3	0.2	n.d		0.6	Moore, 1971
Halite	0.2	0.2			1.2	Moore, 1971
Evaporites (Average Abundance)	n.d	0.2	0.1	2.00		Mernagh and Mieztis, 2008
Gypsum/anhydrite	0.3	0.30	0.2	1.5	1.2	Yesilova , 2021
Range	0.2-0.4	0.2-0.3	0.1-0.2	1-2	0.5-1	
Median	0.3	0.2	0.2	1.5	1.2	

Supplementary Table 2.1: Compilation U, Th, K concentration in sediments from the literature

A

Formation	Depth Interval (m)	Medium Depth (m)	Lithology	Clastics %	Carbonate %	Salt %	Carbonate plus salt %	Crystalline rock %	TC (wt%)	TOC (wt%)	GR (API)	GR_K (%)	GR_Th (ppm)	GR_U (ppm)	Th/U	Th/K	U/K
Mihambia	1996-1999	1997.5	[§] Siltstone	76	9	14	22	0	2.13	1.08	72	1.70	10.72	2.18	4.91	6.31	1.28
	1999-2002	2000.5	[§] Siltstone	85	14	1	15	0	2.22	0.58	70	1.94	11.51	1.57	7.34	5.92	0.81
	2006-2009	2007.5	[§] Siltstone	74	18	7	25	0	3.11	0.97	80	1.95	15.69	0.88	17.87	8.06	0.45
	2012-2015	2013.5	[§] Siltstone	80	15	5	20	0	2.30	0.50	73	1.66	10.43	2.13	4.91	6.28	1.28
	2021-2024	2022.5	[€] Siltymudstone, dolomite	47	29	21	50	0	6.24	2.71	51	1.03	5.84	2.10	2.79	5.66	2.03
	2039-2042	2040.5	[£] Shale	57	17	25	42	0	3.43	1.44	62	1.72	9.75	1.62	6.04	5.67	0.94
	2048-2051	2049.5	* Sandy dolomite	65	13	21	34	0	2.86	1.29	47	1.22	9.02	0.92	9.84	7.38	0.75
	2054-2057	2055.5	* Sandy dolomite	55	25	20	44	0	3.92	0.94	42	0.93	7.17	1.13	6.35	7.70	1.21
	2070-2073	2063.5	Shale	50	26	23	49	0	3.89	0.77	45	0.97	7.01	1.28	5.47	7.23	1.32
	2079-2082	2080.5	[®] Shale, dolomite	52	21	24	45	0	5.33	2.82	54	1.50	9.58	1.22	7.86	6.39	0.81
	2088-2091	2089.5	^h Shales, halite interbeds	46	17	35	52	0	4.27	2.19	44	1.22	7.15	1.48	4.84	5.86	1.21
Nondwa	2106-2109	2107.5	^ç Halite	9	2	88	90	0	0.58	0.28	16	0.26	1.35	0.59	2.27	5.12	2.25
	2112-2115	2113.5	^ç Halite	7	2	91	93	0	0.41	0.18	21	0.42	2.82	0.33	8.48	6.72	0.79
	2124-2127	2125.5	[®] Shales	43	34	20	54	0	7.53	3.45	55	1.39	7.37	1.93	3.81	5.30	1.39
	2130-2133	2131.5	[®] Argillaceous shale	65	12	21	33	0	2.98	1.51	73	1.62	10.21	2.78	3.68	6.30	1.71
Mbua	2143-2146	2144.5	^Δ Shales	65	17	17	34	0	2.66	0.60	91	2.40	18.22	0.76	23.91	7.57	0.32
	2146-2149	2147.5	Claystone	65	33	1	34	0	5.03	1.03	100	2.93	18.03	1.83	9.85	6.15	0.62
	2155-2158	2156.5	^Δ Shales	61	13	25	38	0	2.59	1.03	66	1.83	10.95	1.43	7.64	5.97	0.78
	2167-2170	2168.5	^ç Argillaceous sandstone	59	23	18	41	0	3.29	0.54	60	1.71	9.56	1.20	7.99	5.58	0.70
	2176-2179	2177.5	[°] Sandstone	66	16	17	34	0	2.63	0.65	54	1.81	8.90	1.29	6.87	4.92	0.72

	2195-2198	2196.5	◊ Sandstone	79	6	14	20	0	1.31	0.62	91	3.26	15.22	1.96	7.78	4.67	0.60
	2204-2207	2205.5	∞ Claystone	64	0	34	34	0	1.77	1.97	84	2.83	14.31	1.65	8.67	5.06	0.58
	2207-2210	2208.5	¥ Claystones	62	9	29	38	0	1.76	0.70	76	2.72	10.81	2.23	4.85	3.98	0.82
	2216-2219	2217.5	Claystone	63	11	25	36	0	1.99	0.63	111	3.33	17.31	2.98	5.81	5.20	0.90
	2219-2222	2220.5	# Argillaceous siltstone	64	20	15	35	0	3.12	0.71	79	2.55	11.82	2.26	5.24	4.64	0.89
	2222-2225	2223.5	Clay evaporite	22	8	70	78	0	1.32	0.38	93	3.59	16.12	2.06	7.82	4.49	0.57
	2225-2228	2226.5	Clay evaporite	30	6	64	70	0	1.13	0.37	120	3.89	18.54	2.50	7.41	4.76	0.64
	2231-2234	2232.5	Clay evaporite	31	23	45	69	0	3.38	0.58	119	4.13	18.79	1.85	10.15	4.55	0.45
	2234-2237	2235.5	§ Claystone	67	5	27	32	0	1.49	0.92	80	2.45	9.54	2.18	4.39	3.89	0.89
	2240-2243	2241.5	# Argillaceous siltstone	75	6	18	24	0	1.22	0.48	83	3.12	12.90	0.90	14.40	4.14	0.29
	2246-2249	2247.5	º Sandy siltstone	67	4	28	32	0	0.98	0.46	84	2.60	9.57	1.70	5.63	3.68	0.65
	2277-2280	2278.5	¢ Argillaceous sandstone	47	2	50	52	0	0.82	0.58	59	2.07	7.03	1.35	5.19	3.39	0.65
	2282-2285	2283.5	Argillaceous sandstone	46	0	53	53	0	0.64	0.65	71	2.52	9.81	1.46	6.73	3.90	0.58
	2298-2301	2299.5	‰ Claystone , sandstone interbeds	41	3	55	58	0	0.77	0.38	93	2.85	17.20	1.71	10.05	6.04	0.60
	2301-2304	2302.5	‰ Claystone , sandstone interbeds	69	3	27	30	0	0.81	0.40	78	2.39	11.82	1.94	6.08	4.94	0.81
Basement	2304-2307	2305.5	Biotitic gneiss	0	0	0	0	100	0	0	51	2.00	6.86	1.21	5.67	3.43	0.60
	2307-2310	2308.5	Biotitic gneiss	0	0	0	0	100	0	0	34	1.21	2.85	1.14	2.51	2.37	0.94
	2310-2313	2311.5	Biotitic gneiss	0	0	0	0	100	0	0	36	1.38	2.76	1.43	1.93	2.00	1.04
	2313-2316	2314.5	Biotitic gneiss	0	0	0	0	100	0	0	41	1.45	3.79	1.54	2.46	2.61	1.06
	2316-2320	2318	Biotitic gneiss	0	0	0	0	100	0	0	41	1.64	4.01	1.09	3.69	2.45	0.66
	2320-2323	2321.5	Biotitic gneiss	0	0	0	0	100	0	0	47	2.11	3.16	1.62	1.95	1.50	0.77

2323-2326	2324.5	Biotitic gneiss	0	0	0	0	100	0	0	47	2.04	3.38	1.68	2.01	1.66	0.83
2326-2329	2327.5	Biotitic gneiss	0	0	0	0	100	0	0	46	1.93	3.29	1.34	2.45	1.70	0.69
2329-2332	2330.5	Biotitic gneiss	0	0	0	0	100	0	0	46	1.87	3.38	1.78	1.90	1.81	0.95
2332-2335	2333.5	Biotitic gneiss	0	0	0	0	100	0	0	63	2.02	6.23	1.88	3.32	3.09	0.93
2335-2338	2336.5	Biotitic gneiss	0	0	0	0	100	0	0	43	2.02	2.66	1.40	1.90	1.31	0.69
2338-2341	2339.5	Biotitic gneiss	0	0	0	0	100	0	0	41	2.09	2.72	1.53	1.77	1.30	0.73
2341-2344	2342.5	Biotitic gneiss	0	0	0	0	100	0	0	39	1.64	3.16	1.70	1.85	1.92	1.04
2344-2347	2345.5	Biotitic gneiss	0	0	0	0	100	0	0	44	1.79	2.91	1.85	1.58	1.63	1.03
2347-2350	2348.5	Biotitic gneiss	0	0	0	0	100	0	0	39	1.40	2.67	1.73	1.54	1.90	1.23
2350-2353	2351.5	Biotitic gneiss	0	0	0	0	100	0	0	41	1.59	3.08	1.85	1.66	1.94	1.17
2353-2356	2354.5	Biotitic gneiss	0	0	0	0	100	0	0	45	2.03	4.05	1.29	3.13	2.00	0.64
2356-2359	2357.5	Biotitic gneiss	0	0	0	0	100	0	0	48	2.18	3.53	1.54	2.28	1.62	0.71
2359-2362	2360.5	Biotitic gneiss	0	0	0	0	100	0	0	50	1.99	2.86	2.22	1.29	1.43	1.11
2362-2365	2363.5	Biotitic gneiss	0	0	0	0	100	0	0	35	1.34	2.91	1.43	2.04	2.18	1.07
2365-2368	2366.5	Biotitic gneiss	0	0	0	0	100	0	0	41	1.69	2.64	1.66	1.59	1.57	0.99
2368-2371	2369.5	Biotitic gneiss	0	0	0	0	100	0	0	41	1.81	2.20	2.20	1.00	1.22	1.22
2371-2374	2372.5	Biotitic gneiss	0	0	0	0	100	0	0	35	1.61	2.44	1.52	1.61	1.52	0.94
2374-2377	2375.5	Biotitic gneiss	0	0	0	0	100	0	0	48	1.76	5.25	1.36	3.85	2.99	0.78
2377-2381	2379	Biotitic gneiss	0	0	0	0	100	0	0	36	1.67	2.36	1.49	1.58	1.41	0.89

B

Mbuo	1561-1564	1562.5	Claystone	80	9	9	19	0	1.8	0.7	96	1.83	11.44	2.38	4.81	6.3	1.30
	1564-1567	1565.5	Claystone	70	5	25	30	0	1.3	0.6	83	2.51	15.27	1.17	13.02	6.1	0.47
	1567-1570	1568.5	Claystone	78	3	19	22	0	1.1	0.7	104	2.64	18.43	1.82	10.13	7.0	0.69
	1570-1573	1571.5	Claystone	81	1	17	18	0	1.0	0.9	85	2.08	14.65	2.56	5.72	7.0	1.23
	1579-1582	1580.5	∞ Claystone	71	14	15	29	0	2.0	0.3	66	1.76	14.53	2.00	7.26	8.3	1.14
	1627-1630	1628.5	∞ Claystone	82	6	11	18	0	1.3	0.6	105	2.86	17.93	2.34	7.67	6.3	0.82
	1640-1643	1641.5	∞ Claystone	80	15	5	20	0	2.3	0.5	83	1.94	12.74	2.40	5.31	6.6	1.24
	1652-1655	1653.5	Claystone	82	6	12	17	0	1.3	0.7	102	2.90	19.90	1.99	9.98	6.9	0.69
	1658-1651	1659.5	Claystone	78	4	17	21	0	1.7	1.2	99	2.69	18.65	1.94	9.61	6.9	0.72
	1679-1682	1680.5	∞ Claystone	74	5	21	26	0	1.5	0.8	74	1.89	12.56	1.70	7.40	6.6	0.90
	1695-1698	1696.5	∞ Claystone	79	5	15	20	0	1.2	0.6	73	1.82	15.49	1.23	12.61	8.5	0.68
	1707-1710	1708.5	∞ Claystone	78	13	9	22	0	2.1	0.5	91	1.97	14.90	2.13	7.00	7.6	1.08
	1722-1725	1723.5	Claystone	84	3	13	16	0	0.8	0.5	94	2.33	15.96	2.44	6.53	6.9	1.05
	1731-1734	1732.5	∞ Claystone	84	3	13	16	0	0.8	0.4	86	2.09	17.86	1.25	14.24	8.6	0.60
	1737-1740	1738.5	∞ Claystone	83	5	11	17	0	1.3	0.7	87	2.03	16.92	1.61	10.48	8.3	0.80
	1747-1750	1748.5	Claystone	87	3	10	13	0	1.1	0.7	88	1.77	14.68	2.43	6.04	8.3	1.37
	1783-1786	1784.5	Claystone	83	9	8	17	0	1.6	0.6	96	2.47	15.18	3.26	4.65	6.1	1.32
	1792-1795	1793.5	Claystone	80	11	8	19	0	1.9	0.7	97	2.20	16.47	2.66	6.20	7.5	1.21
	1813-1816	1814.5	Claystone	81	4	15	19	0	1.3	0.8	106	2.49	16.25	3.51	4.62	6.5	1.41
	1823-1826	1824.5	Claystone	80	8	11	19	0	1.6	0.7	99	2.39	20.05	1.83	10.98	8.4	0.76
	1826-1829	1827.5	Claystone	82	6	11	18	0	1.5	0.7	98	2.58	18.96	2.18	8.71	7.3	0.84

	1829-1832	1830.5	Claystone	80	8	11	19	0	1.6	0.7	112	2.89	18.83	3.57	5.28	6.5	1.23
	1835-1838	1836.5	Claystone	85	4	11	15	0	1.0	0.6	104	2.66	18.74	2.51	7.46	7.1	0.95
	1838-1841	1839.5	Claystone	84	1	14	15	0	0.9	0.7	93	2.46	17.12	2.00	8.57	7.0	0.81
	1841-1844	1842.5	Claystone	78	3	19	22	0	0.9	0.5	88	2.13	16.88	1.81	9.31	7.9	0.85
	1856-1859	1857.5	Claystone	81	5	13	18	0	1.3	0.7	107	2.69	18.73	2.15	8.72	7.0	0.80
	1859-1862	1860.5	[∞] Claystone	84	0	15	15	0	0.8	0.8	108	3.05	17.59	2.97	5.92	5.8	0.98
	1865-1868	1866.5	[#] Argillaceous siltstone	79	6	14	20	0	1.8	1.0	74	1.46	12.18	2.32	5.24	8.4	1.60
	1875-1878	1876.5	Claystone	83	4	12	16	0	1.4	0.9	100	2.49	17.82	2.65	6.73	7.2	1.06
	1887-1890	1888.5	[#] Argillaceous siltstone	73	11	15	26	0	2.3	1.0	125	3.41	18.48	4.00	4.61	5.4	1.17
	1892-1895	1893.5	[#] Argillaceous siltstone	71	3	26	29	0	0.8	0.4	48	1.02	10.77	0.55	19.63	10.6	0.54
	1902-1905	1903.5	[¢] Sandstone	79	2	18	20	0	0.8	0.6	61	1.62	11.70	0.85	13.82	7.2	0.52
Basement	1905-1908	1906.5	Biotitic gneiss	0	0	0	0	100	0	0	62	1.90	10.08	1.23	8.22	5.3	0.65
	1908-1911	1909.5	Biotitic gneiss	0	0	0	0	100	0	0	61	1.17	18.63	1.00	18.63	15.9	0.85
	1911-1914	1912.5	Biotitic gneiss	0	0	0	0	100	0	0	49	0.97	7.99	1.75	4.58	8.2	1.80
	1914-1917	1915.5	Biotitic gneiss	0	0	0	0	100	0	0	49	1.63	7.09	1.45	4.91	4.3	0.89
	1917-1920	1918.5	Biotitic gneiss	0	0	0	0	100	0	0	55	1.69	6.89	1.61	4.29	4.1	0.95
	1920-1923	1921.5	Biotitic gneiss	0	0	0	0	100	0	0	44	1.37	4.04	1.69	2.39	2.9	1.23
	1923-1926	1924.5	Biotitic gneiss	0	0	0	0	100	0	0	50	1.51	7.53	0.99	7.58	5.0	0.66
	1926-1929	1927.5	Biotitic gneiss	0	0	0	0	100	0	0	59	1.84	10.36	0.85	12.17	5.6	0.46
	1929-1932	1930.5	Biotitic gneiss	0	0	0	0	100	0	0	49	1.50	7.72	0.93	8.26	5.2	0.62
	1932-1935	1933.5	Biotitic gneiss	0	0	0	0	100	0	0	55	1.67	7.64	1.29	5.92	4.6	0.77
	1936-1939	1937.5	Biotitic gneiss	0	0	0	0	100	0	0	69	1.89	10.63	0.97	10.91	5.6	0.52

1939-1942	1940.5	Biotitic gneiss	0	0	0	0	100	0	0	53	1.68	7.24	1.38	5.24	4.3	0.82
1942-1945	1943.5	Biotitic gneiss	0	0	0	0	100	0	0	65	2.17	11.11	0.70	15.80	5.1	0.32
1945-1948	1946.5	Biotitic gneiss	0	0	0	0	100	0	0	55	1.98	9.06	0.95	9.57	4.6	0.48
1948-1951	1949.5	Biotitic gneiss	0	0	0	0	100	0	0	71	2.16	12.71	1.12	11.36	5.9	0.52
1951-1954	1952.5	Biotitic gneiss	0	0	0	0	100	0	0	57	1.67	9.27	0.87	10.72	5.6	0.52
1954-1957	1955.5	Biotitic gneiss	0	0	0	0	100	0	0	61	2.11	8.54	1.17	7.32	4.0	0.55
1957-1960	1958.5	Biotitic gneiss	0	0	0	0	100	0	0	79	2.46	16.92	0.85	19.85	6.9	0.35
1960-1963	1961.5	Biotitic gneiss	0	0	0	0	100	0	0	115	4.67	12.28	3.17	3.88	2.6	0.68
1963-1966	1964.5	Biotitic gneiss	0	0	0	0	100	0	0	85	3.21	9.48	2.40	3.95	3.0	0.75
1966-1969	1967.5	Biotitic gneiss	0	0	0	0	100	0	0	73	1.65	10.77	1.85	5.81	6.5	1.12
1969-1972	1970.5	Biotitic gneiss	0	0	0	0	100	0	0	51	1.63	5.83	2.03	2.87	3.6	1.25
1972-1975	1973.5	Biotitic gneiss	0	0	0	0	100	0	0	66	1.66	10.82	1.62	6.66	6.5	0.98
1975-1978	1976.5	Biotitic gneiss	0	0	0	0	100	0	0	53	1.31	8.96	1.78	5.04	6.8	1.35
1978-1981	1979.5	Biotitic gneiss	0	0	0	0	100	0	0	53	1.53	8.12	1.47	5.50	5.3	0.96
1981-1984	1982.5	Biotitic gneiss	0	0	0	0	100	0	0	56	1.63	6.05	2.07	2.93	3.7	1.27
1984-1987	1985.5	Biotitic gneiss	0	0	0	0	100	0	0	86	1.84	14.27	2.57	5.56	7.7	1.39
1987-1990	1988.5	Biotitic gneiss	0	0	0	0	100	0	0	97	2.37	18.19	2.68	6.78	7.7	1.13
1990-1993	1991.5	Biotitic gneiss	0	0	0	0	100	0	0	66	1.97	10.90	2.37	4.61	5.5	1.20
1993-1996	1994.5	Biotitic gneiss	0	0	0	0	100	0	0	124	1.57	8.26	0.56	14.73	5.3	0.36

Supplementary Table 2.2: Sediment Composition & SGR data in the lower Jurassic sediments in the Mandawa Basin: A (Mita Gamma-1 well), B (East Lika-1 well). *

Sandy dolomite with thinly claystone interbeds, # Argillaceous siltstone and claystone interbeds, \$ Siltstone, sandstone, shales interbeds, € Silty mudstone, dolomite interbeds, £ Shale argillaceous sandstone interbeds, Φ Shale, dolomite, thin salt interbeds, ⌘ Shales halite interbeds, ¢ Halite, shales interbeds, θ Shale, thin

argillaceous sandstone, dolomite interbeds, $\varnothing$ Argillaceous shale sandstone interbeds, Δ Shale, dolomite interbeds, ¢ Argillaceous sandstone, claystone interbeds, $\diamond$
 Sandstone with thinly shale interbeds, ∞ Claystone, argillaceous siltstone interbeds, ¥ Claystone, sandy siltstone interbeds, § Claystone, sandstone interbeds, φ
 Sandy siltstone, claystone interbeds, g Claystone, sandstone interbeds.

Chapter Three

Formation	Depth Interval (m)	Medium Depth (m)	Lithology	Clastics %	Carbonate %	Salt %	Carbonate plus salt %	TC (wt%)	TOC (wt%)	TOC_C btf (wt%)	TOC_Sf (wt%)	TOC_Cbt+Sf (wt%)	GR (Al)	DT (μs/ft)	ILD (Ohm-m)	RHOB (g/cc)	PEF (b/e)	GR_K (wt%)	GR_Th (ppm)	GR_U (ppm)	Th/U
Mihambia	1996-1999	1997.5	[§] Siltstone	76	9	14	22	2.13	1.08	1.18	1.25	1.39	72	63	6.5	2.62	3.45	1.70	10.72	2.18	4.91
	1999-2002	2000.5	[§] Siltstone	85	14	1	15	2.22	0.58	0.68	0.59	0.69	70	64	9.1	2.64	3.45	1.94	11.51	1.57	7.34
	2006-2009	2007.5	[§] Siltstone	74	18	7	25	3.11	0.97	1.18	1.04	1.29	80	62	11.5	2.63	3.26	1.95	15.69	0.88	17.87
	2012-2015	2013.5	[§] Siltstone	80	15	5	20	2.30	0.50	0.58	0.52	0.62	73	63	14.5	2.62	3.19	1.66	10.43	2.13	4.91
	2021-2024	2022.5	[€] Silty mudstone, dolomite	47	29	21	50	6.24	2.71	3.84	3.42	5.43	51	58	13.1	2.61	3.84	1.03	5.84	2.10	2.79
	2039-2042	2040.5	[£] Shale	57	17	25	42	3.43	1.44	1.73	1.92	2.47	62	59	168.8	2.80	4.15	1.72	9.75	1.62	6.04
	2048-2051	2049.5	* Sandy dolomite	65	13	21	34	2.86	1.29	1.49	1.63	1.96	47	59	20.3	2.71	3.04	1.22	9.02	0.92	9.84
	2054-2057	2055.5	* Sandy dolomite	55	25	20	44	3.92	0.94	1.25	1.17	1.69	42	55	34.6	2.78	3.03	0.93	7.17	1.13	6.35
	2070-2073	2063.5	Shale	50	26	23	49	3.89	0.77	1.05	1.01	1.52	45	56	83.5	2.83	4.11	0.97	7.01	1.28	5.47
	2079-2082	2080.5	[®] Shale, dolomite	52	21	24	45	5.33	2.82	3.56	3.71	5.12	54	73	96.9	2.80	3.92	1.50	9.58	1.22	7.86
Nondwa			[®] Shales, halite interbeds																		
	2088-2091	2089.5		46	17	35	52	4.27	2.19	2.65	3.37	4.60	44	70	2384	2.24	4.75	1.22	7.15	1.48	4.84
	2106-2109	2107.5	[©] Halite	9	2	88	90	0.58	0.28	0.29	2.32	2.90	16	67	10592	2.41	2.79	0.26	1.35	0.59	2.27
Upper Mbuo	2112-2115	2113.5	[©] Halite	7	2	91	93	0.41	0.18	0.19	2.01	2.56	21	67	32157	2.10	2.61	0.42	2.82	0.33	8.48
	2124-2127	2125.5	[®] Shales	43	34	20	54	7.53	3.45	5.23	4.29	7.43	55	67	38197	2.12	4.18	1.39	7.37	1.93	3.81
	2130-2133	2131.5	[®] Argillaceous shale	65	12	21	33	2.98	1.51	1.72	1.91	2.25	73	69	2142	2.68	4.59	1.62	10.21	2.78	3.68
	2143-2146	2144.5	^Δ Shales	65	17	17	34	2.66	0.60	0.72	0.72	0.90	91	76	255	2.43	4.77	2.40	18.22	0.76	23.91
	2146-2149	2147.5	Claystone	65	33	1	34	5.03	1.03	1.54	1.03	1.56	100	74	88.5	2.64	4.70	2.93	18.03	1.83	9.85
	2155-2158	2156.5	^Δ Shales	61	13	25	38	2.59	1.03	1.18	1.37	1.65	66	64	10.1	2.59	3.98	1.83	10.95	1.43	7.64
	2167-2170	2168.5	[€] Argillaceous sandstone	59	23	18	41	3.29	0.54	0.70	0.65	0.91	60	61	22.2	2.72	4.41	1.71	9.56	1.20	7.99

Lower Mbuo	2176-2179	2177.5	◊ Sandstone	66	16	17	34	2.63	0.65	0.78	0.79	0.99	54	61	43.3	2.62	3.50	1.81	8.90	1.29	6.87
	2195-2198	2196.5	◊ Sandstone	79	6	14	20	1.31	0.62	0.65	0.72	0.77	91	77	7.3	2.44	3.44	3.26	15.22	1.96	7.78
	2204-2207	2205.5	∞ Claystone	64	0	34	34	1.77	1.97	1.97	3.00	3.00	84	79	6.1	2.45	4.86	2.83	14.31	1.65	8.67
	2207-2210	2208.5	¥ Claystones	62	9	29	38	1.76	0.70	0.77	0.98	1.12	76	72	6.6	2.44	4.34	2.72	10.81	2.23	4.85
	2216-2219	2217.5	Claystone	63	11	25	36	1.99	0.63	0.71	0.84	0.99	111	80	8.0	2.54	4.62	3.33	17.31	2.98	5.81
			#										1								
	2219-2222	2220.5	Argillaceous siltstone	64	20	15	35	3.12	0.71	0.89	0.83	1.09	79	68	15.4	2.61	3.92	2.55	11.82	2.26	5.24
	2222-2225	2223.5	Clay evaporite	22	8	70	78	1.32	0.38	0.41	1.24	1.67	93	71	9.5	2.62	2.50	3.59	16.12	2.06	7.82
	2225-2228	2226.5	Clay evaporite	30	6	64	70	1.13	0.37	0.40	1.02	1.24	120	82	5.9	2.48	3.27	3.89	18.54	2.50	7.41
	2231-2234	2232.5	Clay evaporite	31	23	45	69	3.38	0.58	0.76	1.07	1.87	119	76	8.9	2.57	2.96	4.13	18.79	1.85	10.15
	2234-2237	2235.5	§ Claystone	67	5	27	32	1.49	0.92	0.97	1.27	1.36	80	67	6.5	2.52	4.18	2.45	9.54	2.18	4.39
			#																		
	2240-2243	2241.5	Argillaceous siltstone	75	6	18	24	1.22	0.48	0.51	0.58	0.63	83	71	7.3	2.51	3.16	3.12	12.90	0.90	14.40
	2246-2249	2247.5	φ Sandy siltstone	67	4	28	32	0.98	0.46	0.48	0.64	0.68	84	70	4.6	2.45	3.38	2.60	9.57	1.70	5.63
			€																		
	2277-2280	2278.5	Argillaceous sandstone	47	2	50	52	0.82	0.58	0.59	1.16	1.21	59	67	3.1	2.44	3.62	2.07	7.03	1.35	5.19
	2282-2285	2283.5	Argillaceous sandstone	46	0	53	53	0.64	0.65	0.65	1.39	1.39	71	67	3.6	2.43	3.54	2.52	9.81	1.46	6.73
	2298-2301	2299.5	‡ Claystone , sandstone interbeds	41	3	55	58	0.77	0.38	0.39	0.85	0.92	93	70	11.5	2.57	3.75	2.85	17.20	1.71	10.05
	2301-2304	2302.5	‡ Claystone , sandstone interbeds	69	3	27	30	0.81	0.40	0.41	0.54	0.57	78	76	3.7	2.45	4.40	2.39	11.82	1.94	6.08

Supplementary Table 3.1: TOC, sediment composition and well log discrete data in the Lower Jurassic sediments from the Mita Gamma-1 well, used in this study

Number of samples		Total organic carbon (wt %)				Sediment composition (wt%)				Well logs data							
		TOC_raw (wt%)	TOC_Cbtf (wt%)	TOC_Sf (wt%)	TOC_Cbtf+Sf (wt%)	Clastics %	Carbonate %	Salt %	Carbonate plus salt %	GR (API)	DT (μs/ft)	ILD (Ohm-m)	RHOB (g/cc)	PEF (b/e)	GR_K (%)	GR_Th (ppm)	GR_U (ppm)
n=10	Minimum	0.50	0.58	0.52	0.62	47.2	8.7	1.3	14.9	41.6	55.0	6.5	2.6	3.0	0.9	5.8	0.9
	Maximum	2.82	3.84	3.71	5.43	84.5	29.3	25.1	50.1	79.9	73.3	169	2.8	4.1	1.9	15.7	2.2
	Average	1.31	1.65	1.63	2.22	64.1	18.6	16.0	34.6	59	61	46	3	4	1	10	2
n=3	Minimum	0.18	0.19	2.01	2.56	6.9	1.9	35.0	52.3	15.7	66.8	2384	2.1	2.6	0.3	1.4	0.3
	Maximum	2.19	2.65	3.37	4.60	45.5	17.3	90.9	92.9	44.1	70.0	32157	2.4	4.8	1.2	7.2	1.5
	Average	0.89	1.04	2.57	3.35	20.6	7.2	71.3	78.5	26.9	67.9	15044	2.2	3.4	0.6	3.8	0.8
n=6	Minimum	0.54	0.70	0.65	0.90	43.0	12.2	0.7	33.0	54.8	60.8	10	2.1	4.0	1.4	7.4	0.8
	Maximum	3.45	5.23	4.29	7.43	65.5	34.0	25.0	53.6	99.8	76.0	38197	2.7	4.8	2.9	18.2	2.8
	Average	1.36	1.85	1.66	2.45	59.7	22.1	16.8	38.9	74.0	68.5	6786	2.5	4.4	2.0	12.4	1.7
n=16	Minimum	0.37	0.39	0.54	0.57	22.1	-1.7	14.2	20.0	53.9	61.4	3.1	2.4	2.5	1.8	7.0	0.9
	Maximum	1.97	1.94	3.08	3.00	79.4	23.2	69.6	77.5	119.8	82.5	43	2.6	4.9	4.1	18.8	3.0
	Average	0.65	0.71	1.06	1.22	55.9	7.6	35.9	43.5	85.9	72.2	9.5	2.5	3.7	2.9	13.1	3.0
n=35	Minimum	0.41	0.18	0.19	0.52	0.6	0.0	0.7	14.9	15.7	55.0	3.1	2.1	2.5	0.3	1.4	0.3
	Maximum	3.45	5.23	4.29	7.43	84.5	34.0	90.9	92.9	120	82	38197	3	5	4	19	3
	Average	0.98	1.20	1.46	1.90	55.9	13.2	30.0	43.2	71.2	68.0	2470	2.5	3.8	2.1	11.2	1.6

Supplementary Table 3.2: Minimum, maximum and average total organic carbon (TOC), sediment composition as well as well log data in the Lower Jurassic sediments from the Mita Gamma-1 well.

Chapter Four

MITA GAMMA WELL-1

Formation	Number of samples			TOC [%wt.]		S1 [mgHC/gRock]		S2 [mgHC/g]		HI [mgHC/gTOC]		OI		Tmax [°C]		PI	
	Standard samples	Organo facies		Standard samples	Organo facies	Standard samples	Organo facies	Standard samples	Organo facies	Standard samples	Organo facies	Standard samples	Organo facies	Standard samples	Organo facies	Standard samples	Organo facies
Mihambia	n=10	n=2	Minimum	0.6	2.31	0.04	0.26	0.18	2.51	29	108	25	25	437	437	0.09	0.09
			Maximum	2.82	2.82	0.44	0.44	4.91	4.91	174	174	165	165	447	444	0.19	0.1
			Average	1.28	2.6	0.14	0.35	1.33	3.71	74	141	107	95	444	441	0.12	0.10
Nondwa	n=2	n=2	Minimum	1.58	1.58	0.26	0.26	2.04	2.04	135	135	78	78	434	434	0.12	0.12
			Maximum	3.08	3.08	0.88	0.88	6.07	6.07	196	196	246	246	441	441	0.13	0.13
			Average	2.33	2.33	0.57	0.57	4.06	4.06	166	166	162	162	438	438	0.13	0.13
Upper Mbuo	n=2	n=1	Minimum	0.94	1.26	0.06	0.19	0.42	1.40	44	111	148	209	436	443	0.12	0.14
			Maximum	1.26	1.26	0.19	0.19	1.40	1.40	111	111	209	209	443	443	0.14	0.14
			Average	1.1	1.26	0.13	0.19	0.91	1.40	78	111	179	209	440	443	0.13	0.14
Lower Mbuo	n=7	n=1	Minimum	0.38	1.78	0.03	1.74	0.13	3.20	28	179	155	179	439	n.c.*	0.16	0.35
			Maximum	1.78	1.78	1.74	1.74	3.20	3.20	179	179	283	179	441		0.35	0.35
			Average	0.76	1.78	0.29	1.74	0.65	3.20	59	179	218	179	440		0.24	0.35

EASTLIKA-1 WELL

Mihambia	n=11	No sample	Minimum	0.31	n.c.*	0	n.c.*	0.01	n.c.*	4	n.c.*	56	n.c.*	429	n.c.*	0.04	n.c.*
			Maximum	1.09		0.04		0.64		58		186		477		0.67	
			Average	0.42		0.02		0.01		17		110		445		0.37	
Nondwa	n=10	n=2	Minimum	0.48	1.52	0	0.03	0.13	1.07	21	70	33	43	430	431	0.03	0.04
			Maximum	1.67	1.62	0.04	0.04	1.23	1.24	73	73	89	46	436	431	0.14	0.04
			Average	1.17	1.6	0.03	0.04	0.6	1.15	46	72	56	45	432	431	0.07	0.04
Upper Mbuo	n=6	No sample	Minimum	0.52	n.c.*	0.01	n.c.*	0.05	n.c.*	7	n.c.*	74	n.c.*	429	n.c.*	0.1	n.c.*
			Maximum	0.92		0.03		0.27		25		152		437		0.34	
			Average	0.68		0.02		0.15		18		98		433		0.2	
Lower Mbuo	n=4		Minimum	0.39	n.c.*	0.01	n.c.*	0.07	n.c.*	14	n.c.*	70	n.c.*	432	n.c.*	0.19	n.c.*
			Maximum	0.74		0.04		0.16		26		166		435		0.46	
			Average	0.6		0.02		0.12		20		109		434		0.31	

MBUO-1 WELL

Mihambia	n=8	n=2	Minimum	0.78	2.2	0	0.03	0.11	0.94	13	58	39	42	433	433	0.03	0.03
			Maximum	2.21	1.6	0.03	0.03	1.44	1.44	65	65	80	55	438	433	0.14	0.04

			Average	1.2	1.9	0.02	0.03	0.54	1.19	37	62	55	49	434	433	0.08	0.03
Nondwa	n=4	n=1	Minimum	0.7	1.65	0.01	0.11	0.11	4.49	12	220	59	80	428	439	0.03	0.03
			Maximum	2.04	1.65	0.11	0.11	4.49	4.49	220	220	112	80	439	439	0.27	0.03
			Average	1.22	1.65	0.04	0.11	1.25	4.49	68	220	79	80	433	439	0.12	0.03
Upper Mbuo	n=6	n=1	Minimum	0.8	1.88	0.04	0.11	0.3	1.69	25	89	47	49	430	441	0.06	0.06
			Maximum	1.88	1.88	0.11	0.11	1.69	1.69	89	89	110	49	441	441	0.23	0.06
			Average	1.24	1.88	0.06	0.11	0.61	1.69	45	89	69	49	433	441	0.13	0.06
Lower Mbuo	n=3	No sample	Minimum	0.58	n.c.*	0.06	n.c.*	0.21	n.c.*	30	n.c.*	214	n.c.*	442	n.c.*	0.19	n.c.*
			Maximum	0.71		0.07		0.34		52		285		444		0.25	
			Average	0.65		0.07		0.26		39		251		443		0.22	

MBUO CORE																	
Lower Mbuo	n=17	n=4	Minimum	0.58	3.13	0	0.29	0.02	2.21	2	69	10	13	442	442	0.06	0.06
			Maximum	4.83	4.83	0.62	0.62	8.62	8.69	233	233	41	17	468	446	0.38	0.08
			Average	2.42	4.04	0.15	0.50	1.72	8	44	162	22	14	449	445	0.21	0.08
NONDWA OUTCROP																	
Nondwa	n=8	n=3	Minimum	1.7	2.53	0.01	0.02	0.56	1.29	33	49	8	9	429	436	0.01	0.01
			Maximum	4.62	3.79	0.52	0.03	6.58	6.58	142	142	121	121	437	437	0.07	0.01
			Average	2.64	3.65	0.08	0.02	1.71	3.24	54	81	27	47	434	437	0.02	0.01

Supplementary Fig. 4: The average TOC and rock pyrolysis data for organofacies and standard samples in the Lower Jurassic sediment from Mita Gamma-1, Mbuo-1, and East Lika-1 wells and outcrop samples.