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Department of Earth Science and Engineering

Qatar Carbonates and Carbon Storage Research Centre

Palaeoenvironmental control on multi-scale vertical and
lateral heterogeneities within a Triassic - Jurassic carbonate
ramp system (Musandam Peninsula)

A thesis submitted in fulfilment of the requirements for the degree of
Doctor of Philosophy and the Diploma of Imperial College

By

Martin Robert Hönig

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Abstract

Sedimentologic heterogeneities in carbonate systems are caused by various processes during and after deposition and can have a profound influence on reservoir properties relevant for hydrocarbon recovery or carbon storage. This study is focussed on examining and quantifying sedimentologic heterogeneities on an equatorial, Triassic-Jurassic, Middle Eastern, epeiric carbonate ramp (Musandam Peninsula, UAE and Oman) at the regional scale (~ 45 km), the 'inter-well' scale (~ 1.9 km) and the metre- to tens of metres scale (~120 m). The stratigraphic ages of the examined strata were refined using a combination of the existing biostratigraphic framework and chemostratigraphic analysis (stable carbon and radiogenic strontium isotopes). Results suggest that the regional architecture and carbonate factory evolution across the Triassic-Jurassic boundary was mainly governed by relative sea level changes and not affected by ocean acidification in the equatorial region. Lithofacies types were deposited and are preserved as facies mosaics on the shallow-marine carbonate ramp, despite an overall 'layer-cake' type bedding pattern. Geostatistical analysis reveals that facies have a probability of 60 % of being laterally discontinuous over distances between 220 and 1100 m on the shallow-marine carbonate ramp. The comparison with other preserved facies mosaics as well as with modern analogous systems suggests that the large amount of lateral facies variability, also observed within individual beds at the metre-scale, is a result of the combination between locally heterogeneous carbonate production as well as of coeval mixing and re-sedimentation processes. The combination of laterally heterogeneous facies types and vertical exposure stages is likely the result of constant autocyclic mixing and re-shuffling of grains within the shallow-marine EOD and allocyclic changes in relative sea level. This study highlights the capability of neritic carbonate successions to archive palaeoceanographic changes in deep time as well as to preserve small-scale facies mosaics.

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1. Thesis introduction

Carbonate systems are influenced by a variety of processes. The complex interplay of different mechanisms can result in heterogeneous distribution of lithofacies types and geobodies or lithosomes (a uniform mass of rock lithographically different from the surrounding rocks) on ancient as well as modern carbonate ramps and platforms. These architectures can be preserved as heterogeneous carbonate strata in outcrops as well as in the subsurface. The study of these heterogeneities can lead to a better understanding of depositional processes that govern their distribution. Furthermore lithologic heterogeneities can have implications for reservoirs that host hydrocarbons or are potential storage sites for CO₂ sequestration, if for example lithofacies types template petrophysical properties or geological units template flow units. Heterogeneities in carbonate systems occur at a range of scales, from regional sequences to the pore-scale. Especially medium-scale heterogeneities in the subsurface are hard to detect with widely used imaging techniques. Lateral facies termination and merging can occur in between wells, and vertical stacking patterns thinner than ~ 20 m cannot be resolved using seismic data. Studying outcrops as analogues for these subsurface reservoirs allows the gathering of valuable information on dimensions and architectures that can be implemented into, or used to guide reservoir or seismic models and subsequently lead to improved recovery of hydrocarbons and effective sequestration of CO₂.

Additionally, shallow-marine carbonate systems are very sensitive to environmental changes and have the potential to archive a history of climatic patterns or global cycles. Palaeoenvironmental studies of the Mesozoic and older periods often rely exclusively on such carbonate archives as much of the deep-sea record has been subducted. Using these neritic carbonates as archive for global change does however present some pitfalls. Most commonly the sections include hiatuses, if the shallow-marine carbonates were frequently exposed, or sedimentation was halted due to a carbonate factory crisis, potentially induced by environmental changes. Early as well as late diagenetic processes pose a further difficulty when analysing these sedimentary archives, especially when measuring isotopic proxies.

If a meaningful palaeoenvironmental signal is extracted, much can however be learnt regarding carbon cycle perturbations and biotic responses to environmental changes from shallow-water carbonate archives. Additionally, some climate perturbations that occurred in past geologic periods potentially serve as analogues for modern-day climatic changes linked to greenhouse gas emissions.

This PhD project has three main objectives: (i) to characterise and quantify the architecture of Late Triassic to Middle Jurassic carbonate systems analogous to formations in the subsurface of Qatar, from the regional to the metre-scale; (ii) to refine the stratigraphic ages of the examined strata that were previously constrained through biostratigraphy, mainly by using carbon isotope and strontium isotope stratigraphy; and (iii) to understand processes that led to the observed architectures at a range of scales, including global palaeoenvironmental controls and local hydrodynamic regimes.

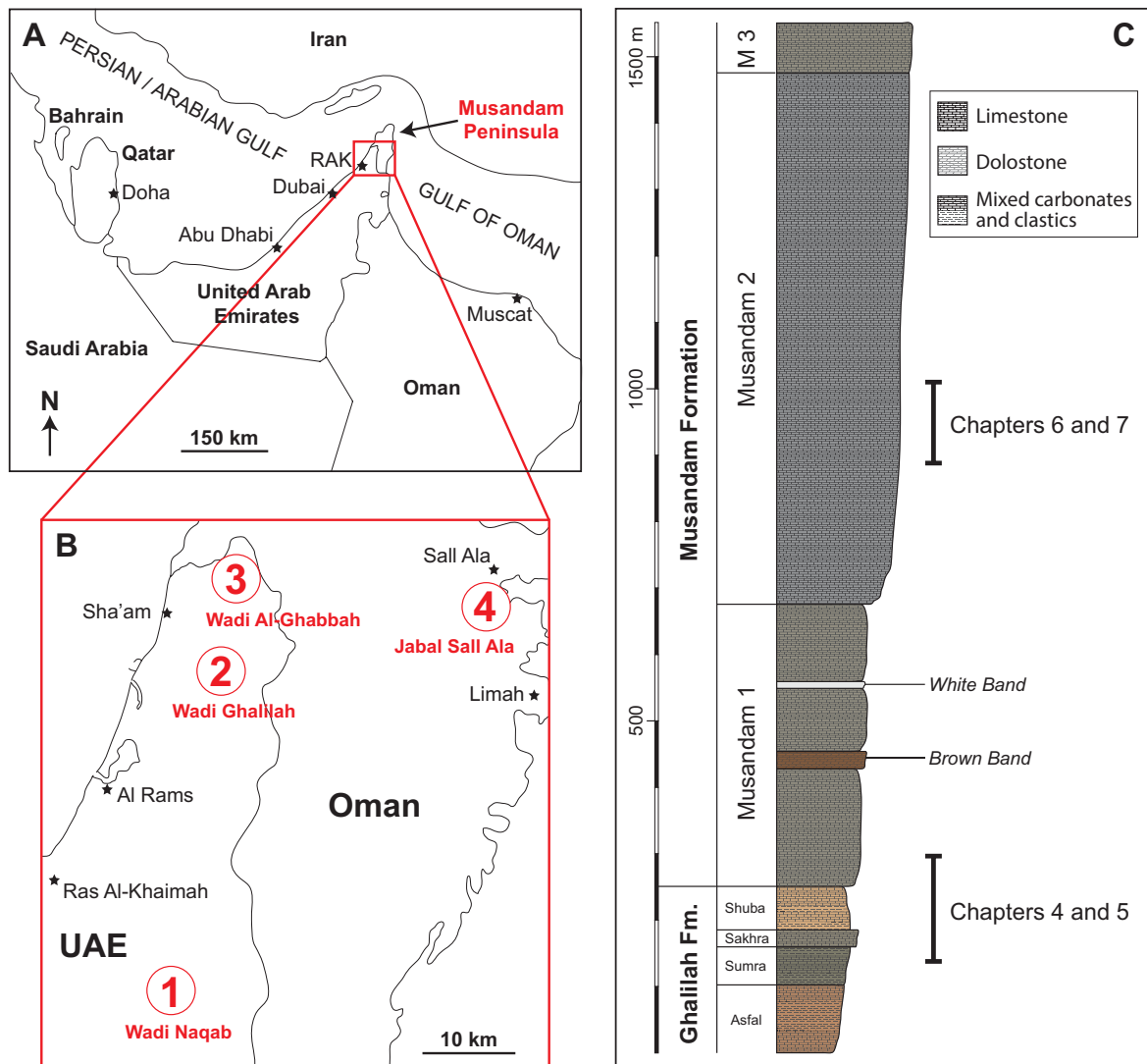


Fig. 1.1 – A, B) Overview maps of the study area, indicating the four specific study sites. **C)** Composite lithostratigraphic section of the Upper Ghailah and the Musandam Formations on the Musandam Peninsula. The lithostratigraphic positions of the main chapters 4 to 7 are indicated.

In order to accomplish the set objectives, a well-exposed series of outcrops on the Musandam Peninsula (northern United Arab Emirates and northern Sultanate of Oman) were studied (Fig. 1.1). The examined strata are suitable analogues for subsurface applications in Qatar (and other parts of the Middle East) in terms of stratigraphic age (Fig. 1.2), depositional environment, lithofacies types and vertical facies distribution.

Age [Ma]	Epoch		Formations					
			Musandam Peninsula	UAE (subsurface)	Qatar			
					On shore	Off shore		
~145	CRETACEOUS	Upper	Juwaiza/ Fukairi and Fika Ausaq	Aruma	Halul Laffan Mishrif			
		Lower	Wasia M/M Nahr Umr Thamama	Wasia M/M Nahr Umr Thamama	Ahmadi	Ahmadi Khatiyah		
	JURASSIC	Upper	Musandam	Sahtan	Darb Diyab Tuwaiq Dhruma	Hith		
						Qatar	Arab A-D	
		Darab				Jubailah		
		Diyab				Hanifa		
		Middle				M 2	Araej	U. Araej
								Uwainat
								L. Araej
		Lower				M 1	Marrat U. Minjur	Izhara
Marrat	Hamlah							
~201	TRIASSIC	Upper	Elphinst. Ghalilah Milaha	L. Minjur	Minjur	Gulailah		
		Middle	Ghail	Gulailah /Jilh	Jilh			
		Lower	Rus al Jibal Hagil	Sudair	Suwei		Sudair	
	PERMIAN	Upper	Bih	Khuff	Khuff			
		Middle	no outcrop	Gharif/ Hauschi	Gharif			
		Lower						

Fig. 1.2 – Stratigraphic column and associated formations of the Musandam Peninsula, the subsurface of the United Arab Emirates (M/M: Maaddud/ Mishrif Formations) and Qatar (modified from Searle et al., 2014 and references therein; incorporating information from Alsharhan and Nairn, 1994; de Matos, 1997; Maurer et al., 2008).

This thesis is organised in the following way:

Chapter 2 provides a concise literature review on carbonates as well as on the current understanding of the Triassic-Jurassic boundary and the associated mass extinction. In **chapter 3** the Musandam Peninsula is introduced as well as the different study sites. Additionally, information on the geological background of the Musandam Peninsula is given, including the structural evolution, main formations and studied lithostratigraphic units. **Chapter 4** is an adaptation of a journal article that is currently being revised for the journal *Gondwana Research* (Hönig, John and Manning , under revision, “*minor revisions*”) and focuses on the regional evolution of the Upper Triassic and Lower Jurassic sequences on the Musandam Peninsula and links to palaeoenvironmental changes. Cédric John assisted during fieldwork, discussed results with the first author and reviewed earlier versions of the manuscript. Christina Manning assisted with sample preparation for strontium isotope measurements and performed the strontium isotope measurements. **Chapter 5** focuses on the Triassic-Jurassic boundary at one specific study site and discusses the findings from the Musandam Peninsula in a global context with regards to the Triassic-Jurassic mass extinction and associated global environmental perturbations. The chapter is an adapted version of a manuscript that will be submitted to a journal (Hönig, John, Adatte and Heimhofer). Cédric John assisted during fieldwork, discussed results with the first author and reviewed an earlier version of the manuscript. Thierry Adatte performed the XRD analysis and Ulrich Heimhofer performed the organic carbon isotope analysis. **Chapter 6** contains an adapted version of a published journal article in the journal *Marine and Petroleum Geology* (Hönig and John, 2015) and deals with Middle Jurassic inter-well scale heterogeneities at one of the study sites. Cédric John assisted during fieldwork, discussed results with the first author and reviewed earlier versions of the manuscript. In **chapter 7** the same Middle Jurassic strata are examined at a smaller, individual bedset scale to analyse lithofacies changes in shallow-marine carbonates at the metre scale. This chapter has been submitted to the industry sponsors as a manuscript and is subsequently intended for submission to the journal *Sedimentology*. Cédric John assisted during fieldwork, discussed results with the first author and reviewed earlier versions of the manuscript. **Chapter 8** contains the main conclusions drawn from the project and a

paragraph on considerations on how to potentially build upon the findings of this project.

2. Scientific background

2.1 Carbonate ramp depositional systems and facies distribution

Carbonate depositional systems are different from siliciclastic-dominated systems in several ways. The most fundamental difference, while sediment accumulation in both systems is strongly controlled by physical processes, is the fact that most carbonate material is the product of biological (induced or controlled) precipitation (Schlager, 2005). This precipitation is in turn controlled by a variety of mechanisms, often intimately linked (Fig. 2.1). These include the biological evolution of biota as well as their ecological requirements such as temperature, trophic levels and salinity (Pomar et al., 2012). Pomar and Kendall (2007) identify that constantly evolving biology in a carbonate depositional system is identified as the fundamental difference between clastic and carbonate systems. Hydrodynamic changes affect both systems but carbonates will often react differently (Fig. 2.2).

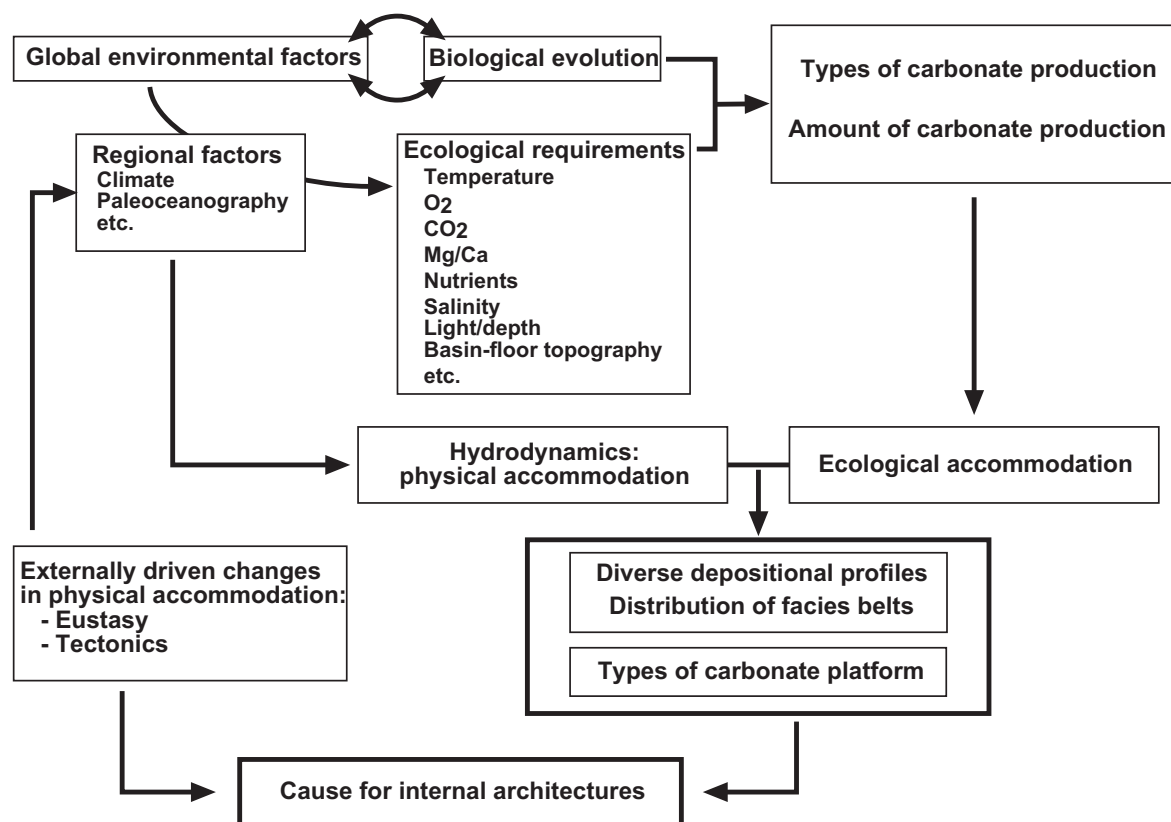


Fig. 2.1 - Flow diagram of the physical and biological relationships that control carbonate accumulation on carbonate platforms (reproduced from Pomar and Kendall, 2007).

The term ecological accommodation is used to illustrate that the accumulation of carbonates is strongly dependent on the capacity of organisms to produce and to accumulate sediments above the hydrodynamic threshold (Pomar and Kendall, 2007). Schlager (2005) summarises four main patterns, linked to principles of carbonate production as well as hydrodynamic processes, that are characteristic of carbonate systems: (i) carbonate factories commonly produce buildups that are elevated locally, this occurs at various scales; (ii) carbonate factories thrive in shallow water and tend to be flat-topped close to the sea level; (iii) carbonate systems are often bound together by stiff rims formed by framebuilders, carbonate production is high at the rims and lower on the interior part of the system (also referred to as the “bucket principle”) and (iv) carbonate slopes tend to be relatively steep, owed to early and strong lithification.

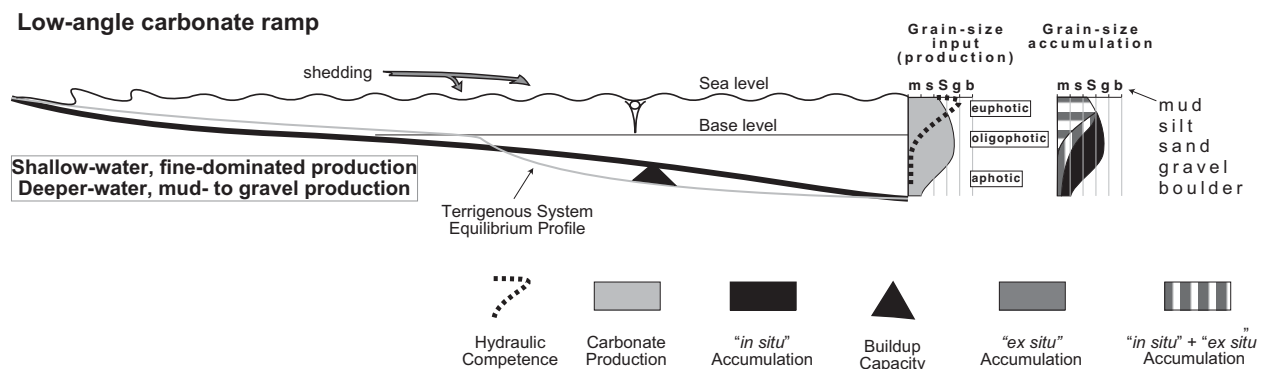


Fig. 2.2 – Depositional profile typical of low-angle carbonate ramps, dominated by physical accommodation (modified from Pomar and Kendall, 2007).

The term carbonate ramp was introduced by Ahr (1973) and defined as inclined platform extending basinward, but lacking a pronounced slope, as an alternative to a “shallow shelf” depositional model. This definition was subsequently extended by the differentiation between purely homoclinal and distally steepened ramps (Read, 1985). The ramp depositional model was subdivided into broad depositional zones, commonly present on carbonate ramps, ranging from the inner, to the mid and the outer ramp and ending in the basin (Fig. 2.3) (e.g. Burchette and Wright, 1992). Various refined depositional models have been proposed since to account for the differences in the distribution of facies belts such as the different positions of high-energy

lithofacies types or the presence/absence of protected lagoons (Schlager, 2005). A recent subdivision of the ramp depositional was put forward by Pierre et al. (2010) and differentiates between grainy, mainly ooid-dominated systems, and muddy, ooid-free ramp systems.

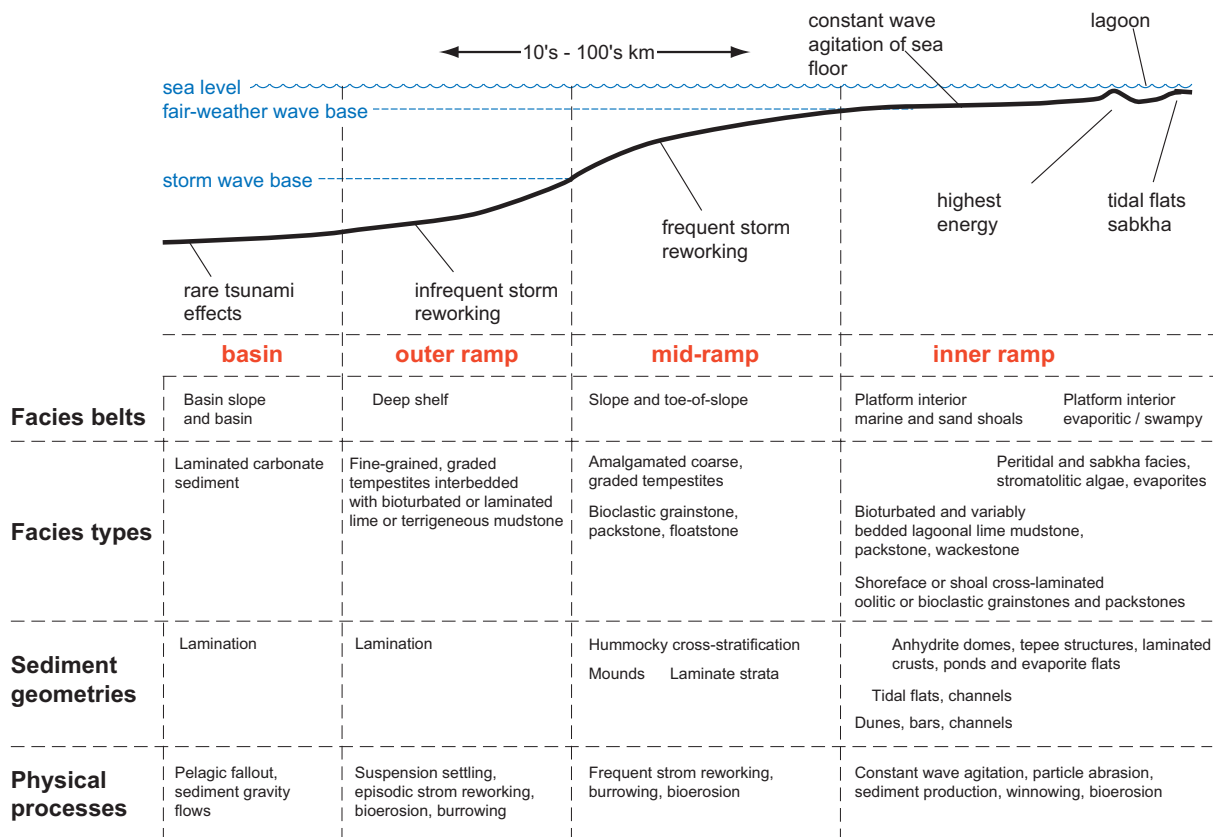


Fig. 2.3 – Carbonate ramp depositional model including main facies belts, facies types, sediment geometries and main physical processes. All the indicated boundaries are gradual (modified from Schlager, 2005; after Wilson 1975 and incorporating information from Burchette and Wright, 1992; James, 1997; Immenhauser, 2009).

The inner part of a carbonate ramp comprises a generally shallow-marine depositional environment, above the fair weather wave base (Wilson, 1975; Burchette and Wright, 1992; Schlager, 2005). Most of the facies heterogeneity encountered on carbonate ramps is commonly present on the inner, to mid-ramp. As the depositional profile reaches deeper water on the outer ramp towards the basin, sediments become more homogeneous and pelagic mud, mainly exported from the neritic realm, is the dominant facies type (Wilson, 1975; Schlager, 2005). The inner, shallow-marine parts of carbonate

ramps contain three broad facies belts, ranging from the tidal or supratidal complex, through sublittoral sandsheets to fringing banks or skeletal carbonates (Read, 1985). Especially the outer two facies belts, excluding the supratidal- to tidal environment, commonly contain a large variety of facies types as well as sedimentary bodies composed of different grains and formed by hydrodynamic processes as well as carbonate growth patterns (Wilson, 1975; Read, 1985; Schlager, 2005). A variety of different sedimentary bodies, also termed geobodies or lithosomes, can be encountered, especially in the shallow-marine depositional environment of ramps (as well as platforms), including amongst others mounds, shoals, tidal and channel bars and wedges (e.g. Burchette and Wright, 1992; Jung and Aigner, 2012).

Although lithofacies types appear to be ordered as belts at the ramp-scale, LaPorte (1967) was amongst the first to recognise that facies types within the shallow-marine depositional environment can lack the organisation as facies belts but rather appear as mosaics. An increasing amount of outcrop studies is emerging that examine facies distribution on ancient carbonate systems, partly aiming at reducing uncertainties in analogous subsurface reservoirs. Shallow-marine zones on carbonate ramps form economically important hydrocarbon reservoirs throughout the world, for example in the Jurassic Smackover Formation in the USA or the Jurassic Arab Formation in the Middle East (e.g. Read, 1985). Some outcrop based studies suggest that ancient, shallow-marine carbonates are preserved in a 'layer-cake' type architecture where pinch out and facies heterogeneities are rare (e.g. Borkhataria et al., 2005; Zeller et al., 2011; Pratt et al., 2012; Bendias et al., 2013), whereas in other examples large amounts of facies heterogeneity is documented, even on a scale of ~100 to 1000 m (sometimes termed "inter-well scale") and especially within shallow water depths. These studies were performed in a variety of different regions and periods, for example in the Mesozoic strata of the eastern Tethys (Triassic: Obermaier et al., 2015; Cretaceous: Immenhauser et al., 2004; Homewood et al., 2008; Sena and John, 2013), the Jurassic of Morocco (Verwer et al., 2009; Amour et al., 2011; Christ et al., 2011) as well as in the western Tethys (e.g. Triassic: Palermo et al., 2010; Jurassic: Alnazghah et al., 2013; San Miguel et al., 2013; Pomar et al., 2015; Cretaceous: Brandano and Loche, 2014;

Miocene: Brandano, 2003; Tomás et al., 2010), the Devonian of Australia (Frost, 2007) and the Permian of the US (Kerans et al., 1994).

Some modern analogue studies have yielded comparable results regarding the irregular distribution of facies types (e.g. Wilkinson et al., 1999; Rankey and Reeder, 2011; Harris et al., 2015; Purkis et al., 2015a; Purkis and Harris, 2016) although it is debatable on how applicable these are to ancient systems regarding overall system dimension and carbonate factories. Outcrop studies with closer-spaced sections (lateral distance between logs approximately 2.5 - 35 m) are rare and only a few published examples in carbonate strata exist, one from the Permian (Kerans et al., 1994), two from the Triassic (Kostic and Aigner, 2004; Obermaier et al., 2015), two from the Jurassic (Samankassou et al., 2003; Bádenas et al., 2010 – two logs only) and one from the Cretaceous (Adams et al., 2011), of which most show closely spaced facies heterogeneities within shoal bodies with inclined, and wedge-shaped deposits. Facies distribution at a similarly small scale from a modern analogue, the Florida Keys, reveals a relatively ordered pattern with facies closely linked to relative sea level. This relationship is however not necessarily applicable to facies in the subsurface or at outcrop (Bosence et al., 1985; Bosence, 2008) and the applicability to ancient systems again remains debatable. The modern Arabian Gulf coast is one place from where carbonate facies heterogeneity on the small scale of 10's to 100 m has been reported (Purkis et al., 2005; Kendall and Alsharhan, 2011).

2.2 The Triassic-Jurassic boundary

The Triassic-Jurassic (T-J) boundary was marked by one of the “big five” mass extinctions of the Phanerozoic (Raup and Sepkoski, 1982). The extinction affected marine taxa as well as terrestrial animals and plants. The following groups were mostly affected in the marine realm: bivalves, brachiopods, ammonites, gastropods, echinoderms, corals, sponges, foraminifera, ostracods, radiolarian and conodonts (Hallam, 2002). Several studies have examined the mass extinction and it has emerged that a substantial extinction at the T-J boundary was preceded by several smaller extinctions during the Late Triassic (Hallam, 2002; Tanner et al., 2004).

The cause for the mass extinction has equally been the subject of several studies. Initially a meteor impact was put forward, similar to the Cretaceous-Paleogene mass extinction (Olsen et al., 1987), supported in part by elevated iridium levels measured in T-J boundary sections (Olsen et al., 2002). More recently the elevated iridium levels, occurring in several sections worldwide, were however connected with volcanic activity as well as enrichment due to local redox conditions (Tanner et al., 2008, 2016).

It has been put forward that the opening of the Central Atlantic Ocean took place around the T-J boundary, despite analytical uncertainties (Fig. 2.4) (Marzoli et al., 1999). The spreading of the oceanic crust is associated with extensive Central Atlantic Magmatic Province (CAMP) volcanism (e.g. Marzoli et al., 2004; Cirilli et al., 2009). Carbon cycle perturbations recognised in the sedimentary record through the measurement of carbon isotopes have been linked to CAMP associated CO₂ degassing (e.g. Hesselbo et al., 2002). Some records confirm elevated CO₂ levels at the T-J boundary, for example through the stomatal density and the stomatal index of fossil leaf cuticles (e.g. McElwain et al., 1999). A biocalcification crisis associated with the T-J boundary is evident in the form of a carbonate-barren boundary interval, for example in the western Tethys and at the T-J GSSP type locality in the Kendelbach Formation in Austria (e.g. Kuerschner et al., 2007; Ruhl et al., 2009).

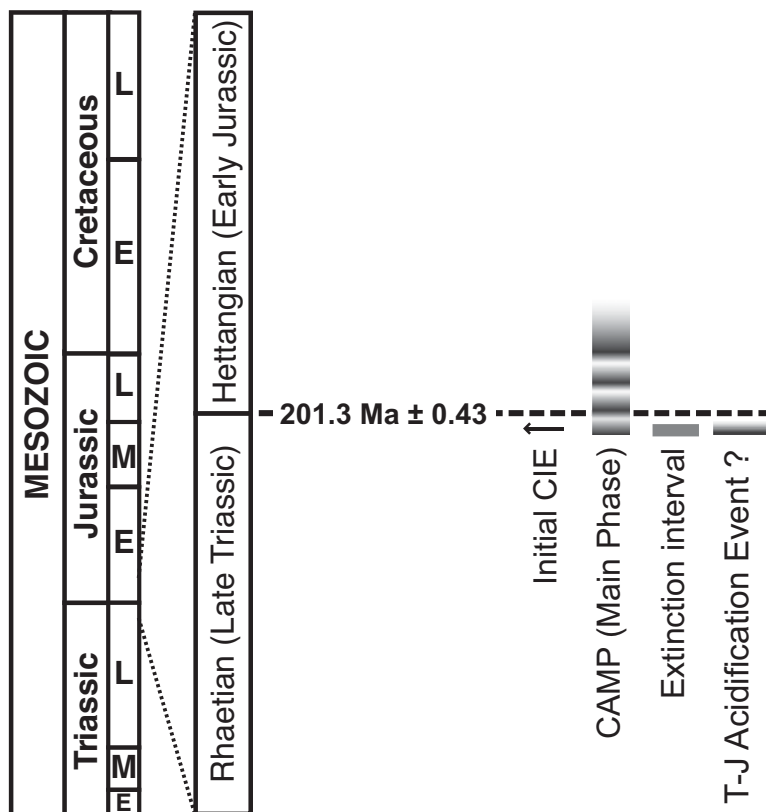


Fig. 2.4 – The Mesozoic timescale with expanded Rhaetian and Hettangian Periods highlighting events around the Triassic-Jurassic boundary (modified from Greene et al., 2014 and references therein).

This biocalcification crisis in the marine realm has been linked to potential ocean acidification, in turn induced by the dramatically elevated CO_2 levels from the CAMP associated volcanic activity (Fig. 2.5) (e.g. Greene et al., 2012). Contrasting evidence from pedogenic carbonates does not support elevated CO_2 levels at the T-J boundary. Alternative models to an induced ocean acidification event include global warming, triggered by other volcanic gases and potentially amplified by a release of CH_4 from gas hydrates, ultimately acting as main kill mechanism of the T-J mass extinction (Tanner et al., 2001; Beerling and Berner, 2002; Kiessling and Simpson, 2011; Hönisch et al., 2012).

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2.3 Stable carbon isotope stratigraphy

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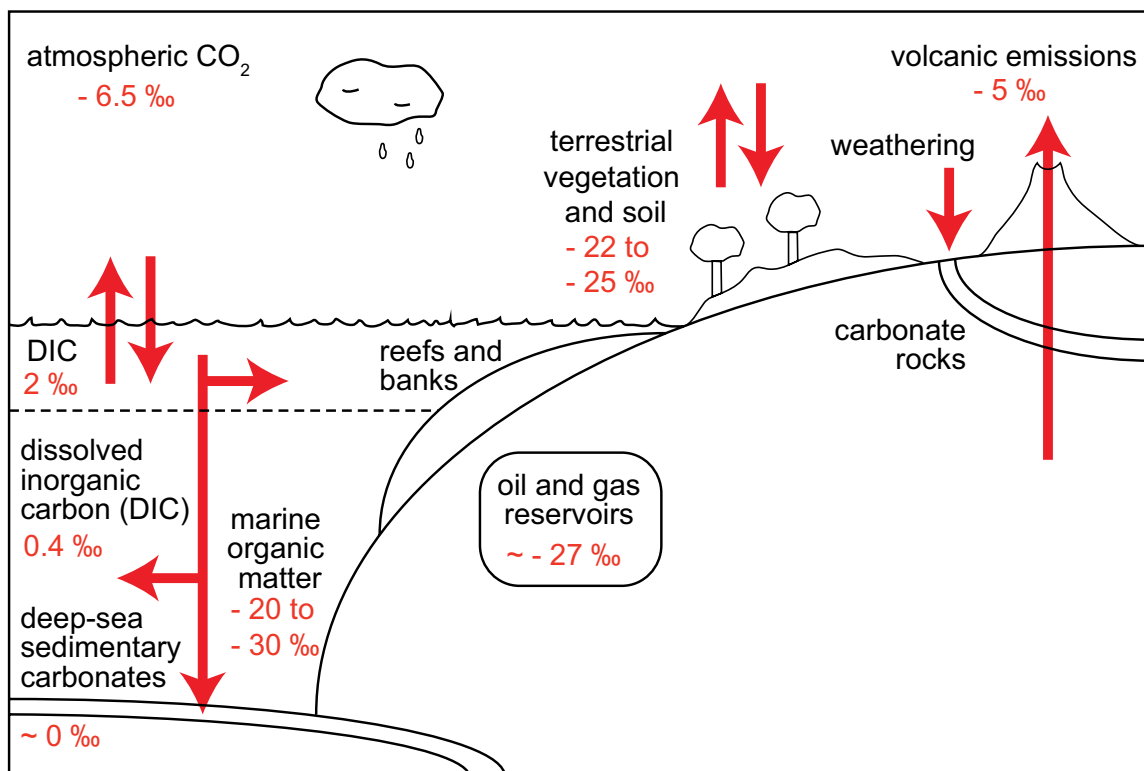


Fig. 2.6 – Carbon reservoirs of the present carbon cycle with their carbon isotopic composition (modified from Saltzman and Thomas, 2012; after Dunkley-Jones et al., 2010).

Fluctuations in the $\delta^{13}\text{C}$ signal of carbonates were firstly recognised in samples from the Deep Sea Drilling Project (e.g. Shackleton and Kennett, 1975) and have been used more widely ever since. Scholle and Arthur (1980) and Weissert (1989) were amongst the first to realise the potential of $\delta^{13}\text{C}$ measured from bulk samples as a correlation tool between pelagic stratigraphic sections. While this method was previously applied mainly to pelagic and hemipelagic settings, due to a more complete sedimentary record, newer studies have also focused on neritic sections (e.g. Föllmi and Godet, 2013; Krencker et al., 2014). The challenges here are to find enough usable material to perform the analysis on, as for example micritic limestone, and to make sure that the sedimentary record is as complete as possible. Multiple pitfalls have however been mentioned by various authors, when interpreting a neritic carbon isotope signal. One concern is the vulnerability of the shallow part of a carbonate platform to be exposed during the time of deposition. When carbonate production is subsequently shut down, the carbon isotope curve will not depict the carbon cycle perturbations during this interval. Sea level

fluctuations will also cause different mixture rates with meteoric water, which will also alter the isotopic signal (Immenhauser et al., 2003). Föllmi et al. (2006) pointed out, that changes in the type of carbonate factory (e.g. a shift from a photozoan dominated to a heterozoan dominated ramp) can cause changes in the isotopic composition of the precipitated carbonates. The same can be observed when mineralogical changes between aragonite and calcite occur (Swart, 2008). Swart (2008) suggests that aragonite is primarily produced in the shallow marine setting and will cause the carbon isotope signal of periplatform sediments to vary accordingly. One way that has been suggested to tackle these uncertainties is to compare carbon isotope measurements from bulk carbonates and the organic matter in these carbonates, to assess whether a carbon cycle perturbation has potentially influenced different carbon pools (e.g. Bodin et al., 2016).

3. Study area and geological setting

The outcrops studied and sampled as part of this project are situated on the Musandam Peninsula. The southern part of Musandam Peninsula is part of the United Arab Emirates (UAE) and the northern area is part of the Sultanate of Oman (Fig. 1.1A). The peninsula is covered by the Musandam Mountains, which form the northeastern extension of the Oman Mountains. The Triassic and Jurassic outcrops of the Musandam Peninsula were surveyed during a reconnaissance trip in December 2012 and during the subsequent three field expeditions (February 2013, January 2014, December 2015) outcrops in Wadi Naqab, Wadi Ghalilah, Wadi Al-Ghabbah (all UAE) and on Jabal Sall Ala (Oman) were studied (Fig. 1.1B).

The Musandam Mountains are composed mainly of Permian and Mesozoic carbonates (Fig. 3.1). Strata on the eastern side are of Permian and Triassic age, showing a general younging trend from east to west. The centre of the peninsula is mainly composed of Jurassic sequences at the surface and the Cretaceous sequences mainly crop out in the northeast. The entire Upper Permian to Lower Cretaceous shelf carbonate sequence of the Musandam Peninsula was deposited in a time of relative tectonic quiescence above a regional unconformity (Fig. 3.2) (Blendinger et al., 1990). The Late Carboniferous to Early Permian in the northern Oman area (including the Musandam Peninsula) was marked by alluvial plain and coastal deposition (Glennie et al., 1974).

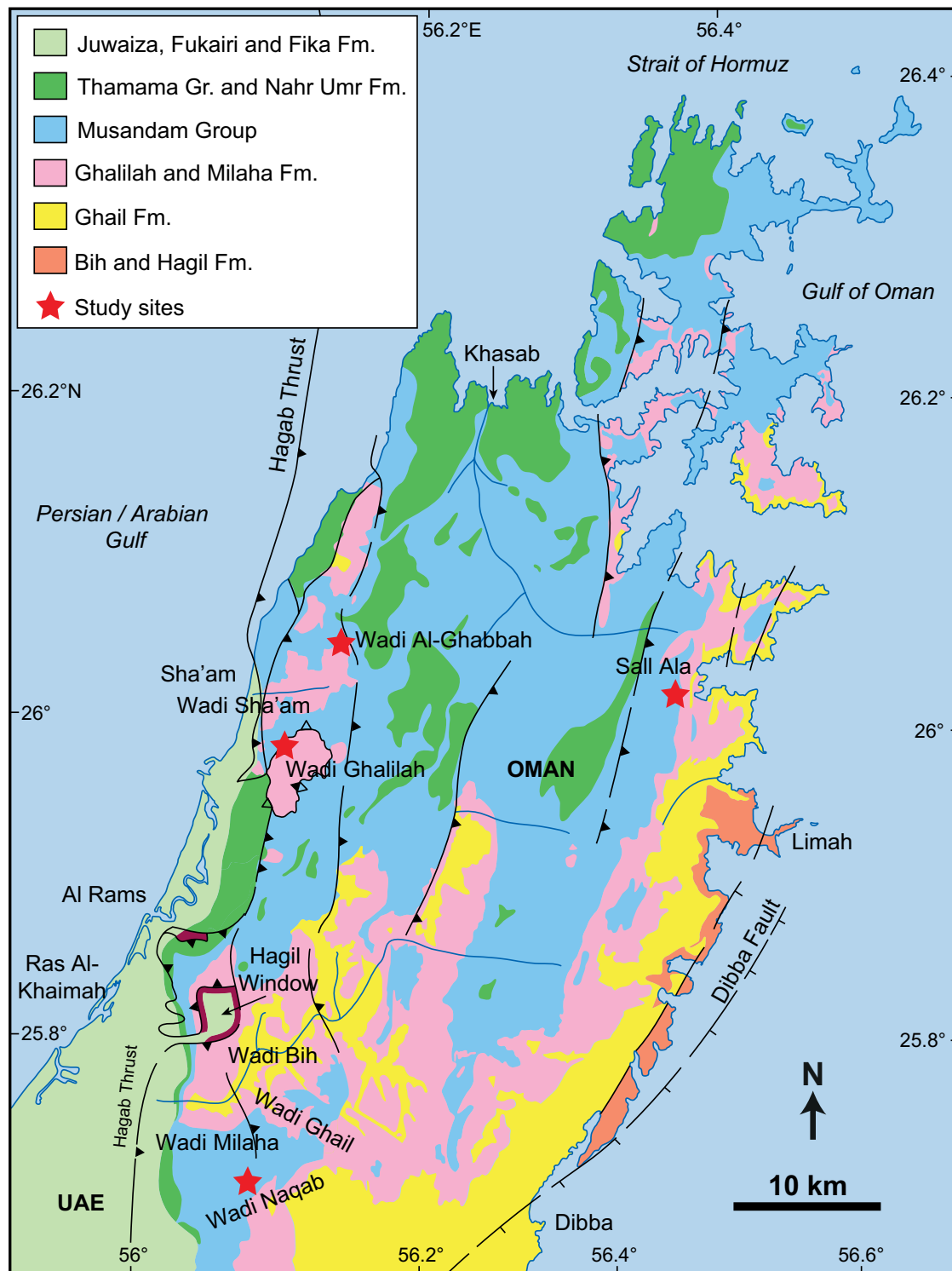


Fig. 3.1 – Geological map of the Musandam Peninsula with the four main study sites (modified from Searle et al., 2014 and references therein).

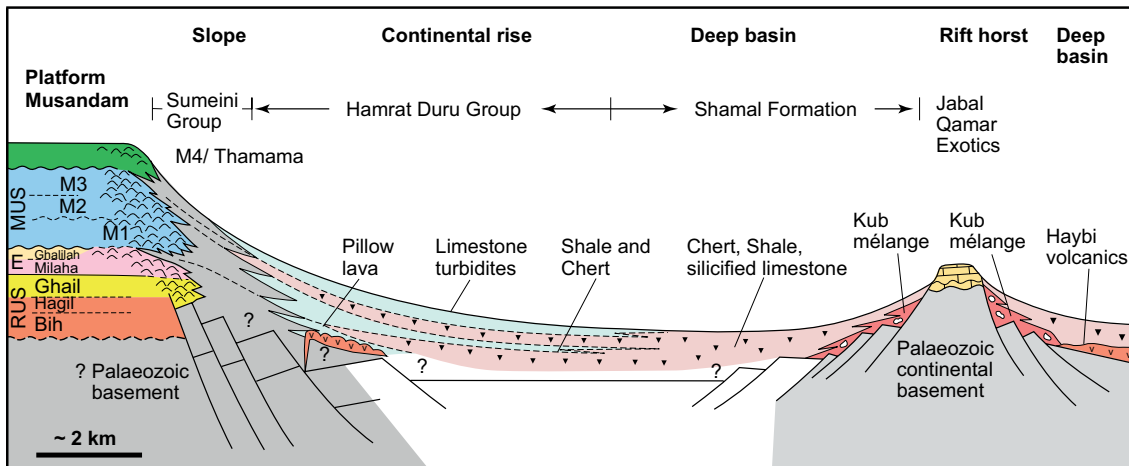


Fig. 3.2 – Reconstruction of the main thrust sheets on the Musandam Peninsula (modified from Searle et al., 2014; after Searle et al., 1983.)

Blending et al. (1990) put forward a ‘simple shear model’ to explain the configuration of the overlying Upper Permian units. Due to isostatic uplift of the Oman margin, that commenced during the Late Carboniferous, the siliciclastic Upper Carboniferous to Lower Permian sediments in the northern part of Oman were eroded. Extension of the lithosphere was caused by low-angle normal faults. Irregular uplift caused asymmetrical rifting in conjunction with coeval extensional tectonics, caused by sea floor spreading (Fig. 3.3).

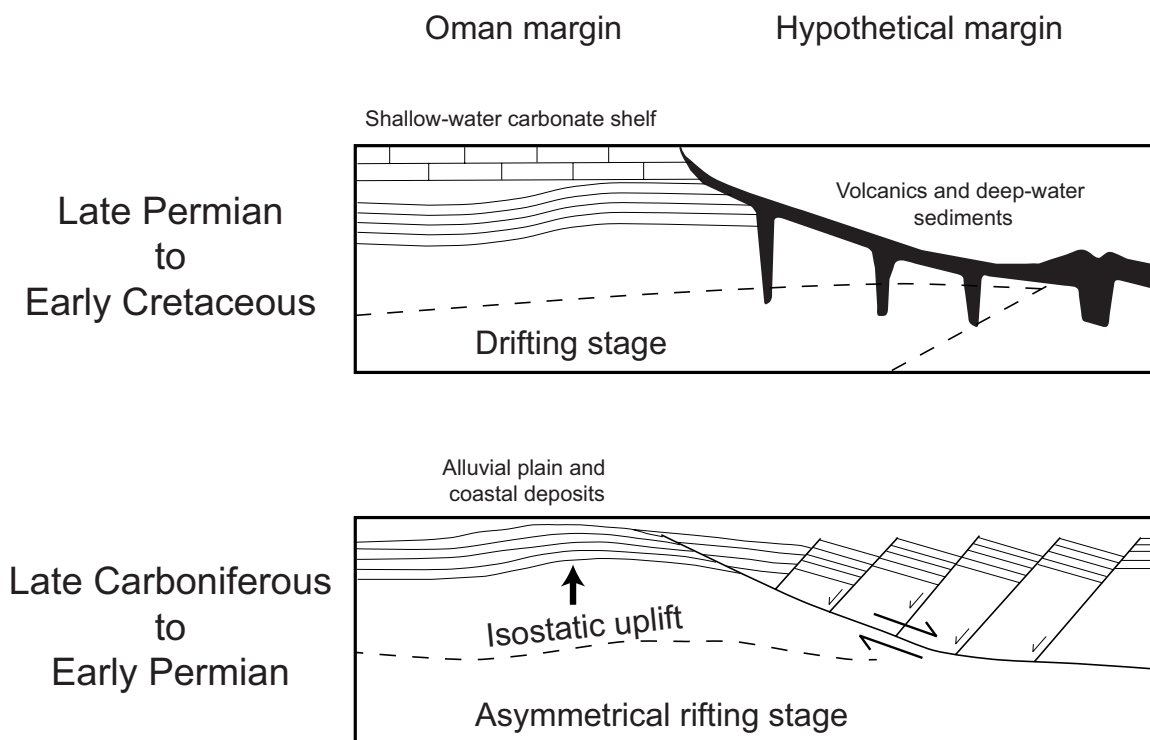


Fig. 3.3 – Simple shear model as possible explanation for the geodynamic history of northern Oman (modified from Blendinger et al., 1990).

This phase was followed by a prolonged period of relative tectonic quiescence during the Triassic and Jurassic when the shelf carbonates studied here accumulated in northern Oman and the Musandam Peninsula. This period of quiescence was interrupted during the Late Cretaceous when the Semail Ophiolites were obducted, beginning with the subduction of Triassic-Jurassic oceanic crust. The subduction ended approximately 70 Ma ago during the Late Cretaceous and was followed by ~ 30 Myrs of stable, shallow-marine conditions (Fig. 3.2). A second phase of crustal shortening that commenced during the Oligocene to Early Miocene (~ 33 - 20 Ma ago) followed. The third and final shortening phase started during the Pliocene (~ 5 Ma ago) and marks the continental collision into the Iranian Plate (Searle et al., 2014). The different shortening events are reflected in the present day buckled configuration of the upper lithospheric crust on the Musandam Peninsula (Fig. 3.4).

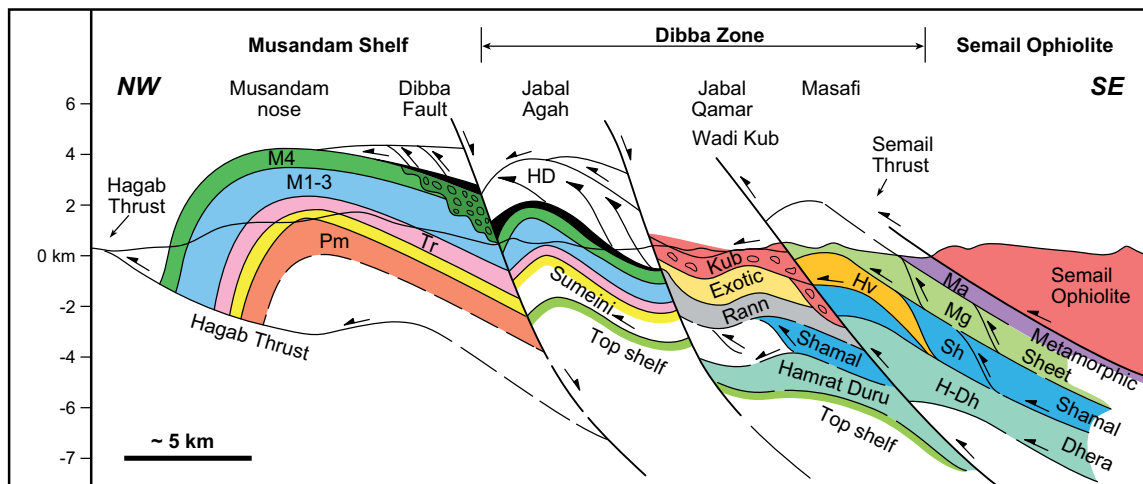


Fig. 3.4 – Simplified geological cross-section of the Musandam Peninsula (modified from Searle et al., 2014; after Searle et al., 1988).

4. Regional development of an equatorial carbonate platform across the Triassic-Jurassic boundary

4.1 Abstract

The Triassic-Jurassic boundary is marked by one of the 'big five' mass extinctions of the Phanerozoic. This boundary event was accompanied by several carbon cycle perturbations, potentially induced by the opening of the Central Atlantic and associated volcanism, and accompanied by an ocean acidification event. Continuous carbonate successions covering this interval of environmental change are however rare. Here data from a shallow-marine equatorial mixed carbonate-siliciclastic succession is presented, that was studied on a regional scale. Four sections that are 48 km apart were examined on the Musandam Peninsula (United Arab Emirates and Sultanate of Oman). The system was analysed for its sedimentology, vertical and lateral facies changes, and stable carbon and oxygen isotopes. Strontium isotope analysis was used to determine the position of the Triassic-Jurassic boundary horizon. The studied ramp experienced an episode of demise during the Late Triassic, followed by a restricted microbialite dominated ramp, containing large amounts of siliciclastic facies. During the Latest Triassic the diverse carbonate factory revived and flourished across the Triassic-Jurassic boundary. No clear evidence for a biocalcification crisis or an ocean acidification event across the Triassic-Jurassic boundary is visible. Lateral facies heterogeneities can be observed across the studied interval, attributed to hydrodynamic activity, including tropical storms, crossing the extensive shelf area. Although evidence for synsedimentary tectonic activity is present, the vertical stacking pattern is largely controlled by changes in relative sea level. The refined chronostratigraphy accompanied by the detailed environment of deposition analysis allows for a refinement of the regional palaeogeography. The neritic equatorial carbonate ramp has archived a negative carbon isotope excursion preceding the Triassic-Jurassic boundary that has also been reported from other study sites. The lack of evidence for a biocalcification crisis across the equatorial Triassic-Jurassic boundary indicates that the Tethys did not experience a distinct global acidification event.

4.2 Introduction and aims

The Triassic-Jurassic transition is accompanied by one of the major mass-extinctions of the Phanerozoic, and is marked by drastic biotic and environmental changes (e.g. Hesselbo et al., 2007) preceded by a carbon cycle perturbation (Hesselbo et al., 2002). This perturbation of the carbon cycle is synchronous with the eruption of the Central Atlantic Magmatic Province (CAMP), producing large amounts of CO₂ and associated volcanic volatiles (Cohen and Coe, 2002; Marzoli et al., 2004). Ocean acidification as the result of the exhalation of CO₂ has been suggested as the main cause for the mass extinction and the associated crisis (Hautmann et al., 2008; Greene et al., 2012, Hönisch et al., 2012; Ikeda et al., 2015). The Arabian Peninsula was covered by an extensive carbonate platform during most of the Mesozoic. The sedimentary rocks of the Arabian Platform have the potential to serve as excellent archive for palaeoenvironmental change. It has however been postulated that a significant stratigraphic gap exists between the Triassic and the Jurassic sequences on most of the Arabian Peninsula, due to erosion or non-deposition (e.g. Al-Husseini, 1997; Sharland et al., 2001; Ziegler 2001). In Oman it has therefore only been possible to study the Upper Triassic shallow-water platform (Bernecker, 2005; 2007), the Liassic shallow-water platform, unconformably overlying the Triassic (Bendias and Aigner, 2015) or the deep-sea record of the Upper Triassic to Lower Jurassic interval (Blendinger, 1988; Blechschmidt et al., 2004). The Musandam Peninsula hosts the stratigraphically most complete shallow carbonate platform section across this time interval on the Arabian Platform, although the exact stratigraphic boundary positions and hiatuses within the Late Triassic and Early Jurassic sequences in this region still remain unclear (Maurer et al., 2008; 2015). The Musandam Peninsula was located at an equatorial position during the Late Triassic to Early Jurassic interval (Tanner et al., 2004; Golonka, 2007). A continuous sedimentary record across the Triassic-Jurassic boundary is rare (Greene et al., 2012 and references therein), and so the carbonate platform of the Musandam Peninsula offers a unique opportunity to examine a carbonate platform at an important palaeogeographic location, potentially affected by major environmental changes across the Triassic-Jurassic boundary.

The aims of this study are: (1) to construct a robust chronostratigraphic framework for the Upper Triassic to Lower Jurassic carbonate sequences on the Musandam Peninsula and (2) to examine the evolution of an equatorial carbonate system across an interval of major global environmental change. These findings are then used to evaluate implications for the regional palaeoceanography and assess factors controlling the development of the carbonate system at the Triassic-Jurassic boundary.

4.3 Study area and geological setting

Four sections were selected for this study that are located on the Musandam Peninsula (Fig. 4.1A). The southern part of Musandam Peninsula is part of the United Arab Emirates (UAE) and the northern area is part of the Sultanate of Oman. The Musandam Mountains on the peninsula form the north western extension of the Oman Mountains and contain a well preserved, ~3 km-thick succession of shallow-water carbonates (Maurer et al., 2009). The eastern margin of the Arabian Plate was tectonically passive during the Permian and the Mesozoic. The overall relative tectonic quiescence was interrupted by two events, one during the Late Cretaceous when ophiolites were obducted on the Arabian plate, and the second one during the Late Oligocene – Early Miocene when the Oman and Musandam Mountains were formed during an initial collision phase of the Zagros orogeny (Glennie, 2005; Searle et al., 2014). From the Pliocene onwards the eastern margin of the Arabian platform was in an active continental setting and the Arabian plate is currently being subducted beneath the Eurasian Plate (Sharland et al., 2001; Searle et al., 2014).

The area of the Musandam Peninsula was located at an equatorial position during the Late Triassic to Early Jurassic interval (Tanner et al., 2004; Golonka, 2007). The area was covered by a shallow-marine carbonate platform during the Late Triassic and Early Jurassic (Ziegler, 2001) (Fig. 4.1B). To the north, more open-marine conditions prevailed. Terrigenous conditions existed to the south, with the coastline towards the west running roughly parallel to its present day location (Ziegler, 2001). Most of the UAE, Oman and Qatar were characterised by terrigenous to marginally marine environments, where shallow-marine clastics, evaporites and coastal and deltaic sediments were deposited

(Al-Husseini, 1997; Ziegler 2001).

The four sections are 48 km apart (Fig. 4.1A). The “Wadi Naqab section” lies on the northern side of Wadi Naqab, 13 km southeast of the city of Ras-Al-Khaimah (GPS coordinates of the log base: N 25°43’5.7”; E 56°05’14.1”). The “Wadi Ghalilah section” is located on the western flank of Wadi Ghalilah (GPS coordinates of the log base: N 25°58’23.1”; E 56°05’40.5”). The “Wadi Al-Ghabbah section” forms the northern extension of Wadi Sha’am, to the northeast of the city of Sha’am and is in close proximity to the border with Oman to the north (GPS coordinates of the log base: N 26°03’13.3”; E 56°08’32.6”). The “Jabal Sall Ala section” is located on the seaward facing flank of the mountain by the village Sall Ala in the Omani part of the Musandam Peninsula (GPS coordinates of the log base: N 26°01’46.6”; E 56°22’44.8”).

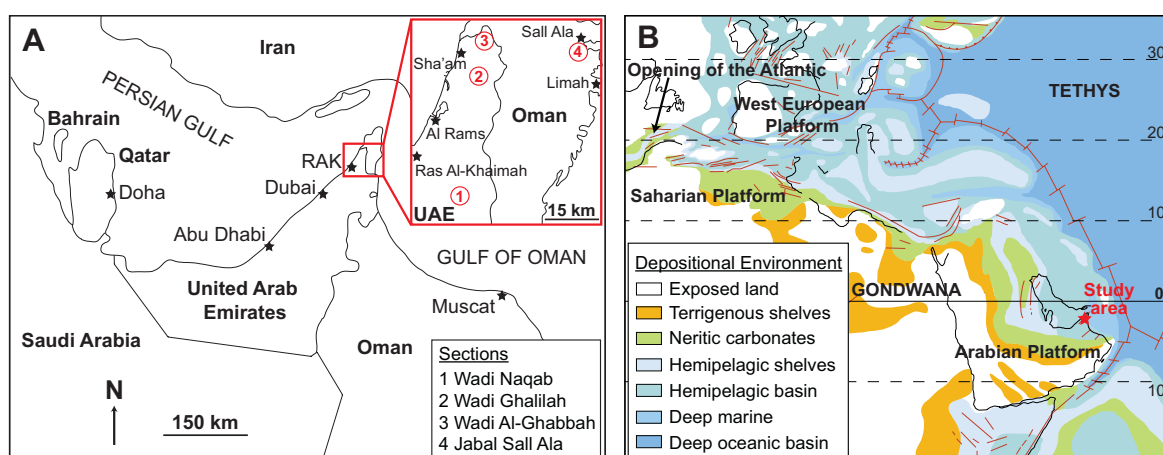


Fig. 4.1 - A) Location map of the study area. **B)** Palaeogeographic setting of the Tethyan realm during the Early Jurassic (modified from Pierre, 2006; after Thierry and Barrier, 2000).

4.4 Methods

The four sections were logged bed-by-bed using a Jakob's staff, and examined in the field for their sedimentary features, fossil assemblages and texture. Textural classification of carbonate facies followed Dunham (1962) and Embry and Klovan (1971). Palaeocurrent directions were measured from a cross-stratified bedset in the basal Musandam Limestone in Wadi Naqab using a geologic compass. Fist-size samples for petrography and stable isotope analysis were taken every 2 to 5 m along the vertical transects, using a

geological hammer, and 51 representative thin sections were made. Half of each thin section was stained after the method described by Dickson (1965), and subsequently analysed using a transmitted light Zeiss Axioskop 40 microscope (plain and polarised light).

Stable isotope measurements were performed in the Qatar Stable Isotope Laboratory at Imperial College London. All measurements were performed on micrite as it is the most common carbonate phase, and thin section analysis could be used to check the purity of the micrite. Where it was not possible to sample the micrite phase, a bulk sample was analysed. A total of 148 sample powders were obtained using an electrical dental drill, carefully avoiding veins, fossils and recrystallized sections. Approximately 100 to 230 μg (one sample with a relatively low CaCO_3 content) of sample powder were dissolved with 105 % orthophosphoric acid at 70 °C in a Kiel IV carbonate device, and the resulting CO_2 gas was measured on a Thermo MAT 253 mass spectrometer. Data were corrected according to measurements of the international standard NBS 19 and an in-house standard (ICM, Imperial College Carrara Marble). Stable isotope values are reported using the standard δ notation relative to the VPDB (Vienna Pee Dee Belemnite) standard. External precision for ICM (one standard deviation) on carbon and oxygen is better than 0.05 ‰ and 0.08 ‰, respectively. The precision for duplicate and triplicate sample measurements (one standard deviation) is better than 0.05 ‰ for carbon and better than 0.14 ‰ for oxygen.

The strontium isotope analyses ($^{87}\text{Sr}/^{86}\text{Sr}$) were carried out at the Royal Holloway University of London on selected oysters and brachiopods. The brachiopods were screened for diagenetic alteration prior to the analysis at Imperial College London using petrographic and cathodoluminescence (CL) microscopy. A CITL Cathodoluminescence Mk5-2 stage mounted on a Nikon Eclipse 50i microscope was used. Operating conditions for the CL microscope were about 270 μA and 14 kV. Samples were acidified for one hour at 80 °C using 5 % HNO_3 solution. Strontium was then separated from the solution using EichromSr-spec resin and then loaded on single Re filaments with a TaF emitter. The $^{87}\text{Sr}/^{86}\text{Sr}$ analyses were determined using the multidynamic procedure of Thirwall (1991) and analysed on an Isotopx Phoenix Thermal Ionisation mass spectrometer. The standard SRM 987 analysed alongside

samples gave a mean of 0.710237 ± 0.000006 (2sd, N=3) (Appendix A.4), within error of the long-term mean of 0.710234 ± 11 (2sd, N=177).

4.5 Chronostratigraphic framework

Consensus seems to prevail that Upper Triassic and a large portion of Lower Jurassic sediments are not preserved on much of the Arabian Platform due to erosion, or were never deposited due to exposure during a sea level lowstand (e.g. Alsharhan and Nairn, 1994; Al-Husseini, 1997; Le Nindre et al., 2003; Bendias and Aigner, 2015). Liassic deposits of Toarcian age occur on parts of the Arabian Shield due to a transgression and the subsequent creation of accommodation (Al-Husseini, 1997). On the Musandam Peninsula, a more complete stratigraphic record of the Triassic and Jurassic is present. The Triassic period is only missing parts of the Carnian and Norian (Maurer et al., 2008). The Jurassic sedimentary record is missing parts of the Hettangian and Sinemurian and parts of the Toarcian, the Aalenian and the Tithonian (de Matos, 1997; de Matos and Walkden, 2000).

The four sections comprise the top Sumra Member, the Sakhra and Shuba Members as well as the basal Musandam Limestone (Fig. 4.2; 4.3A). The former are the upper members of the Ghalilah Formation and the Musandam Limestone forms the Lower Musandam Formation. The three members as well as the underlying Asfal and Sumra Members were originally described by Hudson (1960) and attributed to the Triassic. Glennie et al. (1974) later revised this and dated the Sakhra and Shuba Members as Early Jurassic, based on a single *Orbitopsella* specimen, thus extending the Ghalilah Formation into the Jurassic. This was supported by Maurer et al. (2008; 2015) based on the last occurrence of Rhaetian corals in the uppermost Sumra Member and the absence of fossils in the Sakhra and Shuba Members, which Maurer et al. (2008; 2015) attribute to the end-Triassic mass extinction. De Matos et al. (1994) and de Matos (1997) studied the Jurassic stratigraphy in Wadi Naqab extensively and, based on the macrofossil assemblage described previously by Hudson and Jefferies (1961) and the stratigraphic study by Metwally and Ali (1992), identified the contact between the Shuba Member and the basal Musandam Formation as the Triassic-Jurassic boundary at Wadi

Naqab. It is noteworthy that Metwally and Ali (1992) assign an Early Jurassic age to the “cliff-forming limestones” (presumably the basal Musandam limestone and not the Sakhra limestone, since the latter is only approximately 25 m thick (Hudson, 1960)) based on a macrofossil assemblage containing the ammonoid *Tragophylloceras numismale* and the echinoid *Scaptodiadema*. The ammonite *T. numismale* points to a Pliensbachien age in the Euroboreal realm (Meister et al., 2012). It is furthermore noted that Maurer et al. (2008), who focused on the Permian to Triassic sequences of the Musandam Peninsula, place the Triassic-Jurassic boundary based on the last occurrence of *Retiophyllia* corals in the Upper Sumra Member, as both the Sakhra Member and most of the Shuba Member do not contain any age-diagnostic fauna. De Matos (1997) performed a detailed biostratigraphic examination of the Upper Ghalilah and the Musandam Formations in Wadi Naqab and neighbouring wadis and did not confirm the *Orbitopsella* find from Glennie et al. (1974) in the Shuba Member. The Triassic-Jurassic boundary was placed within the uppermost Shuba Member based on the presence of abundant Liassic *Balanocrinus subteroides* crinoids in the basal Musandam Limestone and an assemblage of Triassic shark teeth (*Acrodus cf. lateralis*) and, most likely, Triassic bivalves (*Pseudoplacunopsis* sp., *Plicatula radiata*, *Bakevillia* sp.) encountered in a correlatable section examined in Wadi Milaha (~ 8 km north of Wadi Naqab) in the Upper Shuba Member (de Matos, 1997).

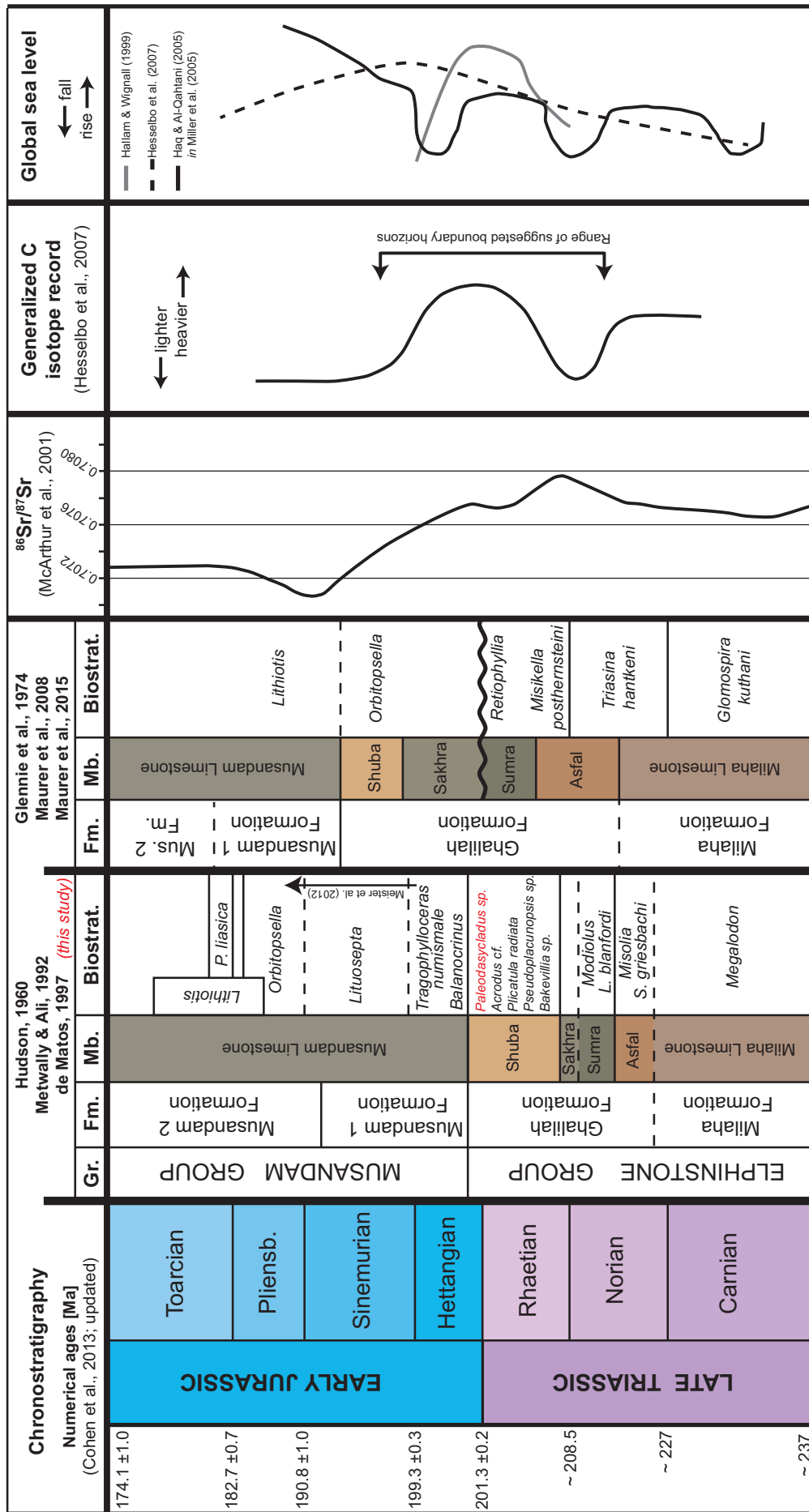


Fig. 4.2 - Late Triassic – Early Jurassic chronostratigraphy and different biostratigraphic frameworks from the Musandam Peninsula combined with chemostratigraphic data (strontium and carbon isotopes) and different sea level curves.

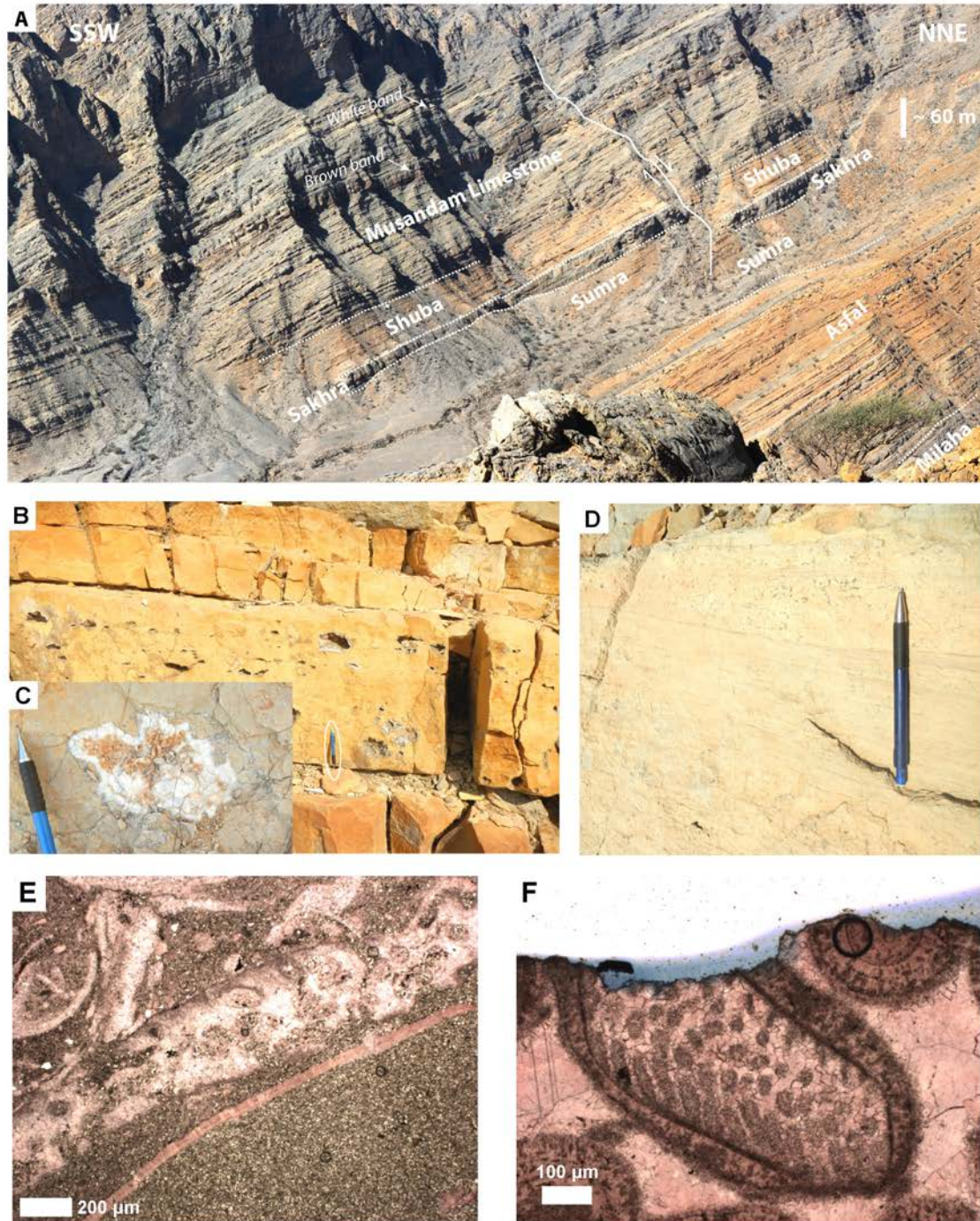


Fig. 4.3 - **A)** Panoramic view of the studied formations in Wadi Naqab. **B)** Abundant vugs within a brown dolo-mudstone bed. **C)** White calcite cement nodule with pink dolomite cement in the centre (both from within the Shuba Member at Wadi Al-Ghabbah) **D)** Whitish microbial laminite overlain by fenestral limestone (from the Shuba Member at Wadi Naqab). **E, F)** Dasycladalean algae *Paleodasycladus* sp. (from the Upper Shuba Member at Wadi Naqab).

4.6 Results

4.6.1 Lithofacies associations

Nine different lithofacies associations (LF 1 to 9) were differentiated (Table 4.1). LF 1 comprises fenestral mud- and wackestones, containing peloids and bioclast fragments. Quartz grains as well as 2-5 cm long mudclasts are common in this facies. In some of these beds, subangular vugs with a diameter of up to ca. 15 cm are present (Fig. 4.3B). Some of these vugs are filled with nodular white cement, occasionally with a pink cement nucleus. Some of these vugs have irregular rims, resembling cauliflower (Fig. 4.3C). LF 2 comprises boundstones and mudstones containing laminae with fenestrae (Fig. 4.3D). The microbial laminites occur commonly as continuous beds and occasionally occur with domal structures, thus forming stromatolites. Tepee-structures are common within these laminites (Fig. 4.4A). The mudstones of LF 1 and 2 are frequently dolomitized. Two distinct types of crystalline dolomite occur. A fine-grained type with fine rhomb-shaped dolomite crystals (10 μm in diameter) cemented together and a coarser type where large dolomite rhombs (50 μm in diameter) float in a calcite cement matrix. LF 3 and LF 4 comprise wackestones, mudstones and packstones, containing few specimens of bivalves, gastropods, peloids, ostracods, echinoderms and dasycladalean algae fragments (*Paleodasycladus sp.*, Fig. 4.3E, F). The wackestone facies is in some cases channelized (Fig. 4.4B). LF 4 is distinguished from LF 3 by the fact that the former contains quartz, and is occasionally partly dolomitised. LF 3 contains no quartz grains. Bedding within the two lithofacies types is often irregular and nodular, and occasional bioturbation occurs. LF 5 has a grainstone texture, typically containing ooids and peloids (Fig. 4.4C). The nuclei of the ooids are mostly bioclasts, though quartz grains are also abundant. Different bivalves, such as oysters are commonly present within this facies association as well as different other bioclasts, including echinoderms. LF 5 additionally frequently contains lithoclasts and aggregate grains (Fig. 4.4D). It is typically cross-stratified as well as cross-laminated, and mud-drapes can be observed (Fig. 4.4E, F, G). LF 6 comprises diverse packstones or sometimes rudstones with a rich fossil content. The fossil content includes complete bivalve shells and fragments, and often oysters, gastropods, benthic foraminifera, ostracods,

echinoderms, peloids and coral fragments. These beds have often been heavily bioturbated. The abundant ichnofossils are commonly *Rhizocorallium* and *Chondrites*. LF 7 is characterised by the presence of corals. These can occur as branching framestones, with the corals preserved in living position (Fig. 4.5A), massive coral heads (Fig. 4.5C, D) or coral debris rudstones (Fig. 4.5E). The matrix of the limestones around the corals is formed of ooids and large gastropods. LF 7 is in places heavily bioturbated and contains cm-sized vugs. The branching corals in the top part of the Sumra Member on the Musandam Peninsula have been identified as *Retiophyllia* corals (Maurer et al., 2008). LF 8 comprises siliciclastic facies of fine to medium sized quartz sandstones, which are typically cemented by calcite (Fig. 4.5B). Occasional dolomite cement is present. These sandstones are often cross-stratified and cross-laminated (Fig. 4.5F), exhibit nodular bedding, and sometimes show flaser-bedding, lenticular bedding as well as wave ripples (Fig. 4.5G) and in one case hummocky cross-lamination (Fig. 4.5H). Other clastic rock types included in this facies association are monomictic and polymictic conglomerates (Fig. 4.5I). LF 9 comprises shales and marls that are very abundant throughout the studied sections. These fine-grained sediments are brownish to grey, red and green in colour and are typically finely laminated. Bioturbation is common, mostly *Thalassinoides*.

Table 4.1 - Description, dimensions and interpretation of the nine observed lithofacies associations (LF); (m: marl; M: mudstone; W: wackestone; P: packstone; G: grainstone; F: floatstone; R: rudstone; B: boundstone).

LF	Facies	Dunham Texture	Main Components	Sedimentary Features	Depositional Environment	Dimensions
1	Mudflat facies	M M-W	Fenestral, peloids, bioclasts fragments, calcite nodules, quartz, mudclasts	Nodules, lamination, mudclasts	Flood plain, maybe tidal flat	Thickness: 0.1 – 3.8 m Max. lateral extent: < 8 km
2	Microbial laminite	B M	Microbial mats, stromatolite, fenestral, often dolomitised	Dolal structures, tepees, mud-streaks, bioturbation (<i>Skolithos</i>)	Flood plain, maybe tidal flat or intertidal zone	Thickness: 0.15 – 2 m Max. lateral extent: 46 km
3	Restricted fauna	M W W-P	Bivalves, gastropods, peloids, echinoderms, lithoclasts, ooids	Irregular, nodular bedding, bioturbation	Restricted platform	Thickness: 0.05 – 2.7 m Max. lateral extent: 30 km
4	Quartz-rich restricted fauna	W W-P	Bivalves, quartz, gastropods, ostracods, echinoderms, peloids partly dolomitised	Irregular, nodular, channelised	Restricted platform	Thickness: 0.1 – 1.2 m Max. lateral extent: 23 km
5	High-energy facies	G	Ooids, peloids, bivalves, oysters, lithoclasts, aggregate grains, echinoderms, dasycladacean algae	Cross-lamination and – stratification, mud-drapes	High-energy platform; open marine platform; intertidal zone	Thickness: 0.15 – 26 m Max. lateral extent: 46 km
6	Diverse fauna	P R	Bivalve fragments, gastropods, foraminifera, ostracods, echinoderms, oysters, peloids, coral fragments	Bioturbation (e.g. <i>Rhizocorallium</i> , <i>Chondrites</i>)	Open marine platform	Thickness: 0.05 – 3.7 m Max. lateral extent: 46 km
7	Coral facies	R B	Large massive corals, grainy ooidal matrix, gastropods	Bioturbation, vugs	Subtidal zone; open platform	Thickness: 0.2 – 0.9 m Max. lateral extent: 46 km
8	Siliclastic facies	-	Quartz, calcite-cemented, occasional dolomite cement, conglomerate (monomictic and polymictic)	Cross-lamination and – stratification, nodular bedding, flaser-bedding, red Fe/Mn nodules, lenticular bedding, wave ripples, hummocky cross-stratification	Intertidal zone; open platform	Thickness: 0.1 – 3.8 m Max. lateral extent: 30 km
9	Shale and marl	m	Marls brown-grey, red, green, laminated shales, nodules	Lamination, nodules, bioturbation, (e.g. <i>Thalassinoides</i>)	Deep, open platform	Thickness: 0.1 – 3.8 m Max. lateral extent: 30 km

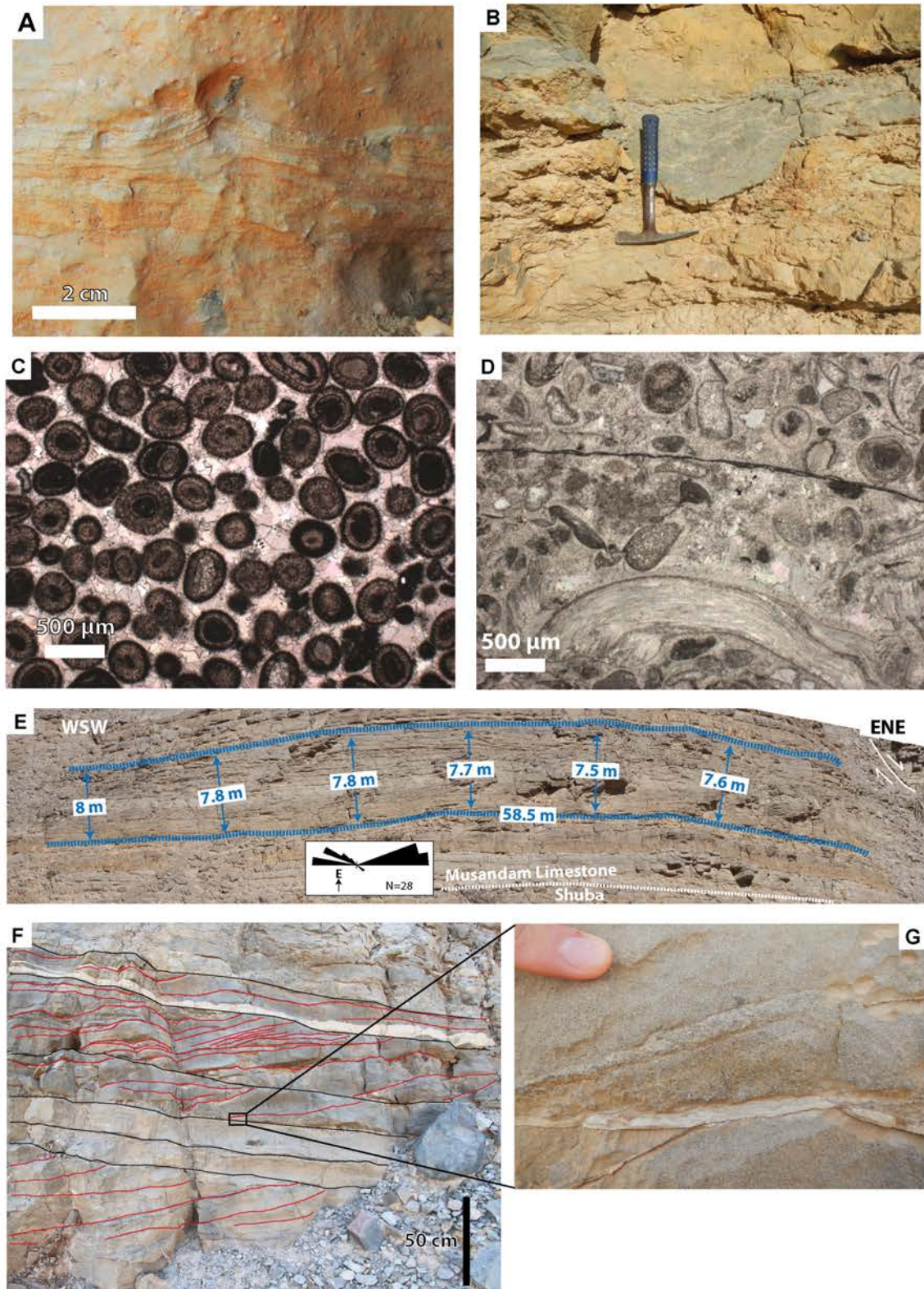


Fig. 4.4 - A) Tepee structure within a microbial laminite bed (from the Shuba Member at Wadi Naqab). **B)** Channelized bioclastic wackestone (from the Shuba Member at Wadi Ghalilah). **C)** Stained thin section image of an ooidal grainstone. **D)** Thin section image of a bioclastic grainstone. **E)** Detailed panorama of the oolitic, cross-stratified dune at the base of the Musandam Limestone, including a rose diagram of the palaeocurrent measurements. **F)** Detailed view of the oolitic dune shown in B. Black lines indicate bedding planes. Red lines

indicate cross beds. **G)** Detailed view of at the contact between a bedding plane and a cross bed with a mudstringer in between (all from Wadi Naqab).

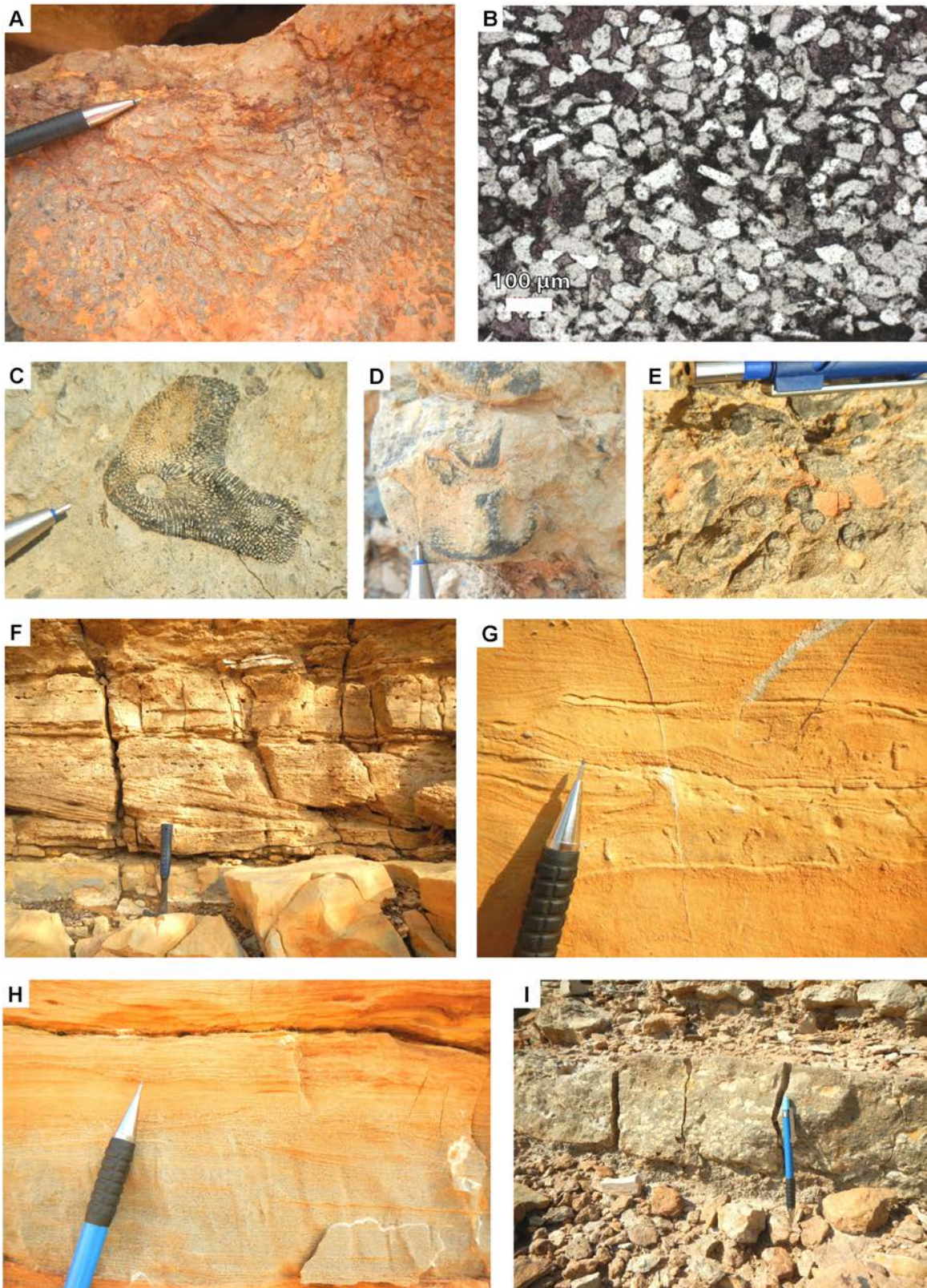


Fig. 4.5 - A) Branching coral in living position from the Shuba Member in Wadi Ghalilah. **B)** Stained thin section image of calcite-cemented quartz sandstone. **C)** Large coral head, top Sumra Member at Wadi Naqab. **D)** Branching corals from the top Sumra Member at Wadi Naqab. **E)** Well-preserved coral fragments (from the top Sumra Member at Jabal Sall Ala (all corals *Retiophyllia* (?), sensu Maurer et al., 2008). **F)** Cross-laminated sandstone from the top of the Shuba Member at Wadi Al-Ghabbah. **G)** Lenticular bedding in a sandstone bed at the top of the Shuba Member at Wadi Al-Ghabbah. **H)** Hummocky cross-stratification in a sandstone bed at the top of the Shuba Member at Wadi Al-Ghabbah. **I)** Monomictic conglomerate from the Shuba Member in Wadi Ghalilah.

4.6.2 Vertical stratigraphic stacking pattern

The base of the measured Wadi Naqab section is in the upper part of the Sumra Member (Fig. 4.6). It is formed of bioclastic limestones (LF 6) and coral float- and rudstones (Fig. 4.5C, D, E) (LF 7) interbedded with marl and shale beds (LF 9). The Sumra member is overlain by a ~ 26 m thick oolite unit (LF 5), which forms the Sakhra Member. The oolite shows abundant cross bedding and cross laminations. The top of this oolitic grainstone unit is marked by red staining, as well as scours and fractures, containing red mudstone and oolite clasts (Fig. 4.7A). The oolitic grainstone is overlain conformably by the Shuba Member. The Lower Shuba Member is dominated by microbial and fenestral limestones (LF 2), fine sand- and siltstone beds (LF 8), interbedded with marl and shale layers (LF 9). The upper half of the Shuba Member contains mainly bioclastic and oolitic limestones interbedded with marls (LF 3, 5, 6 and 9). The uppermost beds of the Shuba Member are condensed, pale brown to yellowish, ooidal grainstones, rich in oysters and crinoids (LF 5 and 6). The Shuba Member contains 19 discontinuity surfaces, which are marked by mineral crusts, red staining and occasionally by abundant bioturbation. Two of these surfaces contain desiccation cracks (Fig. 4.7B). The basal part of the overlying Musandam Limestones consists of a 7 m thick cross-stratified very heterogenous (Fig. 4.4E, F) ooidal grainstone unit (LF 5), containing one crinoidal packstone bed (LF 6), dm thick layers of bioclastic pack- and grainstones (LF 6) and frequent 5 – 10 cm thick mudstringers (Fig. 4.4G). The measured palaeocurrent directions from this cross-stratified interval are NNE-

SSE trending (Fig. 4.4E). The upper part of the section is marked by dm to 1.5 m thick mud- to wackestone beds (LF 3 and 6).

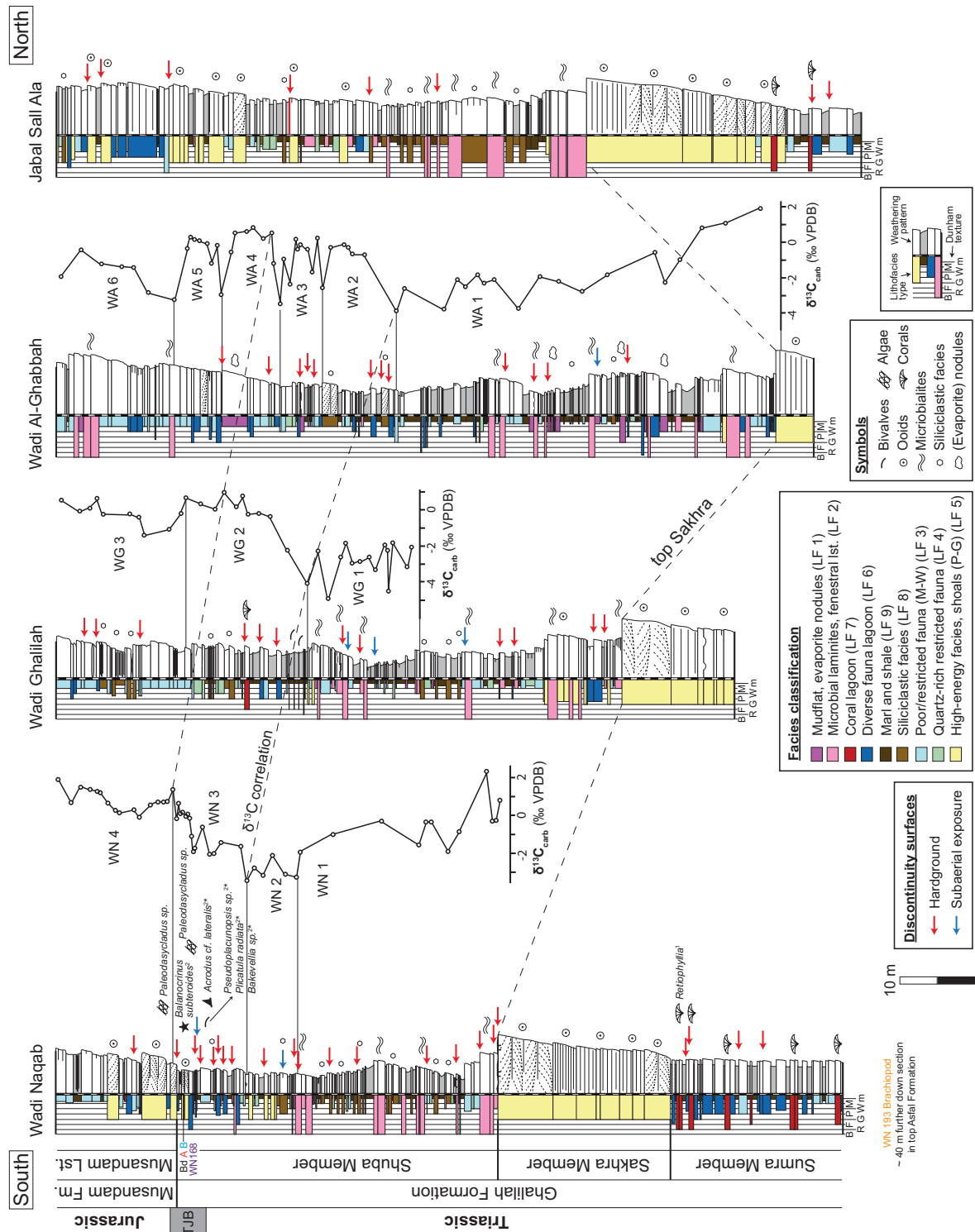


Fig. 4.6 - Detailed sedimentological sections from the four study sites, including stable carbon isotope curves. The chronostratigraphic boundaries for the Wadi Naqab section are placed in accordance with strontium isotope results obtained in this study and biostratigraphic markers

(positions of samples used for Sr isotope analysis are indicated on the Wadi Naqab log; ¹ from Maurer et al., 2008; ² from de Matos, 1997; ^{2*} from de Matos, 1997 - observed in Wadi Milaha; m: marl; M: mudstone; W: wackestone; P: packstone; G: grainstone; F: floatstone; R: rudstone; B: boundstone).

The oolites of the Sakhra Member form the basal part of the Wadi Ghalilah section (LF 5) (Fig. 4.6). The Sakhra Member contains one channelized bed. The channels are approximately 30 cm deep and filled with laminated mudstones (Fig. 4.7C). The cliff-forming Sakhra Member is overlain by the Shuba Member, which comprise in its lower part interbedded microbial boundstones (LF 2), bioclastic pack- to grainstones (LF 6) and ooidal grainstone beds (LF 5). The upper part of the Shuba Member contains microbial boundstones, fenestral mudstones (LF 2), interbedded with fine sand- to siltstones, one monomictic conglomerate bed (Fig. 4.5I) (LF 8) and marl and shale layers (LF 9). One of the fenestral mudstone beds contains small normal-fault blocks (55 m upsection, Fig. 4.7D). The uppermost Shuba Member contains bioclastic limestone beds with several thin oyster packstone layers (LF 6), interbedded with marls and shales (LF 9). The topmost of these beds is a coral float- to rudstone (LF 7), with some large, massive corals in growth position (Fig. 4.5A). The coral limestone is overlain by several siliciclastic and quartz-rich limestone beds (LF 8 and 4). The onset of the overlying Musandam Limestone is placed where the beds are thicker and are more resistant to weathering. The lower part of the Musandam Limestone forms the top part of the logged section and comprises mudstone (LF 3 and 4) and sand- and siltstone beds (LF 8). The Shuba Member logged in Wadi Ghalilah contains 12 discontinuity surfaces of which three are marked by desiccation cracks. The Lower Musandam Limestone contains three such discontinuity surfaces.

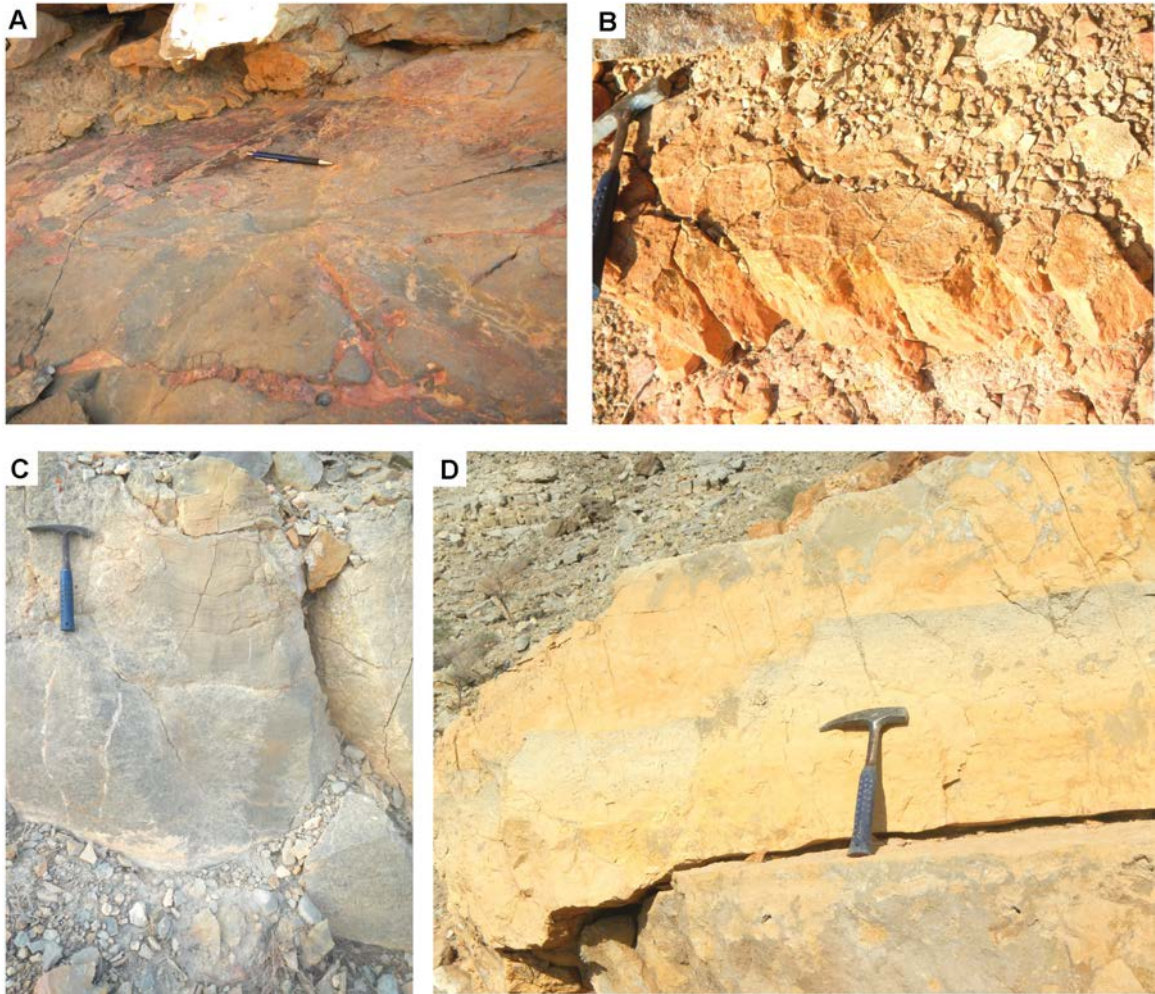


Fig. 4.7 - A) Karstified and stained top of the Sakhra Member oolite, Wadi Naqab. **B)** Desiccation cracks on a bedding surface from the Shuba Member in Wadi Ghalilah. **C)** Karst infill within the Sakhra Member oolite at Wadi Ghalilah. **D)** Synsedimentary normal fault within a fenestral mudstone bed of the Shuba Member in Wadi Ghalilah.

The base of the measured Wadi Al-Ghabbah section corresponds to the top at the Sakhra Member, which forms a prominent ledge and consists of ooidal grainstone (LF 5) (Fig. 4.6). The basal part of the overlying Shuba Member consists of 12 m of microbial boundstones (LF 2), mudstones (LF 3) and few bioclastic limestones (LF 6). This interval is overlain by 30 m of shales and marls (LF 9), which are interbedded with dm to metre-thick limestone (LF1, 2 and 3). The mudstones of LF 1 in this interval are characterised by abundant vugs with diameters of around 10 cm, which are sometimes filled with white cement. The upper half of the Shuba Member is dominated by marl and shale beds (LF 9). These are interbedded with mainly metre-thick beds of mudstone (LF 3) and bioclastic wacke- and packstones (LF 6). A noticeable 1.05 m thick

cross-stratified sandstone bed, which is overlain by a thin polymictic conglomerate, lies within this interval (60 m upsection, both LF 8). The conglomerate is only present in depressions within the top of the underlying sandstone. It consists of grey and ochre coloured mudclasts with diameters between 1 and 5 mm. Approximately 6.5 m up section a second prominent, 2.35 m thick, cross-stratified and cross-laminated sandstone bed (Lf 8) is present (67 m upsection, Fig. 4.5F). In its upper half it contains a horizon with lenticular bedding and internally laminated wave ripples (Fig. 4.5G). This bed is in parts bioturbated internally and on its surface. The top surface of the bed is a condensed layer, rich in bioclasts and coloured pale brown- to yellowish with red stained patches. The overlying Musandam Limestone above the Shuba Member is marked by thicker beds and contains an echinoid-rich wackestone (LF 3) and two very quartz-rich beds (LF 4) in its basal part. These two beds are overlain by thickening-upwards mudstones (LF 1, 2 and 3). The Shuba Member in Wadi Al-Ghabbah contains 11 discontinuity surfaces of which one is marked by desiccation cracks.

The Jabal Sall Ala section consists of limestone beds (LF 3 and 6) interbedded with shales and marls of the Sumra Member in its basal part (Fig. 4.6). The uppermost Sumra Member contains two coral float- to rudstone beds with well-preserved branching corals (Fig. 4.5C, D, E) (LF 7) as well as two ooidal grainstone beds (LF 5). The overlying Sakhra Member is 25 m thick, and contains 5 thickening upwards, partly cross-stratified and cross-laminated ooidal grainstone beds (LF 5). The overlying Shuba Member contains microbial boundstones (LF 2) and silt- to fine sandstones (LF 8) in its lower half, interbedded with few thin shale and marl layers (LF 9). The upper half of the Shuba Member contains different types of limestone beds (LF 3, 4, 5 and 6), interbedded with shales and marls (LF 9). The basal Musandam Limestone is marked by a cross-stratified ooidal grainstone bed (LF 5), overlain by ooidal grainstone and by a relatively thick (> 3 m) bioclastic packstone bed (100 m upsection, LF 5 and 6). The Shuba Member of the Jabal Sall Ala section contains three discontinuity surfaces.

4.6.3 Stable carbon isotope record

The majority of $\delta^{13}\text{C}$ values from the three sections lie between -3 and 1 ‰ (Fig. 4.6). The $\delta^{18}\text{O}$ values lie between -7 and -1 ‰. A table containing all carbon and oxygen isotope results can be found in appendix A.5-A.7. A cross-plot of stable isotope results from the three measured sections shows no significant trends between $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ (Fig. 4.8). The dataset could however be divided into two groups, one with $\delta^{18}\text{O}$ values of between 0 and -2.7 ‰ and a second group with more depleted $\delta^{18}\text{O}$ values between -3 and -8 ‰. The $\delta^{13}\text{C}$ record generated from samples taken from Wadi Naqab ranges from the Upper Sakhra Member to the basal Musandam Limestone.

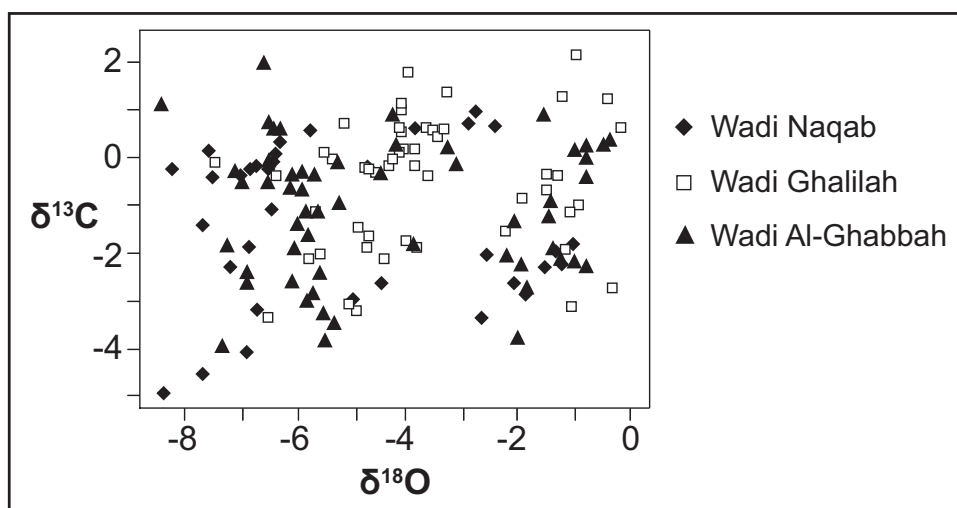


Fig. 4.8 - Crossplot between all $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ measurements from the three measured sections.

4.6.4 Strontium isotope stratigraphy and chronostratigraphy at Wadi Naqab

$^{87}\text{Sr}/^{86}\text{Sr}$ isotope analysis was performed on oysters and a single brachiopod from Wadi Naqab (the positions of the samples are indicated in Fig. 4.6 and 4.10A). Only specimens that display little or no luminescence were selected for the Sr analysis as this implies low Mn^{2+} and Fe^{2+} concentrations (Boggs and Krinsley, 2006). The brachiopod sample (WN 193 Br3) was taken from the top of the Asfal Member (the lowest member of the Ghalilah Formation) approximately 40

m stratigraphically below the base of the studied section. The specimens are thicker than 500 μm and the foliated texture is well-preserved (Fig. 4.9A, B).

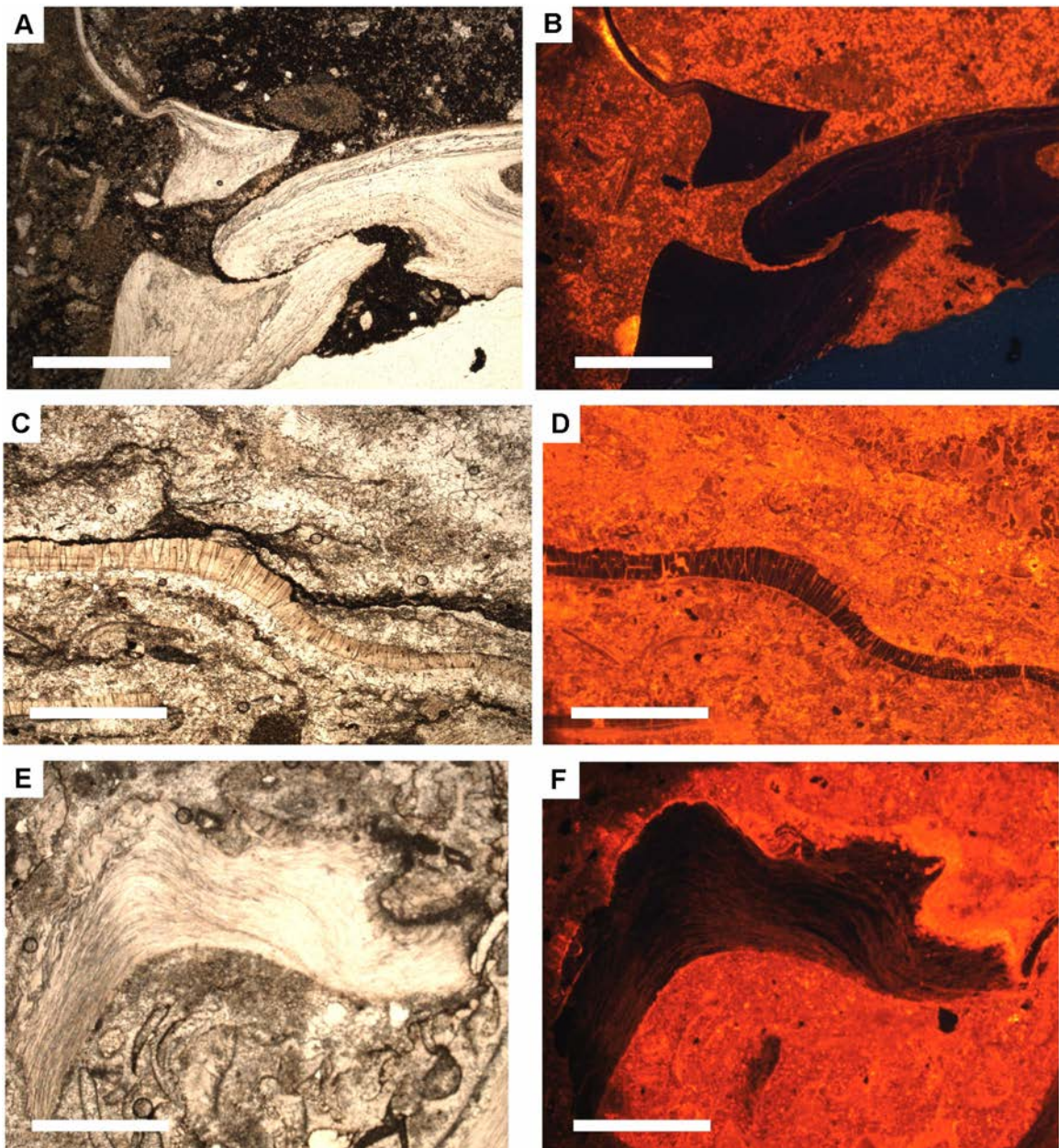


Fig. 4.9 - **A)** Thin section image of a brachiopod from sample WN 193. **B)** Cathodoluminescence image of A. **C)** Thin section image of an oyster from sample WN 168 Oy. **D)** Cathodoluminescence image of C. **E)** Thin section image of an oyster from sample WN TJ B. **F)** Cathodoluminescence image of E. The scale bar in all images is 500 μm long.

The sample WN 168 Oy was taken from a fossil-rich bioclastic rudstone in the top of the Shuba Member. The selected specimen is relatively thin and contains luminescent cracks (Fig. 4.9C, D) Samples WN TJ B2 and WN TJ B3 were

taken from an ooidal grainstone bed, rich in oysters and crinoids, two beds above. The selected specimens have relatively thick shells and have a well-preserved internal foliation (Fig. 4.9E, F). A table containing all strontium isotope results can be found in Appendix A.4. The Asfal Member was unequivocally deposited during the Late Triassic and is therefore younger than ~232 Ma (de Matos, 1997; Maurer et al., 2008; 2015). Using the LOWESS best fit-curve by McArthur et al. (2001), the strontium isotope analysis yields in a numerical maximum age of ~208.3 Ma for the brachiopod sample, which corresponds to the Rhaetian (Fig. 4.10B) (Cohen et al., 2013). The bed containing sample WN 168 Oy was deposited significantly above the bed from which the brachiopods were taken and is thus younger in age. Its measured $^{87}\text{Sr}/^{86}\text{Sr}$ value yields in a minimum stratigraphic age of ~203.8 Ma and a maximum age of ~208 Ma. Due to the shape of the LOWESS best-fit curve an unequivocal age determination using the $^{87}\text{Sr}/^{86}\text{Sr}$ value is not possible around the inferred Triassic-Jurassic boundary (dated at ~201.3 Ma, Cohen et al., 2013). The two samples WN TJ B2 and WN TJ B3, taken from the same bed, show overlapping $^{87}\text{Sr}/^{86}\text{Sr}$ values. The numerical maximum age according to the LOWESS best-fit curve is ~201.5 Ma and the minimum age is ~195.8 Ma. This theoretically implies an age difference between the bed with the two samples WN TJ B and the bed containing WN 168 Oy, separated by one bed, of between 2.3 to 12.2 Myr (Fig. 4.10B). Based on the texture of sample WN 168 Oy showing luminescent cracks this sample is tentatively excluded from the age determination (Fig. 4.9C, D). Using the numerical age derived from the radiogenic strontium isotope analysis alone does thus not allow an unambiguous placement of the boundary.

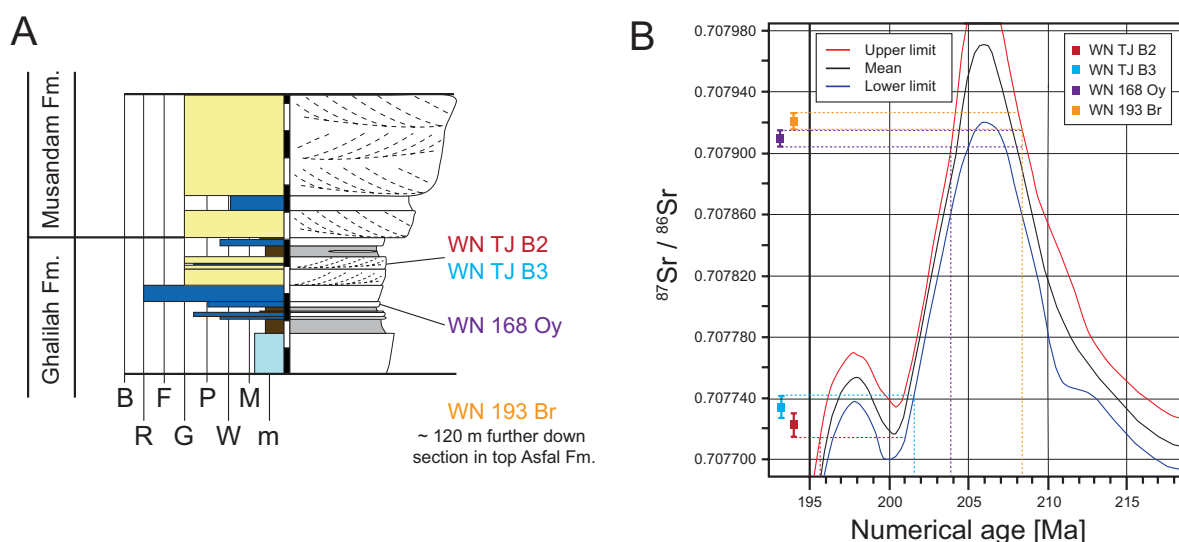


Fig. 4.10 - A) Detailed sample positions selected for Sr isotope analysis from the Wadi Naqab section (see Fig. 4.6 for the facies colour code and Dunham texture abbreviations). **B)** Measured $^{87}\text{Sr}/^{86}\text{Sr}$ data (left side of the plot) used for numerical age determination via the Late Triassic to Early Jurassic part of the LOWESS best-fit curve (McArthur et al., 2001). Values from the LOWESS best-fit curve within the 95 % confidence level are shown.

Despite often being incomplete and prone to diagenetic alteration, the sedimentary sections from this neritic setting contain distinct shifts and has most likely archived the characteristic “initial” negative carbon isotope excursion, the following positive shift and potentially the “main isotope excursion” (Hesselbo et al., 2002) (Fig. 4.11). This is complemented by the biostratigraphic observations made by de Matos (1997) with the last occurrence of Triassic shark teeth and bivalves (see section 4.5 and Fig. 4.6) in the uppermost Ghalilah Formation. The dasycladalean algae described here (Fig. 4.3E, F) could only be identified at the genus level and do therefore not allow for a detailed age assignment. The genus *Paleodasycladus* has however been frequently reported from the Liassic in the western Tethys (e.g. Barattolo and Bigozzi, 1996; Di Stefano et al., 1996). The Triassic-Jurassic boundary interval is therefore placed at the transition zone between the Shuba Member and the Musandam Limestone, above the horizon containing the last Triassic taxa, based on the combined strontium-derived numerical age, the carbon isotope pattern as well as available biostratigraphic markers.

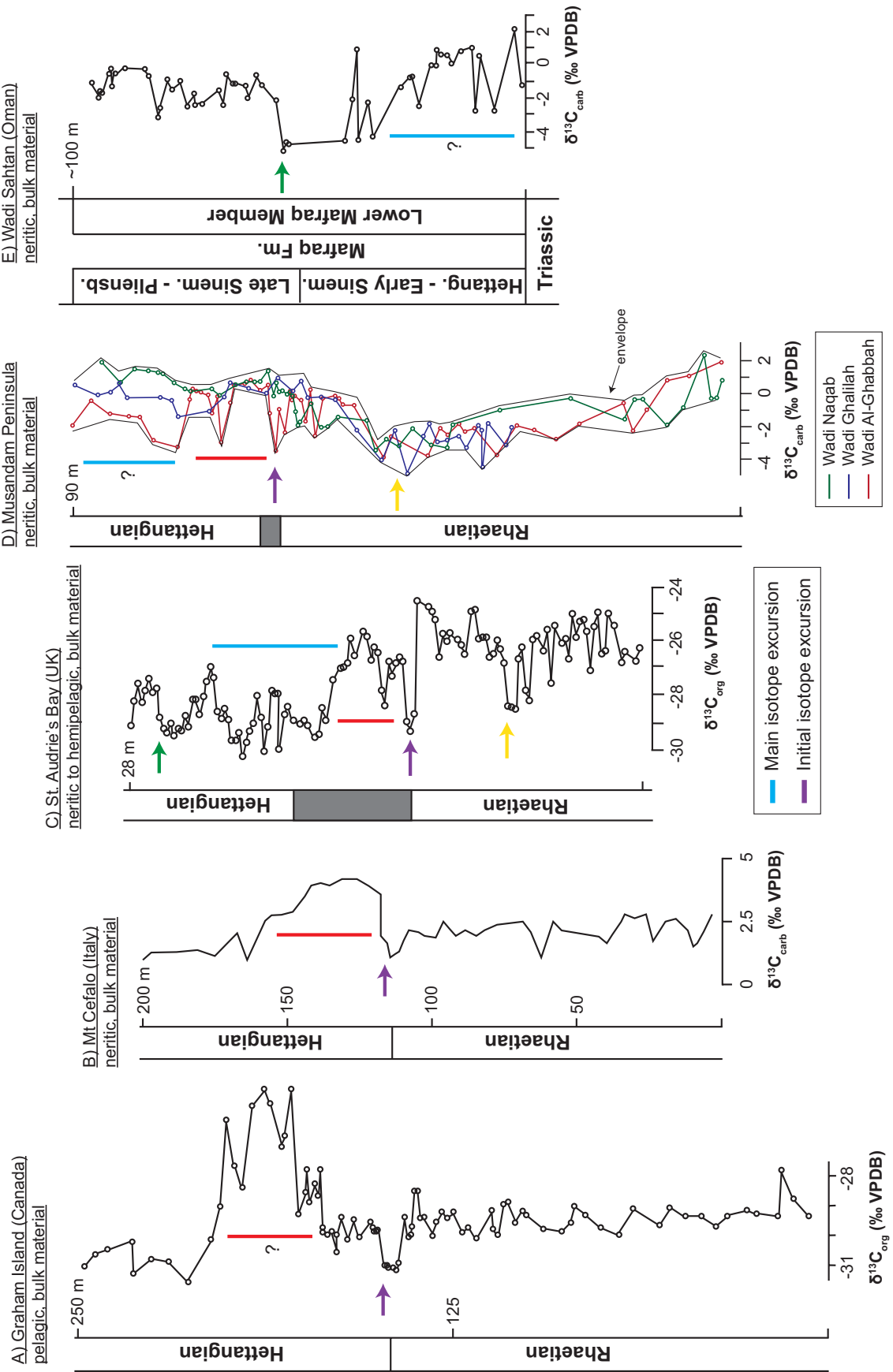


Fig. 4.11 - Comparison of selected $\delta^{13}\text{C}$ reference curves compared to the composite $\delta^{13}\text{C}$ curve from the Musandam Peninsula. Potential correlation patterns are marked with coloured arrows and lines. The “initial” and “main” carbon isotope excursions (Hesselbo et al., 2002) are indicated (A: Williford et al., 2007; B: Bachan et al., 2012; C: Hesselbo et al., 2002; E: Bendias and Aigner, 2015).

4.7 Discussion and implications

4.7.1 Environments of deposition (EOD)

Based on field observations and microfacies analysis, palaeoecological and hydrodynamic interpretations, five main environments of deposition (EODs) were identified: (i) floodplain or tidal flat, to intertidal zone; (ii) restricted, shallow-marine ramp; (iii) high-energy, shallow-marine ramp; (iv) open marine ramp and (v) subtidal middle ramp (Table 4.1). LF 1 and LF 2 were deposited on mudflats or tidal flats within the intertidal, shallow-marine realm. Tepee structures within the microbialite layers and desiccation cracks marking some of the discontinuity surfaces, which frequently cap these facies types, serve as evidence for regular subaerial exposure. The cauliflower-shaped vugs (Fig. 4.3B, C) filled with white calcite cement that are present in some of these beds are interpreted as former evaporite nodules, which were dissolved and later filled with calcite cement. The precipitation of evaporites, such as anhydrite indicates a very shallow, intertidal and at least at times restricted or supratidal environment (e.g. Warren and Kendall, 1985). The occurrence of intertidal fenestral limestones and occasional domal stromatolites, interbedded with subtidal shales and limestones resembles the vertical stacking pattern of Lofer cycles, with the fenestral and microbial limestone facies being typical tidal flat deposits (e.g. Fischer, 1964; Satterley and Brandner, 1995; Enos and Samankassou, 1998). The finely laminated marls and shales (LF 9) are also mud- or tidal flat facies, since they are interbedded with the intertidal facies. The sandstones of LF 8 that contain wavy, and lenticular bedding (Fig. 4.5G) as well as cross-stratification (Fig. 4.5F) were most likely all deposited within the shallow-marine realm or even on mud or tidal flats. LF 3 and 4 contain little or no bioclasts, displaying a low faunal diversity, typically either bivalves, gastropods or echinoderms. LF 4 contains quartz grains, while LF 3 lacks

quartz. These facies types represent a relatively restricted environment, either deposited within the shallow-marine or subtidal open ramp. No sedimentary structures could be observed to allow a more differentiated interpretation. The ooidal grainstones of LF 5 were deposited within the high-energy, shallow-marine realm. The facies contain no micrite and are typically cross-stratified and contain various bioclast fragments, such as oysters and other, unidentified bivalves, gastropods, echinoderms and lithoclasts. The coral bearing facies (LF 7) are attributed to a shallow- to more open-marine, high-energy carbonate ramp, as the corals are frequently associated with ooidal, grainy matrices and were likely to form reefs within the shallow-marine environment. The open marine EOD is dominated by lithofacies associations with diverse fossil assemblages such as pack- and rudstones of LF 6, containing bivalves, gastropods, benthic foraminifera, ostracods, echinoderms, oysters, peloids and coral fragments. The ichnofossils *Rhizocorallium* and *Chondrites* point to a slightly deeper marine EOD, compared to the restricted and fossil-lean deposits of LF 3 and 4. The hummocky cross stratification in one sandstone bed in the top Shuba Member at Wadi Al-Ghabbah is interpreted as being produced by storm-induced currents. These sands were most likely deposited between storm- and fair-weather wavebase. The finely laminated shales and marls of LF 9, often exhibiting nodular bedding and bioturbation (typically *Thalassinoides*) are deposits of a relatively deep, calm environment, and thus represent the deepest deposits within the studied sections.

4.7.2 Sequence stratigraphic framework

The sequence stratigraphic scheme used in this study follows concepts proposed by Strasser et al. (1999). The workflow allows for a descriptive analysis of vertical stacking patterns in shallow environments, especially for mixed, carbonate-clastic systems. The high-resolution analysis groups units into sequences based on vertical facies trends and interpreted bathymetric changes. This has proven useful when age constraints are relatively loose and the exact duration of sequences is difficult or impossible to assess, and has been applied in various studies (e.g. Immenhauser et al., 2004; Amour et al., 2013; Gomez and Astini, 2015). Two different scales of sequences were

recognised within the studied strata based on the cyclical, vertical facies evolution. Small-scale sequences are typically a few metres to tens of metres thick, and the large-scale sequences have thicknesses of maximum 30 m. When the vertical facies evolution indicates deepening-upwards, the sequence is interpreted to record increasing sediment accommodation and a relative sea level rise, while a relative shoaling-upwards sequence is interpreted to have recorded decreasing accommodation and a relative sea level fall. Sequence boundaries were placed at surfaces containing evidence for breaks in sedimentation, such as discontinuity surfaces, some showing evidence for subaerial exposure such as desiccation cracks and palaeokarsts, or where breaks in the trend of relative sea level change are evidenced. The application of the described workflow allows the identification of sequences for the four studied sections as follows (Fig. 4.12): The measured Wadi Naqab section is divided here into seven sequence sets. Each of these sequence sets consists of a deepening and a shoaling-upwards sequence. The top of the Sakhra Member oolites is strongly karstified and thus for example forms a sequence boundary (Fig. 4.7A). De Matos (1997) has previously described the Jurassic succession in Wadi Naqab. The sequence “He1”, corresponding to the basal Musandam Limestone (de Matos, 1997), is approximately equivalent to the top two and a half sequences in the new sequence stratigraphic scheme presented here. The Wadi Ghalilah section consists of six and a half sequence sets. The Wadi Al-Ghabbah section consists of eight complete sequence sets. The basal oolitic unit marks the top of a shoaling-upwards sequence and the uppermost beds of the section mark the beginning of a deepening-upwards sequence. The Jabal Sall Ala section consists of seven sequence sets. Sequences were grouped into larger order sequences in a composite sequence stratigraphic framework for all four studied sections (Fig. 4.12). This results in a shoaling-upwards sequence ranging from the Sumra Member to the top of the Sakhra Member. The overlying Shuba Member and Musandam Limestone are interpreted as a deepening upwards – shallowing-upwards sequence set, with the relatively deepest deposits within the top Shuba Member.

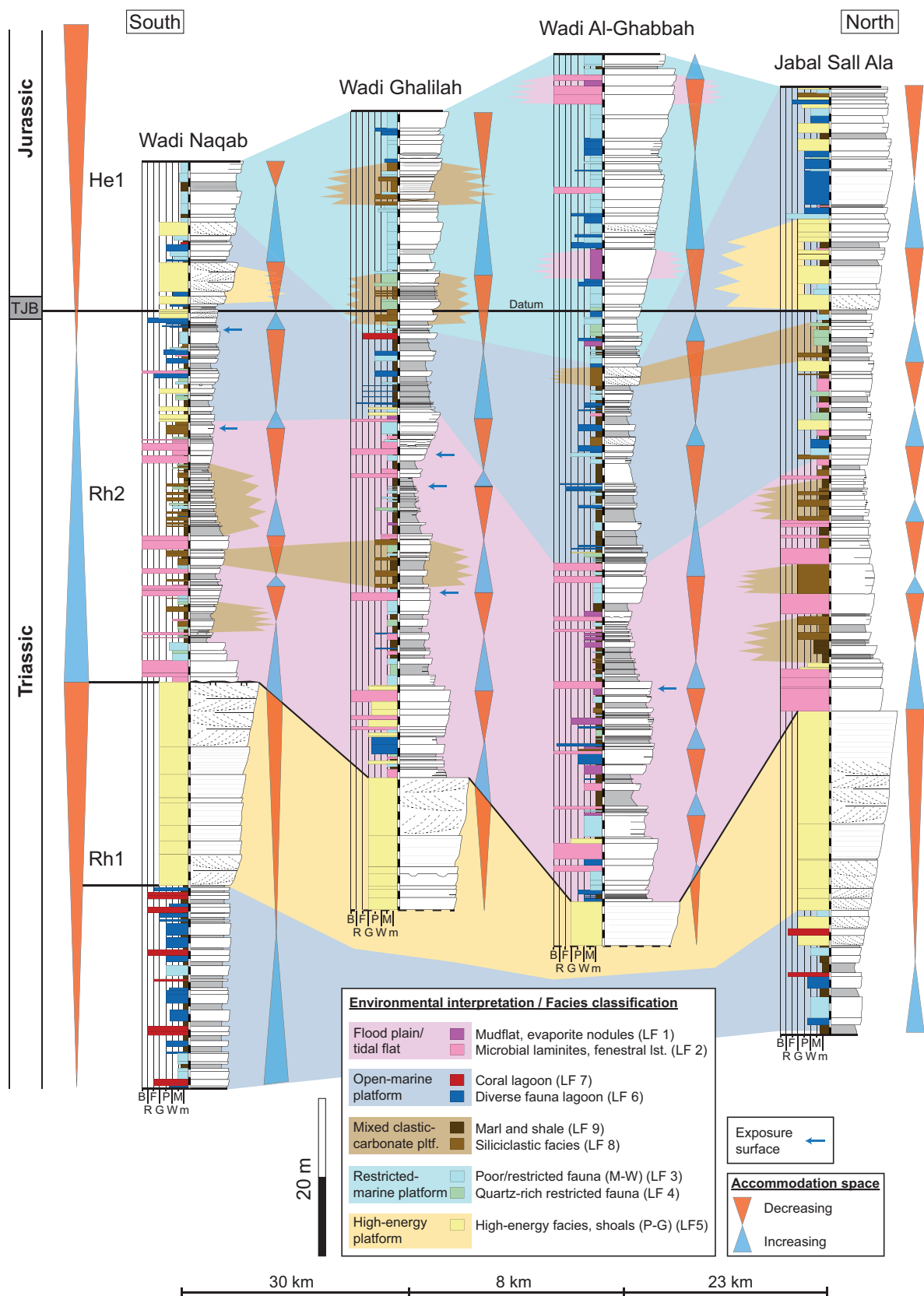


Fig. 4.12 - South to North correlations between environments of deposition across the Musandam Peninsula between the studied sections, based on chemo- and lithostratigraphic

correlations. Medium- and large-scale sequences are indicated (see Fig. 4.6 for the Dunham texture abbreviations).

4.7.3 Regional correlation

The four sections are correlated based on a combined stratigraphic approach. The Wadi Naqab section is used as regional reference section as its chronostratigraphic age has been constrained by bio- and isotope stratigraphy. All three $\delta^{13}\text{C}$ curves show broad negative trends in their lower parts followed by a positive shift with amplitudes of 3.5 – 4 ‰. The relatively heaviest value at Wadi Naqab (WN 3 – WN 4, Fig. 4.6) comes from the basal Musandam Limestone and the Triassic-Jurassic transition zone. The heaviest $\delta^{13}\text{C}$ values following the 3.5 – 4 ‰ shifts at Wadis Ghalilah (WG 2 – WG 3, Fig. 4.6) and Al-Ghabbah (within WA 4, Fig. 4.6) are interpreted to correspond to the heaviest value recorded at Wadi Naqab. The heavy $\delta^{13}\text{C}$ values at Wadis Ghalilah and Al-Ghabbah occur within the basal Musandam Limestone. The Jabal Sall Ala section is correlated to the Wadi Naqab section at the Lower Musandam Limestone. This combined $\delta^{13}\text{C}$ and lithostratigraphy correlation line is used as datum in the regional correlation panel and corresponds to the Triassic – Jurassic boundary horizon (Fig. 4.12).

The chronostratigraphic framework established for the four sections allows for regional correlations and examination of inferred EODs, as deduced from lithofacies associations, as well as the regional correlation of sequences. The broad EODs are continuous across the entire study window. It changes from high-energy, open marine environment to the shallow-marine, intertidal to tidal flat zone. This tidal- or mudflat EOD contains an abundant siliciclastic fraction, especially in the eastern Wadi Naqab and Jabal Sall Ala sections. The environment then shifts to open-marine, represented by diverse bioclastic limestones of the Upper Shuba Member. The Lower Jurassic is marked by open-marine and high-energy facies in the eastern sections while the western sections (Wadis Ghalilah and Al-Ghabbah) contain facies types deposited on a restricted part of the carbonate ramp. Siliciclastic as well as tidal- or mudflat facies are present within this restricted EOD that are however not continuous between the western sections. The overall thickness of the Shuba Member as

well as of the recorded EODs present within that member changes markedly laterally. While the Shuba Member is approximately 50 m thick in the eastern sections (Wadi Naqab and Jabal Sall Ala sections) it is ~ 60 m thick at Wadi Ghalilah and ~ 75 m thick in Wadi Al-Ghabbah. The Wadi Al-Ghabbah section also contains more sequences than the other sections and is therefore regarded as the stratigraphically most complete out of the four sections. The general thickening towards the west could be the result of larger sediment accommodation in the west, potentially induced by an overall extensional tectonic regime, as evidenced by synsedimentary normal faults in a bed within the Shuba Member in Wadi Ghalilah (Fig. 4.7D), or by differential subsidence and the subsequent creation of more accommodation towards the west. Conversely local erosion in the eastern locations could account for the missing cycles. This is supported by the abundance of erosional surfaces at Wadi Naqab. The Jabal Sall Ala section however lacks these surfaces (Fig. 4.6). A locally different palaeotopography with a palaeolow towards the west could be postulated but is at present not possible to assess, as the base of the underlying Sakhra Member could not be measured in Wadi Al-Ghabbah.

4.7.4 Implications for palaeogeography

The broad palaeobathymetric configuration during the Rhaetian and Hettangian of the northeastern Arabian Peninsula shows a deepening towards the (Neo)-Tethyan ocean in the northeast (Fig. 4.13A). The area of the Musandam Peninsula was part of a large bay, covered by a shallow-marine carbonate platform, with shallower clastics and evaporites being deposited in more proximal settings towards the Arabian Craton (Al-Husseini, 1997; Ziegler, 2001). Based on the findings of this study, a differentiation between the Late Triassic and the Early Jurassic is necessary as the EOD distribution markedly changed through time (Fig. 4.12). The Shuba Member contains multiple exposure surfaces as well as some evaporite beds in its lower half across the entire study area. The shallow-marine clastic and evaporite EOD must have extended at least to the northern end of the Musandam Peninsula during parts of the Rhaetian (Fig. 4.13B). This is supported by facies descriptions acquired from cores onshore and offshore of Abu Dhabi, where clastic deposits have been

reported (Loufti and Sattar, 1987; Hassan, 1989). This shallow-marine clastic and evaporites EOD potentially extended as far as Iran during the Late Triassic according to facies descriptions from Alavi (2004). Since equivalent subsurface data from the Persian Gulf is scarce, this remains a hypothesis at present. The sedimentary record from the Musandam Peninsula is proposed as being more complete than that of the western Arabian Peninsula (Maurer et al., 2008) and a Late Triassic – Early Jurassic stratigraphic gap is frequently mentioned (e.g. Sharland et al., 2001). The Triassic and Jurassic facies distribution on the Musandam Peninsula is more heterogeneous than that of the underlying Shuba Member. It comprises oolitic dunes in the eastern sections and sandbodies that are in one case connected between the logged sections. Evaporites as well as an abundant siliciclastic component are present in the western sections (Wadis Ghalilah and Al-Ghabbah). The broad EODs are however continuous across the study area and switch to a shallow- to open marine carbonate system, containing no evidence for subaerial exposure. The western sections however contain more restricted EOD facies types (Fig. 4.12). The palaeogeographic map for the Early Jurassic is thus similar to previously published maps (Fig. 4.13C). A closer look at the EOD and facies distribution reveals heterogeneity at the scale of tens of kilometres, similar to what has been previously established for the hundreds of metre to kilometre scale in shallow-marine carbonates from Middle Jurassic strata on the Musandam Peninsula (Hönig and John, 2015).

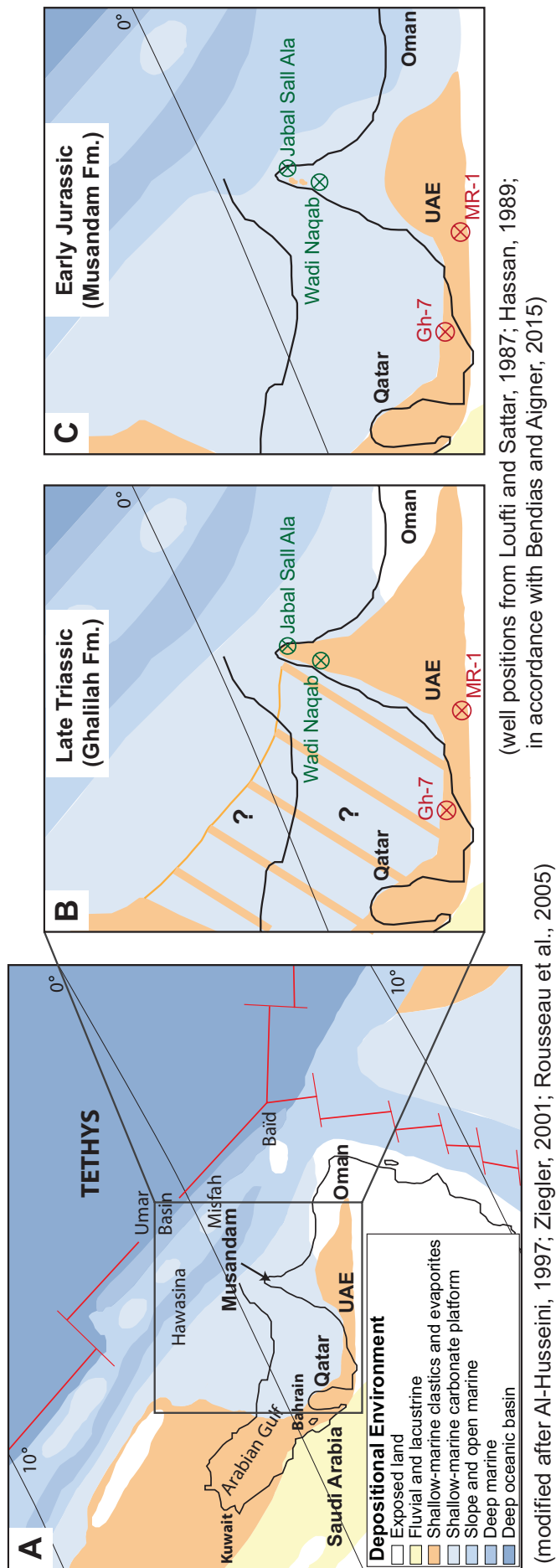


Fig. 4.13 - A) Palaeogeographic map of the Arabian Peninsula for the Late Triassic - Early Jurassic interval. **B)** Detailed and refined palaeogeographic map of the north-eastern Arabian Peninsula, incorporating published well data from Abu Dhabi during the Late Triassic. **C)** Detailed and refined palaeogeography during the Early Jurassic.

4.7.5 Controls on the regional development of the platform

The overall EOD across the study area shifts from a shallow-marine carbonate system with diverse facies associations containing corals, different bioclasts and ooids to a microbialite-dominated mud- or tidal flat with abundant siliciclastic material and frequent exposure events, back to a deeper, more open-marine carbonate system, which contains only one local exposure surface. The demise of the ooid-dominated high-energy carbonate ramp, followed by an intertidal microbialite-dominated system was either induced by changes in oceanographic conditions or by a drop in relative sea level (James, 1997; Pomar and Hallock, 2008). The top of the Sakhra Member oolite at Wadi Naqab is marked by a karstified surface, which favours the model of a major relative sea level drop followed by a hiatus lasting long enough to form this karstified surface. There is however no evidence for karstification of the top Sakhra present within the other sections, despite the drastic facies and EOD change. At Wadi Ghalilah one horizon within the Sakhra Member oolite exhibits a strongly mottled surface as well as a palaeokarst. An abrupt drop in relative sea level, followed potentially by an omission phase is postulated. Restricted oceanographic conditions favouring microbialite growth, such as increased salinity (e.g. Vennin et al., 2015 and references therein) might have prevailed locally on the extensive shelf area. The Shuba Member contains bioclastic and ooidal limestones in its lower part in the western sections (Wadis Ghalilah and Al-Ghabbah), so restricted conditions could have not prevailed across the entire shelf initially, following the relative sea level drop causing the demise of the Sumra and Sakhra Member carbonate system. Environmental stress might have increased during the deposition of the Shuba Member, favouring microbialite growth.

The second switch, back to a healthy carbonate system and slightly preceding the Triassic-Jurassic boundary is the result of a relative sea level

deepening, evidenced by deeper, more open-marine facies types as well as the absence of regional subaerial exposure within the Upper Shuba Member and the Musandam Limestone. It could have only been caused by either an increased subsidence rate or by a eustatic rise.

Two orders of cycles have been recognised within the studied strata. An even higher-frequency cyclicity in the Liassic of Wadi Naqab as well as the nearby Oman Mountains was previously interpreted as the result of orbital forcing (de Matos, 1997; Walkden and de Matos, 2000; Bendias and Aigner, 2015). The observed medium- and large-scale cycles also represent changes in relative sea level, most prominently marked by various exposure stages, of which some from within the Lower Shuba Member are continuous across the entire region (Fig. 4.12). Although autocyclic variations might have played a role in the production of the stratigraphic pattern, they are not able to produce long lasting stages of exposure (Burgess, 2001). The inferred oscillations of relative sea level on the investigated scale were either induced by eustasy or tectonic processes. Although the eastern margin of the Arabian Plate is considered a passive margin during much of the Mesozoic (Searle, 1988; Glennie, 2005), evidence for synsedimentary extension is present within the Shuba Member (Fig. 4.7D). An overall extensional regime with multiple subsiding fault blocks provides an explanation for the larger accommodation available in the north-western part of the study area, as evidenced by overall larger cycle thicknesses within the Wadi Al-Ghabbah section. This would have furthermore resulted in a variable palaeotopography across the shelf.

The tidal flat, the open-marine and the restricted marine EODs contain lenses of marl and shale deposits that are mostly discontinuous over the entire study area (Fig. 4.12). This has also been reported for the subsurface of the emirate of Abu Dhabi. The age-equivalent reservoir formations that lie in the immediate proximity in the UAE are the Minjur and Marrat Formations onshore, and the Gulailah, Hamlah and Izhara Formations offshore. The published sedimentological data on these units is still relatively scarce. Marls and shales are especially abundant within the Marrat Formation, where gamma ray peaks often cannot be correlated over more than a few tens of kilometres (Taher et al., 2012). The shale beds are especially prominent within the shallow, restricted EOD in the Lower Shuba Member (Fig. 4.12). Clastic material is transported

from the hinterland and the interfingering marl and shale deposits are thus a function of clastic supply onto the carbonate-dominated ramp and the availability of accommodation with calm hydrographic conditions, especially to deposit the fine grained sediments. This clastic supply from the hinterland could be induced by increased continental runoff at times. Consequently climatically induced changes could also have played an important role in the distribution of clastic deposits on the carbonate ramp. The coarser silt- and sandstones contain evidence for both current, possibly tidal, and occasional storms crossing the shelf. It is postulated that these coarser grained and well-sorted sandbodies would have been shifted across the shelf by different hydrodynamic processes and deposited wherever hydrodynamic and bathymetric conditions favoured deposition. The coarser material can be found in the studied sections irrespective of the interpreted relative sea level, as they are also present within the more open-marine Jurassic strata, unlike the shales and marls. A climatically induced clastic input, i.e. a switch from humid to arid conditions, would have resulted in the entire termination of clastic input. The finer clastic components during the Late Triassic and Early Jurassic were most likely transported to more distal parts of the shelf, where hydrodynamic quiescence prevailed. The Lower Jurassic deep-sea record of the Eastern Oman Mountains contains turbidite successions and silt- and sandstones, supporting the assumption of constant siliciclastic runoff (Blendinger, 1988; Blechschmidt et al., 2004).

4.7.6 Responses to global changes

Observing evidence for an ocean acidification event is however difficult, especially in deep time (Greene et al., 2012). The global carbon cycle perturbations, in the form of stable carbon isotope excursion around the Triassic-Jurassic boundary, have been recorded by the shallow carbonate sections of the Musandam Peninsula (Fig. 4.11). The overall sedimentation is mostly continuous across the Triassic-Jurassic boundary interval on the Musandam Peninsula. The Wadi Naqab section is an exception, with a local subaerial exposure surface just below the Triassic-Jurassic transition marked by a very condensed interval. Carbonate sedimentation remains largely constant

across the boundary and the studied sections provide no evidence for a biocalcification crisis. A reduced carbonate content or increase in the amount of siliciclastic material, as reported from coeval continuous carbonate successions around the Triassic-Jurassic boundary in Eastern Europe (Pálffy et al., 2001; Rožič et al., 2009; Korte and Kozur, 2011) and Western Europe (e.g. Felber et al., 2015) could also not be observed on a regional scale.

The major carbonate system changes observed within the studied section are most likely caused by changes in relative sea level. The analysed sequences allow for a construction of long and a short-term relative sea level curves (Fig. 4.14). The constrained chronostratigraphic age allows for a comparison with other sea level curves to establish potential eustatic controls. The Triassic-Jurassic transition broadly falls into a period of eustatic sea level rise (Fig. 4.2). This is also the case for the Musandam Peninsula. Eustatic changes are thus a potential mechanism to produce the vertical stratigraphic architecture. It is noteworthy that both main observations (no evidence for a biocalcification crisis and a relative sea level highstand) associated here with the Triassic-Jurassic boundary at Wadi Naqab would also be valid if the Triassic-Jurassic boundary lies at the transition between the Sumra Member and the Sakhra Member, despite a marked carbonate facies change at this transition (Fig. 4.6, 4.12).

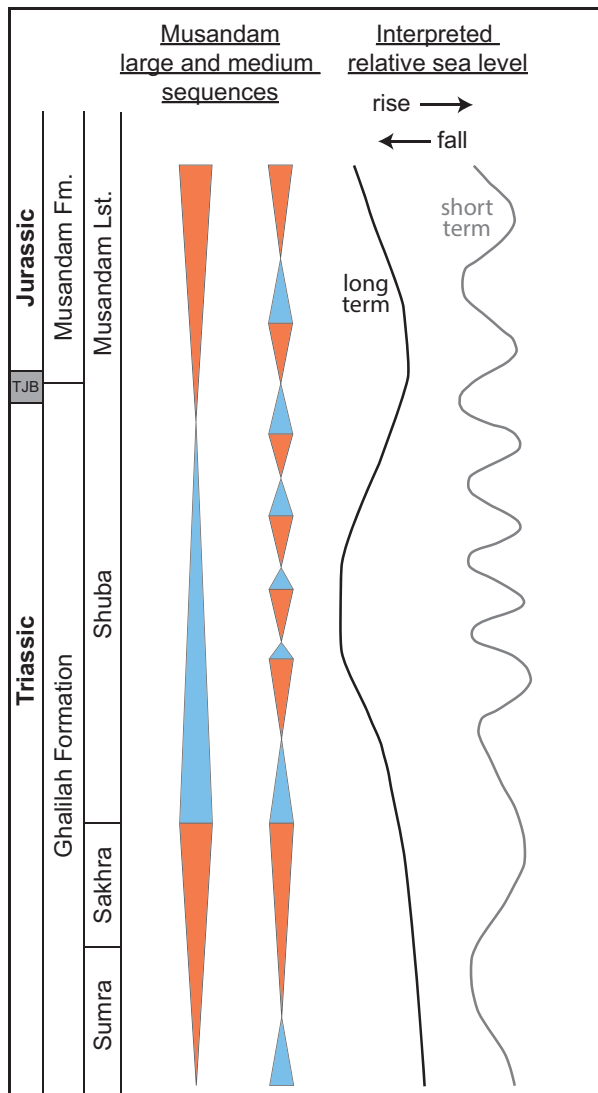


Fig. 4.14 - Chronostratigraphic framework from the Musandam Peninsula with a relative sea level reconstruction, based on the interpreted accommodation space changes.

4.8 Conclusions

A shallow-marine equatorial mixed, siliciclastic carbonate ramp was studied on a regional scale on the eastern Arabian Platform. Strontium isotope analysis on well-preserved oysters and brachiopods allows the placement of a Triassic-Jurassic boundary horizon, while acknowledging previous biostratigraphic studies. The studied system changes on a regional scale from an open-marine, diverse-fauna carbonate ramp, to an intertidal microbialite dominated system with abundant siliciclastic material and regional exposure stages during the Late Triassic. The Latest Triassic is marked by a revival of an open-marine, diverse-fauna carbonate ramp, which succeeds into the Early Jurassic. All sediments

were deposited within the neritic part of the shallow-marine ramp. The sections thus serve as shallow-marine carbonate archive, which is regionally continuous across the Triassic-Jurassic boundary. A combination of stable carbon isotope stratigraphy and lithostratigraphy allows the regional correlation between the different study sites. The acquired carbon isotope record furthermore correlates well to global coeval reference curves, which highlights the capability of neritic carbonate successions in the Middle East to archive global environmental changes, and merits further exploration.

The findings of the study have led us to modify the regional palaeogeographic map of the region in accordance with available subsurface data. This implies that especially the distribution of shallow-marine, clastic facies types during the Late Triassic could extend further north in the region than previously thought. The findings of this study thus might provide information on the regional distribution of potential reservoir and/or seal facies distribution.

The observed vertical carbonate system changes on the Musandam Peninsula were most likely caused by eustatic sea level changes. Lateral heterogeneities were governed by hydrodynamic action, including occasional tropical storms, on the extensive shelf, as well as the local palaeotopography. A comparison between the constructed regional sea level curve and global eustatic sea level curves reveals that the studied system was controlled by eustatic changes, accompanied by a locally extensional tectonic regime. The ramp development across the Triassic-Jurassic transition was governed by a transgression, as observed at other Tethyan localities. No clear evidence for a biocalcification crisis marking the Triassic-Jurassic boundary, potentially induced by a global acidification event, is present regionally within the studied succession.

5. Climatic and environmental changes across the Triassic-Jurassic boundary recorded on an equatorial carbonate ramp

5.1 Abstract

The Triassic-Jurassic boundary is marked by one of the 'big five' mass extinctions of the Phanerozoic. Events that lead to the mass extinction are debated but are likely linked to coeval rapid volcanic exhalation of CO₂ and other associated volatiles, from the Central Atlantic Magmatic Province. This can be viewed as analogue for modern atmospheric CO₂ increase. Ocean acidification has been suggested as the main kill mechanism of the mass extinction in the marine realm. Here an equatorial shallow-marine record across the Triassic-Jurassic boundary is examined to assess events associated with the Triassic-Jurassic boundary. The results indicate that CO₂ levels were elevated during the Latest Triassic potentially accompanied by elevated atmospheric temperatures and an equatorial relative sea level rise. It is speculated that this is a reflection of a eustatic rise caused by elevated atmospheric temperatures, induced by an enhanced greenhouse climate.

5.2 Introduction and aims

The Triassic-Jurassic (T-J) boundary, at ~201 Ma is marked by one of the big five mass extinction events in Earth's history. The mass extinction was potentially caused by an ocean acidification event linked to an increase of atmospheric CO₂ levels (Hautmann et al., 2008; Greene et al., 2012, Hönisch et al., 2012; Ikeda et al., 2015). The extinction during the Latest Triassic was gradual in nature, as opposed to a relatively instantaneous disappearance of taxa (Hallam, 2002). The intensity and the rapid release of CO₂ can be viewed (with caution, Kemp et al., 2015) as a promising analogue for modern atmospheric CO₂ increase and stresses the importance of studying the associated driving mechanisms, similar to the Palaeocene Eocene Thermal Maximum (PETM).

The mass extinction was preceded by a carbon cycle perturbation as indicated by negative excursions in organic and inorganic carbon isotope

records worldwide during the Latest Triassic (Hesselbo et al., 2002). This perturbation of the carbon cycle was accompanied by the eruption of the Central Atlantic Magmatic Province (CAMP), producing large amounts of CO₂ and associated volcanic volatiles (Cohen and Coe, 2002; Marzoli et al., 2004). Although the exact amount of released CO₂ remains unclear, roughly a doubling of atmospheric CO₂ within a span of 10 to 20 kyrs has been suggested (Ruhl et al., 2011). The large increase in CO₂ has also been inferred from stomatal density and the stomatal index of fossil leaf cuticles (McElwain et al., 1999). It has previously been suggested that the potential ocean acidification coupled with increased pCO₂ at the T-J boundary acted as the main “kill mechanism”, in particular affecting carbonate reef-building organisms (Hautmann et al., 2008; Greene et al., 2012; Hönisch et al., 2012; Ikeda et al., 2015). Direct evidence for an ocean acidification event is however difficult to gather in deep time (Greene et al., 2012). The Permian-Triassic mass extinction is tied to an ocean acidification event, as documented in the marine realm through the use of boron isotopes in bulk carbonates as proxy for ocean water pH (Clarkson et al., 2015), and in the terrestrial realm through the use of vanillic acid to vanillin ratios as proxy for acidification (Sephton et al., 2015). Contrasting evidence from pedogenic calcites across the T-J boundary however, shows no significant increase of atmospheric CO₂ coeval with the T-J mass extinction (Tanner et al., 2001). Alternative models to the ocean acidification event include global warming triggered by other volcanic gases and potentially amplified by a release of CH₄ from gas hydrates ultimately causing the T-J mass extinction (McElwain, 1999; Tanner et al., 2001; Kiessling and Simpson, 2011; Hönisch et al., 2012). Well-studied sections covering the T-J boundary in the western Tethyan realm commonly contain a marl unit following the Late Triassic extinction, especially prominent at the Kendelbachgraben type locality in Austria (“Grenzmergel”) that has frequently been explained as a potential result of a carbonate factory crisis (e.g. Hallam and Goodfellow, 1990; Hesselbo et al., 2007; Kuerschner et al., 2007; Korte and Kozur, 2010). It can be postulated that ocean acidification as the main “kill mechanism” would have affected the entire ocean system, including environmentally sensitive biota in the tropical, equatorial realm. Recent research has added evidence for enhanced volcanic activity through elevated mercury levels across the T-J boundary but questions

ocean acidification as (sole) cause for the wide spread biotic crisis (Thibodeau et al., 2016). Equatorial perspectives on the T-J boundary from the Tethys are however currently scarce. The only other published equatorial archive, situated on the northern flanks of the Himalayas, experienced continuous shallow-marine sedimentation across the T-J boundary and contains evidence for sea level rise starting during the Late Rhaetian and continuing into the Early Jurassic (Yin et al., 1999; Hallam et al., 2000).

In this study a record of the Triassic-Jurassic transition from a low-latitude (tropical) succession of limestones deposited during the Triassic-Jurassic transition is presented. Exploring the response of low-latitude carbonates to the T-J events is crucial, as low-latitudes are less likely to experience a significant warming during a greenhouse episode. The equatorial region experiences the highest amount of solar radiation and is the warmest area on the planet. It hosts biotic communities that are commonly sensitive to environmental changes. If an ocean acidification was the prevailing kill mechanism causing the T-J mass extinction, evidence for this should be visible within the equatorial sedimentary record. On the contrary evidence for atmospheric and ocean warming would be less pronounced at low latitudes due to generally warmer background temperatures.

5.3 Study site and stratigraphic constraints

The studied section is located in the northern Emirate of Ras Al-Khaimah in the Musandam Mountains. The section is located on the northern side of Wadi Naqab, 13 km southeast of the city Ras Al-Khaimah (GPS coordinates of the log base: N 25°43'5.7"; E 56°05'14.1"). The section was described in detail for its sedimentology and depositional environment (Hönig et al., *under revision*). The sediments were deposited on an area of an expansive epeiric ramp that was situated within a few degrees of the equator (Fig. 5.1). The western Arabian Peninsula experienced a generally summer-wet, tropical climate during the Late Triassic and Early Jurassic with high mean surface air temperatures, moderate to high levels of precipitation (Sellwood and Valdes, 2006; 2008) and a medium to high degree of annual cloud cover (Sellwood and Valdes, 1997). The region is furthermore listed as a site that might have been close to coastal

upwelling zones (Price et al., 1995). The position of the T-J boundary interval was refined using a multi-proxy approach including chemostratigraphy (radiogenic strontium isotopes and stable carbon isotopes) as well as biostratigraphy (Hönig et al., *under revision*). Although there is an on going debate regarding the exact position of the T-J boundary on the Musandam Peninsula (Maurer et al., 2008; Maurer et al., 2015; Al-Suwaidi et al., 2016) the boundary interval is placed here at the transition between the upper part of the Shuba Member and the Lowest Musandam Limestone (Fig. 5.2). As the stable carbon isotope record and the biostratigraphic framework do not allow an unambiguous placement of the boundary, its position chosen here is based largely on the available biostratigraphic markers and the radiogenic strontium isotope data.

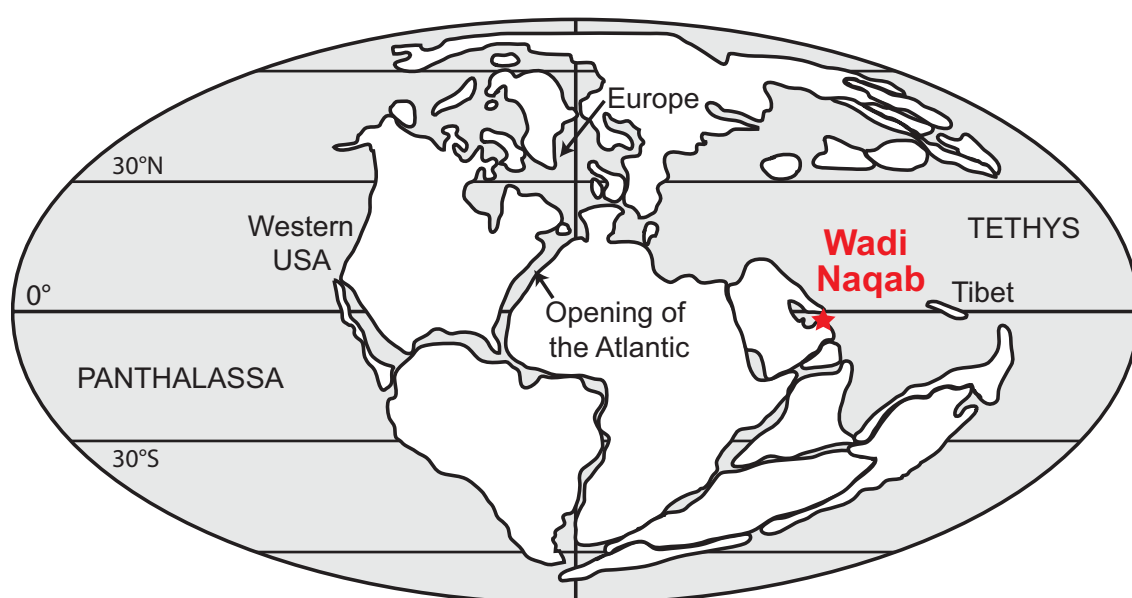


Fig. 5.1 – Global plate tectonic map across the Triassic-Jurassic boundary (modified from Tanner et al., 2004; Golonka, 2007).

5.4 Methods

5.4.1 Carbonate stable oxygen and carbon isotopes

Measurements of stable isotopes of carbon and oxygen were performed in the Qatar Stable Isotope Laboratory at Imperial College London. The measurements were performed on micrite where possible, and on bulk samples

for grainy facies, as micrite is the most common carbonate phase in the sample set and thin section analysis could be used to check the purity of the micrite. A total of 78 sample powders were obtained using an electrical dental drill, carefully avoiding veins, fossils and recrystallized sections. Approximately 100 to 150 μg of sample powder were dissolved with 105 % orthophosphoric acid at 70 °C in a Kiel IV carbonate device, and the resulting CO_2 gas was measured on a Thermo MAT 253 mass spectrometer. Data were corrected according to measurements of the international standard NBS 19 and an in-house standard (ICM, Imperial College Carrara Marble). Stable isotope values are reported using the standard δ notation relative to the VPDB (Vienna Pee Dee Belemnite) standard. External precision for ICM (one standard deviation) on carbon and oxygen is better than 0.05 ‰ and 0.08 ‰, respectively. The precision for duplicate and triplicate sample measurements (one standard deviation) is better than 0.05 ‰ for carbon and better than 0.14 ‰ for oxygen.

5.4.2 Carbonate content and organic carbon isotope analyses

A total of 47 powdered bulk rock samples were decarbonated using HCl (1M) prior to the analyses. Subsequently samples were rinsed with H_2O to remove residual acid until neutrality was reached. The carbonate content was calculated by subtracting the weight of the residue from the weight of the original sample powder. The organic carbon isotope measurements were performed at the laboratory of the Leibniz University of Hannover (Germany) on an organic elemental analyser (Thermo Scientific Flash 2000) connected to a mass spectrometer (Thermo Scientific Delta V Advantage MS). The analytical accuracy and reproducibility (better than $\pm$ one SD 0.06 ‰) were checked by measuring replicates as well as in-house standards and the international standard NBS 22. Results are reported using the standard δ notation relative to the VPDB (Vienna Pee Dee Belemnite) standard.

5.4.3 Phosphorous content

The analysis was performed at the laboratory of the University of Lausanne, Switzerland on 28 samples. For this method bulk rock samples were carefully

screened to avoid altered parts and veins and were then crushed using an agate mortar. For the phosphorous content analysis, approximately 100 mg of each sample were used. The powder was mixed with 1 ml of MgNO_3 and then the mixture was dried in an oven for two hours at 45°C . Afterwards the samples were ashed in a furnace at 550°C for two hours. The samples were then left to cool before being mixed with 10 ml of 1N HCl in reaction tubes. These tubes are then shaken for fourteen hours. The produced solutions were afterwards filtered with a $63\ \mu\text{m}$ filter and diluted with 90 ml of deionized water. This diluted mixture was then analysed using the ascorbic acid method, which is described by Eaton et al. (1995). For this analytical method the produced solution was mixed with ammonium molybdate and potassium antimonyl tartrate. This was reacted in an acid medium with orthophosphate to form phosphomolybdic acid. This acid was then reduced with ascorbic acid to form an intense blue colour, the intensity of which is controlled by the Phosphate (PO_4) content of the sample. The colour intensity is measured with a photospectrometer (Perkin Elmer UV/Vis Photospectrometer Lambda 10). Using known standard solutions to calibrate the photospectrometer, the PO_4 concentration of each sample was then determined.

5.4.4 Clay mineralogy

Samples from the same 28 beds used for the phosphorous content analysis were analysed for their clay mineralogical content using X-ray diffraction (XRD) at the laboratories of the University of Lausanne (Switzerland), following the procedures described by Kübler (1987). Samples were air-dried at room temperature and kept overnight in an ethylene-glycol saturated atmosphere. The produced diffractograms were analysed using the software MacDiff 4.2.5 (Petschick, 2000). Quantifications were obtained by measuring the peak heights of the (001) peaks. The clay minerals that were investigated include micas (characteristic peak around $10\ \text{\AA}$), chlorites (characteristic peak around $14.2\ \text{\AA}$), kaolinite (characteristic peak around $7.18\ \text{\AA}$) and smectites (characteristic peak between $15\text{--}20\ \text{\AA}$). As the $7.18\ \text{\AA}$ peak contains both the kaolinite (001) peak and the chlorite (002) peak, the ratio between the chlorite (004) and the kaolinite (002) peaks at approximately $3.52\ \text{\AA}$ and $3.58\ \text{\AA}$ respectively, was

calculated. Clay minerals are reported in relative proportion to each other in order to interpret their relative abundance through time.

5.4.5 Electron microscopy

Two clay-rich samples were selected for scanning electron microscopy (SEM) analysis in order to examine the morphology of the clay minerals. This allows the differentiation between detrital and authigenic clay minerals. The analysis was performed in the Department of Materials at Imperial College London, using a JEOL JSM 6400 scanning electron microscope, fitted with an Oxford Instruments INCA energy dispersive analytical system (EDS) that allows semi-quantitative and qualitative elemental analysis. The thin section samples that were imaged were covered with a gold coating prior to the analysis.

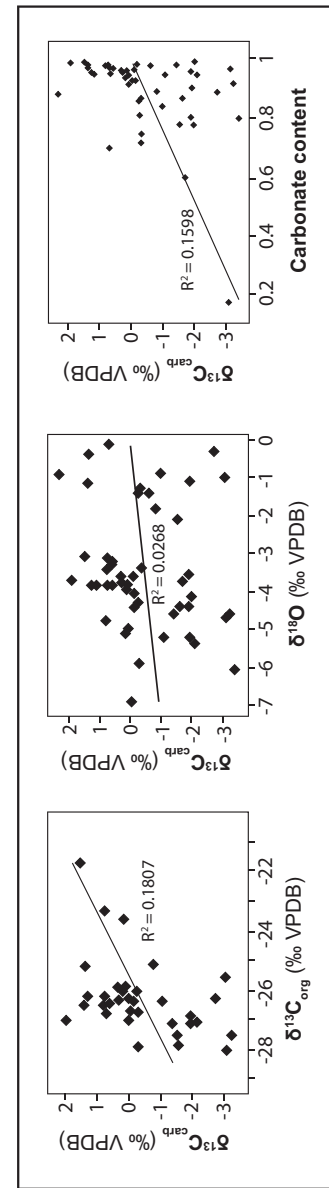
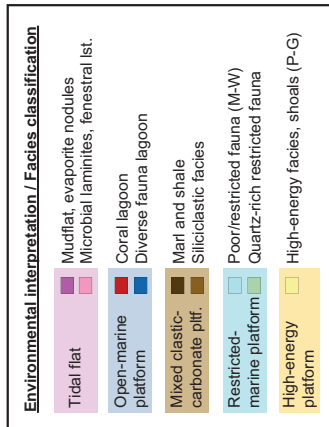
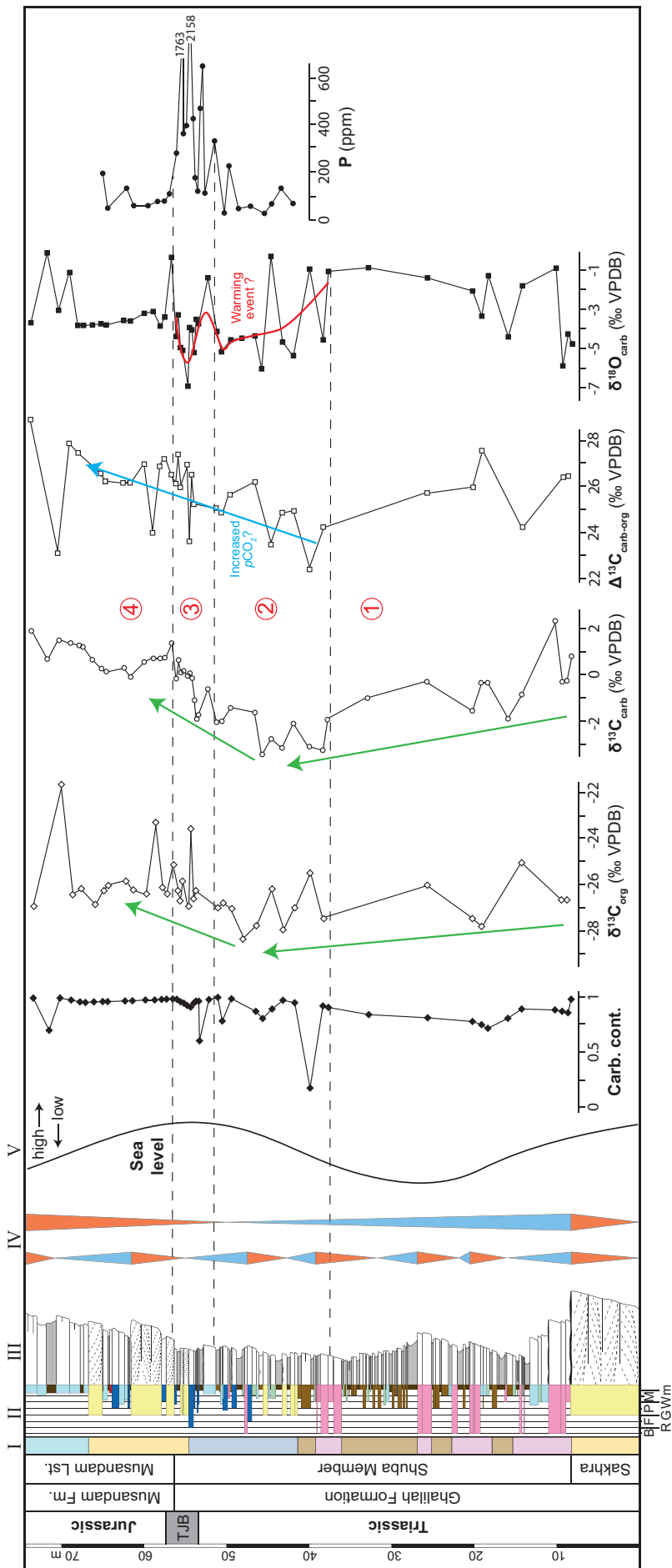


Fig. 5.2 – Stratigraphy, environment of deposition (I), Dunham texture (II), schematic weathering pattern (III), development of accommodation (IV, blue triangle=increasing accommodation, red triangle=decreasing accommodation), interpreted relative sea level curve (V) across the Triassic-Jurassic boundary at Wadi Naqab, coupled with geochemical data.

5.5 Results

The base of the logged section is formed by a thick cross-stratified ooidal grainstone (top Sakhra Member). This oolite is capped by a red-stained and mineralized hardground. The overlying Lower Shuba Member is approximately 45 m thick. Its lower part is dominated by marls and shales, which are interbedded with siliciclastic-rich limestones and silt- to fine sandstone beds. Within this siliciclastic-rich interval 6 intervals of microbial-fenestral limestone are present. The upper part of the Shuba Member (~ 15 m thick) comprises mainly bioclastic and ooidal limestones, interbedded with marls. The overlying basal Musandam Limestone contains cross-stratified ooidal grainstones as well as bioclastic mud- and wackestones.

The carbonate content of the sampled beds is overall high and shows a general increase from the base to the top of the section from ~80 % to almost 100 %, with four samples having a lower carbonate content and not following this overall trend. The stable carbon carbonate isotope curve ($\delta^{13}\text{C}_{\text{carb}}$) contains values between -3.5 and +2.5 ‰. Values broadly decrease up section in the lower half of the curve and then increase towards the top. Carbon and oxygen isotope values show no significant correlation. The carbon isotope curve measured on the organic carbon ($\delta^{13}\text{C}_{\text{org}}$) fluctuates between -28 ‰ and -22 ‰. It resembles the $\delta^{13}\text{C}_{\text{carb}}$ curve as values slightly increase in the lower part of the curve to a minimum of -28 ‰ (Fig. 5.2 (1,2)) before broadly increasing towards the top. The upper part of the curve that shows the overall increasing trend with an amplitude of ~1.5 ‰, contains three sharp positive excursions, each marked by one data point, with amplitudes of 2 to 4 ‰. The broad trends of carbonate carbon and organic carbon isotopes are similar and the data points show some degree of correlation ($R^2 = 0.1807$). The broad trend of $\delta^{18}\text{O}$ values shows a broad rising to -1 ‰ at 35 m up section (Fig. 5.2 (1)), followed by a broad shift to more depleted values around -4 ‰, although containing large

scatter, at the T-J boundary (Fig. 5.2 (2,3)). The curve then remains relatively stable at -3.2 ‰ before displaying marked shifts in the upper most part of the section (Fig. 5.2 (4)). Carbon and oxygen isotope values show no significant correlation, limiting the possibility of a diagenetic overprint. The phosphorous (P) content curve covering the boundary interval shows constant values below and above the boundary between 50 and 200 ppm but displays a sharp increase (from ~50m up section), beginning just before the boundary, to over 2150 ppm (Fig. 5.2 (3)).

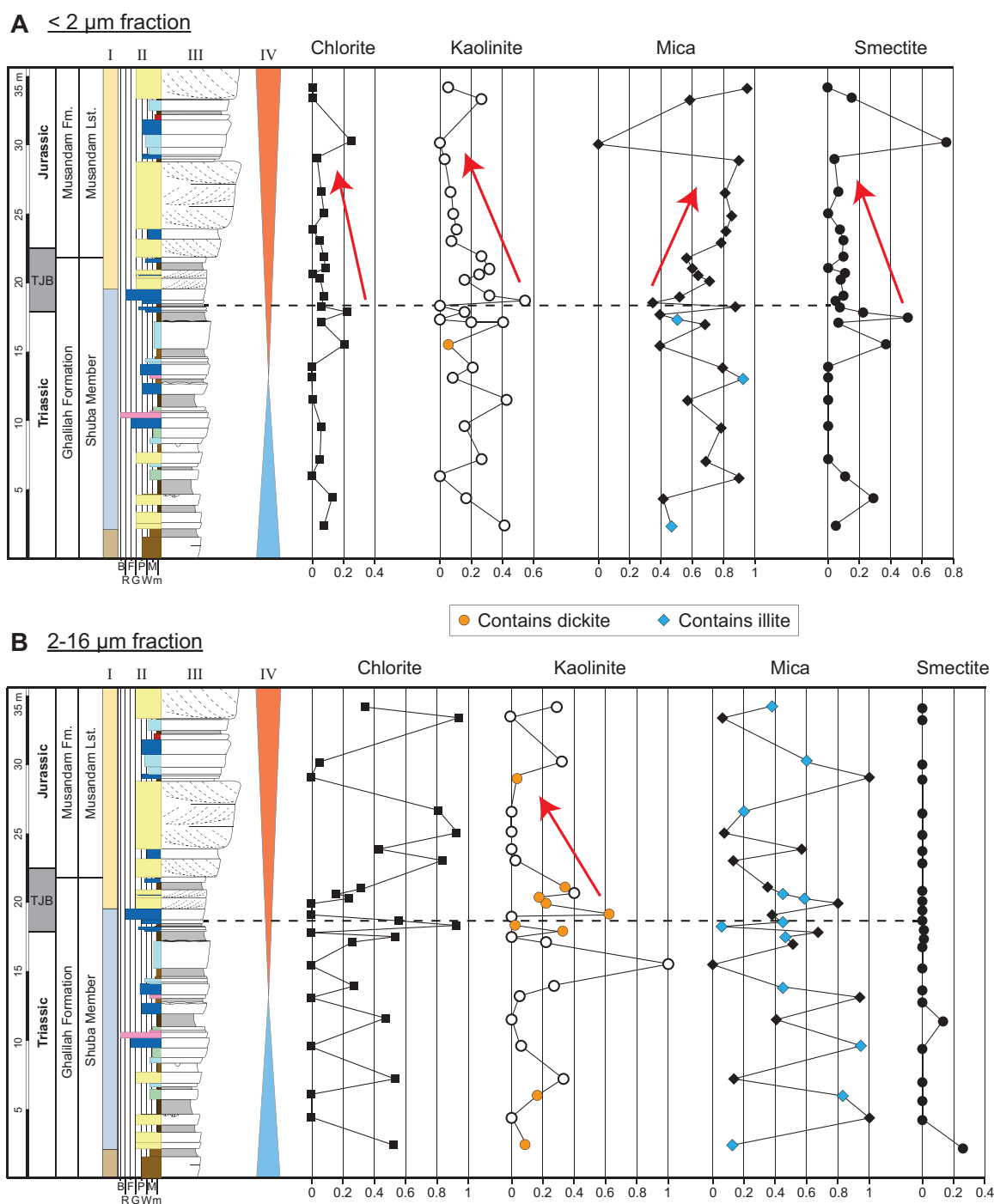


Fig. 5.3 – Clay mineral content across the Triassic-Jurassic boundary at Wadi Naqab, interpreted from X-ray diffractograms (refer to Fig. 2 for I-IV and facies colour codes).

The abundances of four main clay mineral groups (chlorite, kaolinite, mica and smectite) were normalized to examine the relative abundances and the proportions of these mineral groups across the T-J boundary (Fig. 5.3). Diffractograms for the $< 2 \mu\text{m}$ and the 2 to $16 \mu\text{m}$ grain size fraction were analysed separately. The $< 2 \mu\text{m}$ grain size fraction (Fig. 5.3A) is dominated by

micas (mostly 40 – 100 %). Chlorites and smectites fluctuate between 0 and 25 %, with the three exceptions where smectites make up more than 40 % of the small clays. The kaolinite content fluctuates between 0 and 55 %. It is highly variable in the basal part of the Shuba Member with a maximum in its upper part (55 %). Values then increase steadily across the Upper Shuba Member and the Musandam Limestone. Corresponding chlorite and smectite contents are likewise decreasing in the upper half of the section, while the mica content is increasing. The abundances of chlorites and micas in the 2 - 16 μm grain size fraction are highly variable across the studied section and fluctuate between 0 and 100 %. Smectite contents above 10 % were only measured in two samples within the lower part of the section. The kaolinite content of the section fluctuates around 30 % with a maximum (100 %) in the upper part of the Shuba Member. Values decrease in the upper half of the section and show a trend broadly similar to that of the kaolinite content of the < 2 μm grain size fraction (Fig. 5.3B).

The scanning electron microscope (SEM) analysis reveals that most clay minerals occur as grain-coatings in a fine honeycomb structure (Fig. 5.4A, B, D, E). The clay minerals are partly strongly weathered (Fig. 5.4B) and partly occur as rosettes (Fig. 5.4E). The rosettes are more iron-rich than the more weathered grain-coatings (Fig. 5.4C, F).

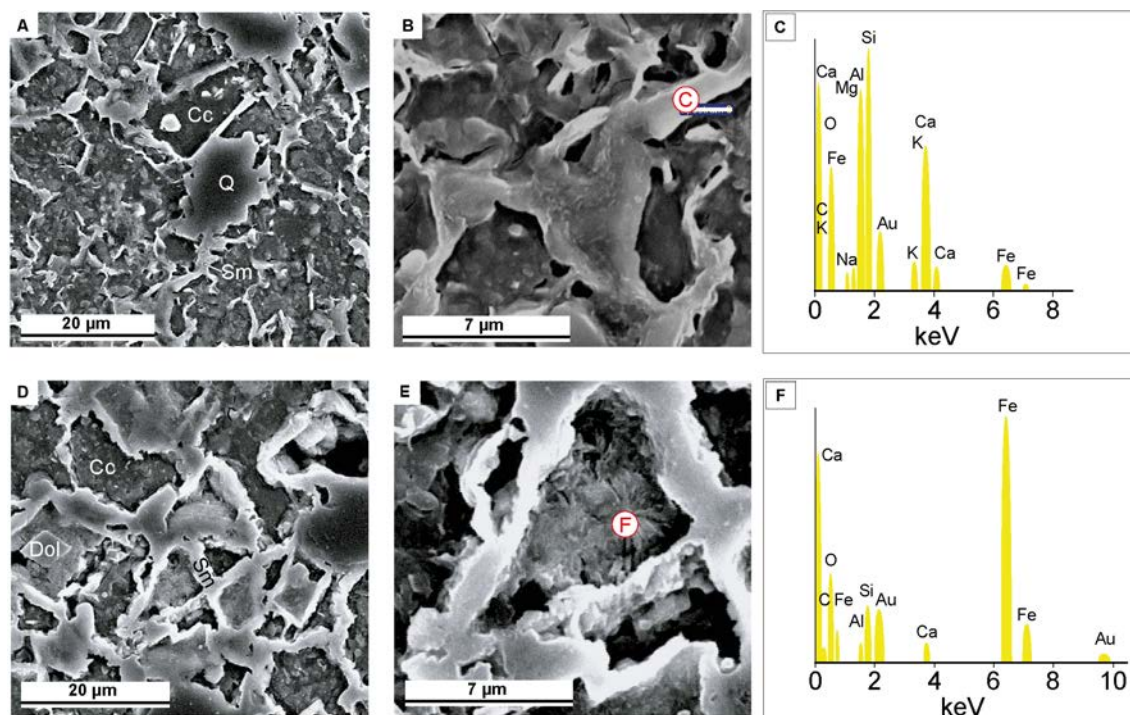


Fig. 5.4 – Scanning electron microscope (SEM) micrographs (**A, B, D, E**) and energy dispersive analytical system (EDS) elemental analysis (**C, F**) of sample WN 168 in the lower part of the Triassic-Jurassic boundary interval (Cc= calcite, Q= quartz, Sm=smectite, Dol=dolomite). **C**) Elemental analysis of grain coating, as indicated in B. **D**) Elemental analysis of clay mineral rosette, as indicated in E.

5.6 Interpretation and discussion

5.6.1 Sedimentary record

The Upper Triassic part of the section is dominated by microbial carbonates deposited on tidal flats, interbedded with marls, shales and fine sandstones. These intertidal deposits are overlain by deeper-water, more open marine bioclastic limestones, which form the uppermost Triassic part of the succession. These are followed by high-energy ramp oolites and restricted-fauna mudstones that were deposited during the Early Jurassic (Fig. 5.2). Several hardgrounds are present within the section, as is typical for such shallow-water deposits (Christ et al., 2015). The progressive deepening from intertidal to open marine, high-energy carbonates during the Late Triassic and into Early Jurassic is interpreted as the result of a transgression, which was followed by a normal regression. The constructed local relative sea level curve based on the

development of accommodation has its maximum across the inferred T-J boundary horizon (Fig. 5.2).

5.6.2 Clay mineralogy

Clay mineral assemblages can serve as proxies for climatic changes. Especially the detrital kaolinite content is typically used as indicator for changes in weathering patterns in the source area linked to more arid versus humid climates (e.g. Adatte et al., 2002; John et al., 2003). Alternative interpretations suggest a link of clay mineral assemblages, including kaolinite to sea level changes (e.g. Worden and Morad, 2003; John et al., 2006). The clay mineral assemblage analysed at Wadi Naqab shows increased kaolinite contents immediately below and within the invoked T-J boundary horizon both in the < 2 and the 2 - 16 μm grain size fraction (Fig. 5.3). The presence of illite in some samples as well as dickite, more common in the 2 - 16 μm grain size fraction, suggests the possibility of a burial diagenetic origin of at least a proportion of the clay minerals (Worden and Morad, 2003). The iron-rich smectite rosettes (Fig. 5.4E, F) are a further indicator of formation caused by diagenesis rather than a purely detrital origin (Hein et al., 1979). The weathered appearance of the clay coatings (Fig. 5.4A, B, D) however is indicative of a detrital origin, but the presence of authigenic clays cannot be excluded.

If the increased amount of kaolinite, roughly corresponding to the Triassic-Jurassic boundary, is indeed of detrital origin it would suggest increased humidity and a more tropical climate prevailing for a relative short time span. A similar clay mineral pattern with increased kaolinite contents accompanying the T-J transition followed by a decrease in kaolinite has also been reported from a western Tethyan section in Austria (Zajzon et al., 2012) and was interpreted as proxy for increased humidity and the intensification of terrigenous input at the T-J boundary (Pálffy and Zajzon, 2012). This would contradict previous assumptions that the Late Triassic, especially in the equatorial realm, was marked by a general aridification (Tanner et al., 2004). The overall decrease of the kaolinite content in the upper half of the section at Wadi Naqab occurs approximately coeval to the onset of a fall in relative sea level (Fig. 5.3). This relative sea level fall was caused either by the filling of accommodation subsequent to the preceding sea level rise, tectonic uplift or a

eustatic drop. During a sea level highstand more kaolinite originating from continental weathering processes is mobilised and incorporated into shallow-marine sediments (Worden and Morad, 2003). A following sea level drop would result in a more distal shoreline and a reduction in kaolinite as well as other detrital clay minerals. It thus remains a possibility that the clay mineral assemblage at Wadi Naqab is a reflection of local sea level changes rather than changes in climatic patterns.

5.6.3 Carbon cycling and temperature

The good overlap between the organic and the carbonate carbon isotopes across the studied section implies a primary origin of the observed pattern. No mechanism is currently known that could alter both signals and cause the similarity in the trends (Bachan et al., 2012). As the shift from a negative to a positive trend in both carbon isotope curves occurs within an interval that contains similar facies types deposited within the same depositional environment, a lithological control on this feature can also be excluded (Fig. 2 (2)). Since the entire section was furthermore located in a very proximal setting on the carbonate ramp, it is unlikely that the observed isotopic patterns are a result of distal mixing between organically and inorganically sourced carbon, preserved in the carbonate carbon isotope signal (Oehlert et al., 2012). It is therefore postulated that the observed trends reflect primary isotopic changes of the carbon reservoir. The question arises whether this is a reflection of atmospheric and thus global changes, or simply a reflection of isotopic changes in the local water mass.

The sediments that have archived the isotopically light carbon signals in the lower half of the section at Wadi Naqab were deposited under very shallow-marine conditions (Fig. 5.2 (1,2)). Their enrichment in ^{12}C could be the result of inflow of isotopically light meteoric water and ^{12}C enriched organic carbon in the form of plant material from the relatively close continent. The subsequent positive shift (Fig. 5.2 (2,3)) in both carbon isotope curves could be a reflection of the transgression that is evident through the deepening of the depositional environment. In this model, put forward mainly to explain positive shifts in the carbonate carbon signal, the enrichment with ^{13}C is caused by the influx of

isotopically heavy water from the open ocean that is subsequently incorporated into the precipitated carbonates (Immenhauser et al., 2003). By extension the organic carbon record might become enriched with ^{13}C due to the reduced inflow of terrestrial, isotopically light, organic matter owing to the more distal position with regards to the coastline.

The alternative to this localised explanation is the enrichment of the atmosphere with isotopically light carbon, causing negative trends in both carbon isotope curves. The progressively falling $\delta^{13}\text{C}$ values during the Late Triassic (Fig. 5.2 (1,2)) have also been measured in records from the United Kingdom and Greenland (Hesselbo et al., 2002) as well as from Montenegro (Črne et al., 2010) and were interpreted as evidence for the injection of isotopically light CO_2 into the atmosphere through the initiation of volcanic activity. If isotopically light carbon that was injected into the atmosphere during the Latest Triassic originates from the CAMP, a relatively long term, and maybe slow CO_2 injection can be invoked (Ruhl and Kürschner, 2011; Lindström et al., 2015) as well as potentially ongoing and repeated activity (Lindström et al., 2012), rather than a rapid and short lived pulse (Marzoli et al., 2004; Cirilli et al., 2009; Ruhl et al., 2011).

The following stepwise positive excursion, maybe the return to stable background values following the long lasting negative shift (Fig. 5.2 (2,3)) with an amplitude of 2.5 (organic carbon) and 4 ‰ (carbonate carbon) could be the result of increased carbon burial, whereby basinal areas that have since been subducted could have acted as sinks for the excess organic matter. Multiple sections from Italy also contain a positive excursion of both carbonate and organic carbon $\delta^{13}\text{C}$ (Bachan et al., 2012). Although the positive excursion is not expressed in sections globally, as it is not seen in the United Kingdom for example, it has been interpreted as being the result of increased carbon burial, induced in part by increased weathering rates and a rise in temperature, which in turn could have been induced by rising levels of atmospheric CO_2 (Bachan et al., 2012). The elevated phosphorous content observed at Wadi Naqab, corresponding to the highest values of both organic and carbonate carbon isotopes (Fig. 5.2 (3)), implies high rates of phosphorous burial, which can be interpreted as the result of increased primary productivity (Föllmi, 1996) or potentially as proxy for increased river discharge induced by a wetter climate.

This serves as further evidence for the likelihood of high rates of increased carbon burial associated with the T-J boundary. The explanation for the carbon isotope record at Wadi Naqab with the onset of the positive shift of organic and carbonate carbon isotopes occurring in the Upper Triassic part of the section implies that atmospheric CO₂ levels were elevated during the Late Triassic preceding the T-J boundary. The rise of atmospheric CO₂ levels distinctly before the T-J boundary has previously been inferred from Rhaetian pedogenic carbonate nodules from North America, accompanied by a marked temperature increase (Cleveland et al., 2008). Carbon isotope fractionation in organic matter relative to the carbonate ($\Delta^{13}\text{C} = \delta^{13}\text{C}_{\text{carb}} - \delta^{13}\text{C}_{\text{org}}$) has previously been used to infer atmospheric CO₂ levels (e.g. Kump and Arthur, 1999; Hermoso et al., 2012). The $\Delta^{13}\text{C}$ curve at Wadi Naqab shows a marked positive shift across the T-J boundary (Fig. 5.2 (2,3,4)). If this serves as reliable proxy then it is an additional indication for elevated atmospheric CO₂ levels during the Late Triassic, well before the T-J boundary and continuing well into the Early Jurassic with duration of 1 Myrs and more.

Elevated temperatures as a result of increased atmospheric CO₂ and other greenhouse gases have frequently been suggested as driver or major contributor to environmental and biotic changes associated with the T-J boundary (e.g. Cleveland et al., 2008; Bachan et al., 2012; Hönisch et al., 2012; Tanner et al., 2016). The use of carbonate oxygen isotope values to estimate palaeotemperatures is a problematic task, as oxygen isotopes from carbonates are more susceptible to diagenetic alteration than stable carbon isotope values (e.g. Krencker et al., 2014). If the oxygen isotopes at Wadi Naqab do represent changes in palaeoseawater temperatures then the negative excursion (Fig. 5.2 (2,3)) corresponding to the Latest Triassic could imply a warming phase, possibly induced by a coeval elevated CO₂ level.

5.7 Conclusions

The equatorial record studied at Wadi Naqab contains no evidence for a biocalcification crisis, potentially induced by ocean acidification, associated with the T-J boundary. The changes in the vertical stacking pattern and clay mineral assemblages at Wadi Naqab can be explained by a relative sea level rise

during the Late Triassic with the maximum relative sea level corresponding to the T-J boundary horizon. The amplitude of the sea level rise can be approximated to lie in the order of at least 5 to 50 m (based on the fossil assemblages). Carbon and oxygen isotope data suggest elevated atmospheric CO₂ levels and potentially elevated ocean temperatures during the Latest Triassic accompanying this rise in relative sea level. It can be invoked that elevated atmospheric CO₂, potentially sourced from a relatively early onset of CAMP associated degassing, caused global temperatures to rise and might have resulted in a global sea level rise (Fig. 5.5). This is in line with the gradual extinction of taxa during the Rhaetian, in contrast to a more or less sudden extinction (Hallam, 2002). Evidence for a regional sea level rise, such as that suggested above, at least in the equatorial western Tethys, is present from another available equatorial T-J boundary study site (Hallam et al., 2000).

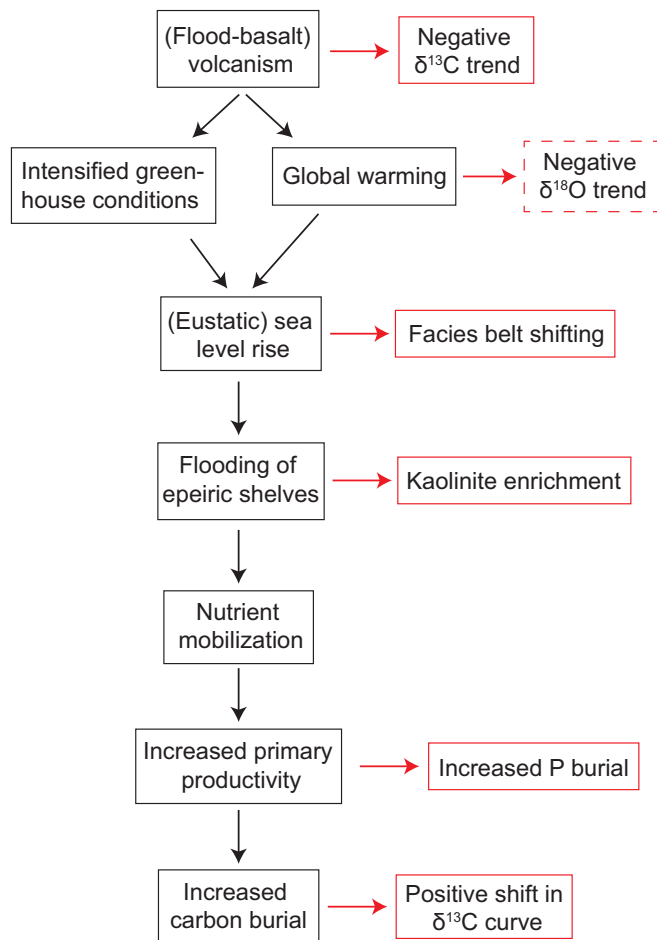


Fig. 5.5 – Flow chart illustrating events during the Late Triassic and Early Jurassic (black boxes) and evidence present at Wadi Naqab (red boxes) (modified from Föllmi, 1996).

6. Inter-well scale sedimentological and isotopic heterogeneities within a Jurassic carbonate ramp (UAE) and implications for reservoirs in the Middle East

6.1 Abstract

Carbonate rocks are major hydrocarbon reservoirs in the Middle East and across the world. Capturing the spatial distribution and dimensions of carbonate sedimentary facies is thus of interest for reservoir modelling and for the understanding of process sedimentology. Here data from a Middle Jurassic carbonate ramp is presented, outcropping on the Musandam Peninsula (Wadi Naqab, Northern United Arab Emirate of Ras-Al-Khaimah), which serves as an analogue for the hydrocarbon bearing units in the subsurface of the Middle East with respect to stratigraphic age, lithofacies types, vertical stacking patterns and palaeogeographical setting. Seven closely spaced sections were logged across a 1900 x 1200 x 120 m fault-bounded outcrop offering a pseudo-3D view of the stratigraphy. Sections were correlated bed-by-bed by walking key surfaces and/or by using photopanel, to produce a detailed lithofacies map. Stable isotopes (carbon and oxygen) were measured from micrite samples in order to provide chemostratigraphic constraints. Ten lithofacies types were identified in thin sections and hand specimens, ranging from marls and mudstones to grainy oncolites and peloidal-oidial grainstones. Beds are organised in a 'layer-cake' architecture, but individual bed thicknesses vary laterally. Lateral merging between different lithofacies types within beds is very common, resulting in a complex facies mosaic at the scale of the outcrop. Lithofacies types have a maximum probability of 40 % of being continuous between logged sections with the highest frequencies of lateral variations occurring in the open, shallow part of the ramp, where maximum wave and current energy are indicated by facies. The vertical stratigraphic stacking pattern of the succession is characterised by metre-scale shoaling upwards parasequences. These are frequently capped by hardgrounds, interpreted as composite surfaces with depleted carbon and oxygen isotope values proving evidence for frequent exposure during relative sea-level lowstand, followed by flooding and formation of a condensed hardground. The carbon isotope curves show a large amount of vertical 'isotope

heterogeneity' reflected by meteoric diagenesis around hardground surfaces, and these are not always traceable laterally between the closely spaced neritic sections due to bias in sampling. However, if considering only the positive envelop of the carbon isotope curve and thus filtering out the meteoric diagenetic events, carbon isotope chemostratigraphy becomes possible and confirms a Bajocian age for the section. The main conclusion of the study is that maximum heterogeneity of facies at the scale of 100 meters or less is expected in an environment of deposition with high wave energy. This has important implications for reservoir applications since rock facies often template flow units.

6.2 Introduction

Acquiring information on heterogeneities in carbonate systems helps to understand processes that govern sediment distribution within the carbonate depositional realm, and is thus crucial for subsurface applications. Multiple modern analogue studies have shown that carbonate facies distribution in the shallow marine environment can be very heterogeneous (Wilkinson et al., 1999; Rankey and Reeder, 2011; Harris et al., 2015; Purkis et al., 2015a). The controlling factors on this distribution include, amongst others, sea-level fluctuations and climatic changes, changes in local hydrography and wind-energy, but also changes in carbonate precipitating biota, which are in turn controlled by their ecological requirements such as temperature, ocean chemistry and nutrient availability (Pomar and Kendall, 2007; Strasser and Védérine, 2009; Harris et al., 2015). Although modern analogues can deliver information on heterogeneities in carbonates, it remains unknown how comparable systems such as the Florida shelf or the Bahamas platform are to the vast epeiric carbonate ramps that existed, for example on the Arabian plate during much of the Palaeozoic and Mesozoic. While physical processes that govern carbonate accumulation will have remained unchanged, the main carbonate producing biota have changed through time (e.g. Pomar and Hallock, 2008), and the scale and configuration of the buildups is often very different.

Outcrop analogue studies play a crucial part in filling the gap between data provided by low vertical resolution seismic surveys, and the high-vertical resolution, 1D core analysis and are often better analogues regarding

lithofacies types and the entire carbonate system, in comparison to modern analogues. Compared to siliciclastic systems however, the amount of quantitative carbonate-analogue data is still relatively scarce (Aigner et al., 2007). New studies have started to close this gap and also paved the way for more outcrop analogue research at the inter-well scale: for instance, carbonate outcrop analogue studies that include the systematic lateral mapping of facies and key stratigraphic surfaces exist from the Jurassic of Morocco (Amour et al., 2011; Christ et al., 2011), from Spain (Alnazghah et al., 2013) and from Oman and the United Arab Emirates, in the proximity of Middle Eastern reservoirs on Permian and Triassic (Zeller et al., 2011; Ahmadzamri et al., 2014) and Cretaceous outcrops (Immenhauser et al., 2004; Vaughan et al., 2004; Sattler et al., 2005; Grélaud et al., 2006; 2010; Rameil et al. 2010; Adams et al., 2011; Sena and John, 2013). No comparable studies have been published yet on outcrop-analogues for Jurassic reservoirs of the UAE, Qatar or Saudi Arabia, despite the fact that Jurassic carbonate reservoirs in the Middle East are of major economic importance (e.g. Alsharhan and Nairn, 1994, 2003; Sudarsana et al., 2009).

Analogue studies using outcrops sometimes suggest that ancient carbonates are preserved in a 'layer-cake' type architecture where pinch out and facies heterogeneities are rare (e.g. Borkhataria et al., 2005; Zeller et al., 2011; Pratt et al., 2012; Bendias et al., 2013), whereas in other examples large amounts of facies heterogeneity is documented, even on a relatively small scale and especially within shallow water depths (e. g. French and Kerans, 2004; Sattler et al., 2005; Bádenas et al., 2010; Amour et al., 2011; Christ et al., 2011; Alnazghah et al., 2013; Sena and John, 2013; Brandano and Loche, 2014). In this study, we focus on inter-well scale facies distribution within outcrops of Middle Jurassic shallow-water ramp deposits at Wadi Naqab, Emirate of Ras-Al-Khaimah (UAE). Our work in Ras-Al-Khaimah on a relatively small inter-well scale (1900 x 1100 x 120 m) outcrop allows us to capture sedimentological as well as isotopic heterogeneities, and provides the first published evidence of the complexity of facies distribution within Jurassic carbonate ramps of the Arabian Peninsula. The objectives of this study are to 1) provide new and quantitative constraints on the bedding and facies

heterogeneities in an inner to middle ramp in the Middle East; 2) explore the links between facies distribution and process sedimentology in this setting, i.e. what controls the facies heterogeneities within carbonate ramps of the Middle East?; and 3) assess the vertical and lateral heterogeneity as well as usefulness of carbon isotopes as a chemostratigraphic constraint in this type of setting.

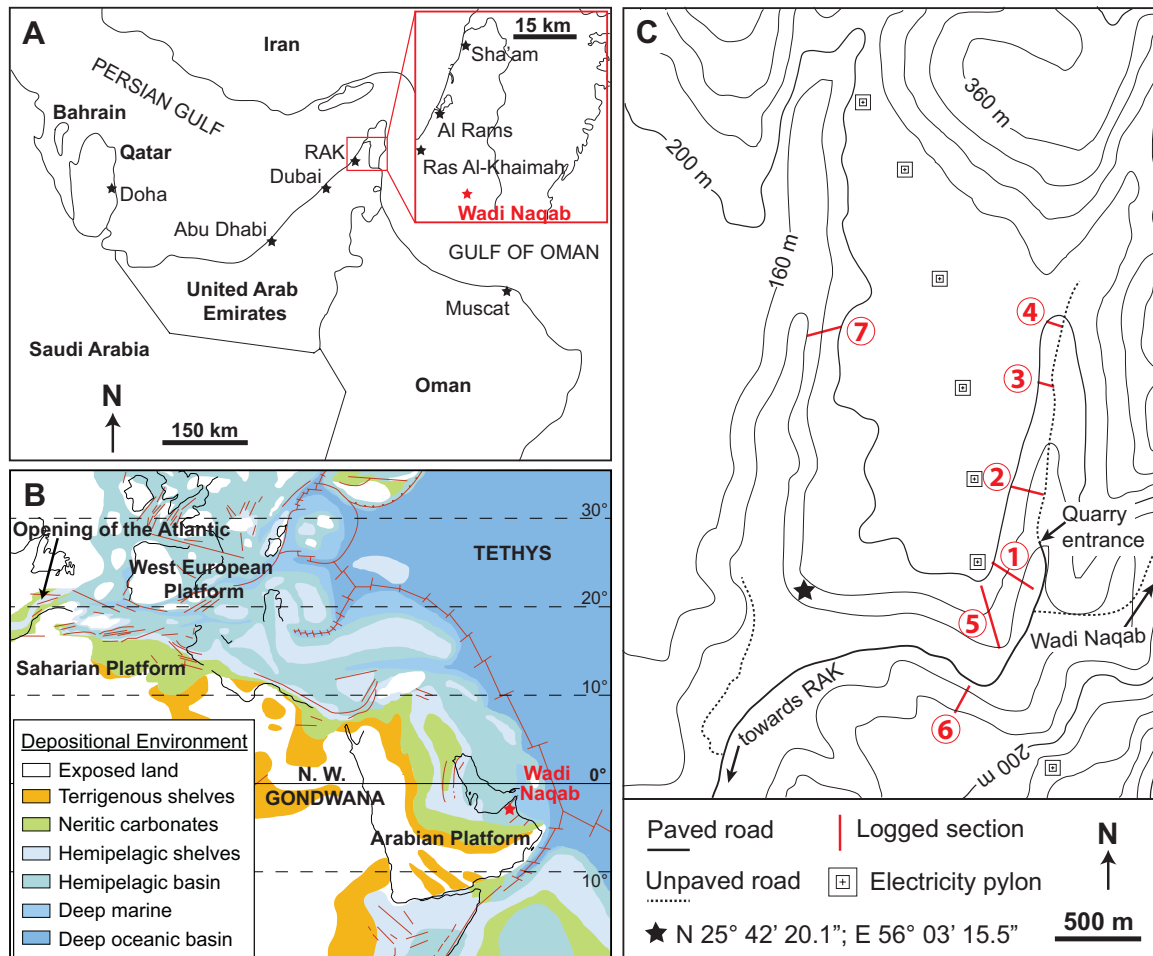


Fig. 6.1 – A) Location map of the study area. **B)** Palaeogeographical setting of the Tethyan realm during the Jurassic (modified from Pierre, 2006; after Thierry and Barrier, 2000). **C)** Location of the logged sections in Wadi Naqab.

6.3 Geological setting

The outcrop is located close to the entrance of Wadi Naqab, which lies 13 km southeast of the city Ras Al-Khaimah in the Musandam Mountains (GPS coordinates of the outcrop: N 25° 42' 20.1"; E 56° 03' 15.5").

The Musandam Mountains form the northern extension of the Oman Mountains and are composed of ~3 km-thick shallow-water carbonates (Maurer et al., 2009). The eastern margin of the Arabian Plate was tectonically passive during the Permian and the Mesozoic, but this period of quiescence was interrupted by two tectonic events, one during the Late Cretaceous when ophiolites were obducted, and the second one during the Pliocene and Pleistocene when the Oman and Musandam Mountains were formed (Searle, 1988; Glennie, 2005). From the Cenozoic onwards the eastern margin of the Arabian platform is in an active continental setting and the plate is being subducted beneath the Eurasian Plate (Sharland et al., 2001).

The Jurassic section on the Musandam peninsula has a stratigraphic thickness of approximately 1300 m and forms the Musandam Formation, which is divided into the Musandam Formations 1 to 3, for the Lower, Middle and Upper Jurassic respectively (Fig. 6.2). The Jurassic sedimentary record is near enough complete except for parts of the Toarcian and the Aalenian and the Tithonian (de Matos, 1997; de Matos and Walkden, 2000). The rock units were deposited in shallow water conditions on the northern margin of the Arabian Peninsula, with an open marine connection to the Neo-Tethys Ocean (Fig. 6.1B). This small oceanic basin was connected to the Palaeo-Tethys and was located between the Afro-Arabian platform in the southwest and a microcontinent in the northeast (de Matos, 1997; Glennie, 2005). The entire sedimentary package in Wadi Naqab extends in age from Permian in the eastern end of the wadi to Cretaceous at the wadi entrance in the west. The Bajocian section alone expands over 550 m (de Matos, 1997). Bajocian-aged strata partly crop out in a 'block' bound to the south and west by wadis and on its eastern side by an abandoned quarry. The Bajocian age was assigned by de Matos (1997) based on benthic foraminifera, stromatoporids, algae and on a single, equivocally classified ammonite. The studied stratigraphic interval can be examined in pseudo three-dimensions (Fig. 6.1C). The Cretaceous package overlies the Late Jurassic rocks at Wadi Naqab with a faulted contact, but no other obvious unconformity is visible within the sedimentary succession (de Matos, 1997).

Chronostratigraphy		Form. Qatar / Abu Dhabi		Form. Musandam Peninsula	
		On shore	Off shore		
UPPER JURASSIC	Oxfordian	Diyab	Lower Jubailah	MUSANDAM GROUP	Musandam 3 Formation
			Hanifa		
MIDDLE JURASSIC	Callovian	Araej	Upper Araej		Musandam 2 Formation
	Bathonian		Uwainat		
	Bajocian		Lower Araej		
			Izhara		
LOWER JURASSIC	Aalenian	Marrat	Elphinst. Group		Musandam 1 Formation
		Hamlah			

Fig. 6.2 – Jurassic stratigraphy of the Musandam Peninsula and of selected subsurface examples (modified from Alsharhan and Nairn, 1994; de Matos, 1997).

6.4 Materials and methods

6.4.1 Field and petrographical analysis

Seven sections were measured around the Wadi Naqab ‘block’ outcrop, thus giving insight on the 3D distribution of facies (Fig. 6.1C). The stratigraphic thickness of sections ranges between 70 and 133 m. Four sections were logged on the eastern face of the outcrop within the abandoned quarry. One section was measured on the southern side of the outcrop and one on the other side of the wadi road also to the south of the 3D outcrop. The seventh corresponding

log was produced on the western face of the outcrop. In total the study window covers an area of 1900 by 1200 m (Fig. 6.3). The Bajocian-aged rocks were logged bed-by-bed and examined in the field for their sedimentary features, fossil assemblages and texture. Textural classification of facies followed Dunham (1962) and Embry and Klovan (1971). The position of each section was mainly determined by the accessibility along the outcrop and the possibility to log it. A surface was chosen as reference datum (hardground II, top of the thick oncolite) as it was easily traceable in the field and on photo panoramas. Due to the folding of the strata only one section could be measured on the western side of the outcrop. Physical lateral tracing was undertaken wherever it could be safely done, and this was supplemented by careful analysis of photo panoramas. Fist-size samples for petrography and carbon and oxygen stable isotope analysis were taken every 2 to 5 m across the vertical transects, using a geological hammer (268 samples in total). At Imperial College London 100 representative thin sections were made. Half of each thin section was stained after the method described by Dickson (1965), and subsequently analysed using a transmitted light Zeiss Axioskop 40 microscope (plain and polarised light).

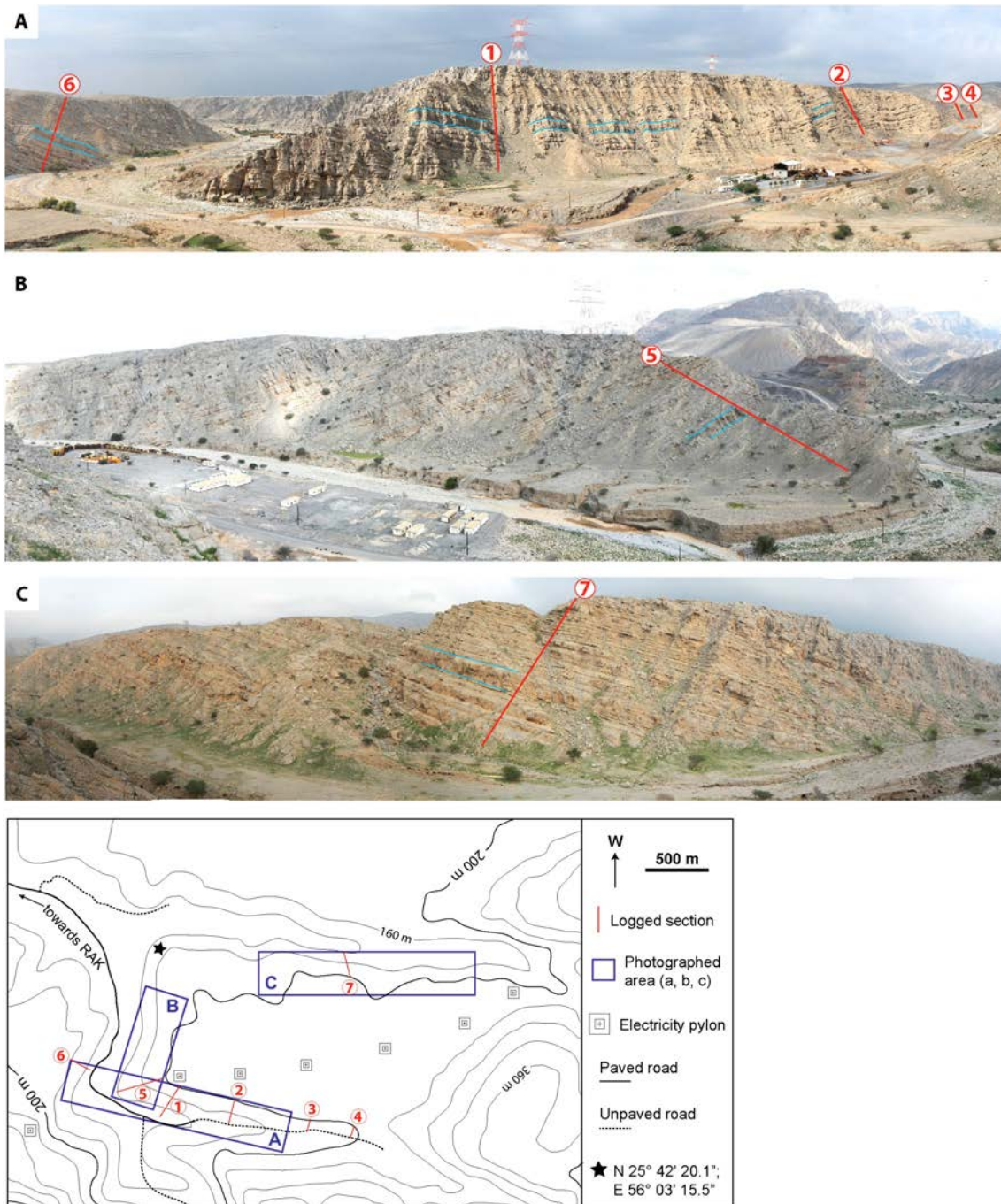


Fig. 6.3 – Photopanel of the studied ‘block outcrop’. The top (hardground II = reference datum) and bottom of the thick oncolite marker bed are marked with blue lines. The map shows the location of each photopanel.

6.4.2 Markov-chain analysis

A Markov-chain analysis was carried out on the acquired stratigraphic data to test whether some degree of ordering is present between the different lithofacies types at the outcrop. A transition frequency matrix and a transition

probability matrix for both vertical and lateral facies transitions were produced. A χ^2 analysis, that compares the observed facies distribution to a calculated, expected random distribution, was subsequently performed. For vertical upwards facies transitions, including discontinuity surfaces, only facies changes and no transitions from a facies into itself were taken into account. This resulted in 81 degrees of freedom (here $(m-1)^2$, where m is the number of possible states). A total of 770 upwards transitions were counted and the null hypothesis, stating that a facies type is not dependent on the underlying facies, could be rejected at the 99.9 % significance level. Lateral facies transitions were considered between neighbouring sections on either side of each section, also counting transitions from a facies into itself (i.e. lateral facies continuity between logged sections). The lateral continuity of discontinuity surfaces was not taken into account in this analysis. The null hypothesis was also rejected for lateral facies transitions at the 99.9 % significance level for 64 degrees of freedom and a total of 1062 facies transitions.

6.4.3 Geochemical analysis

Measurements of stable isotopes of carbon and oxygen were performed in the Qatar Stable Isotope Laboratory at Imperial College London. The measurements were performed on micrite where possible, and on bulk samples for grainy facies, as micrite is the most common carbonate phase in the sample set and thin section analysis could be used to check the purity of the micrite. A total of 164 sample powders were obtained using an electrical dental drill, carefully avoiding veins, fossils and recrystallised sections. Approximately 100 to 150 μg of sample powder were dissolved with 105 % orthophosphoric acid at 70 °C in a Kiel IV carbonate device, and the resulting CO_2 gas was measured on a Thermo MAT 253 mass spectrometer. Data was corrected according to measurements of the international standard NBS 19 and an in-house standard (ICM, Imperial College Carrara Marble). Stable isotope values are reported using the standard δ notation relative to the VPDB (Vienna Pee Dee Belemnite) standard. External precision for ICM (one standard deviation) on carbon and oxygen is better than 0.05 ‰ and 0.08 ‰, respectively. The precision for

duplicate and triplicate sample measurements (one standard deviation) is better than 0.05 ‰ for carbon and better than 0.14 ‰ for oxygen. To assess whether samples were potentially suited for chemostratigraphic correlations using $\delta^{13}\text{C}$, the ' $\delta^{18}\text{O}$ filter' was applied (Metzger and Fike, 2013): $\delta^{13}\text{C}$ data are excluded if the corresponding $\delta^{18}\text{O}$ value is more negative than the average $\delta^{18}\text{O}$ value of all measurements minus one standard deviation (in this case, if the $\delta^{18}\text{O}$ value is more negative than -4.0 ‰).

6.5 Results

6.5.1 Field and petrographical results

6.5.1.1 Lithofacies types

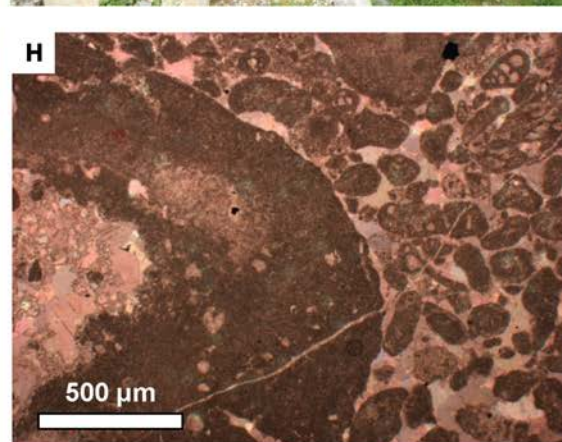
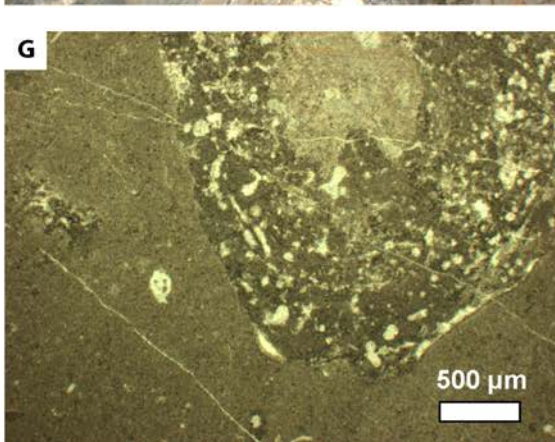
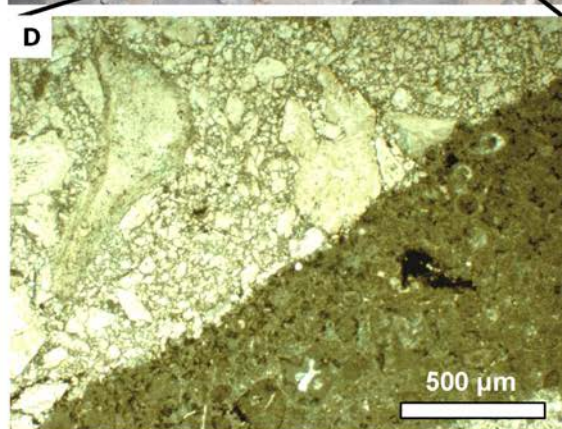
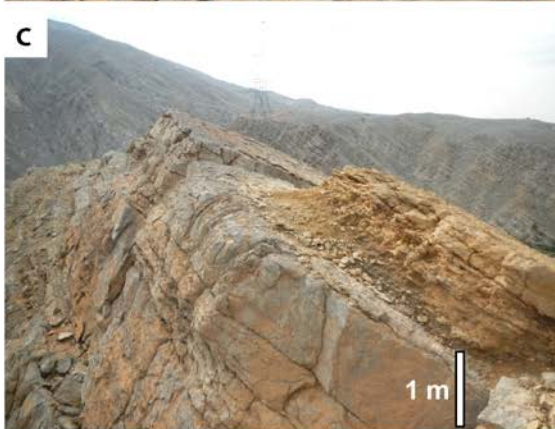
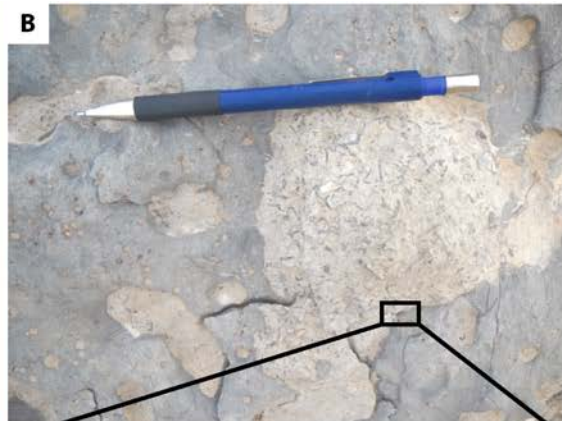
Within the logged stratigraphic interval, ten lithofacies types (LF) were identified (Table 6.1). LF 1 consists of marls and mudstones with few or no bioclasts. Marls are identified through their very recessive weathering pattern and are either finely laminated or heavily bioturbated (*Thalassinoides*) throughout the studied interval. Brachiopods are present but only in a single marl bed. LF 2 comprises nodular mud- to wackestones, which often occur in thin (one decimetre or less) and wavy beds with a rubbly appearance (Fig. 6.4F). Beds of LF2 often contain lenses and laminae of bivalve packstones. Mudclasts and gastropods are occasionally present as well as quartz and glauconite grains. LF 3 and 4 have a wacke- to packstone texture and are characterised by varying amounts of peloids, bivalves and gastropods (Fig. 6.4K). The predominance of peloids (LF3) or bioclasts (LF4) is the key distinguishing factor between the two facies. Ooids and other coated grains as well as benthic foraminifera, ostracods and solitary corals are occasionally present (Fig. 6.4L). The size of the peloids typically varies between 50 to 300 μm . The bioclastic wacke- and packstones of LF 5 are composed of the same bioclasts as LF 3 and 4 but lack peloids. Facies types LF 6 and 7 consist of oncoids, other coated grains, gastropods, bivalves and micritic nodules as well as peloids. The oncoidal floatstone LF 6 has a lime mud matrix (Fig. 6.4I), whereas LF 7 has a grainy,

pack- and grainstone matrix composed of peloids, ooids and other coated grains (Fig. 6.4G, H). The oncoids in both facies types vary in size and have diameters of up to 5 cm. The nuclei of oncoids are mostly bioclasts such as bivalve fragments, and occasionally lithoclasts. The oncoidal limestones form the thickest beds within the succession and are often heavily bioturbated and exhibit wavy, decimetre-spaced bedding. LF 8 and 9 are mainly grainstones with rare patches of muddier matrix (packstones in those areas). In LF 8 peloids are the dominant grain constituent, while LF 9 is comprised of mainly ooids. LF 8 can be distinguished from LF 3 (peloidal bioclastic limestone), as LF 8 is foremost a grainstone and LF 3 occurs as packstone. The main components in LF 8 and LF 9 are accompanied by minor different bioclasts, such as bivalves, gastropods, and benthic foraminifera (Fig. 6.4J). In the peloidal grainstones of LF 8, the peloids vary significantly in size with diameters from ca. 200 to 1000 μm , often with a bimodal distribution. Ooids often show signs of strong micritisation. Their nuclei are commonly benthic foraminifera and lithoclasts. Cross-stratification is rarely visible within these two lithofacies types. Lithofacies LF 10 is a coral boundstone comprising branching and solitary corals. This lithofacies type is only visible in two relatively thin beds within the studied strata.

Table 6.1 – Description, dimensions and interpretation of the ten observed lithofacies types (LF); (m: marl; M: mudstone; W: wackestone; P: packstone; G: grainstone; F: floatstone; R: rudstone; B: boundstone).

LF	LF Name	Texture	Main Components	Sedimentary Features	Dimensions	Interpretation
1	Marl / mudstone	m M	None (brachiopods in some marls)	Occasional bioturbation (mainly <i>Thalassinoides</i> in m), lamination	Thickness: 0.05 – 2 m Maximum lateral extent: 1900 m	Representing intervals of low sedimentation rate; bioturbation points to oxic conditions; brachiopod bearing marls deposited off-shoal/on the middle ramp
2	Mudstone (nodular/t hinly bedded/ pseudobrecciated)	M M-W (P)	Pockets and lags of bivalve P; mudclasts, gastropods; quartz; glauconite	Rubbly and often heavily bioturbated, often nodular, resulting in a pseudobrecciated appearance	Thickness: 0.15 - 3.3 m Maximum lateral extent: 1900 m	Often overlying hardgrounds; related to stratigraphic condensation; nodules are the result of concretionary cementation; oxic conditions if bioturbated
3	Peloidal bioclastic limestone	W W-P P	Peloids, bivalves, gastropods, coated grains, ooids benthic foraminifera, ostracods	Bioturbation, occasionally: calcite veins with distinct organic/oil smell	Thickness: 0.05 - 4.5 m Maximum lateral extent: 1600 m	Relatively low to moderate energy deposits, peloids are always the main constituent and mostly of fecal origin, large gastropods point to a potentially restricted environment
4	Bioclastic peloidal limestone	W-P P F	Bivalves, gastropods, benthic foraminifera, echinoids, peloids	Bioturbation, wavy bedding	Thickness: 0.1 - 3.2 m Maximum lateral extent: 250 m	Moderate energy deposits; bioclasts are the main constituent but peloids are present
5	Bioclastic limestone	W W-P P F	Bivalves, gastropods, benthic foraminifera, coated grains, corals	Bioturbation, wavy bedding, as W often nodular, in parts 'biobrecciated'	Thickness: 0.1 -5.7 m Maximum lateral extent: 1900 m	Moderate or low energy deposits, depending on the fossil assemblage; no peloids are present
6	Oncoidal (microbial) floatstone	F F-R R	Oncoids, coated grains, gastropods, bivalves, micritic nodules, peloids	Bioturbation; wavy, dm-spaced bedding	Thickness: 0.6 - 10.5 m Maximum lateral extent: 930 m	(Restricted) low energy environment; oncoids up to 5 cm in diameter; nuclei are mostly bioclasts (bivalves), rarely lithoclasts
7	Oncoidal floatstone with grainy matrix	F F-R (P/G)	Oncoids, peloids, ooids, coated grains, gastropods, bivalves	Bioturbation; wavy, dm-spaced bedding	Thickness: 0.5 - 9.5 m Maximum lateral extent: 1900 m	(Restricted) low, to moderate energy environment, where oncoids can mix with ooids and peloids that produce the

						characteristic, grainy matrix
8	Peloidal- ooidal limestone	P P-G G	Peloids, ooids, intraclasts, bivalves, gastropods, echinoids, benthic foraminifera	Bioturbation, occasional cross- stratification, sometimes ooids are limonitic	Thickness: 0.2 – 3.1 m Max. lat. ext.: 500 m	High energy deposits; peloids vary significantly in diameter, mostly micritised grains, rarely of fecal origin; preserved as bars or bows or sheets
9	Ooidal grainstone	G P	Ooids, bivalves, gastropods, benthic foraminifera, small oncoids	Occasional cross- stratification; bioturbation, burrowing and boring (if condensed hardground surface); HG's abundant on top of these beds; pockets of bioclastic P common	Thickness: 0.1 – 4.1 m Maximum lateral extent: 1600 m	High energy deposits, potentially deposited in shoals or bars, ooids often with micritised rim, most likely developed in lower energy environments; nuclei often bioclasts (e.g. benthic foraminifera)
10	Coral boundstone	B R	Branching corals, solitary corals	None	Thickness: 0.2 – 0.7 m Maximum lateral extent: 970 m	Branching corals form small patch reefs, most likely within the open- marine, shallow to middle ramp



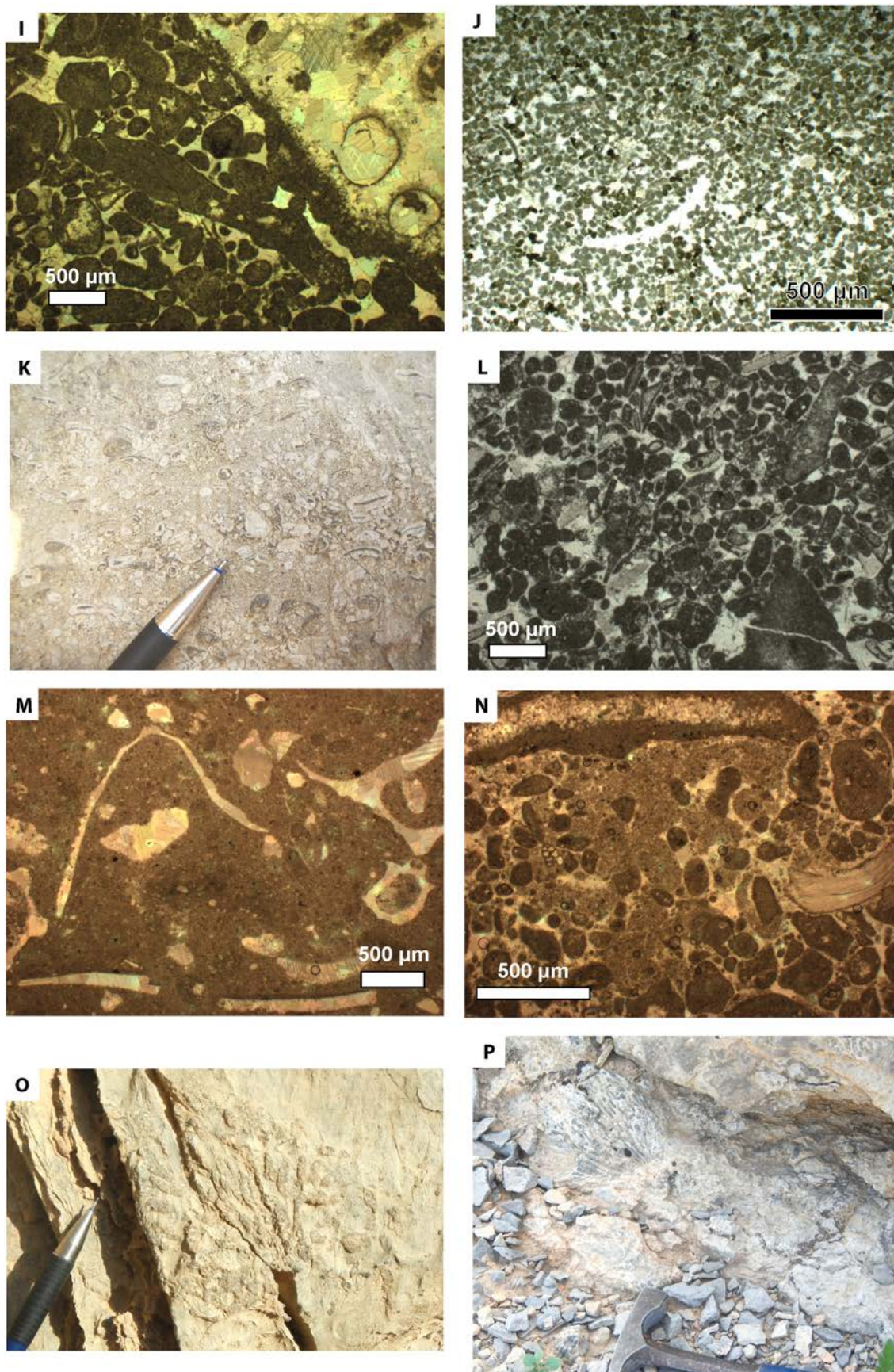


Fig. 6.4 – A) Red stained and strongly cemented discontinuity surface. B) Bored and scoured marine flooding surface. Note that the infill of the scours is selectively dolomitised (D - thin

section image). **C)** Massive carbonate beds capped by marine flooding surface, overlain by recessively weathering rubbly marl and mudstone (LF 1 and 2). **E)** Subtidal rhythmites within section 1. Note person for scale. **F)** Thinly bedded, rubbly mudstone (LF 2). **G)** Thin section image of an oncolite with a muddy matrix (LF 6). **H)** Stained thin section image of an oncolite with a grainy, peloidal grainstone matrix (LF 7). **I)** Thin section image of a coarsely grained peloidal grainstone containing a large bioclast (LF 8). **J)** Thin section image of a finely grained peloidal grainstone (LF 8). **K)** Oncolite with a grainy matrix (LF 7). **L)** Thin section image of an ooidal grainstone, containing echinoderm fragments (LF 9). **M)** Thin section image of a bioclastic mud- to wackestone (LF 5). **N)** Thin section image of an ooidal-peloidal pack- to grainstone with few bioclasts (LF 3). **O)** Branching coral from the upper part of section 2 (LF 10). **P)** Coral boundstone from the top of section 7 (LF 10).

6.5.1.2 Discontinuity surfaces

We identified prominent stratigraphic surfaces, which are interpreted as hardgrounds based on four main criteria: (i) strong resistance to weathering, potentially indicative of pronounced cementation (Fig. 6.4C); (ii) red to brownish staining of the upper part of the bed (Fig. 6.4A); (iii) extensive bioturbation along the top of the bed (Fig. 6.4B), including burrowing and boring where burrows and scours are occasionally selectively dolomitised (Fig. 6.4D) and (iv) mineral crusts and nodules on the bedding planes. Two thirds of the surfaces are continuous over the study area (named using the Roman numerals I-V, Fig. 6.5), and thus extend over a minimum of 1900 x 1200 m. The remaining surfaces are either only visible in one section or extend for 220 to 1100 m (named using Arabic numerals 1-8, Fig. 6.5).

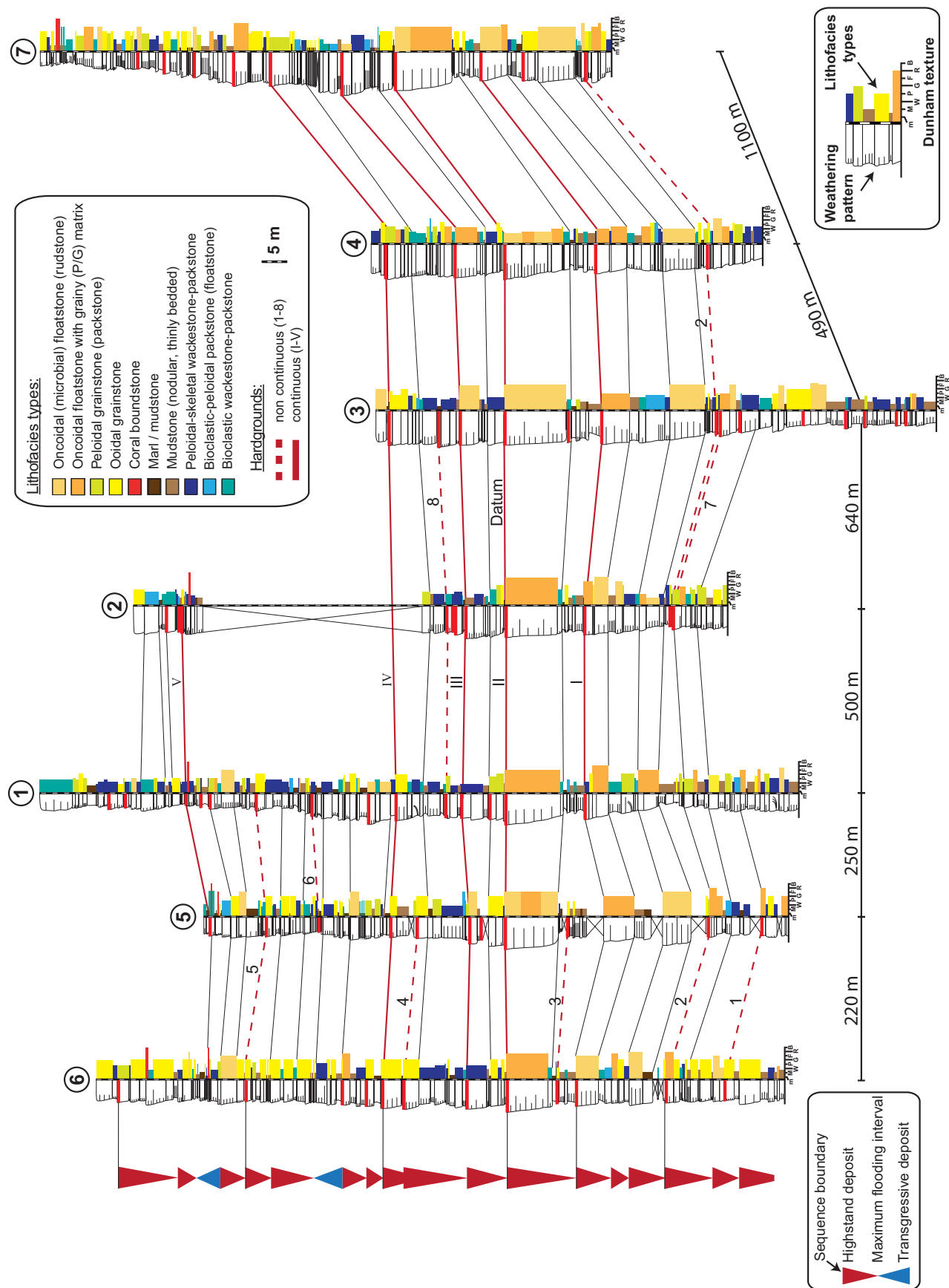
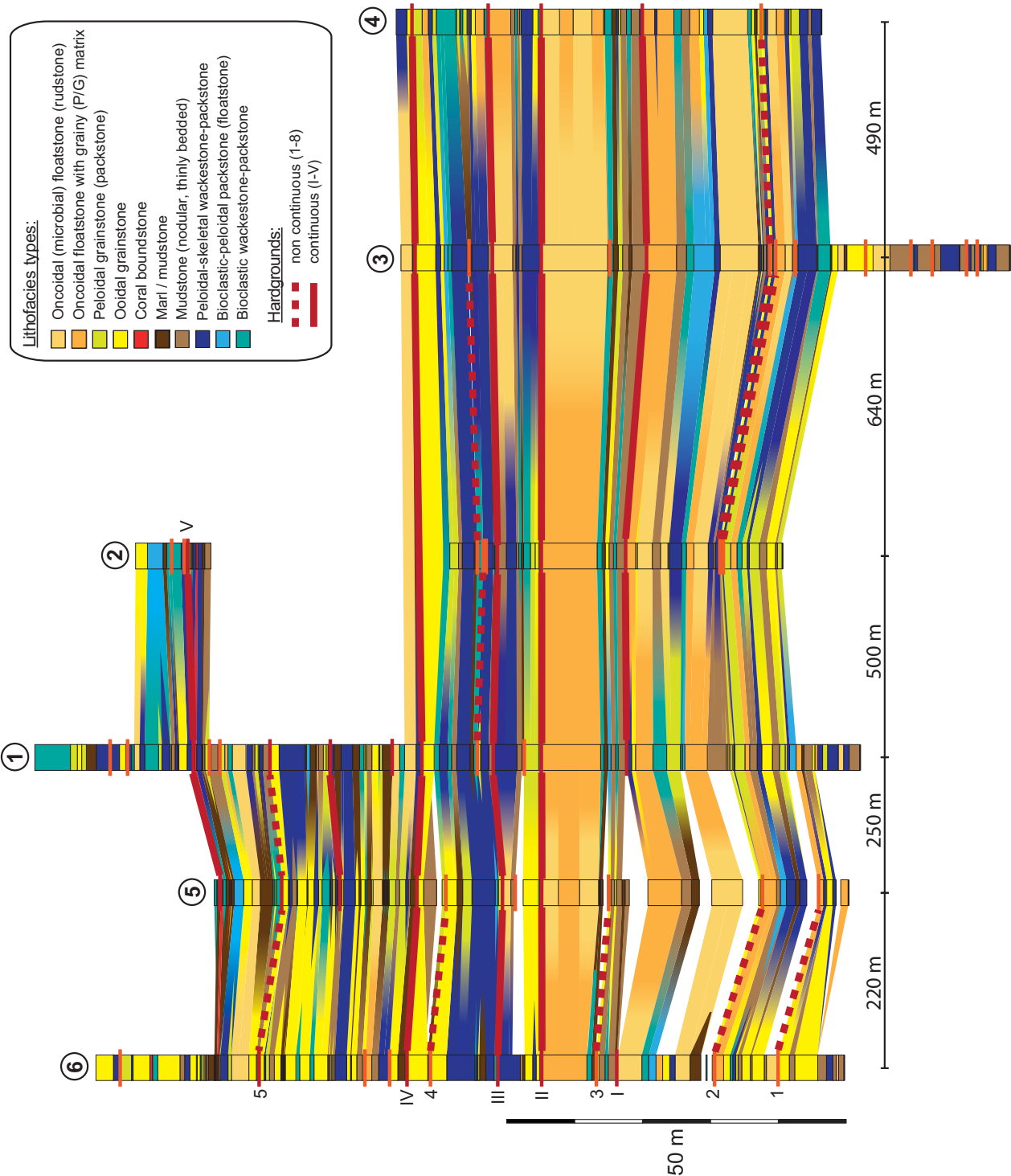


Fig. 6.5 – Measured stratigraphic sections with lateral correlation; (m: marl; M: mudstone; W: wackestone; P: packstone; G: grainstone; F: floatstone; R: rudstone; B: boundstone).

Hardgrounds are indicated in red, divided into continuous (I-V) and discontinuous ones (1-8). Hardground II corresponds to the datum.

A



B

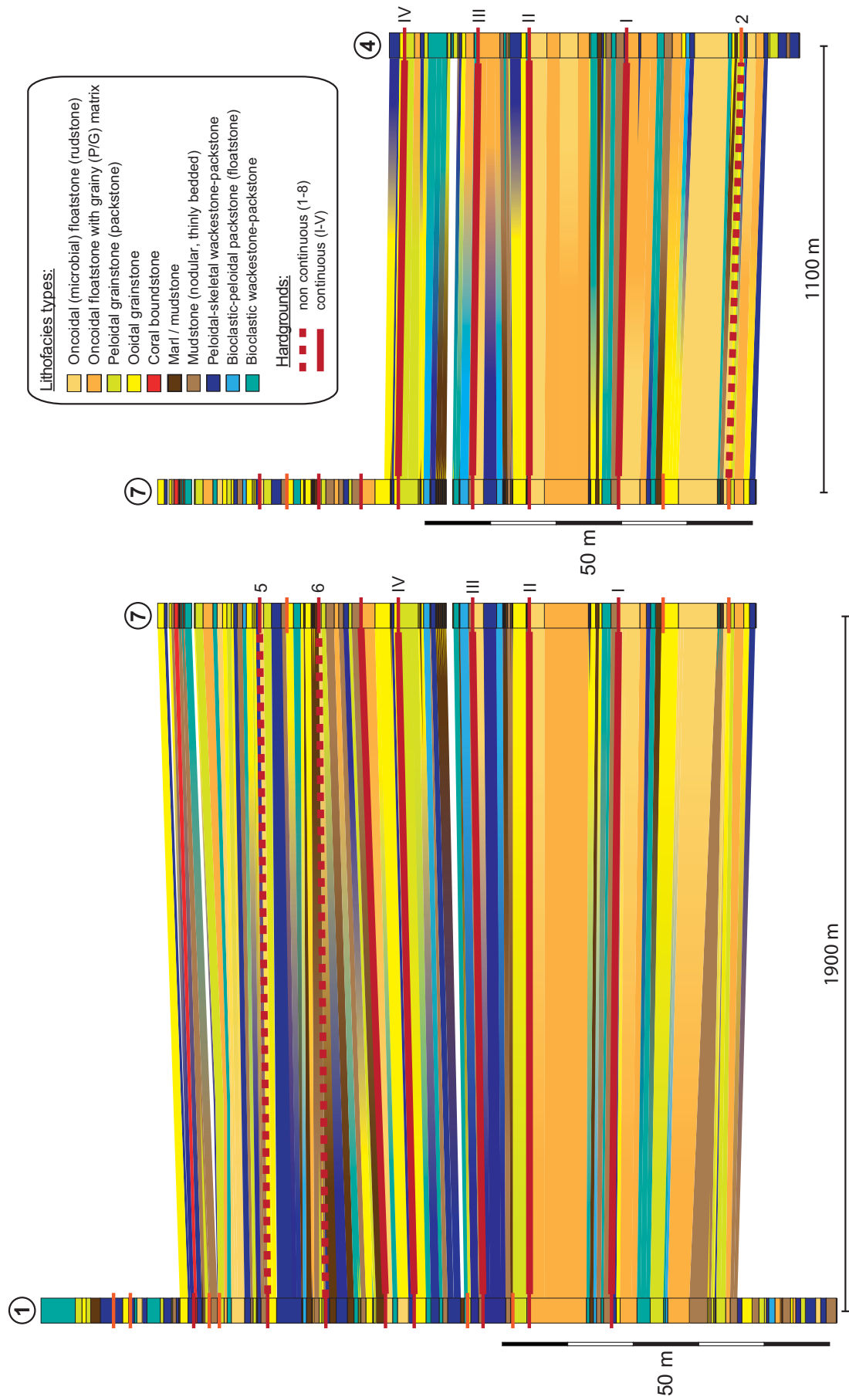


Fig. 6.6 (A, B) – Lithofacies maps of the outcrop between the measured sections. Hardgrounds are indicated in red, divided into continuous (I-V) and discontinuous ones (1-8).

6.5.1.3 Vertical and lateral facies architecture

At the scale of the outcrop, an overall stratigraphic trend from peloidal-oidial limestones at the base to oncolites towards again peloid and ooid bearing beds at the top of the section is observed in all sections (Fig. 6.5). The Markov-chain analysis for the vertical lithofacies architecture reveals that the vertical stacking pattern is not random (Table 6.2A, B). Facies transitions where ooidial grainstones overlie marls and mudstones occur frequently within the studied strata and are thus more likely to occur. Hardgrounds can be found mostly on top of oncolites and ooidial grainstone beds, while they almost never occur on top of marls and mudstones. The hardgrounds are almost exclusively overlain by muddy, recessively weathering lithofacies types. All the vertical facies transitions with relatively high probabilities of occurring (~20% and higher) are visualised in Fig. 6.7A.

All beds are organised in a 'layer-cake' type architecture. A single instance of lateral bed pinch was observed, in the lower part of the outcrop between logs one and two. Lateral changes in bed thicknesses are however frequently observed. For example, the thick oncoid bearing unit in the centre of the logged sections varies in stratigraphic thickness between ~8 and 11 m from south to north. Lateral changes in lithofacies types within a single bed are also apparent. The lateral merging between different lithofacies types can be observed between most logs (Fig. 6.6A). The lateral facies merging that can be observed along the N-S 2D section of the outcrop (logs 1-4) is assumed to also exist between logs 1 and 7 and logs 4 and 7, as facies at log 7 are different than logs 1 and 4. Thus, stratigraphic correlations result in facies merging, rather than in facies pinch out.

The Markov-chain analysis for the lateral lithofacies architecture reveals that lateral lithofacies transitions do not occur randomly (Table 6.2C, D). In contrast to the vertical stacking, continuity of the lithofacies types between the different logs was also recorded. The calculated lateral transition probabilities show that, apart from the bioclastic and peloidal limestone (LF 4), all lithofacies

types have a probability of at least 19% of being laterally continuous between two logs, and thus over a distance of at least 220 m. Furthermore, lateral transitions between the two types of oncolite (LF 6 and LF 7) are very likely, as well as between the marls and mudstones (LF 1 and LF 2, Fig. 6.7B). Other lateral lithofacies relations are more complex. An ooidal grainstone (LF 9) for example is likely to be laterally continuous, but also has a high probability of merging into a peloidal pack- or grainstone (LF 3 or LF 8, Fig. 6.7B). Some lateral facies transitions are very unlikely to occur, as for example the lateral merging between an oncolite (LF 6 or LF 7) and a mudstone or a marl (LF 1 or LF 2). Oncoid bearing beds are relatively thick compared to other lithofacies types and are mostly continuous over the entire study area. These lithofacies types (LF 6 and LF 7) display the only example of a relationship between bed thickness and lateral facies continuity (Fig. 6.6A).

Table 6.2 - Markov-chain analysis for vertical upward and lateral lithofacies and hardground transitions of the seven logged sections. Refer to Fig. 6.5 for the facies colour code and to Table 6.1 for facies descriptions. The red colour indicates hardgrounds. Tables **A** (counted frequencies) and **B** (calculated probabilities) show vertical upward transitions from one facies to another, including hardgrounds. Only upward facies transitions and no transitions from a facies into itself at bedding planes are considered. Tables **C** (counted frequencies) and **D** (calculated probabilities) show lateral transitions from one facies to another between neighbouring sections, and to both sides, also showing transitions from a facies into itself (i.e. lateral facies continuity between logged sections). Discontinuity surfaces were not taken into account.

A) Vertical observed transition frequency matrix (770 transitions; $\chi^2=167.78$)

	Onc F	Onc G	M m	M nod	Pel G	Oo G	Pel P	BioPel P	Bio W	Hardground
Onc F	-	4	3	7	4	6	8	0	6	12
Onc G	6	-	2	8	7	6	6	3	2	11
M m	7	2	-	10	3	27	16	3	19	1
M nod	7	10	9	-	12	24	22	3	13	1
Pel G	3	1	8	11	-	10	14	1	9	9
Oo G	6	6	31	16	11	-	23	2	11	21
Pel P	6	7	18	24	14	20	-	4	10	13
BiPe P	0	0	4	3	1	4	3	-	3	2
Bic W	3	9	10	20	4	14	6	3	-	6
Hardground	2	0	16	24	5	6	15	0	8	-

B) Vertical observed transition probability matrix

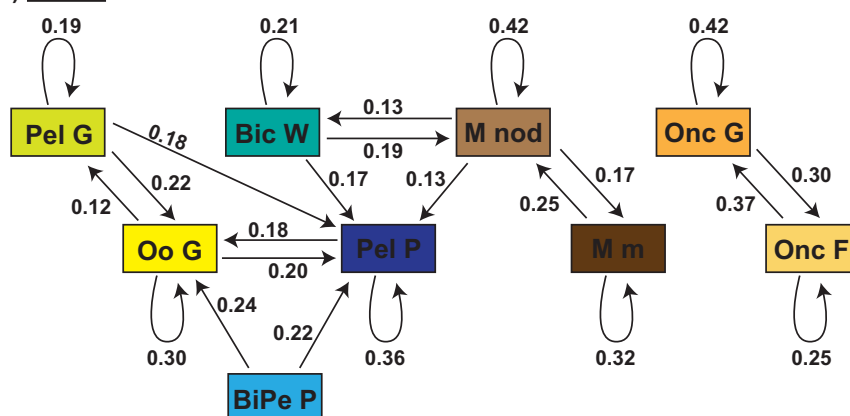
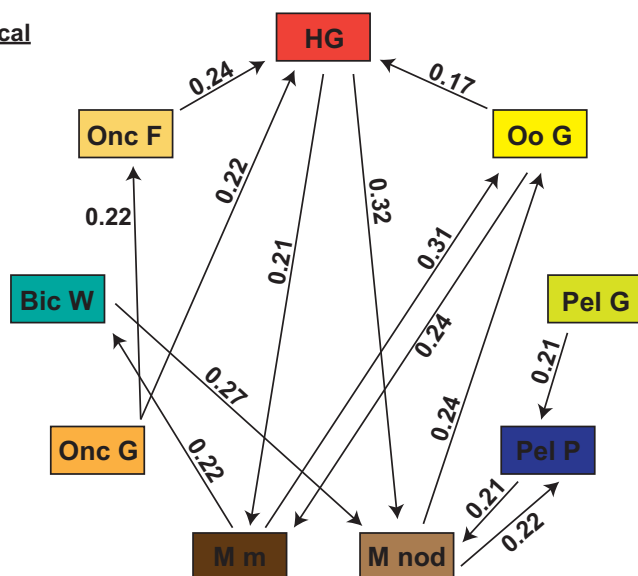
	Onc F	Onc G	M m	M nod	Pel G	Oo G	Pel P	BioPel P	Bio W	Hardground
Onc F	-	0.08	0.06	0.14	0.08	0.12	0.16	0	0.12	0.24
Onc G	0.12	-	0.04	0.16	0.14	0.12	0.12	0.06	0.04	0.22
M m	0.08	0.02	-	0.11	0.03	0.31	0.18	0.03	0.22	0.01
M nod	0.07	0.10	0.09	-	0.12	0.24	0.22	0.03	0.13	0.01
Pel G	0.05	0.02	0.12	0.17	-	0.15	0.21	0.02	0.14	0.14
Oo G	0.05	0.05	0.24	0.13	0.09	-	0.18	0.02	0.09	0.17
Pel P	0.05	0.06	0.16	0.21	0.12	0.17	-	0.03	0.09	0.11
BiPe P	0.00	0.00	0.20	0.15	0.05	0.20	0.15	-	0.15	0.10
Bic W	0.04	0.12	0.13	0.27	0.05	0.19	0.08	0.04	-	0.08
Hardground	0.03	0.00	0.21	0.32	0.07	0.08	0.20	0.00	0.11	-

C) Lateral observed transition frequency matrix (1062 transitions; $\chi^2=703.21$)

	Onc F	Onc G	M m	M nod	Pel G	Oo G	Pel P	BioPel P	Bio W
Onc F	18	27	1	2	7	8	8	1	1
Onc G	27	38	0	0	3	12	7	2	2
M m	1	0	36	28	6	11	12	1	16
M nod	2	0	28	70	9	10	21	3	22
Pel G	7	3	6	9	18	20	17	4	9
Oo G	8	12	11	10	20	52	35	10	15
Pel P	8	7	12	21	17	35	72	9	19
BiPe P	1	2	1	3	4	10	9	4	7
Bic W	1	2	16	22	9	15	19	7	24

D) Lateral observed transition probability matrix

	Onc F	Onc G	M m	M nod	Pel G	Oo G	Pel P	BioPel P	Bio W
Onc F	0.25	0.37	0.01	0.03	0.10	0.11	0.11	0.01	0.01
Onc G	0.30	0.42	0.00	0.00	0.03	0.13	0.08	0.02	0.02
M m	0.01	0.00	0.32	0.25	0.05	0.10	0.11	0.01	0.14
M nod	0.01	0.00	0.17	0.42	0.05	0.06	0.13	0.02	0.13
Pel G	0.08	0.03	0.06	0.10	0.19	0.22	0.18	0.04	0.10
Oo G	0.05	0.07	0.06	0.06	0.12	0.30	0.20	0.06	0.09
Pel P	0.04	0.04	0.06	0.11	0.09	0.18	0.36	0.05	0.10
BiPe P	0.02	0.05	0.02	0.07	0.10	0.24	0.22	0.10	0.17
Bic W	0.01	0.02	0.14	0.19	0.08	0.13	0.17	0.06	0.21



6.5.2 Stable isotope measurements

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Filtered oxygen isotope values range between -4.0 and -2.0 ‰ with the majority of results between -3.5 and -2.5 ‰ (Fig. 6.8).

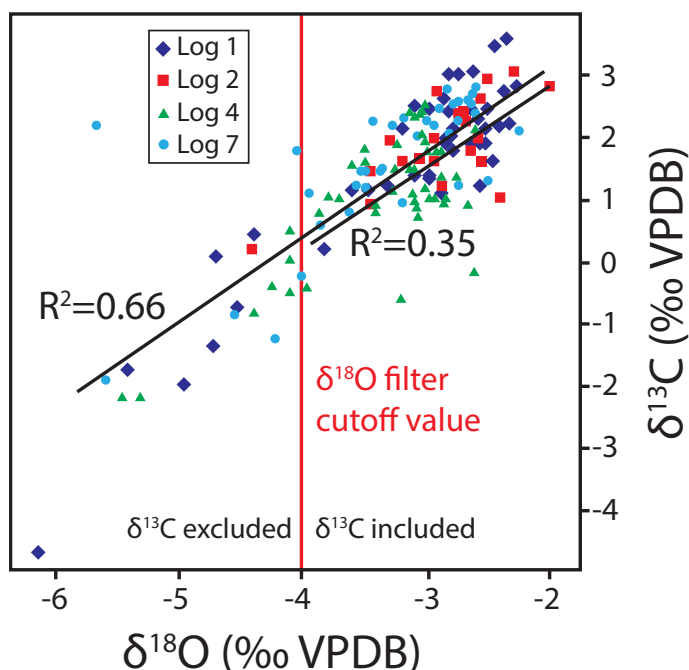


Fig. 6.8 - Crossplot between unfiltered $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ results, with R^2 values calculated with and without the application of the $\delta^{18}\text{O}$ filter (see section 6.4.3). The red line marks the cut-off for the $\delta^{18}\text{O}$ -filter: values to the left of the line are regarded as diagenetically altered and only the “filtered” values to the right of the red line are considered in further discussions.

When the carbon and oxygen isotope measurements are plotted against the respective stratigraphic sections, it is apparent that all four sections are characterised by frequent shifts in isotopic ratios towards more negative values, with amplitudes of up to 3 ‰ (Fig. 6.9). The excursions in carbon and oxygen isotopes are generally recognised around hardgrounds, which can partly be correlated across the four sections (marked E1-6). This is especially apparent in section 7. The upper half of section 1 shows a positive trend in both carbon and oxygen isotopic values that spans over approximately 40 m of section where mostly micrite of peloidal wacke- and packstone (LF 3) were measured. Although this part of section 1 contains seven hardgrounds, no large shifts are visible in both the carbon and oxygen isotopes. This is contrasted by the lower and upper part of section 1 as well as by the other measured sections. Overall,

the acquired isotope data is very heterogeneous and the overlap of all data points between the four sections for carbon and oxygen isotopes is poor.

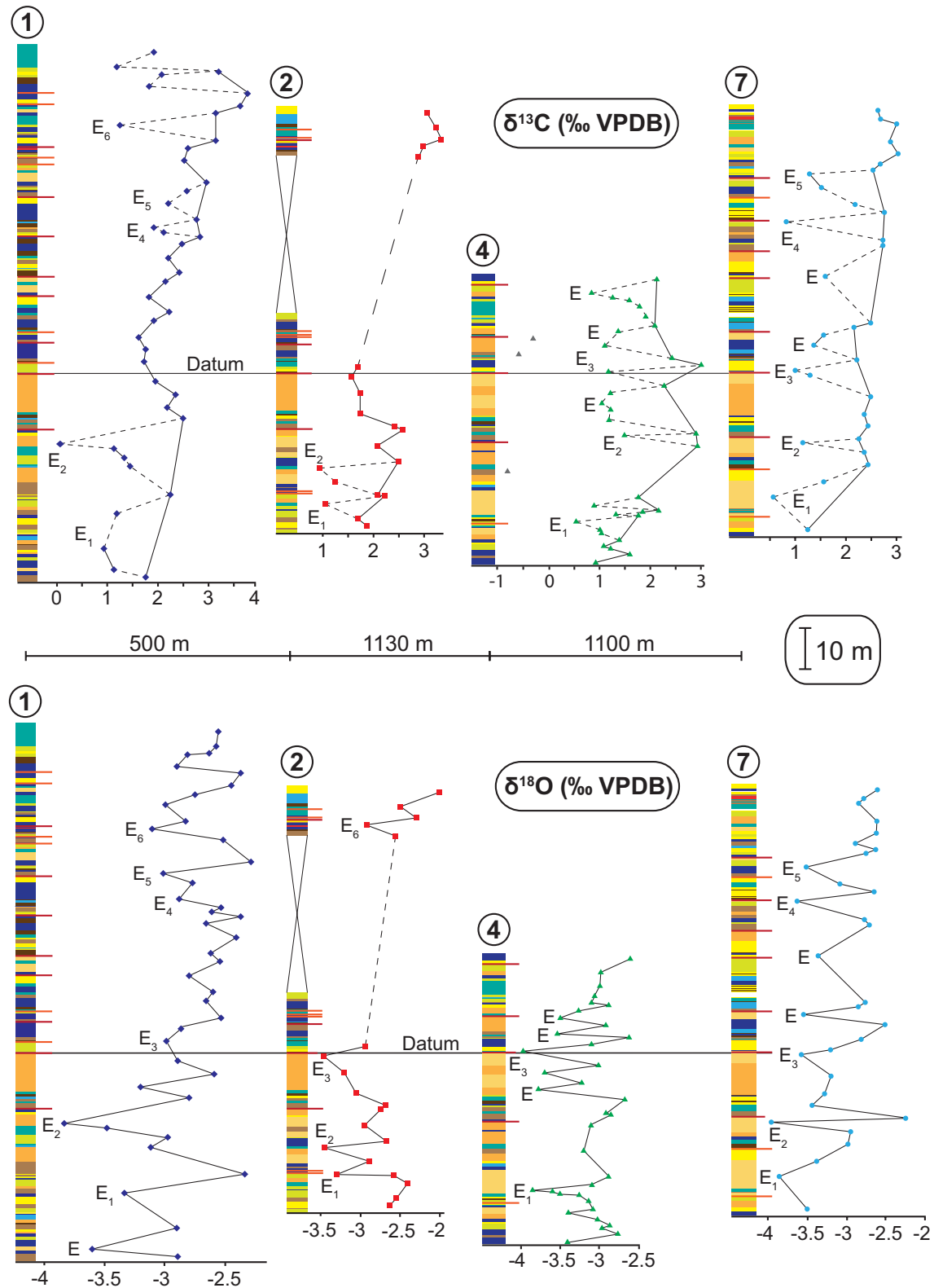


Fig. 6.9 – Filtered stable isotope data from four sections plotted against stratigraphic height. The $\delta^{18}\text{O}$ filter is applied and $\delta^{13}\text{C}$ results with respective $\delta^{18}\text{O}$ values below 4.0 ‰ have been

excluded. Values in grey are regarded as outliers. Average isotope values and standard deviations are indicated for each log. Letters E indicate interpreted exposure stages where $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ curves show negative excursions. Some exposure stages (E_1 - E_6) can be correlated between the sections. Data points with an inferred oceanic $\delta^{13}\text{C}$ signal are connected with a solid line. Connections with a dashed line represent data points interpreted as the result of subaerial exposure during the Middle Jurassic. Hardgrounds within the logged sections are indicated with red lines. Refer to Fig. 6.5 for the facies colour code.

6.6 Interpretation and discussion

6.6.1 Depositional environment

Evidence from the different lithofacies suggests deposition within the intertidal and subtidal regime. Clear markers for subaerial exposure, such as rhizoturbation, epikarst and meteoric cements (e.g. Rameil et al., 2011) are lacking. The shallowest, oncolite facies indicates a shallow and restricted environment and a relatively low hydrodynamic level at most, but no emergence. The oncoid bearing lithofacies, based on their relatively large diameters represent the shallow part of the intertidal to subtidal zone or a lagoon with potentially restricted conditions (Hamon and Merzeraud, 2008; Brigaud et al., 2009). The firmgrounds or hardgrounds visible throughout the interval of study (Fig. 6.5) are interpreted as unconformities linked to marine flooding surfaces based on evidence for early lithification and sediment starvation. Marine borings indicate an early diagenetic origin (Sattler et al., 2005; Rameil et al., 2011). Field evidence for subaerial exposure of any surfaces is lacking but the stable carbon and oxygen isotopes show more negative values around many of the recognised surfaces, possibly indicative of precipitation from meteoric fluids depleted in ^{18}O (Fig. 6.9, see section 6.6.5 below). Many of the marine flooding surfaces could thus be former exposure surfaces, overprinted by marine diagenesis during the subsequent transgression and are now composite surfaces, as observed in other carbonate ramp settings (Rameil et al., 2011).

Although evidence for a barrier is absent regionally (Al-Saad and Sadooni, 2001), a relatively local grain shoal could be postulated. Oncoidal limestones with a grainy matrix were deposited where low-energy sediments

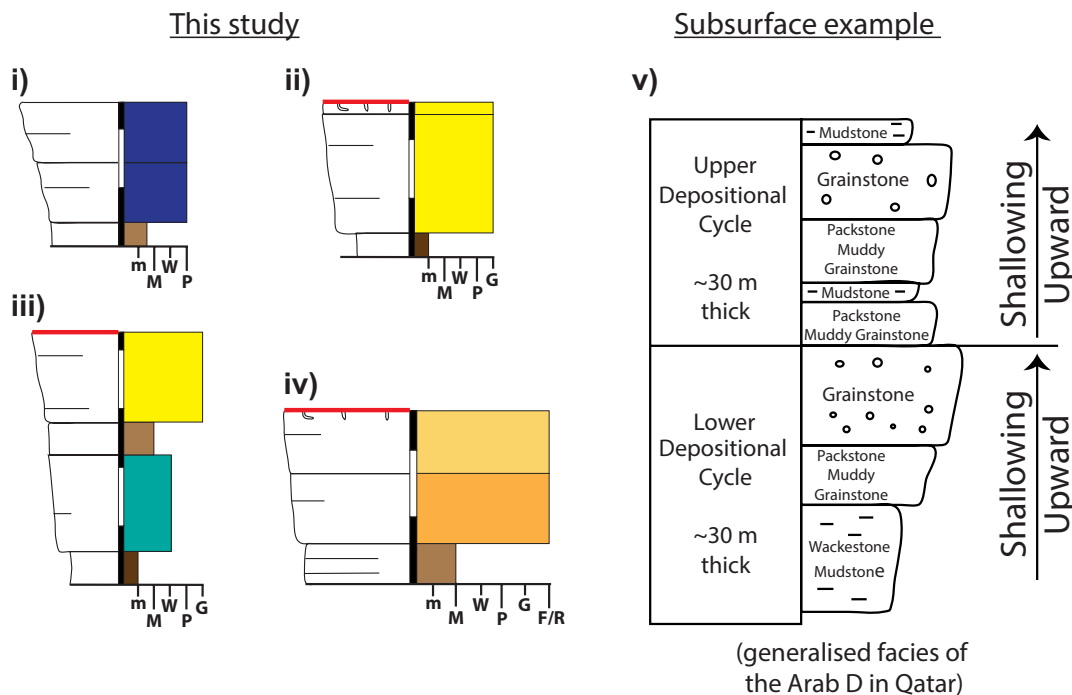
were mixed with reworked grains from adjacent, higher energy zones. All the peloidal-bioclastic limestones are considered to be deposits of moderate energy, based on the relative diverse grain associations and the abundance of lime mud. The pack- and grainstones consisting solely of peloids and/or ooids are attributed to the high-energy zone, and were likely deposited in shoals or bars. However, cross-stratification was only observed in a single bed in the ooidal grainstone facies. The branching corals preserved in boundstone textures are interpreted as a small patch reef and are assigned to the outer shallow-ramp to mid-ramp (Brigaud et al., 2009; Christ et al., 2011). The deposits interpreted to have recorded the deepest water-depth are mudstones and marls, occasionally containing brachiopods but otherwise fine-grained and indicative of low-energy levels.

6.6.2 Controls on lateral and vertical facies architecture at Wadi Naqab

The overall lateral distribution of lithofacies types resembles a facies belt pattern where the environments of deposition (EODs, Fitch et al., 2014) are relatively extensive laterally, with the muddy facies as deepest deposits, the more grainy facies in the higher energy shallow to middle ramp, and the oncolites representing the shallowest deposits (Fig. 6.10B). The lithofacies maps of the outcrop as well as Markov-chain analysis however point to large amounts of facies heterogeneity, and show that some lithofacies types are more frequently continuous across the study window than others. This is particularly apparent within the shallow, high-energy ramp EOD where facies types show the smallest probability of being continuous (Fig. 6.10B). The facies types deposited here, such as the peloidal, bioclastic wacke- and packstones as well as ooidal and peloidal grainstones are likely to laterally merge into each other (Fig. 6.7B, 6.10A). By contrast, sediments deposited in less agitated water, either more distally on the middle ramp or more proximally within a sheltered lagoon, have a higher probability of being continuous over 100's to 1000's of metres (Fig. 6.7B, 6.10B). Thus, relative sea level is the main controlling factor for facies distribution and results in facies belts only at the scale of the EODs (Rankey, 2004). At the facies scale however, waves and currents will have the highest impact on the (re-) distribution of carbonates, especially within the high-

energy, shallow ramp. It is proposed that the constant reworking of different lithofacies types by hydrodynamic processes of varying intensity (we renounce distinguishing between fair and storm weather waves for this shallow setting in accordance with findings by Peters and Loss, 2012) produced the heterogeneous facies mosaic that is preserved at Wadi Naqab. The relatively small grain size of ooidal and peloidal sands (mostly $<200\text{ }\mu\text{m}$), combined with the absence of physical evidence for tide activity, furthermore leads to the interpretation that the shallow ramp, where most of the lithofacies heterogeneity is present, was dominated by waves, rather than by tidal currents (Rankey, 2013).

A) Vertical stacking patterns



B) Conceptual depositional profile

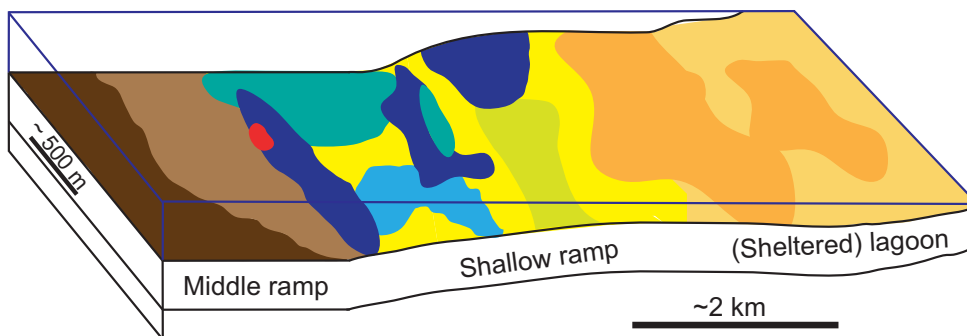


Fig. 6.10 - A) Shallowing-upwards parasequences (i - iv) with a high probability of occurring, based on vertical upwards transition probabilities as derived from Markov-chain analysis of the

logged sections for. Red lines indicate hardgrounds. The subsurface example (i) indicates a similar vertical facies architecture within an Arab-D reservoir (modified from Wilson, 1991). **B)** Conceptual depositional profile of the carbonate system showing adjacent facies types as indicated by lateral transition probabilities, derived from Markov-chain analysis of the logged sections. Refer to Fig. 6.5 for the facies colour code.

The vertical stacking pattern shows high transition probabilities of certain lithofacies types that produce shoaling-upwards cycles, which are frequently capped by hardgrounds or composite surfaces (Fig. 6.7A). These surfaces are typically overlain by marls and mudstones of the transgressive stage (Fig. 6.7A). We note that different lithofacies types on the carbonate ramp are partly distributed in a mosaic pattern, thus facies as proxies for changes in palaeowater-depth have to be regarded with some caution (*sensu* Purkis et al., 2015b; Harris et al., 2015). However, a typical parasequence or shoaling-upwards parasequence at Wadi Naqab thus starts with marl or mudstone, overlain either by peloidal packstone, ooidal grainstone or oncolites. The oncolites and the ooidal grainstones are frequently capped by a composite surface or hardground, as revealed by the Markov-chain analysis (Fig. 6.10A). Nineteen such small-scale parasequences (~3 – 8 m thick) were identified (Fig. 6.5). These shoaling-upwards parasequences often do not contain a complete facies transition from relatively deep to relatively shallow deposits and instead terminate with shallow- to mid ramp high-energy facies types (Fig. 6.5, 6.10A). They are thus “facies-incomplete” shoaling-upwards parasequences, which indicates, at least in part, a rapid fall in relative sea-level subsequently terminating carbonate production and often leading to exposure (Soreghan and Dickinson, 1994).

Rather than interpreting the occurrence of different orders of change visible in the sedimentary record as cyclicity, which would assume a good control on the timing of the deposits, we simply acknowledge the rhythmical pattern in the studied strata. Multiple studies have previously discussed whether peritidal cyclicity or rhythmicity observed in carbonates is the result of auto- or allocyclic processes (see e.g. Bosence et al. 2009, for a detailed review). Various allocyclical controlling factors such as relative sea level change due to tectonic pulses, glacioeustasy, climatic or oceanographical changes can be

considered as possible mechanisms to produce the stratigraphic pattern observed at Wadi Naqab. The Arabian Plate is considered to have been a passive margin during the Jurassic, so tectonically forced relative sea level change is unlikely. The peritidal cyclicity in the Lower Jurassic of Wadi Naqab, stratigraphically below our window of study, where omission surfaces with clear field evidence for subaerial exposure are present, has previously been attributed to eustatic changes induced by orbital forcing (Walkden and de Matos, 2000). Since stratigraphic age constraints in the Middle Jurassic of Wadi Naqab are relatively poor, it is not clear if the observed discontinuity surfaces are visible on a regional scale and can be correlated to the regional maximum flooding surfaces of Sharland et al. (2001). However, no mechanism other than a drop in relative sea level can be invoked to have caused subaerial exposure stages.

Another potential climatically induced effect on the vertical stratigraphic pattern at Wadi Naqab is input of nutrients and clastic material in this shallow ramp setting, through continental runoff. Siliciclastic material in the form of quartz grains is present within some of the mudstones that form the primary transgressive deposit of several of the shoaling-upwards parasequences. This might indicate that sea level rose during more humid periods (Strasser and Védérine, 1999) and that climatic changes played a role in controlling vertical rhythmicity. Global climate modelling suggests that precipitation was possible all year round during the Jurassic along the equator (Sellwood and Valdes, 2008), and strong cyclical storm events can be postulated for the equatorial climate that could have eroded unconsolidated sediment on top of hardgrounds. The neritic carbonate factory that was dominated by non-skeletal grains and benthic organisms would have thrived in meso- to oligotrophic conditions (James, 1997; Halfar et al., 2006), suggesting that periods of increased influx of nutrients could have favoured argillaceous marl deposition and suffocated the neritic carbonate factory. The majority of lime mud in this type of setting is however believed to be produced within the neritic carbonate factory (Pomar and Hallock, 2008), indicating that this factory was only brought to a standstill for relatively short intervals, although the elapsed time during exposure and during flooding stages is virtually impossible to calculate, when the age constraints remain relatively loose. Sedimentation would have only resumed on the shallow ramp after a

storm passed. Deeper, mid- to outer ramp settings could have still accommodated for marl and mudstone deposition. Indications for more arid conditions, for example in the form of evaporites, are absent. Early meteoric carbon and oxygen isotope signals associated with hardgrounds serve as further evidence for periods of humid climate.

In conclusion, the observed vertical stratal pattern is most likely a result of changes in relative sea level and thus of allocyclic origin, with the added complexity that the high-energy environment of deposition of the shallow ramp was also very heterogeneous and likely influenced by climatic changes.

6.6.3 Comparison to other ancient ramp settings and facies mosaics

The observations made in Wadi Naqab are in line with other studies that have demonstrated complex heterogeneities within carbonate systems at the inter-well scale. The preservation of a Jurassic facies mosaic from a similar subtidal setting is visible in data from the Amellago Formation in Morocco (Amour et al., 2011; Christ et al., 2011). This ramp contains a similar vertical rhythmic architecture interpreted as the result of rapid basement subsidence and the continuous creation of accommodation (Christ et al., 2011), but no evidence for subaerial exposure. Of all the identified discontinuity surfaces, two thirds are laterally continuous over the study area in Wadi Naqab, and thus extend over a minimum of 1900 x 1200 m. The remaining surfaces are either only visible in one section or extend for 220 to 1100 m. This is also comparable to the grainy part of the Amellago ramp (Christ et al., 2011). Another Jurassic analogue study (Kimmeridgian, Spain) presented by Alnazghah et al. (2013) reveals downdip facies heterogeneity along a ~1.3 km long 2D transect of a carbonate ramp and several accretionary units with frequent examples of lateral facies merging. Oxfordian shallow carbonate platforms from the Swiss Jura Mountains also contain lateral facies variability within only a few metres (Samankassou et al., 2003), mostly linked to coral patch reefs and small changes in relative sea level. Sattler et al. (2005) and Sena and John (2012) show preserved facies mosaics from Cretaceous units in Oman. When the logs measured by Sattler et al. (2005, Fig. 4) are correlated they result in a facies map showing a comparable amount of lateral facies change to Wadi Naqab. The Cretaceous facies mosaic

presented from the Western Tethys by Brandano and Loche (2014) is similarly heterogeneous. These different Mesozoic case studies from across the Tethys illustrate that the sedimentary record is capable of preserving small-scale facies mosaics.

However some detailed outcrop studies have revealed a largely layer-cake architecture with virtually no facies heterogeneity (e.g. Zeller et al., 2011; Pratt et al., 2012). The question thus remains as to why some settings produce and preserve highly heterogeneous facies mosaics while others do not. Many field-based studies, at least in the Middle East, have focussed on larger, regional-scale geometries (e.g. Koehrer et al., 2010; 2011; 2012; Bendias et al., 2013) rather than on an inter-well scale of one to a few kilometres. Additionally, large-scale studies tend to often not go into detail when examining the inter-well scale, and it can be speculated whether small-scale facies mosaics were not preserved (Wright and Burgess, 2005), or were overlooked or have merely not been the particular focus of a given study. The studies that do demonstrate inter-well scale heterogeneities, irrespective of their stratigraphic age, were all performed on shallow parts of open carbonate platforms or ramps. The general facies associations are grain-dominated with or without reef-builders. All studies have concluded that hydrodynamic processes, with their intensity controlled largely by relative sea level, play a major role in producing facies heterogeneities (French and Kerans, 2004; Grélaud et al., 2010; Sena and John, 2013; Alnazghah et al., 2014; Brandano and Loche, 2014). The Jurassic example from Morocco concludes that autocyclic processes could have produced the observed facies pattern, due to the absence of subaerial exposure stages (Christ et al., 2011). A homogeneous example from the Paleozoic (Pratt et al., 2012) on a shallow and grain-dominated platform contains no continuous hardgrounds and shows no evidence for significant changes in relative sea level or oceanographic conditions. Most likely it was deposited during a very stable period with respect to base level changes (Pratt et al., 2012) and is unique compared to other examples, although all come from greenhouse worlds, and all contain evidence for frequent oscillations in relative sea level.

6.6.4 Analogy and applicability to Middle Eastern reservoirs

Shallow marine carbonates are reported as main reservoir rocks from the Jurassic in the Middle East (e.g. Sudarsana et al., 2009) and comprise shoaling-upward cycles (Holail et al., 2006), including in the younger and economically more important Arab-D units (e.g. Wilson, 1991). These cycles are composed of the same facies types as the ones observed in Wadi Naqab, although the stratigraphic thickness of the cycles varies (Fig. 6.10). A high amount of heterogeneity is present within some of the producing fields (e.g. Stanley et al., 2014). Typical Middle and Upper Jurassic reservoir facies types include (laminated) mudstones, wackestones (sometimes slightly argillaceous), peloidal pack- and grainstones (often containing benthic foraminifera) with interbedded marls, clean grainstones (containing ooids, peloids and bioclasts) and oncoidal packstones and are thus comparable to the lithofacies types in this study (Wilson, 1991; Alsharhan and Nairn, 1994; Alsharhan and Whittle, 1995; Al-Saad and Sadooni, 2001; de Matos, 2002; Alsharhan and Magara, 1994; Holail et al., 2006; Sudarsana et al., 2009). The high amount of lateral facies heterogeneities within the high-energy shallow- to mid-ramp revealed in Wadi Naqab, combined with transition probabilities matrices, can provide valid input for reservoir modelling. Dolomites are an abundant reservoir facies throughout the Middle East but are only present in burrows and scours within hardgrounds in the Middle Jurassic strata of Wadi Naqab, which is thus a limiting factor when drawing comparisons between the outcrops and the subsurface. In addition, the distribution of hardgrounds that may act as baffles to fluid flow in the subsurface is not necessarily visible in cores or seismic data, and the dataset from Wadi Naqab thus potentially provides useful constraints.

6.6.5 Carbon and oxygen isotope heterogeneity and chemostratigraphy

Different studies have demonstrated that neritic carbonate systems can be used for carbon isotope chemostratigraphy, including case studies using outcrops from the Middle East (e.g. Vahrenkamp, 1996; Immenhauser et al., 2004; Sena and John, 2013). The $\delta^{13}\text{C}$ value of carbonates is believed to be close to that of the dissolved inorganic carbon and thus represents changes in global carbon

cycling, without much effect caused by temperature and vital effects (Saltzman and Thomas, 2012). The $\delta^{18}\text{O}$ value of certain pristine fossils is widely used as palaeothermometer, or as proxy for diagenesis in bulk rock or micrite phase datasets. However, although the sections at Wadi Naqab are only 500 to 1130 m apart and beds can be traced across the study window, the coeval isotope records show marked scatter and no clear common trend is apparent at first (Fig. 6.9). The isotope values are very variable in all four sections, especially in the lower half of the sections below the reference datum (hardground II). Coeval negative excursions in oxygen and carbon isotope values are attributed to subaerial exposure around hardgrounds and subsequent input of terrestrial organic matter, partly through soil generation and carbonate precipitation from meteorically influenced fluids, resulting in depleted carbon and oxygen isotope values. Six exposure surfaces (marked E1-6) with this characteristic isotope pattern have been identified within the four sections. Further negative and positive shifts with amplitudes of $> 1\text{‰}$, especially between logs 4 and 7, might contain more such events but cannot be clearly recognised (Fig. 6.9). Since the distance between the sections is relatively small, it could be expected to recognise the same exposure stages in all four sections. This is however not the case here and is attributed to the sample resolution (2-5 meters) which is too coarse to fully resolve all hardground surfaces.

What controls the lateral and vertical isotopic heterogeneity at Wadi Naqab? Carbon isotope heterogeneity between closely spaced sections has been reported by Immenhauser et al. (2002) from a Carboniferous carbonate platform and attributed to the mixing of water masses with different isotopic signatures, according to the distance to the shelf break. Furthermore Immenhauser et al. (2008) have put forward concepts to further the understanding of carbon isotope heterogeneities from shallow-marine settings and conclude that bathymetry, hinterland carbon flux and changes in relative sea level with all its consequences will have the most profound control over isotopic sea water properties. The burial history for the four measured Wadi Naqab sections can be regarded as identical as well as the bathymetry and the overall lithofacies types, despite the identified facies heterogeneities. The depositional setting for all sections at Wadi Naqab is more or less identical and

the exact configuration of the sections relative to the palaeo-coastline is not known. Thus, explaining the observed isotopic trends at Wadi Naqab by mixing of water masses with different isotopic compositions is difficult given the closely spaced nature of the sections, all deposited on a homoclinal, shallow epeiric ramp and separated by around a kilometre.

As explained earlier, we interpret the frequent negative excursions visible in all of the carbon isotope curves as result of subaerial exposure stages during lowstands, which affected all four sections. When the depleted values caused by early meteoric alteration are excluded, and only the more positive data points are connected, a broad common trend from ~1.5 ‰ to 3.0 ‰ is visible in the dataset (marked by continuous lines in Fig. 6.9). This trend could be either caused by an overall transgression (Immenhauser et al., 2003), a change of the carbonate factory (Föllmi et al., 2006; Swart, 2008), or by burial diagenesis. However, it would be expected that progressive burial would result in a negative trend in oxygen isotope values representing warmer temperatures, which is not the case (Fig. 6.9). Changes in facies due to an overall transgression or change in carbonate factory can also be discarded due to the repetitive stratigraphic pattern. We therefore interpret the overall positive trend in carbon isotope values as a primary, oceanic signal, which is comparable to an inferred Jurassic seawater $\delta^{13}\text{C}$ of around 2 ‰ (Veizer et al., 1999).

The Bajocian age attributed at Wadi Naqab by de Matos (1997) is based on benthic foraminifera, stromatoporids, algae and a single, equivocally classified ammonite. The biostratigraphic markers give some constraints but could benefit from additional chronostratigraphic constraints. Given the lack of an unequivocal reference curve for the Middle Jurassic (Wetzel et al., 2013), using $\delta^{13}\text{C}$ for chemostratigraphy at Wadi Naqab is uncertain. Nevertheless, the broad carbon isotope trend from 1.5 ‰ to 3.0 ‰ can be compared to two of the most commonly cited Bajocian carbon isotope curves, the Tethyan reference sections from the Subbetic Basin (O'Dogherty et al., 2006) and from the Umbria Basin (Bartolini et al., 1999). The $\delta^{13}\text{C}$ trend at Wadi Naqab correlates well with the two reference curves (Fig. 6.11), thus confirming a Bajocian age assignment for this part of the Jurassic section, and even allowing to narrow-down the age to the Early Bajocian.

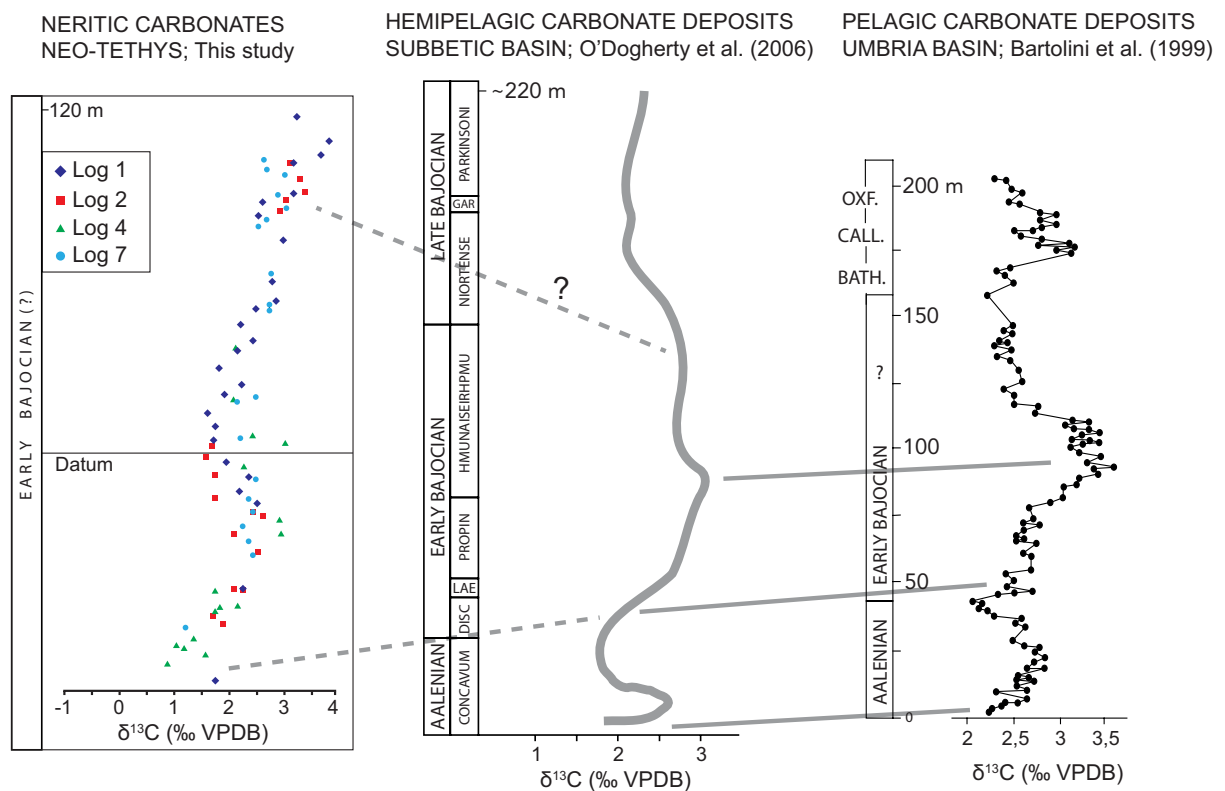


Fig. 6.11 - Composite $\delta^{13}\text{C}$ curve with all data points interpreted as being of oceanic origin from Wadi Naqab (Fig. 6.9) in comparison with Western European Middle Jurassic reference sections. The $\delta^{13}\text{C}$ chemostratigraphy at Wadi Naqab complements the biostratigraphic framework (de Matos, 1997).

7.7 Conclusions

The lithofacies types deposited on the intertidal to subtidal part of the carbonate ramp show a high degree of lateral and vertical variability, despite a largely 'layer-cake' bedding pattern. Lithofacies types have a probability of ~60 % of being discontinuous between logged sections separated by a few hundred to 1130 metres. Facies continuity is smallest within the high-energy shallow ramp and largest in the less agitated sheltered lagoon and middle ramp. The large amount of lateral heterogeneity, facies merging and interfingering is attributed to wave and current action within the higher energy shallow ramp, and it is postulated that this will be the case for most carbonate ramp settings.

The vertical stratigraphic stacking pattern is dominated by shoaling-upwards parasequences, frequently capped by composite surfaces. Evidence

for subaerial exposure is present in the form of depleted carbon and oxygen isotope values accompanying several hardground stages, indicative of early meteoric influence. This vertical cyclicity was most likely caused by frequent changes in relative sea level, superimposed onto the closely spaced facies mosaic, which is still preserved today. When the heterogeneous carbon isotopic signal produced by both meteoric and oceanic processes is carefully disentangled, a meaningful oceanic carbon isotope signature can be extracted from the neritic section, and the section can be attributed to the Early Bajocian stage using $\delta^{13}\text{C}$ chemostratigraphy.

In conclusion, multiple types of heterogeneities are present at Wadi Naqab, both in carbonate facies and isotopic signature. However, this heterogeneity is not random as trends between the amount of heterogeneity, different environments of deposition and the associated lithofacies types are apparent. This has important implications for enhanced hydrocarbon production from intertidal to subtidal carbonates in similar settings: if lithofacies templates petrophysical properties, then heterogeneities in flow units might be present and can be predicted and incorporated into reservoir simulations.

7. Metre-scale facies distribution in shallow-marine carbonate systems: An example from the Middle Jurassic of the United Arab Emirates

7.1 Abstract

A sedimentary bed is classically defined as a distinct layer of sedimentary rock that differs from the overlying and the underlying bed and has a relatively uniform composition. Several outcrop-based studies have however shown that facies types within individual beds can vary laterally on a scale of around 100 m ('inter-well scale'). As the facies type is an important criterion that is used as proxy for the depositional environment and that might template petrophysical properties, this observed facies heterogeneity has implications for both palaeoenvironmental studies and subsurface applications. The aim of this study is to investigate whether sedimentary and facies heterogeneities known to occur at the inter-well scale are also present at the metre to tens of metres scale. A bedset in an outcrop with a well-preserved facies mosaic deposited on the shallow-marine epeiric carbonate ramp (Lower Bajocian, Wadi Naqab, northern UAE) was logged and mapped across a 120 m long curving cliff face. Results reveal a large amount of lateral facies merging at the metre-scale. Lithofacies types have a probability of at least 30 % of being laterally discontinuous between sections spaced between 12 and 18 m apart. The results reported here show the highest amount of lateral facies heterogeneity when compared to other similar studies on ancient geological examples. The case study is also unique in that it reveals intra-bed facies merging in an overall 'layer-cake' architecture, something that has previously not been observed or investigated at this scale. The large amount of facies heterogeneity is attributed to spatially heterogeneous biogenic carbonate production as well as to syndepositional mixing and segregating processes occurring within shallow-marine depositional environments in ancient as well as in modern analogous systems. The existence of small-scale facies mosaics highlights the importance of gathering lateral information when conducting palaeoenvironmental, sedimentological and sequence stratigraphic studies.

7.2 Introduction

A sedimentary bed, also referred to as stratum, is defined as a distinct layer of sedimentary rock that differs from the overlying and underlying beds in one or many ways and that represents more or less uniform depositional conditions (Campbell, 1967; Compton, 1985).

Beds are generally the basic unit of stratigraphic studies and can represent anything from seconds ('event beds') to many thousands or even millions of years. Beds extend laterally from a few metres (typically in fluvial settings) to several kilometres, or even thousands of kilometres in extreme cases (Campbell, 1967; Einsele et al., 1991). A bed has a comparatively uniform composition, in contrast to the underlying and overlying beds (Einsele et al., 1991). Wright and Burgess (2005) however stress that it remains unknown whether a bed might represent a distinct habitat or rather a mixture of environments, averaged over time by different processes. By extension lateral heterogeneity within a bed could either be the result of locally different conditions on the seafloor at the time of deposition, or be caused by a laterally heterogeneous distribution of averaging processes such as for example bioturbation. The presence of lateral heterogeneities in a bed has important implications for two main fields: 1) Stratigraphic and palaeoenvironmental studies tend to examine one-dimensional (vertical) sections. The vertical stacking pattern of beds is often interpreted as the product of past environmental change. Lateral heterogeneity at the ten to hundred metre-scale is something that is mostly not taken into account in this type of study, other than on a relatively large scale for broad environments of depositions (such as transitioning from a near-shore to a basinal setting along a transect). If properties within a bed were to vary laterally on a scale of a few to hundreds of metres, then using a single one-dimensional section as the basis for palaeoenvironmental interpretation might misrepresent the environmental conditions at the time of deposition or the resulting sequence stratigraphic interpretation; and 2) When heterogeneities are present within hydrocarbon reservoirs they can have a strong influence on fluid flow and consequently recovery factors (Borgomano et al., 2008). Lateral as well as vertical sedimentological heterogeneities are thought to have a significant influence on

fluid flow in carbonates (Fitch et al., 2014) but can not be detected using typical subsurface datasets (e.g. well or seismic data). Understanding these heterogeneities and being able to implement them into reservoir or seismic models can lead to improved recovery factors.

There is however a gap in the presently available literature on the lateral continuity of lithofacies types within a single bed on a scale of a few metres to approximately 100 m. Carbonate lithofacies heterogeneities at the micro- to pore scale have been widely studied as they are of importance for modelling flow properties (see e.g. Blunt et al., 2013 for a review). A number of outcrop analogue studies investigating carbonate strata, with sedimentary logs spaced between approximately 100 m and several 100's of metres apart (sometimes referred to as the 'inter-well scale'), have been conducted. The main focus of these inter-well scale studies lies on the acquisition of qualitative and quantitative data in order to reduce uncertainties in subsurface applications and to investigate sedimentological processes at the given scale. Several of these studies have demonstrated that lithofacies types can vary laterally within individual beds between measured sections separated by as little as 100 m, despite a largely 'layer-cake' type architecture. Published examples of this phenomenon as well as high-frequency cycle analysis studies, occurring typically in the shallow-water areas of open carbonate platforms or ramps, include studies from across the world and from different periods (Cambrian: Adams and Grotzinger, 1996; Devonian: Frost, 2007; Permian: Kerans et al., 1994; Triassic: Kostic and Aigner, 2004; Palermo et al., 2010; Obermaier et al., 2015; Jurassic: Hamon and Merzeraud, 2008; Verwer et al., 2009; Bádenas et al., 2010; Amour et al., 2011; Christ et al., 2011; Alnazghah et al., 2013; San Miguel et al., 2013; Hönig and John, 2015; Pomar et al., 2015; Cretaceous: Immenhauser et al., 2004; Homewood et al., 2008; Sena and John, 2013; Brandano and Loche, 2014; Miocene: Tomás et al., 2010). These are complemented by modern analogue studies showing an irregular distribution of facies types (e.g. Wilkinson et al., 1999; Rankey and Reeder, 2011; Harris et al., 2015; Purkis et al., 2015a, b; Purkis and Harris, 2016). Outcrop studies with closer-spaced sections (lateral distance between logs approximately 2.5 - 35 m) are extremely rare (Kerans et al., 1996; Samankassou et al., 2003; Kostic and Aigner, 2004; Bádenas et al., 2010 – two logs only; Adams et al., 2011;

Obermaier et al., 2015) and mostly deal with closely spaced facies heterogeneities within shoal bodies with inclined, and wedge-shaped deposits, thus not representing a pure 'layer-cake' architecture. This however is not representative of many ancient examples of epeiric platforms, where beds were deposited as a 'layer-cake' on shallow-water shelves.

Here new results from a well-studied Middle Jurassic outcrop (Wadi Naqab, Emirate of Ras Al-Khaimah, UAE) are presented, showing a 'layer-cake' type stratigraphic architecture, typical of carbonate ramps in this region (e.g. Al-Husseini, 1997; Zeller et al., 2011; Hönig and John, 2015). At this location lateral lithofacies heterogeneities within individual beds were demonstrated at the 100 m scale (Hönig and John, 2015), and are now examined for heterogeneities at the scale of a few metres. The study focuses on a bedset with a thickness of 1.5 m, which was traced and mapped laterally on a curving cliff face for approximately 120 m, and therefore allows a pseudo-3D assessment of the continuity of the lithofacies types contained within. The dataset is also compared to the few other published examples measured at similar scales, and the underlying processes that control this lateral facies variability that can be preserved in shallow-marine carbonate settings are discussed. This study is the first to systematically target metre-scale heterogeneities on epeiric carbonate platforms.

7.3 Geological setting

The studied outcrop is located on the southern part of Musandam Peninsula, which is the most northern part of the United Arab Emirates (UAE) (Fig. 7.1A). The Musandam Mountains on the peninsula form the northwestern extension of the Oman Mountains and contain a well preserved, ~3 km-thick succession of shallow-water carbonates (Maurer et al., 2009). The eastern margin of the Arabian Plate was tectonically passive during the Permian and the Mesozoic. The overall relative tectonic quiescence was interrupted by two events, one during the Late Cretaceous when ophiolites were obducted on the Arabian plate, and the second one during the Late Oligocene – Early Miocene when the Oman and Musandam Mountains were formed during an initial collision phase of the Zagros orogeny (Glennie, 2005; Searle et al., 2014). From the Pliocene

onwards the eastern margin of the Arabian platform was in an active continental setting and the Arabian plate is currently being subducted beneath the Eurasian Plate (Sharland et al., 2001; Searle et al., 2014).

The area of the Musandam Peninsula was located within a few degrees of the equator during the Middle Jurassic (Tanner et al., 2004; Golonka, 2007). The area was covered by a shallow-marine carbonate platform during the Jurassic (Ziegler, 2001) (Fig. 7.1B). To the north, more open marine conditions prevailed. Terrigenous conditions existed to the south, with the coastline towards the west running roughly parallel to its present day location (Ziegler, 2001). Most of the UAE, Oman and Qatar were characterised by terrigenous to marginally marine environments, where shallow-marine clastics, evaporites and coastal and deltaic sediments were deposited (Al-Husseini, 1997; Ziegler 2001).

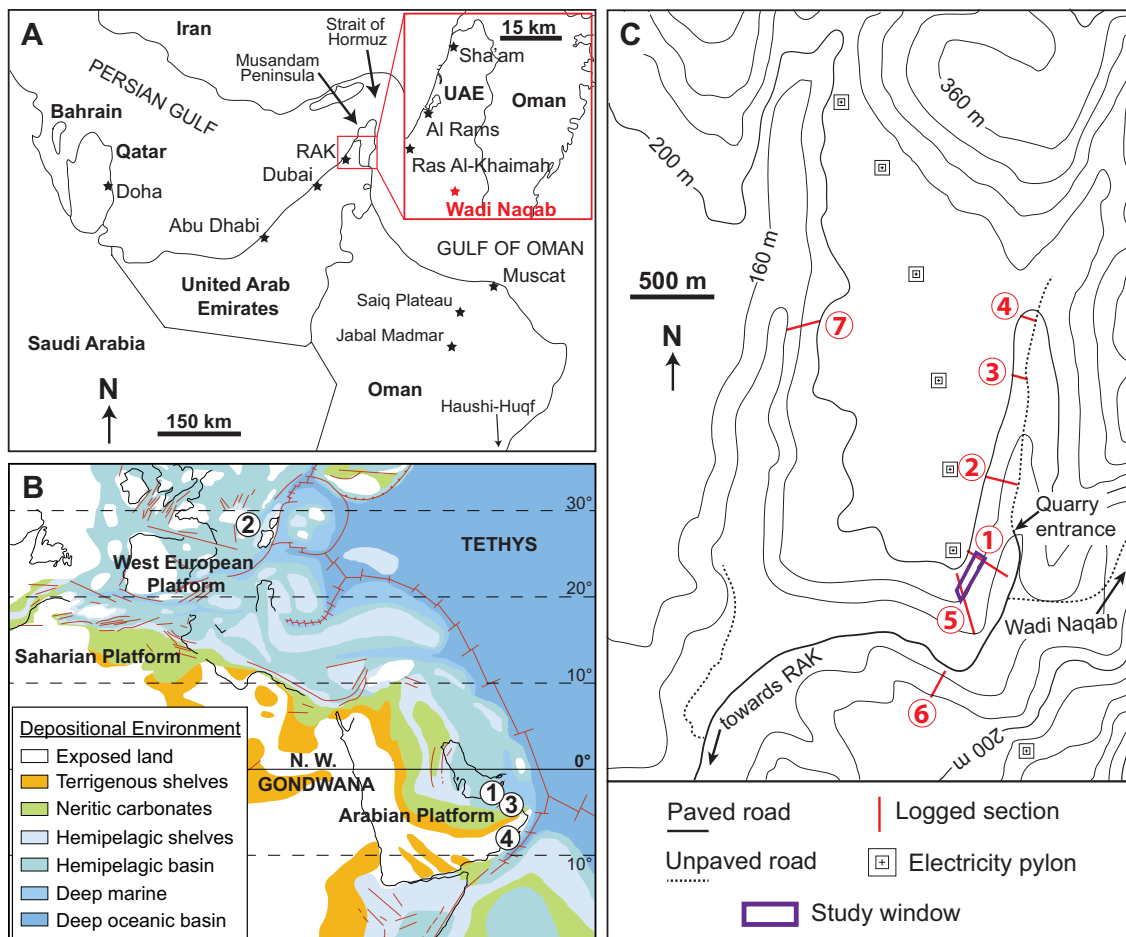


Fig. 7.1 – A) Location map of the study area. **B)** Palaeogeographical setting of the Tethyan realm during the Jurassic (modified from Pierre, 2006; after Thierry and Barrier, 2000). (1) Wadi Naqab; (2) Swiss Jura Mountains; (3) Saiq Plateau and Jabal Madmar; (4) Haushi-Huqf area.

C) Location of the logged sections in the 'block' outcrop at Wadi Naqab (modified from Hönig and John, 2015).

The Jurassic section on the Musandam peninsula has a stratigraphic thickness of approximately 1300 m and forms the Musandam Group. The Bajocian section alone expands over 550 m (de Matos, 1997). Bajocian-aged strata partly crop out in a 'block' bound to the south and west by wadis and on its eastern side by an abandoned quarry (Fig. 7.1C). A Bajocian age was assigned to the examined strata by de Matos (1997) based on biostratigraphy and later refined to a Lower Bajocian age using stable carbon isotope stratigraphy by Hönig and John (2015).

7.4 Materials and methods

7.4.1 Fieldwork, facies and sequence analysis

Ten sedimentary logs, spaced between 12 and 18 m apart, were measured along a 120 metre long transect within a well-exposed bedset, using a Jakob's staff and measuring tape (Fig. 7.1C; 7.2A). The bedset could be followed around a ledge (Fig. 7.2C), and is composed of three thin beds (ca. 10 cm each), overlain by a thicker (ca. 25 cm) bed, which is itself overlain by a ca. 75 cm thick bed (Fig. 7.2D). The selected bedset is situated within Middle Jurassic strata that was studied for inter-well scale sedimentological heterogeneities and is organised in a 'layer-cake type' architecture, between sections 1 and 5 of Hönig and John (2015) (Fig. 7.2B). Fist-sized samples were taken using a geological hammer and 39 thin sections (one in the lower half and three in the thicker, upper half of each section) were made, and subsequently analysed using a transmitted light Zeiss Axioskop 40 microscope (plain and polarised light) at Imperial College London. Textural classification of facies followed Dunham (1962) and Embry and Klovan (1971) and is based on field observations complemented by thin section analyses. The classification of vertical cycles on this very small scale was undertaken in a conventional manner based on facies stacking patterns (e.g. Bádenas et al., 2010; Pöppelreiter et al., 2011; Obermaier et al., 2015). Facies types were used as

proxies for relative water depth based on the depositional model of Hönig and John (2015) with three bathymetric zones: (i) lagoon; (ii) shallow ramp and (iii) middle ramp. Although the vertical scale is very small a cyclostratigraphic analysis was performed in order to compare the results to other studies where a similar analysis of very small cycles is routinely performed, and as a means to evaluate this approach in a location where lateral data is available.

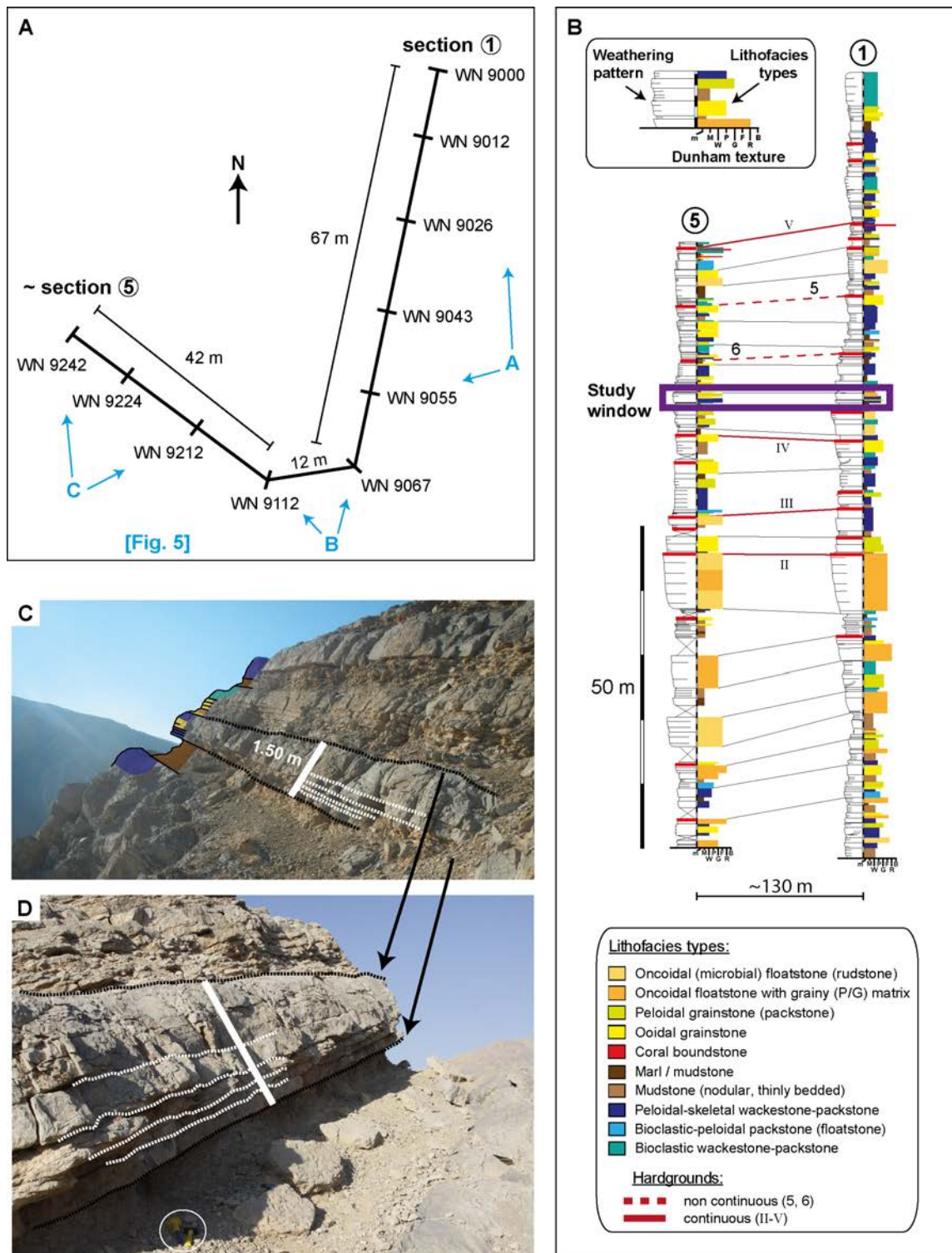


Fig. 7.2 – Overview of the studied bedset at Wadi Naqab. **A)** Configuration of the ten logged sedimentary sections. **B)** Position of the bedset within the sedimentary sections 5 and 1 by Höning and John (2015). **C** and **D)** Photographs illustrating the exposure, weathering pattern and facies types within the studied bedset (Note hammer and chisel for scale in D).

7.4.2 Markov-chain analysis and quantification method

A Markov-chain analysis was carried out in order to extract the probabilities of lateral facies transitions within a bed. Two facies types that contain oncoids were grouped as one facies type for this analysis to reduce the amount of possible states and thus reduce the degrees of freedom to obtain a statistically more robust result, accounting for the relatively small number of facies transitions. A total of 99 lateral facies transitions between the measured sections were recorded, to either side of each section, including lateral transitions into the same facies type (i.e. lateral facies continuity between sections). A Chi-square analysis was subsequently performed and the hypothesis that neighbouring facies are independent of each other could be rejected at the 99.9 % confidence level.

The amount of heterogeneity of individual small-scale correlation panels was quantified by calculating the coefficient of variation (Cv) for the lateral dimensions of all different facies types combined within a given outcrop study using the following equation:

$$Cv = \frac{\sqrt{\sigma^2}}{\bar{x}} \quad (1)$$

where σ is the standard deviation of all the lateral facies dimensions in the given dataset and $\bar{x}$ is the mean thickness of all the lateral facies dimensions in the given dataset. A Cv of zero represents homogeneity and increasing values represent increasing heterogeneity.

7.5 Middle Jurassic facies pattern at Wadi Naqab

7.5.1 Lithofacies types

Six lithofacies types (LF 1-6, following Hönig and John, 2015) occur within the selected bedset (Table 7.1). LF 1 and 2 are oncolites with a float- or rudstone texture. LF1 and 2 also contain gastropods, with individual specimens up to 15 cm long (Fig. 7.3D), bivalve shells (Fig. 7.3C) and occasionally branching corals (Fig. 7.3B). LF 1 has a micritic matrix (Fig. 7.3A), while LF 2 has a packstone or grainstone matrix consisting of peloids, various bioclasts and occasionally ooids (Fig. 7.4B). LF 3 has a grainstone texture and comprises peloids, intraclasts,

bivalves, echinoid fragments, various genera of benthic foraminifera as well as rare ooids (Fig. 7.4C, F). The peloids have equant isopachous cement fringes (Fig. 7.4C [inlet]) and the remaining matrix is formed by granular to blocky calcite cement. LF 4 contains the same grains as LF 3 but has a micritic matrix and thus a packstone texture (Fig. 7.4D). LF 5 occurs as wackestone, packstone or floatstone with bioclasts as the dominant component with only few peloids (Fig. 7.4E). LF 6 is a marly mudstone, has a recessive weathering pattern at outcrop and contains no fossils (Fig. 7.3A).

Table 7.1 – Description and dimensions of the six observed lithofacies types (LF) present in the Wadi Naqab bedset (m: marl; M: mudstone; W: wackestone; P: packstone; G: grainstone; F: floatstone; R: rudstone).

LF	LF Name	Texture	Main Components	Dimensions [m]
1	Oncoidal floatstone	F F-R W	Oncoids, coated grains, gastropods, bivalves, micritic nodules, peloids	Thickness: 0.2 – 0.4 Max. lateral extent: 14.5
2	Oncoidal float- to rudstone with a grainy matrix	F F-R W-P (P/G)	Oncoids, peloids, ooids, coated grains, gastropods, bivalves, corals	Thickness: 0.15 – 0.4 Max. lateral extent: 39
3	Peloidal grainstone	G	Peloids, ooids, intraclasts, bivalves, gastropods, echinoids, benthic foraminifera	Thickness: 0.15 – 0.95 Max. lateral extent: 60
4	Peloidal bioclastic packstone	P	Peloids, bivalves, gastropods, coated grains, ooids benthic foraminifera, ostracods, corals	Thickness: 0.1 – 0.45 Max. lateral extent: 85
5	Bioclastic peloidal wacke- to packstone	W-P P F	Bivalves, gastropods, benthic foraminifera, echinoids, peloids, corals	Thickness: 0.15 – 0.4 Max. lateral extent: 18
6	Marl / mudstone	m M	None	Thickness: 0.05 – 0.05 Max. lateral extent: 30

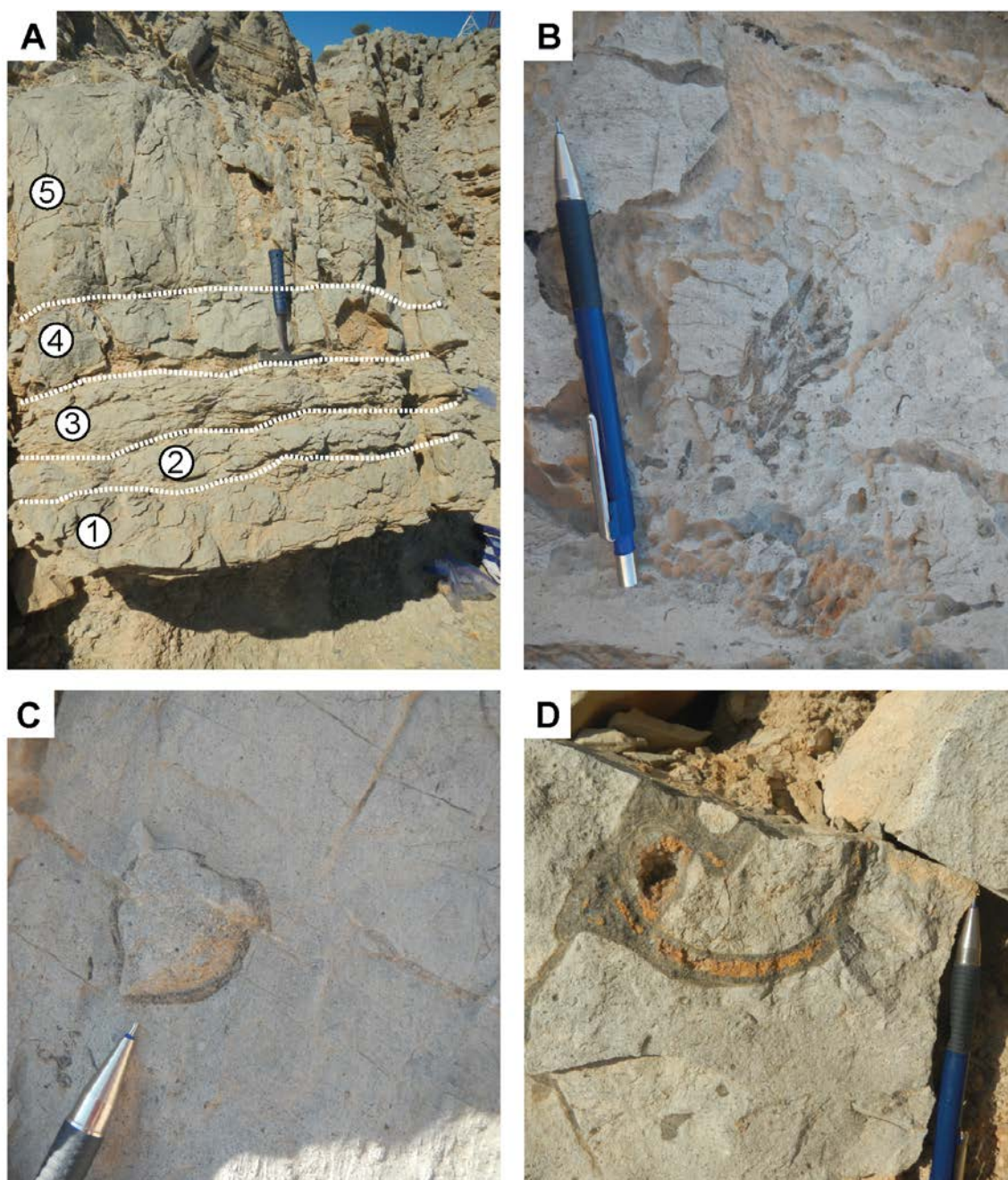


Fig. 7.3 – A) Lower part of the studied bedset at Wadi Naqab. Note the recessive weathering pattern of the three thin marly mudstones intervals at and below the hammerhead (compare bed numbers 1-5 to Fig. 7.5). **B)** Branching coral from the upper part of the bedset between sections WN 9012 and WN 9000. **C)** Bivalve specimen with both valves still attached (between sections WN 9055 and WN 9043). **D)** Large gastropod specimen within the lower part of the upper unit of the bedset (between sections WN 9026 and WN 9012).

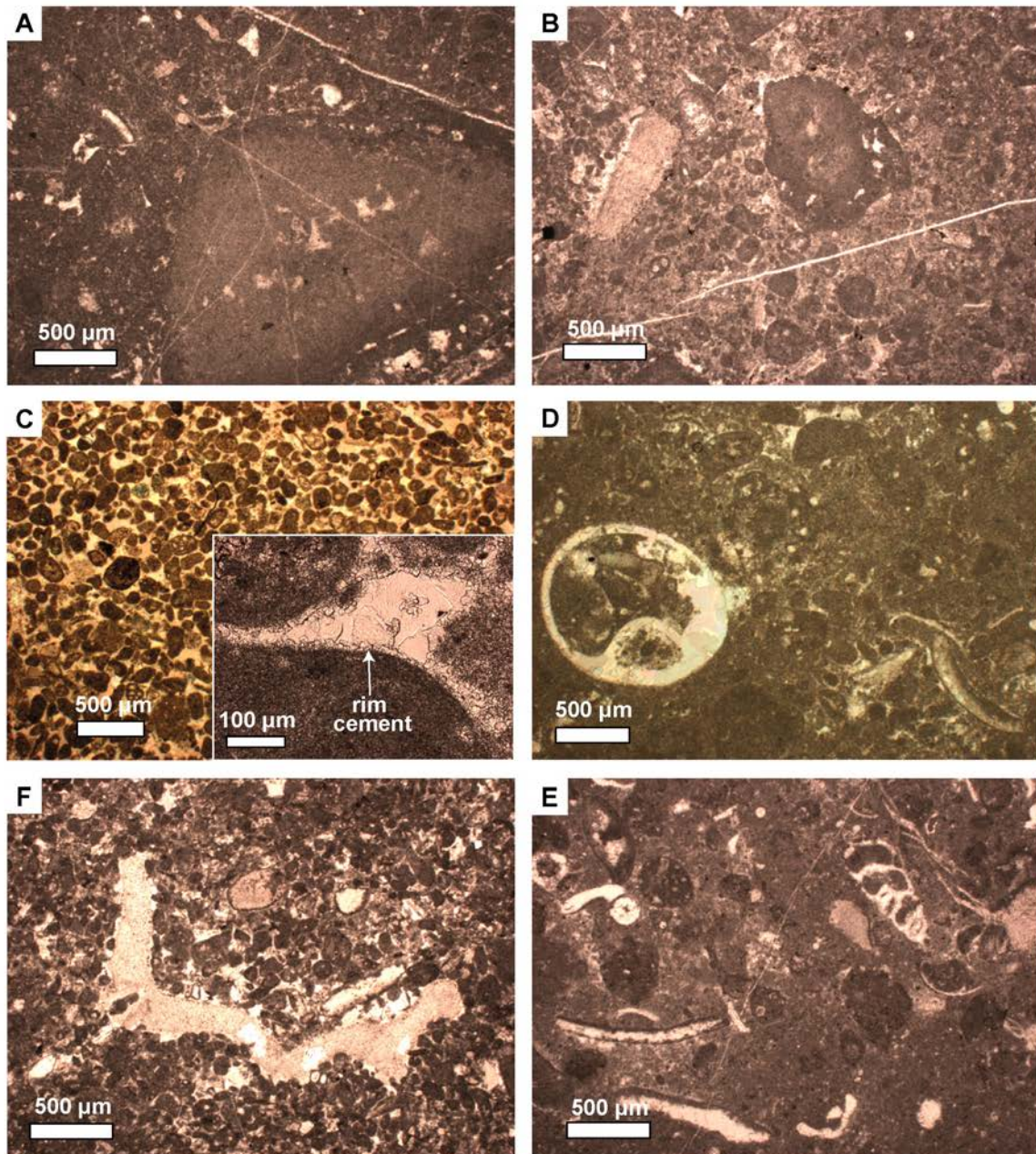


Fig. 7.4 – A) Thin section image of an oncolite with a muddy matrix and bioclast fragments (LF 1, from the upper part of section WN 9242). **B)** Thin section image of an oncolite with a grainy, peloidal grain- to packstone matrix (LF 2). Note the large echinoderm fragment on the left side of the image (from upper part of section WN 9000). **C)** Peloidal grainstone (LF 3, from the upper part of section WN 9026). The inset shows the isopachous rim cement around the peloids. **D)** Peloidal-bioclastic packstone (LF 4, from the upper part of section WN 9012). **E)** Peloidal-bioclastic packstone (LF 4, from the lower upper part of section WN 9026). Note the large sea urchin test fragment. **F)** Bioclastic peloidal wackestone (LF 6, from the upper part of section WN 9055).

7.5.2 Vertical and lateral facies architecture

The ten logged sections can be divided into a lower and an upper part (Fig. 7.5). The lower part is between 40 and 45 cm thick (beds 1, 2 and 3 Fig. 7.5). It comprises peloidal wacke- to packstone (LF 3) and in one instance bioclastic wackestone (LF 5). The lower part appears either as a bedset of three individual beds with sharp and scoured contacts, in two cases separated by thin intercalated mudstone layers. The upper part of the studied bedset (beds 4 and 5, Fig. 7.5) comprises a 20 to 25 cm thick lower bed (bed 4, Fig. 7.5), and a 70 to 85 cm thick upper bed (bed 5, Fig. 7.5). The lower bed mostly contains oncolites (LF 1 and 2), in two cases bioclastic wackestones (LF 6) and in one section a peloidal grainstone (LF 3). The uppermost bed is divided into two halves (bed 5a and 5b, Fig. 7.5) to accommodate for vertical internal facies changes obtained from section analysis. Bed 5 comprises all the different identified lithofacies types, except mudstones (LF 6). The difference in appearance of both the lower and the upper part of the analysed bedset, either as one unit with internal bedding planes or as three or respectively two individual beds, might be dependant on different weathering conditions across the outcrop and therefore not necessarily a reflection of compositional differences. The upper part generally shows the highest amount of lateral facies variability, compared to the lower part (Fig. 7.6). Facies changes between logged sections were marked as transitional changes because no sharp contacts were visible across the outcrop; however, facies identification partly relies on the thin section analysis with the samples coming from the logged sections, and thus on discrete measurements. The facies transition between the two sections furthest apart (WN 9000 and WN 9242, Fig. 7.6) could not be traced as it is not exposed and was interpolated based on the facies transition pattern observed on the exposed part of the strata.

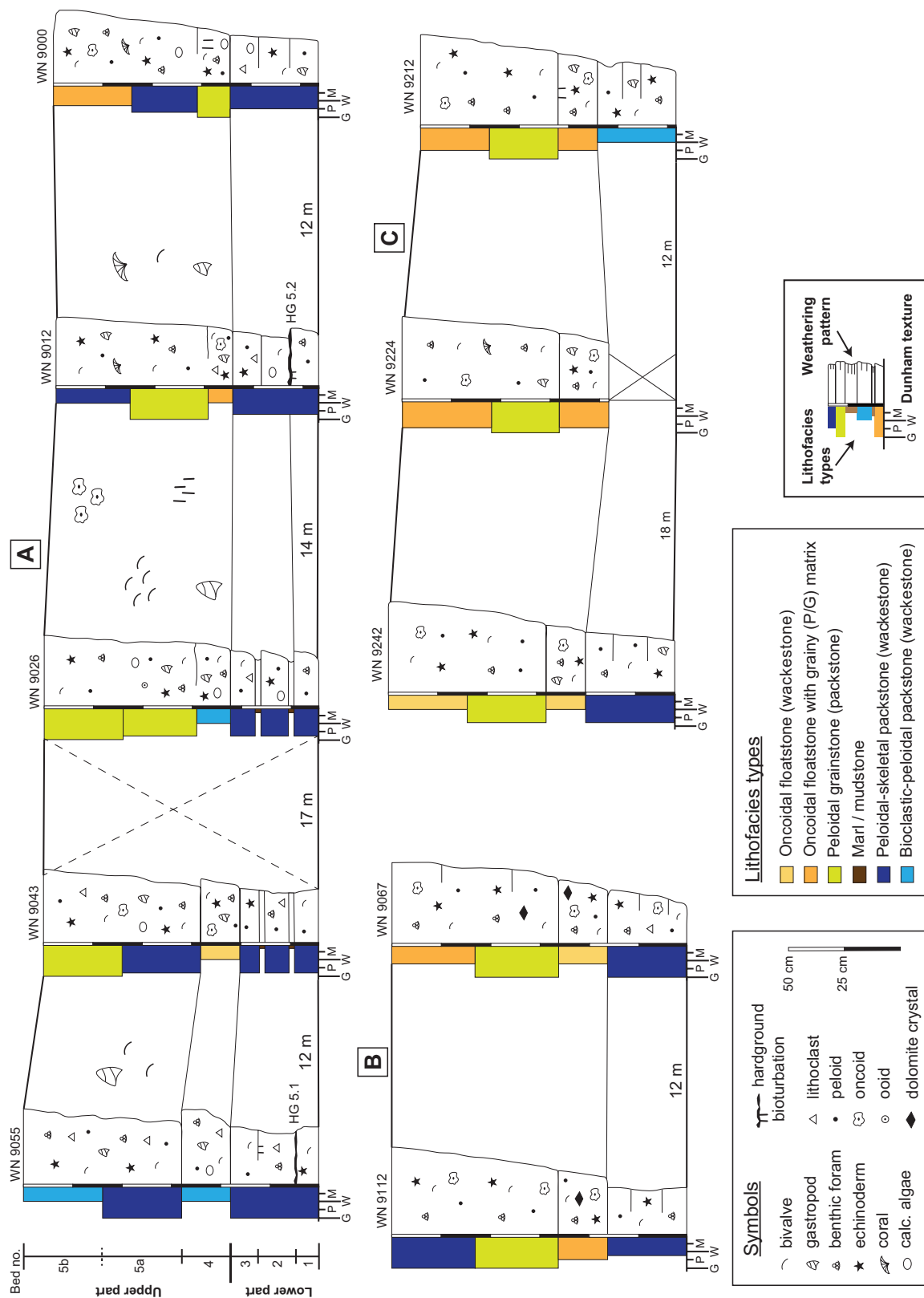


Fig. 7.5 – Detailed sedimentary logs measured within the bedset at Wadi Naqab. Refer to Fig. 7.2 for exact configuration of sections relative to each other.

The Markov-chain analysis (Table 7.2; Fig. 7.6) of the lateral facies transitions reveals that the peloidal pack- and grainstones (LF 3 and 4) have the highest probability of being laterally continuous, followed by the oncolite facies (LF 1 and 2). The bioclastic limestone (LF 5) has a high probability (> 85 %) of being laterally discontinuous and of transitioning into the oncolite facies types (LF 1 and 2) between the logged sections. These transition probabilities match the pattern observed at the 100 to 500 m scale (Hönig and John, 2015). A spatial arrangement of facies is evident, with the sections to the northeast generally dominated by peloidal facies, while the southwestern area contains abundant oncolites, especially in the upper part of the bedset. The lithofacies types between the logged sections were interpolated to create four lithofacies maps, one for each bed (Fig. 7.7). The maps illustrate the facies distribution on the palaeo-seafloor, as preserved in the outcrop. Facies variability is highest for the second and the fourth bed, each resembling a facies mosaic. Beds one and two only contain two and three facies, respectively.

Table 7.2 – Markov-chain analysis results of lateral lithofacies transitions of the ten logged sections at Wadi Naqab. Refer to Fig. 7.5 for the facies colour code and to Table 7.1 for facies descriptions. Table 7.2A (counted frequencies) and 7.2B (calculated probabilities) lateral transitions from one facies to another between neighbouring sections, and to both sides, also showing transitions from a facies into itself (i.e. lateral facies continuity between logged sections).

A) Lateral observed transition frequency matrix (99 transitions; $\chi^2=80.46$)

	Onc	M m	Pel P	BioPel P	Pel G
Onc	14	0	3	5	2
M m	0	4	4	0	0
Pel P	3	4	28	1	4
BiPe P	5	0	1	0	1
Pel G	2	0	5	1	12

B) Lateral observed transition probability matrix

	Onc	M m	Pel P	BioPel P	Pel G
Onc	0.58	0.00	0.13	0.21	0.08
M m	0.00	0.50	0.50	0.00	0.00
Pel P	0.08	0.10	0.70	0.03	0.10
BiPe P	0.71	0.00	0.14	0.00	0.14
Pel G	0.10	0.00	0.25	0.05	0.60

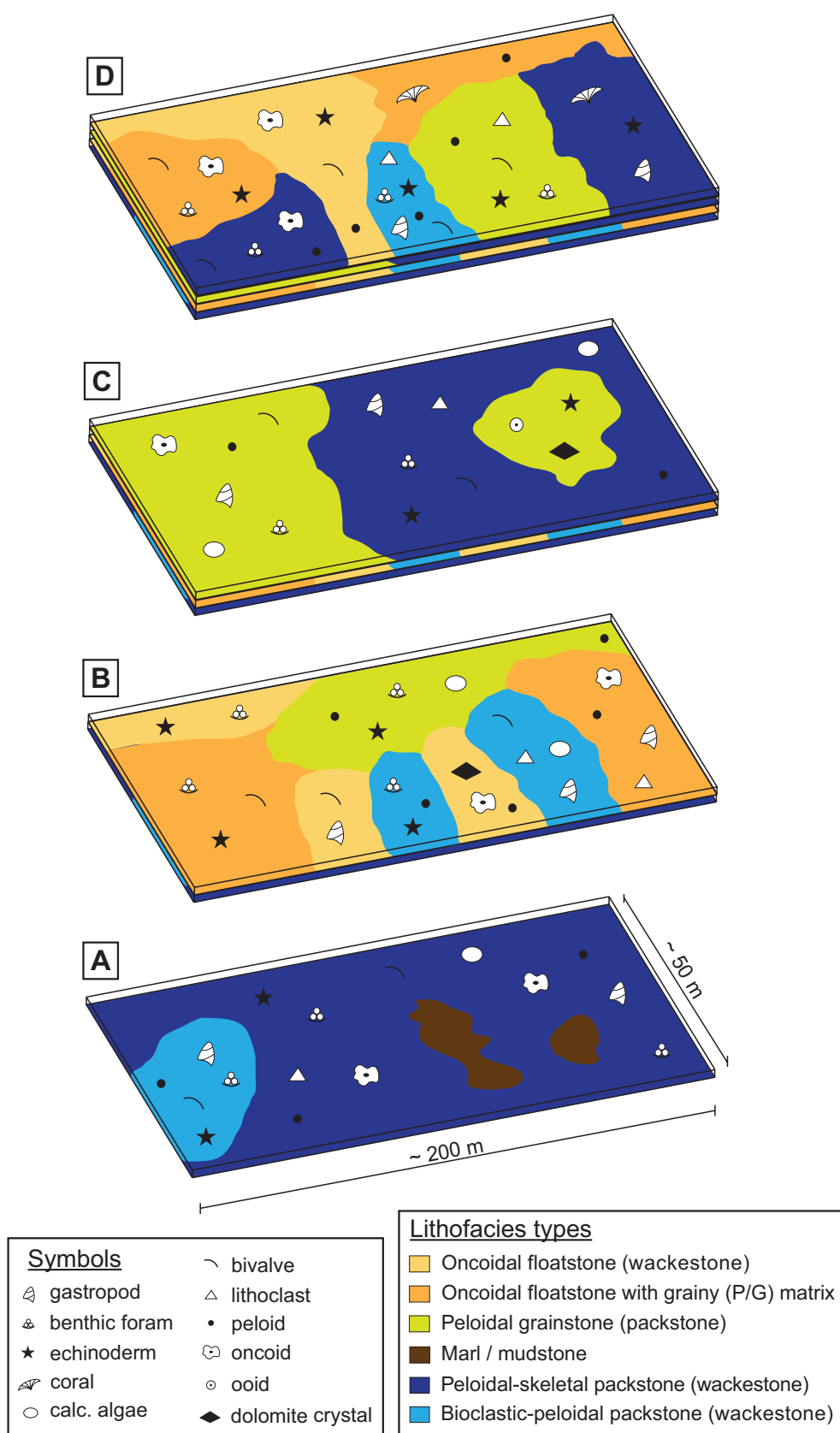


Fig. 7.7 – Facies distribution maps for each unit within the Wadi Naqab bedset, deduced from the facies correlation panel (Fig. 7.6). **A)** Lower part of the bedset (beds 1-3, Fig. 7.5). **B)**

Lowermost bed of the upper part of the bedset (bed 4, Fig. 7.5). **C)** Middle upper part of the bedset (bed 5a, Fig. 7.5). **D)** Uppermost part of the bedset (bed 5b, Fig. 7.5).

7.6 Interpretation and discussion

7.6.1 Environment of deposition at Wadi Naqab

The lithofacies types at Wadi Naqab indicate deposition within the intertidal to subtidal zone of the shallow-marine epeiric carbonate ramp. The oncolite bearing facies (LF 1 and 2) might have been deposited in a lagoonal setting, and the diverse fossil assemblage indicates palaeoenvironmental conditions with open waters. The micrite rich facies, which contains peloids as well as a diverse fossil assemblage (LF 4 and 5), are interpreted as moderate-energy deposits. The marly mudstone (LF 6) is interpreted as having been deposited in the same environment as LF4 and 5 as it only occurs as thin patches within the peloidal-rich facies. The cement-rich grainstones (LF 3) represent zones with relatively high hydrodynamic energy, as indicated by the lack of lime mud. The peloids are presumed to be mainly of fecal origin although some show remains of internal fabric such as benthic foraminifera and are thus the products of micritization processes of bioclasts and potentially of ooids. The isopachous cement fringes are indicative of early-diagenetic precipitation, under phreatic conditions (Fig. 7.4C inset). The remaining coarser calcite cement is a likely product of late-diagenetic, burial-related cementation or recrystallization. The branching corals, especially abundant within the upper unit of the bedset are interpreted as the remains of small patch reefs within the shallow-water environment, because they are concentrated in these intervals and do not form massive structures. The overall good preservation of different fossils such as bivalves with both valves still attached (Fig. 7.3C), large gastropods (Fig. 7.3D), branching corals (Fig. 7.3B) and delicate sea urchin fragments (Fig. 7.4F) is indicative of overall relatively small transport distances of these specific bioclasts. The round carbonate grains could have been transported across longer distances on the homoclinal, low-angle ramp.

Facies maps based on the correlation panels were reconstructed, and show two different types of deposition. Firstly, the maps produced from beds 4

and 5b of the facies correlation panel (Fig. 7.5) show how different facies types occur close to each other as a mosaic on the seafloor (Fig. 7.7B, D), based on the assumption that beds are more or less synchronous depositional units. The second type of facies distribution observed at Wadi Naqab is more homogeneous (Fig. 7.7A, C) and restricted to the lower part and bed 5a (Fig. 7.5). The reconstructed maps contain a maximum of three facies types (LF 3 and 4 as well as LF 4, 5 and 6). These intervals are dominated by peloid rich facies either with a micritic or a cement-dominated matrix. The similarity in grain assemblages indicates an overall homogenous carbonate factory for the entire bedset.

7.6.2 Quantifying metre-scale facies heterogeneity at Wadi Naqab and in other ancient carbonate systems

Statistical measures, such as the coefficient of variation, can provide a quantitative measure for the amount of overall facies heterogeneity for the Wadi Naqab dataset, as well as for other previously published ancient examples of closely spaced sedimentary sections. The coefficient of variation, that has previously not been applied to facies heterogeneities in outcrops, was selected as a measure of heterogeneity in this study as it allows for the comparison of datasets irrespective of their dimensions. Furthermore this coefficient has been suggested as an appropriate method for relatively small datasets with the potential to quantify facies heterogeneity in outcrops, as opposed to the Lorenz Coefficient or the Dykstra-Parsons method (Fitch et al., 2015). The coefficient of variation has the advantage that a single number is obtained that can then be used to compare heterogeneity between the different examples in an objective way. Here, only lateral facies heterogeneity was quantified for two reasons. Firstly because the focus lies on lateral facies changes that cannot be obtained from 1D sections or subsurface core or well log datasets; and secondly because the vertical extent of the different datasets is very small and would allow little insight in comparison to studies that focus purely on 1D vertical sections with much larger thicknesses. Aside from the dataset from Wadi Naqab, a total of four previously published case studies capturing sedimentary features at a

similar scale as presented here were selected for comparison (see Table 7.3 for a summary, and description below).

Table 7.3 – Summary of all discussed case studies of small-scale lateral facies panels following the hierarchical classification scheme proposed by Jung and Aigner (2012) (M: mudstone; W: wackestone; P: packstone; G: grainstone; F: floatstone; R: rudstone; B: boundstone).

No.	Name	Geographic location	Geol. time	Depos. system	Depos. environment	Depos. shapes	Facies types
1	Wadi Naqab	Ras Al-Khaimah (UAE)	Middle Jurassic	Homoclinal ramp	Shallow-marine	None visible	M, W, P, G, F, R
2	Saiq Plateau 240 m	Jabal Al-Akhdar (Oman)	Early Triassic	Homoclinal ramp	Shallow-marine	Shoals	M, W, P, G, R, B
3	Saiq Plateau 12 m	Jabal Al-Akhdar (Oman)	Early Triassic	Homoclinal ramp	Shallow-marine	Individual shoal	W, P, G, B
4	Jura Mountains	Roderis (Switzerland)	Late Jurassic	Carbonate platform	Shallow-marine	Bioherms	W, P, G, R, B
5	Hauschi Huqf	Wadi Jarrah (Oman)	Early Cretaceous	Carbonate platform	Shallow-marine	Individual shoal	M, P, G, F, R
6	Jabal Madmar	Adam Foothills (Oman)	Cretaceous	Carbonate platform	Shallow-marine	Individual shoal	G, F, R
7	Abu Dhabi and Dubai	Western coast UAE	Holocene	Homoclinal ramp	Shallow-marine	Bars and runnels	Mud, grains

7.6.2.1 Lower Triassic of the Saiq Plateau (Sultanate of Oman)

Two small-scale 2D panels were measured by Obermaier et al. (2015) on an epeiric shallow-water carbonate ramp, within a shoal complex. The study was performed on the eastern flank of Jabal Al-Akhdar mountain on the Saiq Plateau. Stratigraphic heterogeneities at different scales were investigated within the Middle Mahil Member, which is of Early to Middle Triassic age. A 240 m long and ~ 1.5 m high panel as well as an adjacent and more detailed 12 m x 1 m panel, both from the same grain-dominated interval are presented (Fig. 7.8). One of the main purposes of the study was to examine the internal, small-scale architecture of shoal and shoal-margin deposits, in order to make recommendations on 1D correlations of sections. The larger 2D panel reveals that most of the lateral facies heterogeneity occurs in the form of pinching and swelling grainstone-dominated intervals that are predominantly cross-bedded (Fig. 7.8A). Grainstones occur as dipping wedges or shingles and are only partly continuous across the 240 m long study window. The more detailed 12 m long panel illustrates the discontinuity of wacke- and packstone facies that pinch out laterally within the studied interval (Fig. 7.8B). Obermaier et al. (2015) attribute the facies heterogeneities, with facies types occurring as wedges rather than as horizontal beds, as being the result of sediment distribution on the sea floor by storms ("event-related") as well as currents, due to the amalgamated and reworked character of the sediments. Furthermore the shingled architecture is attributed to the lateral migration of facies belts induced by relative sea level changes and potentially storms (Obermaier et al., 2015).

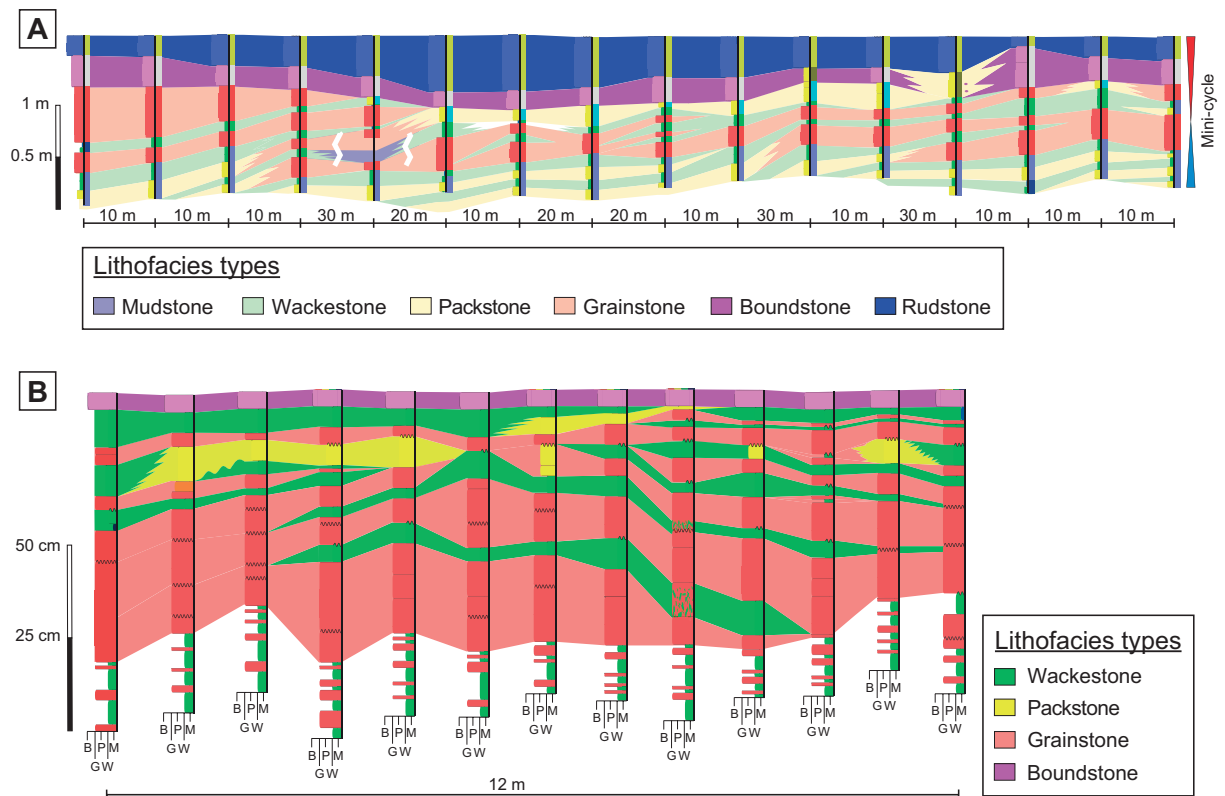


Fig. 7.8 – Facies correlation panels from an individual grainshoal 240 m wide (A) and detailed, 12 m wide (B) (modified from Obermaier et al., 2015; reproduced with permission from Gulf Petrolink).

7.6.2.2 Upper Jurassic of the Jura Mountains (Switzerland)

Samankassou et al. (2003) present 12 sedimentological logs, between 1.2 and 2.2 m high, which were measured on a 70 m long transect in order to construct a composite section. Four of these sections (I to IV, Fig. 7.9 in Samankassou et al., 2003) overlap laterally and are therefore considered for this study (Fig. 7.9). The laterally continuous sections were examined close to the village of Roderis in northern Switzerland, within the Swiss Jura Mountains. The strata are of Upper Oxfordian age and were deposited on a shallow-marine carbonate platform. The four sections consist mainly of ooidal grainstone, coral rudstone, as well as other bioclastic and fenestrael limestones. Facies types merge laterally into each other within only 2 to 3 m. The lateral facies heterogeneity is especially common for the coral rudstone facies, due to the occurrence of laterally discontinuous coral reef bodies. Bed thicknesses change substantially within the study window by as much as ~40 %. Lateral transitions within

individual beds between different fossil assemblages are also reported. Samankassou et al. (2003) explain the lateral facies heterogeneity and mosaic architecture, especially the ones linked to local bioherms, to short-term sea-level fluctuations as well as to an overall dynamic depositional environment.

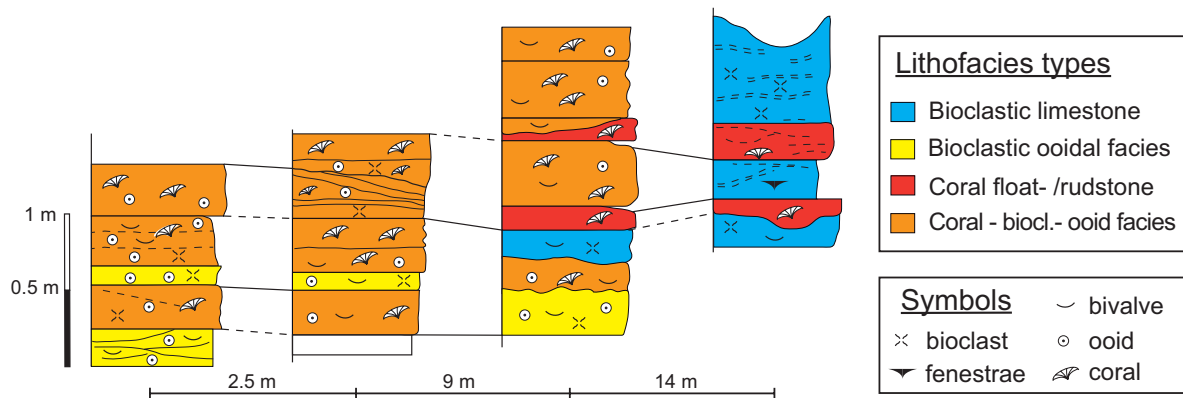


Fig. 7.9 – Closely spaced sedimentary logs from the Oxfordian of the Swiss Jura Mountains (modified from Samankassou et al., 2003).

7.6.2.3 Lower Cretaceous of the Haushi-Huqf area (Sultanate of Oman)

Sena (2013) and Sena and John (2013) examined a subtidal shoal complex in detail on a scale of 52 x 3 m (Fig. 7.10). The 2D panel was measured within the Quishn Formation in strata that is of Early Aptian age. The outcrop is located in Wadi Jarrah on the western flank of the Haushi-Huqf High in western Oman. The sediments were deposited on a sheltered, large carbonate platform under shallow-marine conditions. The studied shoal contains mainly cross-laminated grain- and rudstones as well as bioturbated float- and rudstones, with rudists, peloids and skeletal debris as its main constituents. Lateral facies heterogeneities occur in the form of interfingering cross-laminated- and stratified grain- and rudstones as well as tabular float- and rudstones (Sena, 2013). The shoal is interpreted as a subaqueous dune, deposited within the subtidal regime above fair weather wave base, that migrated approximately parallel to the palaeo-shoreline. This type of dune was permanently subjected to wave and current action, as well as to occasional storm activity (Sena and John, 2013).

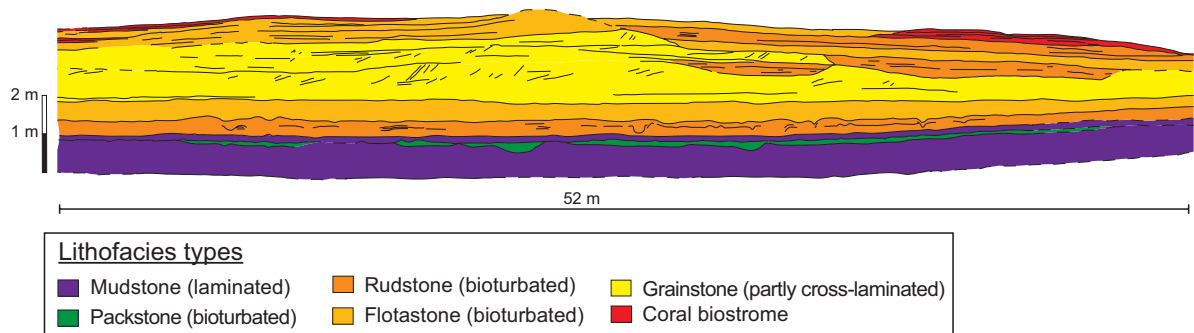


Fig. 7.10 – Facies panel from a grainshoal in the Haushi-Huqf area in south-eastern Oman (modified from Sena, 2013; reproduced with permission from the author).

7.6.2.4 Lower to Upper Cretaceous of Jabal Madmar (Sultanate of Oman)

The Natih Formation (Albian to Turonian age) in Oman has been extensively studied, as it hosts large amounts of hydrocarbons (e.g. van Buchem et al., 2002). Adams et al. (2011) examined the Natih Formation in Jabal Madmar in the Adam Foothills in northern Oman. The goal of the study was to analyse vertical and lateral facies architectures as well as clinoform geometries, partly through using digital outcrop models. The sediments were deposited on a vast shallow-marine carbonate platform. Amongst other things Adams et al. (2011) measured four sedimentary sections between 11 and 34 m apart and each around 8 m high across a shoal complex (Fig. 7.11). The shoal body comprises fine grainstone, bioclastic grainstone and rudist rud- and floatstones. The shoal has a general coarsening upward character and lateral facies merging and interfingering occur mainly between the coarse- and the fine-grained bioclastic grainstone. The architecture of the shoal body reflects its progradational nature with the facies belts oriented parallel to the carbonate platform margin. This is mainly a reflection of the tidal influence on the individual shoal bodies (Adams et al., 2011).

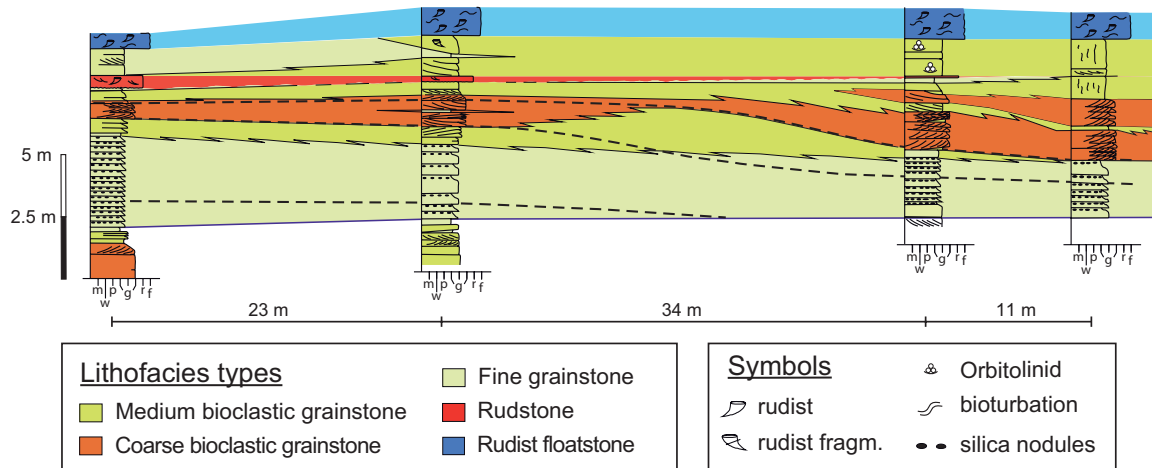


Fig. 7.11 – Closely spaced sedimentary logs and facies correlation panel from a grainshoal in the Adams Foothills at Jabal Madmar, Oman (modified from Adams et al., 2011).

7.6.2.5 Implications of the coefficient of variation in the selected case studies for small-scale heterogeneities in carbonate systems

The coefficient of variation (Cv) ranges between 0.48 and 0.78 for the six case studies (Table 7.4). The Jabal Madmar grain shoal (Fig. 7.11) results in the most homogeneous facies panel (Cv=0.48), while the Wadi Naqab correlation panel is the most heterogeneous one (Cv=0.78, Fig. 7.6), followed by the Saiq Plateau 240 m facies map (Cv=0.73, Fig. 7.8A). The two correlation panels showing the highest degree of heterogeneity are also the most laterally extensive ones among the dataset. It can be speculated that the highest amount of bed-scale lateral heterogeneity between closely spaced sections could best be captured by data gathered over ~ 100 m or more. Individual grain bodies such as the shoals from the Saiq Plateau (12 m panel) (Fig. 7.8B), Jabal Madmar (Fig. 7.11) and the Haushi-Huqf area (Fig. 7.10) have heterogeneous architectures in the form of cross-bedding, pinch-outs, wedges and shingles, and pinch and swell patterns, but facies distribution within individual beds is relatively homogeneous and additionally relatively few facies types are present, compared to the wider correlation panels of Wadi Naqab and the Saiq Plateau (240 m panel). In addition, constraining the intra-bed facies variability was not the main focus of most of the studies compared here, with the exception of the Saiq Plateau and the Wadi Naqab datasets, and so another form of observation

bias is probably present. The sections from the Jura Mountains ($C_v=0.53$) are unique amongst the examined cases as they contain coral buildups, which are very localised (Fig. 7.9). If these buildups are present between closely spaced sections they will have a large impact on the correlation of sections and on the continuity of surrounding facies types. Remains of similar, although maybe smaller, coral patch reefs are present in the form of well-preserved branching corals at Wadi Naqab. They are however not preserved as coral buildups and only occur relatively rarely compared to the Jura Mountains case, and therefore have little influence on the lateral variability of facies within individual beds at Wadi Naqab. All of the examined individual shoal bodies (No. 3, 5 and 6, Table 7.3) are dominated by high-energy facies such as ooidal and bioclastic grainstones (Fig. 7.8B, 10,11), while the larger correlation panels from Wadi Naqab and from the Saiq Plateau (No. 1 and 2, Table 7.3) contain a large amount of lower-energy wacke- and packstones (Fig. 7.5, 7.6, 7.8A). The higher number of facies types, as well as the lateral variability over a longer distance of observation, is therefore reflected in a higher coefficient of variation for these two cases.

Table 7.4 – Measured lateral and vertical facies dimensions from the case studies discussed in this paper. The coefficient of variation was calculated incorporating all facies dimensions of a given case as a measure for heterogeneity.

Case study	Vertical thickness [m]		Lateral thickness [m]					
	Minimum	Maximum	Minimum	Maximum	Median	Mean	Standard Deviation	Coefficient of variation
1) Wadi Naqab	0.1	0.95	6	85	16.5	23.4	18.25	0.78
2) Saiq Plateau 240 m	~ 0.05	0.55	8	170	58	59.2	43.2	0.73
3) Saiq Plateau 12 m	~ 0.01	0.35	1	12	5	5.1	3.3	0.65
4) Jura Mountains	~ 0.05	0.9	~ 1.25	18.5	9.25	10.5	5.5	0.53
5) Hausi-Huqf	~ 0.05	1.9	~ 3.1	52	25.6	25.3	17.4	0.69
6) Jabal Madmar	~ 0.05	3.8	6	68	65	51.2	24.8	0.48

7.6.3 Controls on small-scale facies heterogeneities

7.6.3.1 Carbonate factory variability

Pomar and Kendall (2007) list ecological requirements of biota, water temperature gradients and ocean chemistry, as well as the regional hydrodynamic setting as principle controls on carbonate deposition. In contrast to siliciclastic dominated systems where sediment accumulation is mainly governed by physical accommodation, ecological accommodation is a key controlling factor in carbonate systems and thus on the distribution of facies types on a carbonate ramp or platform (Pomar and Kendall, 2007). Ecological accommodation is defined as the capacity of organisms to produce and accumulate carbonate sediment, where hydrodynamic accommodation is available (Pomar, 2001a, 2001b, Pomar and Kendall, 2007).

The different carbonate grains at Wadi Naqab were all produced within the same shallow-marine environment of deposition (EOD) on the homoclinal carbonate ramp. The facies mosaics encountered at Wadi Naqab (Fig. 7.7B, D) could be the result of local differences in carbonate production, with the distribution of different organisms controlled by the presence of a specific ecological niche as well as competing neighbouring organisms, or the concentration of certain communities within small depressions and highs on the seafloor, as is the case on the Holocene platform interior of the Bahamas (Harris, 1979). The wide coverage of peloids, interpreted as of mostly fecal origin, could indicate that producing organisms such as worms or gastropods, migrated across the entire seafloor within the relatively shallow part of the carbonate ramp. Adams and Grotzinger (1996) studied facies heterogeneities on a Cambrian carbonate platform. Although not focussing on the metre-scale, they assume that very small-scale heterogeneities were governed by changes in local carbonate production, which are in turn controlled by differences in accommodation. This however appears to be an unlikely explanation for the facies mosaics visible on the very gently dipping epeiric ramps (Wadi Naqab and Saiq Plateau, No. 1 and 2, Table 7.3) as accommodation would have been more or less equal on a ~200 m wide transect. A Waltherian facies distribution for these two ancient examples can furthermore be excluded, as it can be assumed, as stated above, that all facies types within the respective systems

were deposited within the same EOD (Middleton, 1973). Facies heterogeneity in the other ancient examples (No. 3, 5 and 6, Table 7.3) commonly occurs in the form of grain shoals, and thus the geometries at these other localities are primarily the result of sediment transport and hydrodynamically controlled accumulation. These latter cases are more comparable to a grain-dominated, siliciclastic system (Rankey et al., 2006; Pomar and Kendall, 2007; Immenhauser, 2009).

Modern analogues contain small-scale heterogeneities, but do not fully help disentangling production-related heterogeneity and coeval transport processes (see example 7, in Table 7.3). The examples of modern carbonate facies, analysed using remote sensing, can only reveal a map view of their distribution, but could serve a useful template for facies distribution in the subsurface if these facies are preserved. Purkis et al. (2005) present the heterogeneous distribution of facies types within individual facies belts on the scale of a few metres to several kilometres parallel to the coastline. The satellite images interpreted by Kendall and Alsharhan (2011) illustrate the heterogeneous morphologies present on a modern coastal carbonate ramp. Bars and runnels composed of carbonate sands have heights of 20 to 40 cm and are 1 to 6 m wide. This scale is similar to the individual facies wedges or shingles within the ancient grain-shoal examples from Oman (No. 2, 3, 5 and 6, Table 7.3). The satellite images however lack detailed information on carbonate grain and facies distribution.

Modern examples nevertheless highlight the complexity of facies distribution in these settings on a very small scale. The distribution of facies types in the Dubai coastal area, where relatively high-resolution facies distribution data is available, is so complex that it is termed random, despite some inter-facies relationships (Purkis et al., 2005). It can be postulated that factors such as the distribution of different carbonate producers and transport both play a role in the distribution of grains in modern carbonate systems (Alsharhan and El-Sammak, 2004). This is supported by a study from the Philippines archipelago where the composition of modern carbonate sediments varies significantly between closely spaced study sites across an approximately 500 m wide transect. Sediment distribution is controlled by localised carbonate

production, as well as other factors related to transport processes (Hewins and Perry, 2006).

What these examples indicate is that a model explaining facies heterogeneity in carbonate strata, based only on changes in local carbonate production, is only applicable for cases where facies heterogeneity is governed by buildups, such as coral or rudist reefs, bioherms or biostromes. This is the case for the Swiss Jura Mountains outcrop (No. 4, Table 7.3), parts of the Iberian Upper Jurassic carbonate ramp in eastern Spain (Alnazghah et al., 2013; San Miguel et al., 2013), or the Holocene isolated carbonate platforms in the South China Sea (Rankey, 2016) where heterogeneity in outcrops and in modern facies mosaics occurs mainly in the form of coral buildups or patch reefs, respectively. In all other cases, hydrodynamic processes must play a significant role in controlling facies distribution.

7.6.3.2 Reworking of sediments on the seafloor

The heterogeneous distribution of different facies types at Wadi Naqab (Fig. 7.7B, D) could be controlled entirely by the transport properties of the grain assemblages. The absence of lime mud in certain facies as proxy for frequent wave agitation might be a result of deposition on small palaeo-highs on the seafloor. The micrite preserved within the grain-rich packstone facies (LF 4) could have accumulated in small palaeo-reliefs or in areas shielded from hydrodynamic energy by palaeo-highs. Mud-dominated facies were thus most likely deposited in local zones of hydrodynamic quiescence, potentially controlled by slight variations in the palaeotopography. Conversely the cement-rich facies were deposited on local but relatively small-relief highs, where wave and current energy prevented the deposition of lime mud.

Lokier et al. (2013) report no clear relationship between geomorphic settings and bioclastic components from the Holocene coastline of the UAE, but rather that grain distribution is attributed to variable degrees of reworking, especially during high-energy events when grains are preferentially accumulated. Facies distribution at a similarly small scale from another modern analogue, the Florida Keys, reveals a relatively ordered pattern with facies closely linked to relative sea level. This relationship is however not necessarily applicable to facies in the

subsurface or at outcrop (Bosence et al., 1985; Bosence, 2008) but highlights the fact that modern systems are not necessarily organised as random mosaics. Analogous to modern systems, it is inferred that sediment transport explains at least some of the observed facies heterogeneity at Wadi Naqab.

But what do the maps produced from the individual beds at Wadi Naqab represent (Fig. 7.7)? Although the beds used to create the facies maps have relatively equal thicknesses (except for bed 4, which is thinner) the respective sedimentation rates and the amount of time represented by individual beds might vary significantly. The lower part of the studied bedset and bed 5a could thus either represent different hydrodynamic settings (e.g. sand flats vs. platform interior) or they represent lower sedimentation rates and more stable conditions where sediments were exposed to mixing processes for a longer period of time, resulting in a relatively averaged and therefore homogeneous grain and facies distribution according to their respective transport properties, compared to the heterogeneous beds that might in fact be the product of heterogeneous production subjected to less mixing through local transport. This is comparable to a modern grain shoal in the Bahamas where sediments deposited on different positions of the shoal complex are subjected to energetically and directionally different flow patterns, which is in turn reflected in differences in the degrees of grain sorting (Sparks and Rankey, 2013). Sediments that have experienced more agitation tend to be well sorted in that specific grain shoal (Sparks and Rankey, 2013). Overall, the heterogeneous facies pattern with lateral facies merging observed at Wadi Naqab is interpreted as being the result of mixing and segregating transport processes occurring during the formation of the different grains, much like the 'kaleidoscope' model explained by Rankey (2016), where elements are constantly changed around as well as mixed.

A question remains, which is why the intra-bed facies mosaic from Wadi Naqab (No. 1, Table 7.3) and the very different individually preserved grain shoals from the Saiq Plateau (No. 2 and 3, Table 7.3) have both originated in very similar settings on vast epeiric, homoclinal ramps? The example from Wadi Naqab, where facies transition within individual beds can be observed within a generally 'layer-cake' type architecture contains overall more muddy sediments than the shoals from the Saiq Plateau. Conversely when encountered, the

grainstones at Wadi Naqab are better sorted than the packstones from the Saiq Plateau. This supports the assumption that transport and re-sedimentation mechanisms control the micrite content rather than a change in carbonate production, as it can be assumed that the carbonate factories and transport mechanisms were very similar in these two cases. The presence of grain shoals preserved on the Saiq Plateau could possibly indicate overall more intense hydrodynamic conditions, compared to Wadi Naqab, or a slightly different position on the epeiric ramp. Obermaier et al. (2015) argue that the preserved shoals from the Saiq Plateau might reflect event-related processes rather than constant agitation. By extension this would support the assumption made above, that the sediments on the Wadi Naqab ramp were more affected by amalgamation and averaging processes. Additionally a Holocene ooid shoal in the Bahamas illustrates that a shoal body even in the very young sedimentary record does not necessarily occur as discrete geobody (Sparks and Rankey, 2013). Transport properties in modern carbonate environments include grain size and buoyancy (Hewins and Perry, 2006). Since these would have been very similar on both carbonate ramps (Wadi Naqab and Saiq Plateau, No. 1,2 and 3, Table 7.3) the exposure time to, and possibly the intensity of hydrodynamic energy is a likely key difference between these two settings. The modern Bahamas carbonate platform, although being an overall smaller depositional system compared to the Mesozoic ramps, contains a comparably heterogeneous facies distribution within the shallow-marine EOD, whereby sandflats, containing mostly packstones, and the adjacent ooidal shoals towards the platform margin occur in relatively close proximity (Harris, 1979). This distribution is controlled largely by mixing processes that occur after the grain production. Changes of this pattern through time, examined using shallow cores, reflects slight sea level changes, a mechanism that could explain the facies distribution changes through time at Wadi Naqab (Harris, 1979; Sparks and Rankey, 2013).

7.6.3.3 Diagenesis and sediment preservation

Diagenetic processes taken into account here include early diagenetic processes in the sense of Perry (2000), for example encrustation,

recrystallisation, microborings, fragmentation, dissolution as well as postdepositional biogenic mixing. The facies distribution maps from Wadi Naqab illustrate that some of the facies heterogeneity is governed exclusively by the presence or absence of micrite and the subsequent filling of the space between the grains by calcite cement (Fig. 7.7C). This is particularly visible at Wadi Naqab within bed 5a (Fig. 7.5, 7.6). This indicates that a significant proportion of the facies heterogeneity at Wadi Naqab, whenever LF 3 and 4 occur adjacent to each other, is governed only by the presence or absence of micrite (vs. cement). The absence of micrite might either be an indicator for deposition at higher energy levels (as discussed above) but it could also be a purely diagenetically controlled phenomenon whereby original micrite was dissolved and the pore space subsequently filled by cement. The dissolution of micrite following burial of the sediment would have most likely affected the entire bed rather than only patches. It is therefore suggested that the calcite cement distribution within individual preserved beds follows the template set by the depositional fabric. This is supported by the presence of early diagenetic cement fringes in the grainstone facies, very similar to the Holocene grainshoals of the Bahamas (Harris, 1979). The Wadi Naqab grainstones might have been deposited as individual grain shoals that could have potentially been oolitic and are now preserved as peloids due to micritization. The cement would have only filled the available pore space between these micritized grains. The less permeable and less porous packstones were less likely to be penetrated by fluids supersaturated with CaCO_3 . The presence of cement appears to be governed by the (syn)-depositionally controlled absence of micrite.

Furthermore the Wadi Naqab (No.1, Table 7.3) facies correlation panel is unique within the presented examples as it contains facies transitions within individual, parallel and horizontal beds (Fig. 7.2C, D, 7.6). The other examples (No. 2-6, Table 7.3) have preserved original sedimentary features, such as cross-bedding, that might have been present at the Wadi Naqab site at the time of deposition but were simply amalgamated or wiped out during early diagenetic and post-depositional processes. Wright and Burgess (2005) discuss the difficult task in assessing what actually forms a sedimentary bed. Constant reworking, erosion and dissolution of sediments after their deposition will inevitably erase many features that were originally present and create a

somewhat averaged archive of the original environment. Perry (2000) comprehensively discusses the processes that influence the preservation of carbonate sediments within a modern reef and concludes, amongst other things, that early diagenetic processes will especially influence grain-rich assemblages but that these processes vary along the depositional slope of the carbonate system. Alteration of grains is especially common in the back-reef, while grains in the fore-reef remain relatively unaltered (Perry, 2000). The inner part of the homoclinal carbonate ramp at Wadi Naqab does differ from these depositional settings and thus prohibits a simple comparison. The sediments at Wadi Naqab as well as from the other studied examples from the geological record (Table 7.3) were subjected to various diagenetic processes. It is however deemed unlikely that these processes are a major reason for the heterogeneity that is preserved in the sedimentary record. Re-sedimentation processes occurring after the formation of sediment grains are of great importance in the configuration of the sedimentary record. Preservation processes and sediment mixing that occurs naturally within the shallow-marine EOD, are however almost impossible to separate from each other in ancient examples.

7.6.4 Cyclostratigraphy at Wadi Naqab using lithofacies

A study of modern analogues demonstrates that facies types in the shallow-marine realm are not necessarily a reliable proxy for water depth (Rankey, 2004; Purkis et al., 2015b), but nevertheless this is an often adopted approach when reconstructing sequence stratigraphic pattern in ancient systems. Here a somewhat rigid approach was taken, given that Wadi Naqab offers a natural laboratory where beds can be traced laterally to reconstruct small-scale “cycles”, and thus allows the study of spatial variation of cycles. This exercise highlights potential pitfalls that the use of small-scale cyclostratigraphy in ancient carbonate ramps bear. The oncolite facies (LF 1 and 2) are considered as the shallowest lagoonal deposits, whereas all the other facies types are interpreted as deposits of the shallow carbonate ramp within slightly deeper waters. The marly mudstone (LF 6) was assigned to the middle to outer ramp environment by Hönig and John (2015), but as it occurs only as small patches within LF 4 it is regarded as shallow-ramp facies here. The resulting

interpretation from the detailed metre-thick cycles (termed sixth-order mini-cycles by Obermaier et al., 2015) vary strongly across the studied bedset (Fig. 7.12). Sections occur either as shallowing-upwards cycles (e.g. section WN 9000 in Fig. 7.12), as deepening upwards cycles (e.g. section WN 9112 in Fig. 7.12), as mixed shallowing-deepening cycles and in one instance the inferred relative sea level remains constant throughout the section (section WN 9055 in Fig. 7.12). These results confirm that the high degree of lateral variability of facies within the shallow-marine realm captured here does not permit lateral correlation of sections based purely on cyclostratigraphy, when facies types are used as proxy for relative water depth.

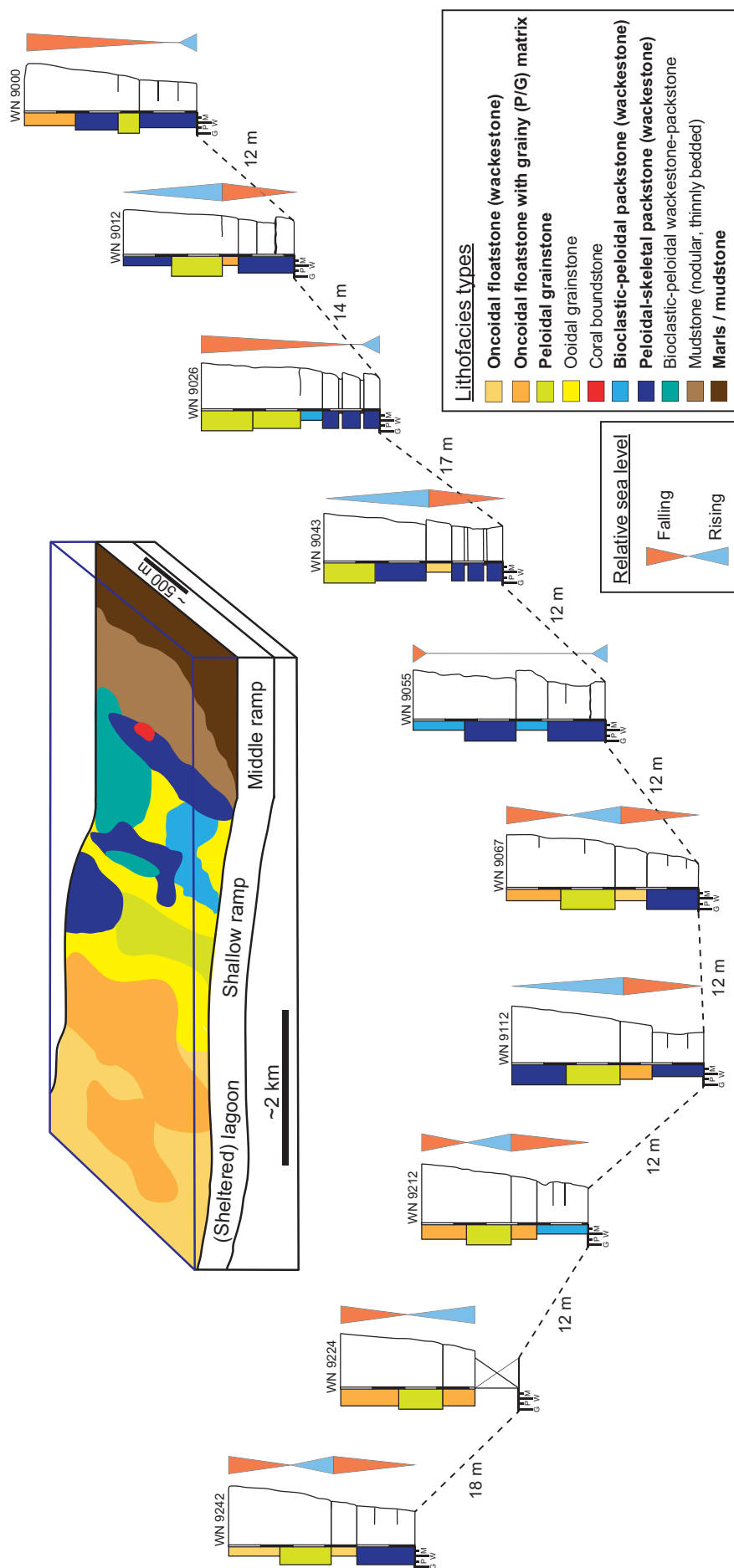


Fig. 7.12 – Sedimentary logs from the Wadi Naqab bedset with metre-scale cycle interpretation. The depositional block model is taken from Hönig and John (2015).

7.6.5 Recommendations for facies and cycle correlations at the metre to inter-well scale

The correlation of 1D sections with little or no information on the architectures and facies pattern between sections is common in subsurface applications, or when studying outcrops that cannot be visibly connected. Metre-scale vertical facies patterns are below seismic resolution and well and core logs do not contain information on lateral variability. Figure 7.13 illustrates how multiple correlation realisations are often possible for different datasets, despite their very different stratigraphic architectures. Several studies have shown that on a lateral scale of approximately 100 to 2000 m, which can already be considered as relatively close inter-well spacings, lithofacies types are not continuous in various shallow-marine carbonate systems, despite a ‘layer-cake’ architecture that might suggest otherwise (e.g. Immenhauser et al., 2004; Frost, 2007; Homewood et al., 2008; Verwer et al., 2009; Palermo et al., 2010; Tómas et al., 2010; Amour et al., 2011; Christ et al., 2011; Alnazghah et al., 2013; San Miguel et al., 2013; Sena and John, 2013; Brandano and Loche, 2014; Hönig and John, 2015; Obermaier et al., 2015). As this study and the examples presented within show, lateral facies variability can also be present at the scale of a few metres on shallow-marine carbonate ramps and platforms, irrespective of whether they are the product of a heterogeneous facies pattern at the time of deposition or a result of diagenesis and preservation processes. To incorporate the complexity at the metre scale in reservoir models, Amour et al. (2013) suggest using a stochastic approach. The present study reveals some trends in lithofacies transitions that might help to guide or give some constraints for well correlations and the population of reservoir models at the scale of a few to ~100 m.

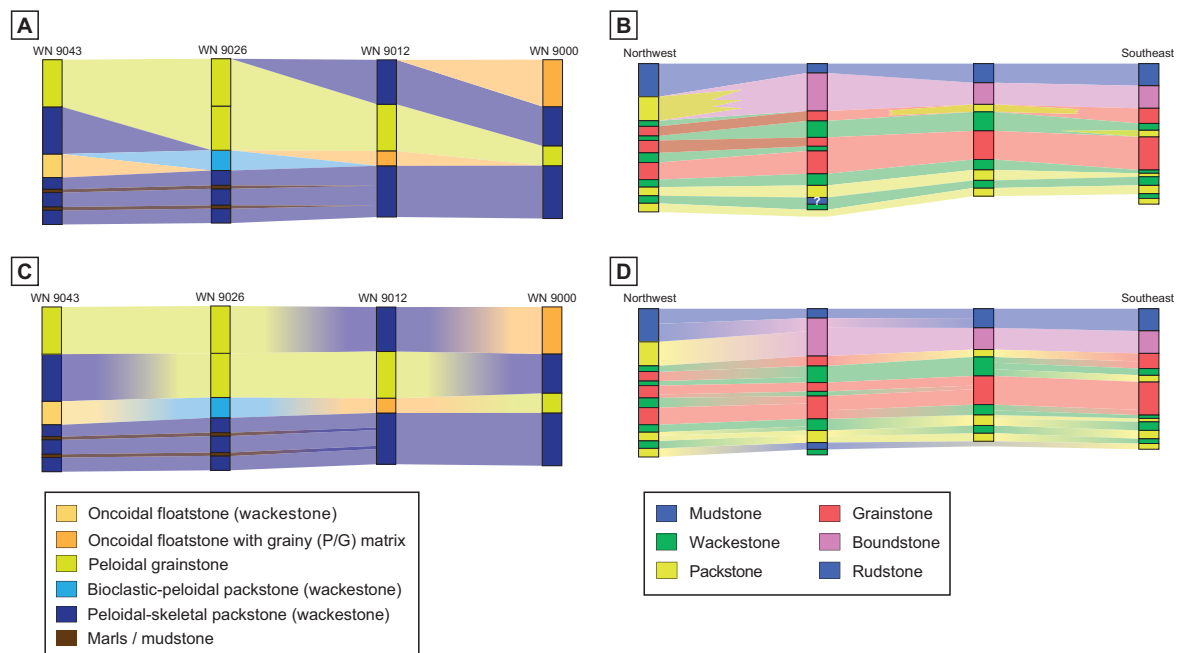


Fig. 7.13 – Correlation panels from four of the sedimentary logs from Wadi Naqab (A, C) and from the 240 m facies panel from the Saiq Plateau (B, D; modified from Obermaier et al., 2015). **A, B)** Correlation using facies shingles and wedges. **C, D)** Correlation using ‘layer-cake’ type bedding and gradual, lateral lithofacies transitions.

The facies mosaic preserved at Wadi Naqab illustrates that lateral correlation based on cyclostratigraphy at the metre scale requires extreme caution, if cycles are determined based on vertical facies changes in the context of inferred Waltherian relationships between facies within the shallow-marine EOD (*sensu* Spence and Tucker, 2007). This is valid for a variety of tectonic settings but especially when differential subsidence has affected the study area (Bádenas et al., 2010). It is recommended that cycles should be assigned only based on facies changes that indicate a change in the broad EOD, for example from inner to mid-ramp, or based on sequence boundaries such as exposure surfaces. The determination of individual cycles at the metre scale is not recommended as lithofacies types deposited within the same EOD do not necessarily follow Walther’s law, or more generally, can vary significantly laterally. Cycles determined at a larger vertical scale can however be very useful for the lateral correlation of sections, despite internal facies heterogeneities (e.g. Bádenas and Aurell, 2010)

Correlations based purely on lithofacies types are likewise challenging as a variety of facies types can occur juxtaposed to each other, even within an individual bed. Whether laterally connected facies bodies simply form sedimentological or geological units or rather act as reservoir and flow units in the sense of Borgomano et al. (2008) needs to be determined for each individual reservoir. Based on the different case studies examined here it is concluded that simple correlations of facies types may not represent the large amount of observed variability (Fig. 7.14). Firstly, the Wadi Naqab correlation panel shows large amounts of facies heterogeneity within individual beds and it can be postulated that this type of variability is equally present in comparable settings. Secondly geometries such as shingles and wedges, especially abundant within individual grainshoals, are a common feature in ancient epeiric carbonate systems and need to be taken into account.

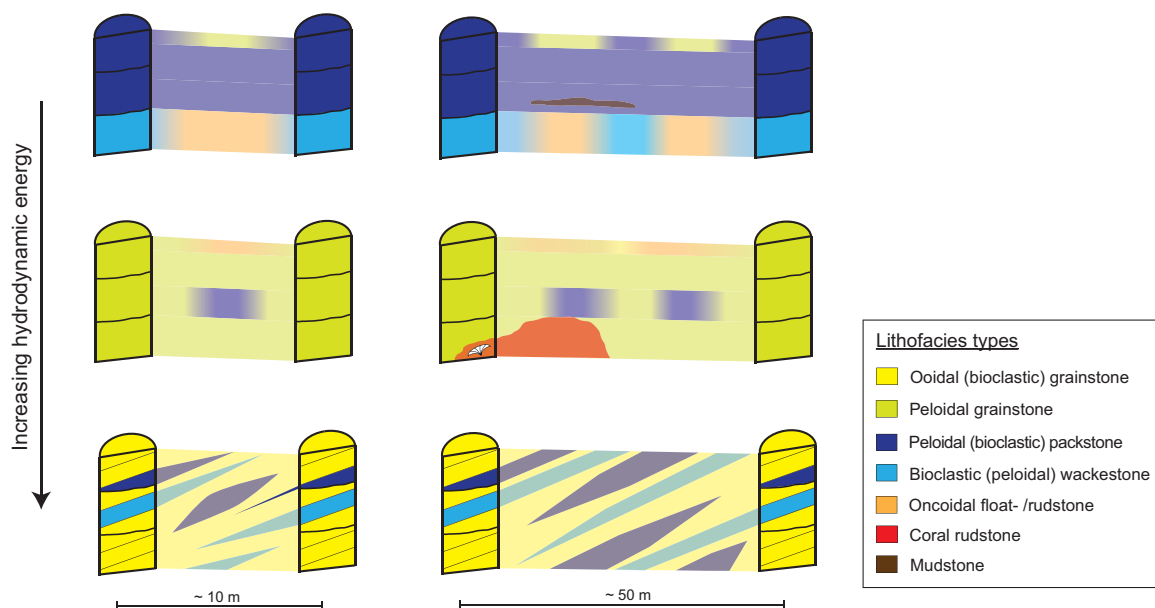


Fig. 7.14 – Recommended lateral facies correlation patterns between 1D vertical sections, based on textural as well as sedimentary features.

7.7 Conclusions

The present study and previously published correlation panels from high-resolution lateral facies analyses within small vertical sections from shallow-marine carbonate systems illustrate the large amount of variability that can be

preserved in these settings. At the metre to several tens of metres scale and within the depositional environment investigated here, facies occur as mosaics. The shallow-marine deposits from Wadi Naqab contain abundant intra-bed facies transitions, while facies heterogeneities from the other settings are governed either by coral buildups (Swiss Jura Mountains) or the heterogeneous architecture of individual shoal bodies (Saiq Plateau, Haushi-Huqf area and Jabal Madmar). The comparison with available data from modern analogous systems suggest that heterogeneous carbonate production as well as re-distribution of grains, occurring more or less coeval to their sedimentation, play a key role in producing closely spaced facies mosaics. Processes that control the re-distribution of sediments such as their transport properties and the intensity and frequency of hydrodynamic processes have an important influence on the sedimentary record of the shallow-marine EOD in ancient as well as in modern systems but are conversely intimately intertwined (Rankey et al., 2006; Rankey and Reeder, 2011). This has implications for the quality and the significance of sedimentary sections for palaeoenvironmental interpretations, as they are highly dependant on these processes that should not be ignored before drawing conclusions.

Based on the findings of this study it is stressed that gathering lateral information is vital when examining shallow-marine carbonate strata, either for palaeoenvironmental or hydrocarbon exploration and production purposes. The use of cyclostratigraphy as only base for lateral correlation should be adopted with great caution as it bears risks due to lateral facies variability.

8. Conclusions and recommendations for future work

8.1 Main conclusions

This outcrop-based study focussed on sedimentological heterogeneities on a Triassic-Jurassic carbonate ramp on the Musandam Peninsula at the regional scale (~ 45 km), the 'inter-well' scale (~ 1.9 km) and at the metre scale (~ 120 m). The study was motivated by the need to better constrain the distribution of lithofacies types on epeiric carbonate ramps in the Middle East, especially on small scales, and to understand the processes that produce the lithofacies distribution. A detailed sedimentological study of a series of well-exposed outcrops on the Musandam Peninsula forms the basis of this study. The field data as well as collected samples were subsequently analysed to extract quantitative information on the distribution of facies types and to perform various geochemical and petrographical analyses in order to constrain palaeoenvironmental conditions that lead to the deposition and preservation of the examined facies pattern.

Several significant results were achieved. The stratigraphic ages of the Triassic and Jurassic sequences on the Musandam Peninsula, that were examined for regional-scale sedimentological heterogeneities, had previously been constrained through biostratigraphy (Hudson, 1960; Glennie et al., 1974; de Matos, 1997; Maurer et al., 2008). The studied strata that were examined on a smaller, 'inter-well' scale and metre-scale, were previously assigned a Bajocian age (de Matos, 1997). Stable carbon isotope stratigraphy allowed a refinement of the stratigraphic age to the Lower Bajocian. This required careful diagenetic and sedimentological screening of the samples used to obtain carbon isotope data as well as the application of the 'oxygen isotope filter' (Metzger and Fike, 2013). The results highlight that shallow-marine carbonate carbon isotope records, preserved in sections that are lithologically relatively homogeneous, have the potential to be used for global correlation purposes. A detailed sedimentological examination is however necessary in order to exclude data that likely represents a meteoric, rather than an oceanic isotopic signal, especially through the identification of possible exposure surfaces. The

excellent lateral correlation of the carbon isotope signal across the studied Lower Bajocian outcrop is a further indication of the good potential of neritic carbonate successions for carbon isotope chronostratigraphy, despite lateral facies changes within the shallow-marine environment of deposition.

The strata that were examined on a regional scale contain very heterogeneous lithologies (i.e. limestones, dolostones, siliciclastic facies) and were biostratigraphically less well constrained than the overlying Middle Jurassic sequence. In order to better constrain stratigraphic boundaries, a combined approach using biostratigraphy, stable carbonate and total organic carbon isotopes as well as radiogenic isotope stratigraphy was adopted, especially to constrain the position of the Triassic-Jurassic boundary. As coeval events across the Triassic-Jurassic transition had a marked influence on the stratigraphic record worldwide through significant environmental changes, the boundary position was key to understanding how different coeval mechanisms influenced the studied equatorial carbonate ramp. Notably, the newest biostratigraphic studies performed on the Musandam Peninsula positioned the Triassic-Jurassic boundary at different positions within the stratigraphic record (boundary between the Ghalilah Formation and the Musandam Formation de Matos 1997; boundary between the Sumra Member and the Sakhra Member, within the Ghalilah Formation, Maurer et al., 2008). Taking all previous biostratigraphic findings into account, combined with the isotope data, supports the placement of the Triassic-Jurassic boundary at the transition between the Ghalilah and the Musandam Formations. The combined approach between carbonate carbon and total organic carbon isotopes and the similar trends preserved by both archives, serves as confirmation for a primary carbon isotopic signal.

As the Musandam Peninsula probably hosts the stratigraphically most complete sections of the Triassic and Jurassic sequences of the Arabian Peninsula (e.g. Maurer et al., 2015), the positioning of the Triassic-Jurassic boundary allows for a refinement of regional Palaeogeographical maps and for the examination of an equatorial carbonate record across this boundary. The findings of this study reveal that there is no evidence for a biocalcification crisis in the eastern equatorial Tethys, in accordance with findings by Hallam et al. (2000) from Tibet. This questions the proposed ocean acidification event as the

proposed kill mechanism of the occurring mass extinction (e.g. Hautmann et al., 2008) that affected mainly marine biota, as evidence for this should be present in the entire Tethys. The results furthermore suggest that the shallow-marine carbonate ramp of the Musandam Peninsula was affected by a broad relative sea level rise that commenced during the Latest Triassic. This was also recorded from the record studied in Tibet (Hallam et al., 2000) and implies that this broad sea level rise was likely not just occurring locally in the area of the Musandam Peninsula. Carbonate factory demise occurring coevally in the western Tethys might have equally been caused by (local) tectonic changes, resulting in relative sea level changes. It remains speculation, whether a global sea level rise might have been the main driver of the Triassic-Jurassic mass extinction, potentially driven by an increase in global temperature, induced by Central Atlantic Magmatic Province related volcanism. The findings presented here do not support the ocean acidification model as the main kill mechanism.

The second main set of results of this project deal with heterogeneities within epeiric platforms, preserved mostly in a 'layer-cake' architecture. The data gathered here – over a range of scales - serves as a case study for facies distribution on extensive, shallow-marine, epeiric carbonate ramps, especially in the Middle East, that might have a profound influence on reservoir properties, both for hydrocarbon exploration and for CO₂ storage.

The findings illustrate that lithofacies types do generally not occur as belts parallel to the coastline in this type of system. This holds true irrespective of the main tidal regime within the neritic realm (e.g. intertidal or subtidal). Facies types preserved within the studied outcrops show a large degree of lateral variability at all studied scales, despite a largely 'layer-cake' type bedding pattern. This is valid for all encountered lithofacies types within the shallow-marine environment of deposition. Both 'conventional' and 'unconventional' reservoir facies types are not continuous over the region. The regional scale study however reveals that the broad environments of deposition (EODs) within the shallow-marine ramp were continuous over at least 45 km. A closer look at the shallow-, open-marine carbonate ramp EOD reveals that facies types within are laterally highly variable at a scale of one to two kilometres ('inter-well')

scale). A Markov chain analysis reveals that facies have a probability of 60 % of being laterally discontinuous over distances between 220 and 1100 m. The lateral variability of facies types in these types of settings can be invoked from similar facies maps at comparable settings, but the facies map presented here is the first of its kind from Jurassic carbonates in the Middle East. The combination of laterally heterogeneous facies types and vertical exposure stages is likely the result of constant autocyclic mixing and re-shuffling of grains within the shallow-marine EOD and allocyclical changes in relative sea level.

Facies variability within the same outcrop at the scale of a few to tens of metres reveals a very similar lateral heterogeneity and sheds more light on the underlying processes. The small-scale facies map presented here is unique as it contains closely-spaced intra-bed facies transitions. Previous studies performed at a similar scale contain an overall less heterogeneous facies pattern (calculated through the coefficient of variation) and only contain information on the heterogeneity within individual shoal bodies. The results presented here at the metre-scale reveal that lateral facies variability within the shallow-marine EOD is likely a result of the combination between locally heterogeneous carbonate production as well as of coeval mixing and re-sedimentation processes. This re-distribution of different sediment types, according to their respective transport properties, is in turn caused by various hydrodynamic processes with varying intensities. The comparison with modern analogues reveals similarities between the sediment distribution and morphologies between the ancient and the modern shallow-marine EOD. It highlights the applicability, at least in some aspects, of modern analogue studies as well as the preservation potential of closely-spaced facies mosaics within ancient carbonate systems.

8.2 Recommendations for future work

Given the vertical extent, the excellent exposure, the existing biostratigraphic framework and the good preservation of an original carbon isotope signal in the studied sequences, the Musandam Peninsula would serve as excellent location for a western Tethys carbon isotope curve, comparable to the compiled curve of Dera et al. (2011), that was produced from fossil hardparts, mainly from the

eastern Tethys. In combination with radiogenic strontium isotope stratigraphy on suitable samples, this would allow for additional global assessment of the various Jurassic carbon cycle perturbations.

The excellent lateral exposure as well as dimensions of the Jurassic outcrops might serve as excellent site for a LiDAR (Light Detection and Ranging) study, as has been performed in other wadis on the Musandam Peninsula in Permo-Triassic strata (Ahmadzamri et al., 2014).

This study presents important quantitative data on facies distribution on shallow-marine epeiric carbonate ramps. It is the first study to compare the lateral facies variability in between different small-scale facies panels using the coefficient of variation as quantitative measure for facies heterogeneity within outcrops. As the database on small-scale (metre- to 'inter-well' scale) facies distribution maps is likely to grow, it would be beneficial to continuously explore statistical measures to quantify heterogeneities. This allows easier implementation into reservoir models through facies transition probabilities matrices as well as an objective and comparable assessment of facies heterogeneity at various scales, through data such as the coefficient of variation.

Modern analogue studies performed at a very small scale, examining the vertical and lateral distribution of facies types within the shallow-marine EOD would be beneficial in order to get a better understanding of the physical and biological processes that control carbonate sediment distribution at the metre scale. Especially studies from the Bahamas are paving the way and are closing in on smaller scales (Harris, 1979; Rankey and Reeder, 2011; Sparks and Rankey, 2013) but are not showing small-scale facies panels, as presented in many outcrop studies.

The discussion dealing with the heterogeneous facies pattern at the metre- to tens of metres scale touches on the topic of sedimentary bed preservation and is a reminder that some fundamental questions regarding sediment preservation remain unsolved. Burgess and Wright (2005) touch on the conundrum, that sedimentologists are still not entirely sure what a sedimentary bed actually represents. The present study reveals some insight with respect to heterogeneous facies mosaics that can be preserved within a bed, but it remains unclear of whether this is a snapshot of a past environment

(much like a frozen image of the seafloor when looking through Rankey's *kaleidoscope* (Rankey, 2016) or rather a kind of *bouillabaisse* averaged through syndepositional and postdepositional mixing and altered through burial and diagenesis. Further research comparing Holocene and older sediments with a lot of attention for processes might shed more light on this fundamental sedimentology question. This thesis is a step in the direction of characterising heterogeneous carbonates at various scales and better understanding the underlying processes.

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Publications from this thesis

Journal articles

- Hönig, M.R., John, C.M., Manning, C. (under revision). Development of an equatorial carbonate platform across the Triassic-Jurassic boundary and links to global palaeoenvironmental changes (Musandam Peninsula, UAE/Oman). Gondwana Research, Manuscript Number GR-D-16-00140.
- Hönig, M.R., John, C.M. (2015). Sedimentological and isotopic heterogeneities within a Jurassic carbonate ramp (UAE) and implications for reservoirs in the Middle East. *Marine and Petroleum Geology* 68, 240–257.

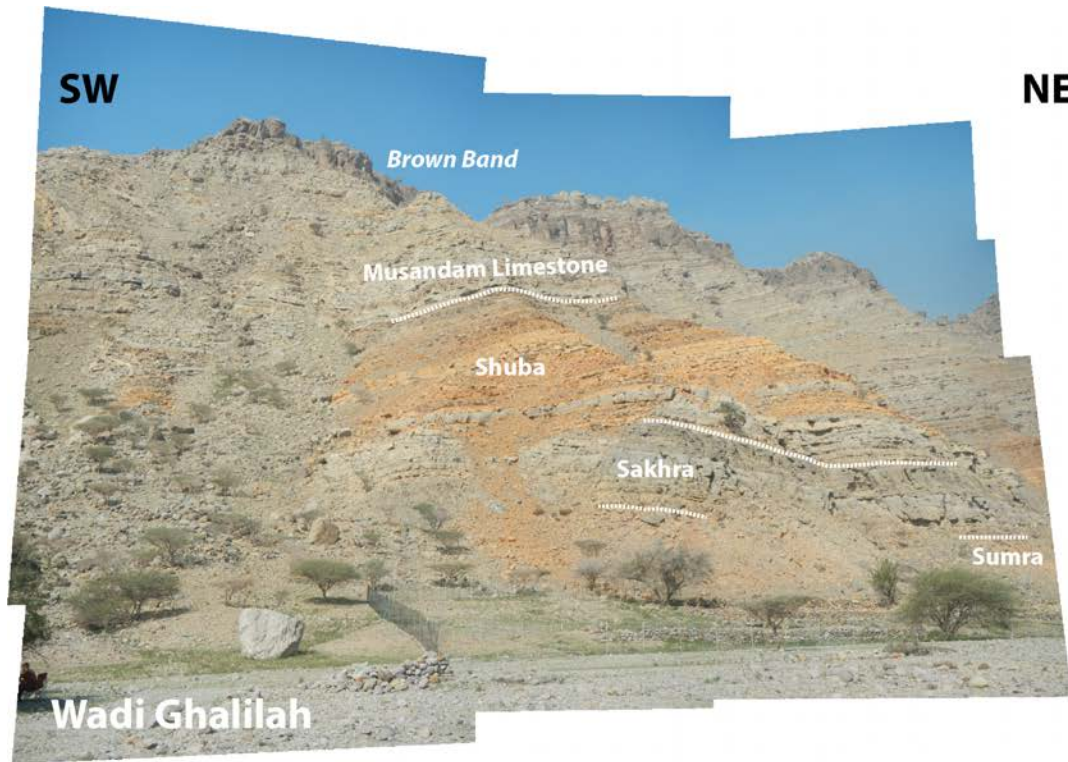
Conference abstracts

- Hönig and John (2016). Regional to inter-well scale sedimentology of a Triassic - Jurassic carbonate ramp (UAE) analogous to Middle Eastern reservoirs. GEO Bahrain, oral presentation.
- Hönig and John (2015). Are shallow-water carbonates a reliable archive for global environmental change? BSRG AGM Keele, poster presentation.
- Hönig and John (2015). Carbonate factory evolution and carbon isotope stratigraphy across the T-J boundary from the Musandam Peninsula (UAE). 5th EAGE Arabian Plate Geology Workshop Kuwait, oral presentation.
- Hönig and John (2014). Equatorial carbon cycling and the evolution of a carbonate factory across the Triassic-Jurassic boundary: evidence from the Musandam Peninsula, UAE. IAS Geneva, oral presentation.
- Hönig and John (2014). Heterogeneities on a Jurassic carbonate ramp (UAE) as analogues for reservoirs in the Middle East. GEO Bahrain, oral presentation.
- Hönig and John (2013). Palaeo-equatorial temperatures and carbon-cycle evolution at the Triassic-Jurassic boundary: A stable isotope perspective from shallow-water carbonates from the UAE. AGU Fall meeting San Francisco, poster presentation (OS13C-1712).
- Hönig and John (2013). Facies distribution on a Middle Jurassic carbonate ramp (United Arab Emirates). IAS Manchester, poster presentation.

Appendices

Appendix A – Chapter 4

A.1 Panorama of study site at Wadis Ghalilah and Al-Ghabbah.



A.2 Panorama of study site at Jabal Sall Ala.



A.3 Corals from the Shuba Member at Wadi Ghalilah.



A.4 Strontium isotope data from Wadi Naqab.

Sample	Material	$^{87}\text{Sr} / ^{86}\text{Sr}$	standard error
WN TJ B2	Oyster	0.707723	0.000007
WN TJ B3	Oyster	0.707734	0.000007
WN 168 Oy	Oyster	0.707909	0.000005
WN 193 Br3	Brachiopod	0.707921	0.000005
SRM 987 97ng	standard	0.710235	0.000007
SRM 987 97ng	standard	0.710221	0.000006
SRM 987 97ng	standard	0.710235	0.000009

A.5 Stable carbon and oxygen isotope data from Wadi Naqab.

Wadi Naqab			
Sample	rel strat height from base	Del18O corr	Del13C corr
WN 139	192.8	-4.76	0.80
WN 140	194.8	-4.28	-0.24
WN 141	196.8	-5.89	-0.31
WN 142	199.8	-0.90	2.30
WN 143	214.8	-1.81	-0.82
WN 144	220.8	-4.39	-1.89
WN 145	229.8	-1.29	-0.31
WN 146	232.8	-3.36	-0.33
WN 147	236.8	-2.07	-1.53
WN 148	256.8	-1.40	-0.29
WN 149	282.8	-0.88	-0.99
WN 151	300.8	-1.08	-1.92
WN 152	302.8	-4.56	-3.23
WN 153	308.8	-0.98	-3.08
WN 154	315.8	-5.35	-2.09
WN 155	320.8	-4.67	-3.13
WN 156	325.8	-0.30	-2.73
WN 157	329.8	-6.05	-3.40
WN 158	332.8	-4.37	-1.62
WN 160	343.8	-4.56	-1.43
WN 161	347.8	-5.19	-1.98
WN 162	349.8	-4.14	-2.01
WN 163	353.8	-1.40	-0.61
WN 164	357.8	-3.74	-1.71
WN 165	358.8	-3.54	-1.88
WN 166	359.8	-5.22	-1.08
WN 167	360.8	-4.04	-0.13
WN 168	361.8	-3.96	0.08
WN 169	362.8	-6.91	-0.03
WN 170	364.8	-5.09	0.17
WN 171	365.8	-4.96	0.08
WN 172	366.8	-3.29	0.64
WN 173	367.8	-4.40	-0.16
WN 174	369.8	-0.38	1.34
WN 175	372.8	-3.40	0.74
WN 176	374.8	-3.84	0.71
WN 177	377.8	-3.13	0.71
WN 178	381.8	-3.21	0.56
WN 179	387.8	-3.60	-0.09
WN 180	390.8	-3.58	0.29
WN 181	398.8	-3.81	0.14
WN 182	400.8	-3.76	0.27
WN 183	404.8	-3.82	0.65
WN 184	409.3	-3.82	1.18
WN 185	410.8	-3.82	1.25
WN 186	414.8	-1.13	1.38
WN 187	419.8	-3.06	1.47
WN 188	424.8	-0.13	0.68
WN 189	431.8	-3.70	1.91

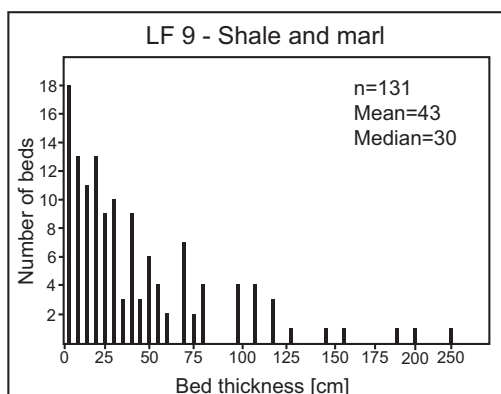
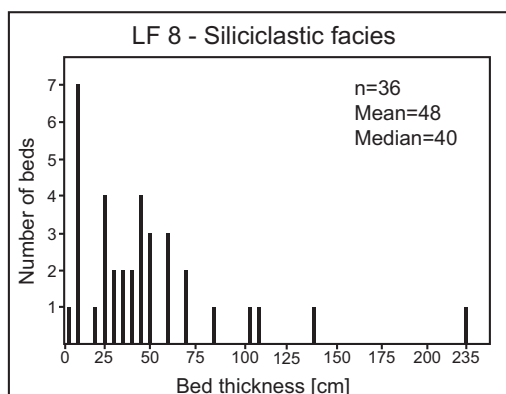
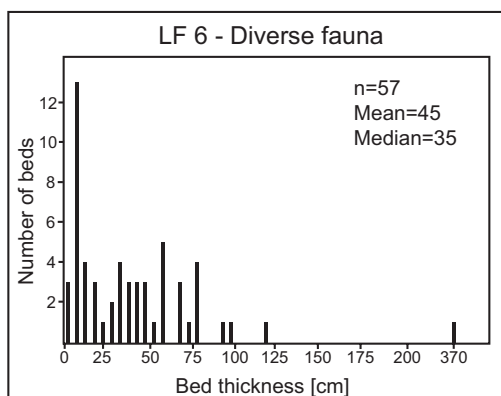
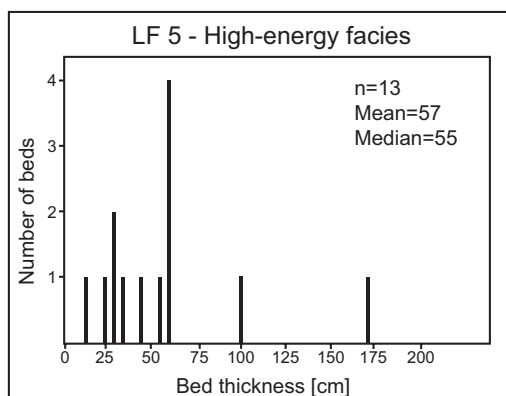
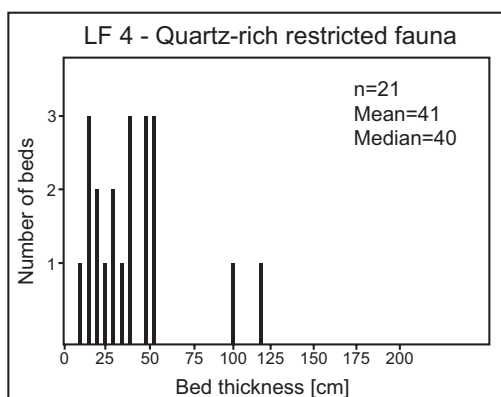
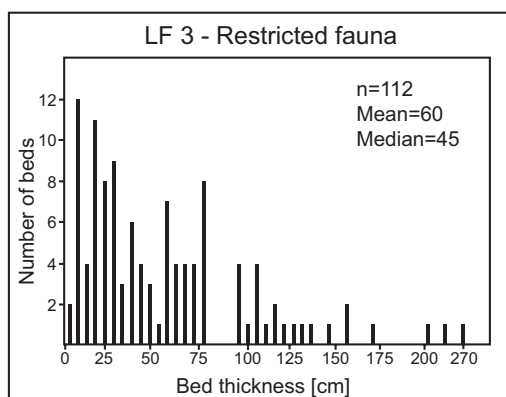
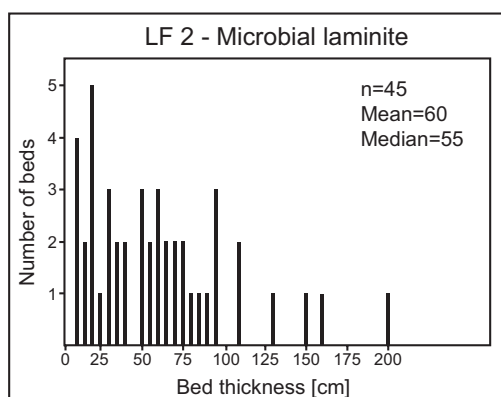
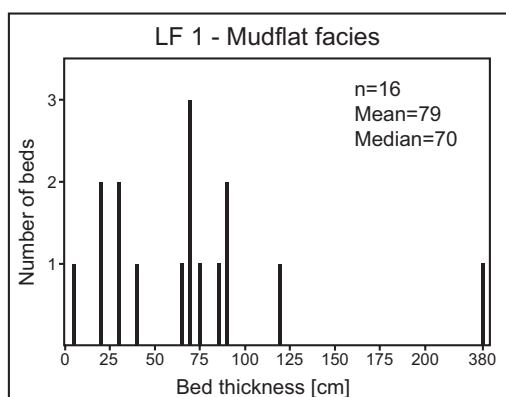
A.6 Stable carbon and oxygen isotope data from Wadi Ghalilah.

Wadi Ghalilah			
Sample	rel strat height from base	Del18O corr	Del13C corr
WG 44	175.16	-2.44	-2.06
WG 45	176.66	-6.72	-3.18
WG 47	182.16	-0.82	-1.82
WG 48	183.66	-7.73	-4.51
WG 49	184.16	-1.04	-2.26
WG 50	185.16	-1.17	-1.96
WG 51	188.66	-2.53	-3.33
WG 52	191.16	-1.91	-2.61
WG 53	194.66	-1.74	-2.86
WG 54	197.66	-4.92	-2.95
WG 55	200.16	-6.86	-1.87
WG 56	202.16	-4.39	-2.63
WG 57	206.66	-8.43	-4.92
WG 58	210.66	-1.35	-2.29
WG 59	214.66	-6.92	-4.08
WG 61	222.16	-7.18	-2.24
WG 25	228.66	-7.00	-0.37
WG 26	233.16	-6.70	-0.20
WG 27	237.16	-6.84	-0.24
WG 28	239.16	-2.76	0.75
WG 29	241.66	-7.61	0.16
WG 30	246.16	-2.66	0.95
WG 31	249.66	-6.38	0.04
WG 32	255.16	-6.28	0.33
WG 33	260.66	-2.27	0.67
WG 34	262.66	-4.67	-0.19
WG 35	267.16	-6.40	-1.09
WG 36	276.66	-7.72	-1.41
WG 37	278.66	-7.54	-0.41
WG 38	282.16	-6.51	-0.23
WG 39	292.16	-8.27	-0.24
WG 40	294.66	-3.76	0.63
WG 41	297.16	-6.34	0.09
WG 42	301.16	-6.41	-0.09
WG 43	308.16	-5.71	0.56

A.7 Stable carbon and oxygen isotope data from Wadi Al-Ghabbah.

Wadi Al-Ghabbah			
Sample	rel strat height from base	Del18O corr	Del13C corr
WS 1	0.5		
WS 2	5	-6.58	1.91
WS 3	13.5	-8.46	1.08
WS 4	19.5	-1.44	0.87
WS 5	25	-1.29	-0.94
WS 6	28.5	-1.86	-2.23
WS 7	31	-6.98	-0.55
WS 8	38	-6.95	-6.60
WS 9	43	-3.84	-1.81
WS 10	49	-1.76	-2.71
WS 11	55	-0.88	-2.16
WS 12	59.5	-1.27	-1.90
WS 13	65	-1.92	-3.69
WS 15	71	-2.13	-2.05
WS 16	73.5	-0.65	-2.26
WS 17	75	-7.23	-1.84
WS 18	78	-6.89	-2.45
WS 19	80	-1.15	-2.10
WS 20	83.5	-5.43	-3.74
WS 21	88	-7.28	-8.27
WS 22	93	-6.88	-2.56
WS 23	95	-7.34	-3.86
WS 24	103	-5.91	-0.67
WS 25	106	-6.09	-0.63
WS 26	107	-7.09	-0.28
WS 27	108	-5.23	-0.11
WS 28	111.5	-4.46	-0.27
WS 29	113.5	-6.07	-2.53
WS 30	114.5	-0.31	0.27
WS 31	116	-5.76	-1.63
WS 32	117	-6.07	-0.39
WS 33	119	-6.49	-0.11
WS 34	119.5	-5.65	-0.40
WS 35	120	-3.21	0.18
WS 36	121.5	-5.56	-2.36
WS 37	123	-5.19	-0.95
WS 38	124	-5.28	-3.43
WS 39	125.5	-5.59	-1.16
WS 40	126	-6.29	0.56
WS 41	128	-4.16	0.22
WS 42	130.5	-4.22	0.87
WS 43	132	-6.47	0.62
WS 44	135	-6.39	0.56
WS 45	136	-6.50	-0.54
WS 46	138.5	-5.78	-2.93
WS 47	139.5	-3.04	-0.16
WS 48	141	-5.80	-1.14
WS 49	142	-0.67	-0.05
WS 50	144	-0.87	0.12
WS 51	145	-0.65	0.20
WS 52	146	-0.25	0.31
WS 53	147	-5.86	-0.35
WS 54	150	-5.48	-3.21
WS 55	156.5	-5.66	-2.79
WS 56	160	-5.96	-1.38
WS 57	163	-1.99	-1.34
WS 58	168	-1.35	-1.21
WS 59	173	-0.66	-0.43
WS 60	178	-6.01	-1.89

A.8 Histograms of bed thicknesses within the Shuba Member and the Musandam Limestone compiled from sections at Wadis Naqab, Ghalilah and Al-Ghabbah.



Appendix B – Chapter 5

B.1 Geochemical data.

Sample	rel strat height	Del18O corr	Del13Ccarb corr	Del13Corg	Carb cont	P cont
WN 139	192.8	-4.76	0.80		0.98	
WN 140	194.8	-4.28	-0.24	-26.69	0.86	
WN 141	196.8	-5.89	-0.31	-26.69	0.87	
WN 142	199.8	-0.90	2.30		0.88	
WN 143	214.8	-1.81	-0.82	-25.09	0.89	
WN 144	220.8	-4.39	-1.89		0.81	
WN 145	229.8	-1.29	-0.31		0.72	
WN 146	232.8	-3.36	-0.33	-27.86	0.75	
WN 147	236.8	-2.07	-1.53	-27.51	0.78	
WN 148	256.8	-1.40	-0.29	-26.04	0.81	
WN 149	282.8	-0.88	-0.99		0.84	
WN 151	300.8	-1.08	-1.92		0.30	
WN 152	302.8	-4.56	-3.23	-27.48	0.90	
WN 153	308.8	-0.98	-3.08	-25.52	0.92	
WN 154	315.8	-5.35	-2.09	-27.02	0.18	
WN 155	320.8	-4.67	-3.13	-27.99	0.95	71
WN 156	325.8	-0.30	-2.73	-26.23	0.97	136
WN 157	329.8	-6.05	-3.40		0.89	67
WN 158	332.8	-4.37	-1.62		0.80	34
WN 159	339.8			-28.41	0.87	59
WN 160	343.8	-4.56	-1.43	-27.09	0.95	52
WN 161	347.8	-5.19	-1.98	-26.85	0.98	230
WN 162	349.8	-4.14	-2.01	-27.05	0.78	36
WN 163	353.8	-1.40	-0.61		0.99	336
WN 164	357.8	-3.74	-1.71		0.98	121
WN 165	358.8	-3.54	-1.88		0.60	646
WN 166	359.8	-5.22	-1.08	-26.32	0.96	472
WN 167	360.8	-4.04	-0.13	-26.64	0.95	126
WN 168	361.8	-3.96	0.08	-23.56	0.93	178
WN 169	362.8	-6.91	-0.03	-26.97	0.92	429
WN 170	364.8	-5.09	0.17		0.93	2158
WN 171	365.8	-4.96	0.08	-25.88	0.94	401
WN 172	366.8	-3.29	0.64	-26.74	0.95	362
WN 173	367.8	-4.40	-0.16	-26.28	0.96	1763
WN 174	369.8	-0.38	1.34	-25.16	0.98	288
WN 175	372.8	-3.40	0.74	-26.45	0.98	111
WN 176	374.8	-3.84	0.71	-26.16	0.98	80
WN 177	377.8	-3.13	0.71	-23.29	0.98	75
WN 178	381.8	-3.21	0.56	-26.42	0.97	66
WN 179	387.8	-3.60	-0.09	-26.25	0.97	64
WN 180	390.8	-3.58	0.29	-25.86	0.97	139
WN 181	398.8	-3.81	0.14	-26.07	0.96	56
WN 182	400.8	-3.76	0.27	-26.29	0.96	198
WN 183	404.8	-3.82	0.65	-26.74	0.96	
WN 184	409.3	-3.82	1.18		0.95	
WN 184	409.3	-3.82	1.18		0.95	
WN 185	410.8	-3.82	1.25	-26.20	0.95	
WN 186	414.8	-1.13	1.38	-26.47	0.96	
WN 187	419.8	-3.06	1.47	-21.67	0.97	
WN 188	424.8	-0.13	0.68		0.70	
WN 189	431.8	-3.70	1.91	-26.95	0.99	

Appendix C – Chapter 6

C.1 Wadi Naqab stable isotope data – section 1.

Sample	rel height	Del 13 C corr	Del 18 O Corr
WN 20	5	1.74	-2.90
WN 21	10	1.19	-3.60
WN 22	13		-5.68
WN 23	22	1.01	-2.90
WN 24	28	-1.95	-4.97
WN 25	34	-4.66	-6.14
WN 26	41	1.24	-3.33
WN 27	47	-0.73	-4.53
WN 28	51	2.21	-2.33
WN 29	63	0.08	-4.71
WN 30	67	1.47	-3.11
WN 31	72	1.36	-2.98
WN 32	77	1.18	-3.48
WN 33	80	0.20	-3.84
WN 34	84	-1.72	-5.42
WN 35	88	0.46	-4.40
WN 36	94	2.42	-2.79
WN 37	100	2.14	-3.20
WN 38	107	2.30	-2.59
WN 39	114	1.92	-2.89
WN 40	119	-1.34	-4.73
WN 41	125	1.73	-2.98
WN 42	132	1.75	-2.87
WN 43	139	1.63	-2.53
WN 44	148	1.92	-2.66
WN 45	153	2.17	-2.60
WN 92	162	1.81	-2.80
WN 93	170	2.12	-2.55
WN 94	175	2.34	-2.62
WN 95	183	2.19	-2.41
WN 96	191	2.41	-2.66
WN 97	195	2.74	-2.37
WN 98	198	2.07	-2.60
WN 99	200	1.91	-2.53
WN 100	205	2.67	-2.88
WN 101	213	2.14	-2.77
WN 102	219	2.46	-3.01
WN 103	225	2.83	-2.28
WN 104	237	2.46	-2.53
WN 105	244	2.48	-3.11
WN 106	248	3.02	-2.82
WN 107	257	1.30	-2.99
WN 108	263	3.02	-2.75
WN 109	268	3.46	-2.45
WN 110	275	3.59	-2.37
WN 111	279	1.84	-2.90
WN 112	285	2.01	-2.82
WN 113	286	3.06	-2.63
WN 114	290	1.23	-2.58
WN 115	298	1.89	-2.55

C.2 Wadi Naqab stable isotope data – section 2.

Sample	rel height	Del 13 C corr	Del 18 O Corr
WN 129	11	1.77	-2.64
WN 130	15	1.62	-2.55
WN 131	23	1.04	-2.41
WN 132	27	2.08	-2.58
WN 116	28	1.95	-3.30
WN 117	35	1.22	-2.89
WN 118	42	0.93	-3.46
WN 119	46	2.32	-2.68
WN 120	55	1.96	-2.95
WN 121	63	2.41	-2.74
WN 122	65	2.24	-2.69
WN 123	72	1.65	-3.06
WN 124	83	1.64	-3.21
WN 125	92	1.50	-3.47
WN 126	97	1.62	-2.94
WN 127	106	0.21	-4.41
WN 133	211	2.65	-2.56
WN 134	217	2.75	-2.92
WN 135	221	3.07	-2.30
WN 136	227	2.96	-2.50
WN 137	235	2.82	-2.01

C.3 Wadi Naqab stable isotope data – section 4.

Sample	rel height	Del 13 C corr	Del 18 O Corr
WN 46b	2	0.86	-3.41
WN 47	6	1.39	-2.77
WN 48	9	1.08	-2.97
WN 49	10.5	0.98	-2.87
WN 50	13	1.22	-3.02
WN 51	16	0.96	-3.41
WN 52	18	0.92	-3.08
WN 53	22	0.54	-4.10
WN 54	24.5	1.52	-3.13
WN 55	25.5	1.17	-3.27
WN 56	27	1.59	-3.60
WN 57	28	1.84	-3.51
WN 58	30	0.81	-3.87
WN 59	34	1.52	-3.10
WN 60	40	1.79	-2.89
WN 61	43	0.07	-4.10
WN 62	45	-0.36	-4.25
WN 63	47	-0.54	-3.21
WN 64	51	-0.78	-4.39
WN 65	59	2.45	-3.05
WN 66	65	2.43	-3.14
WN 67	72	1.07	-2.85
WN 68	64	1.30	-2.91
WN 69	77	1.07	-2.93
WN 70	80	0.94	-2.68
WN 71	85	1.08	-3.79
WN 72	88	1.93	-3.24
WN 73	95	1.05	-3.70
WN 74	99	2.51	-3.01
WN 75	102	2.06	-3.04
WN 76	104	-0.44	-4.11
WN 77	104.2	-0.37	-3.97
WN 78	108	0.99	-3.10
WN 79	112	-0.14	-2.62
WN 80	115	1.20	-3.55
WN 81	118	1.78	-2.93
WN 82	122	1.64	-3.50
WN 83	127	1.54	-3.26
WN 84	131	1.38	-2.88
WN 85	132.5	1.11	-3.11
WN 86	134	0.78	-3.07
WN 87	141	1.81	-2.99
WN 88	144	-2.14	-5.32
WN 89	150	2.00	-3.00
WN 90	154	-2.14	-5.47
WN 91	158	2.18	-2.61

C.4 Wadi Naqab stable isotope data – section 7.

Sample	rel height	Del 13 C corr	Del 18 O Corr
WN 415	1.6	1.20	-3.49
WN 416	2.1	-1.88	-5.60
WN 417	2.7	-1.22	-4.23
WN 418	3.4	0.60	-3.85
WN 419	4.2	1.48	-3.39
WN 420	5.2	2.26	-2.99
WN 421	5.8	2.20	-2.94
WN 422	6.4	1.11	-3.95
WN 423	6.6	2.10	-2.25
WN 424	7.3	2.27	-3.43
WN 425	7.9	2.21	-3.28
WN 426	8.9	2.34	-3.19
WN 427	10.1	1.25	-3.57
WN 428	10	-0.87	-4.55
WN 429	10.4	0.98	-3.20
WN 430	10.9	2.07	-2.82
WN 431	11.7	1.32	-2.50
WN 432	12.3	1.49	-3.54
WN 433	12.7	1.99	-2.85
WN 434	13	2.30	-2.75
WN 435	14.3	-0.23	-4.02
WN 436	14.8	2.22	-5.67
WN 437	15.55	1.52	-3.36
WN 438	16.5	1.81	-4.05
WN 439	17.3	2.53	-2.71
WN 440	17.5	2.54	-2.77
WN 441	18.5	0.83	-3.63
WN 442	19	2.55	-2.64
WN 443	19.5	2.04	-3.08
WN 444	20.35	1.44	-3.51
WN 445	21.1	1.25	-2.74
WN 446	21.3	2.38	-2.64
WN 447	21.7	2.47	-2.89
WN 448	22.2	2.80	-2.61
WN 449	22.9	2.68	-2.60
WN 450	23.8	2.78	-2.85
WN 451	24.1	2.48	-2.78
WN 452	24.6	2.44	-2.60

Appendix D – Chapter 7

D.1 Facies dimensions measured at Wadi Naqab.

LATERAL dimensions	
Facies	width (m)
OncF	9
	9
	12
	14.5
OncGr	27
	39
	12
	13
	6
PelGr	60
	30
	34.5
PelP	18
	85
	26.5
	13
	6
Biocl	18
	12
	12
	15
Mud	30
	30
	30
Median	16.5
Mean	23.40
Stdev	18.25
Cv	0.78

VERTICAL (m)		
Facies	min	max
OncF	0.2	0.4
OncGr	0.15	0.4
PelGr	0.15	0.95
PelP	0.1	0.45
Biocl	0.15	0.4
Mud	0.05	0.05

D.2 Facies dimensions measured from Obermaier et al. (2015) – 240 m panel.

LATERAL	
Facies	width (m)
Mudstone	30
	22
Wackestone	77
	78
	33
	120
	170
	20
	20
	10
	61
	70
	85
	37
	78
Packstone	25
	10
	115
	85
	138
	40
	65
	8
	20
	8
	10
Grainstone	58
	78
	55
	112
	118
	90
	8

Median	58
Mean	59.21
Stdev	43.32
Cv	0.73

VERTICAL (m)		
Facies	min	max
Mudstone	0.05	0.15
Wackestone	0.05	0.2
Packstone	0.05	0.25
Grainstone	0.05	0.55

Obermaier et al., 2015. GeoArabia - **Fig. 14 (C)**

D.3 Facies dimensions measured from Obermaier et al. (2015) – 12 m panel.

LATERAL	
Facies	width (m)
Packstone	5
	3
	1.5
Wackestone	8
	1
	3.8
	2.8
	8
	10
	6
	5
	1
	5
4	
Grainstone	12
	5
	3.8
	4.5
	7
	2
	1
	5
	1.8
	12
10	

VERTICAL (m)		
Facies	min	max
Packstone	0.01	0.1
Wackestone	0.01	0.1
Grainstone	0.02	0.35

Obermaier et al., 2015. GeoArabia - **Fig. 12**

Median	5
Mean	5.13
Stddev	3.31
Cv	0.65

D.4 Facies dimensions measured from Adams et al. (2011).

LATERAL	
Facies	width (m)
Fine Gr	68
	6
	64
	26
Med Biocl	66
	68
	68
Coarse Biocl	68
	16
Rudst	62

VERTICAL (m)		
vertical (m)	min	max
Fine Gr	0.1	3.8
Med Biocl	0.1	2
Coarse Biocl	0.2	1.3
Rudst	0.05	0.7

Median	65
Mean	51.2
Stddev	24.82
Cv	0.48

Adams et al., 2011. Petroleum Geoscience - **Fig. 14**

D.5 Facies dimensions measured from Sena (2013).

LATERAL Facies	width (m)
Packstone	6.5 6.9 25.6
Rudstone	52 7.4 13.7 26.7
Floatstone	52 37.8
Grainstone	52 36 28.6
Coral biost	3.1 16.6 14.5

Median	25.6
Mean	25.29
Stdev	17.43
Cv	0.69

VERTICAL Facies	(m) min	max
Packstone	0.05	0.25
Rudstone	0.05	0.25
Floatstone	0.05	0.65
Grainstone	0.3	1.9
Coral biost	0.05	0.2

Sena, 2013. PhD Thesis ICL - Fig. 3.5

D.6 Facies dimensions measured from Samankassou et al. (2003).

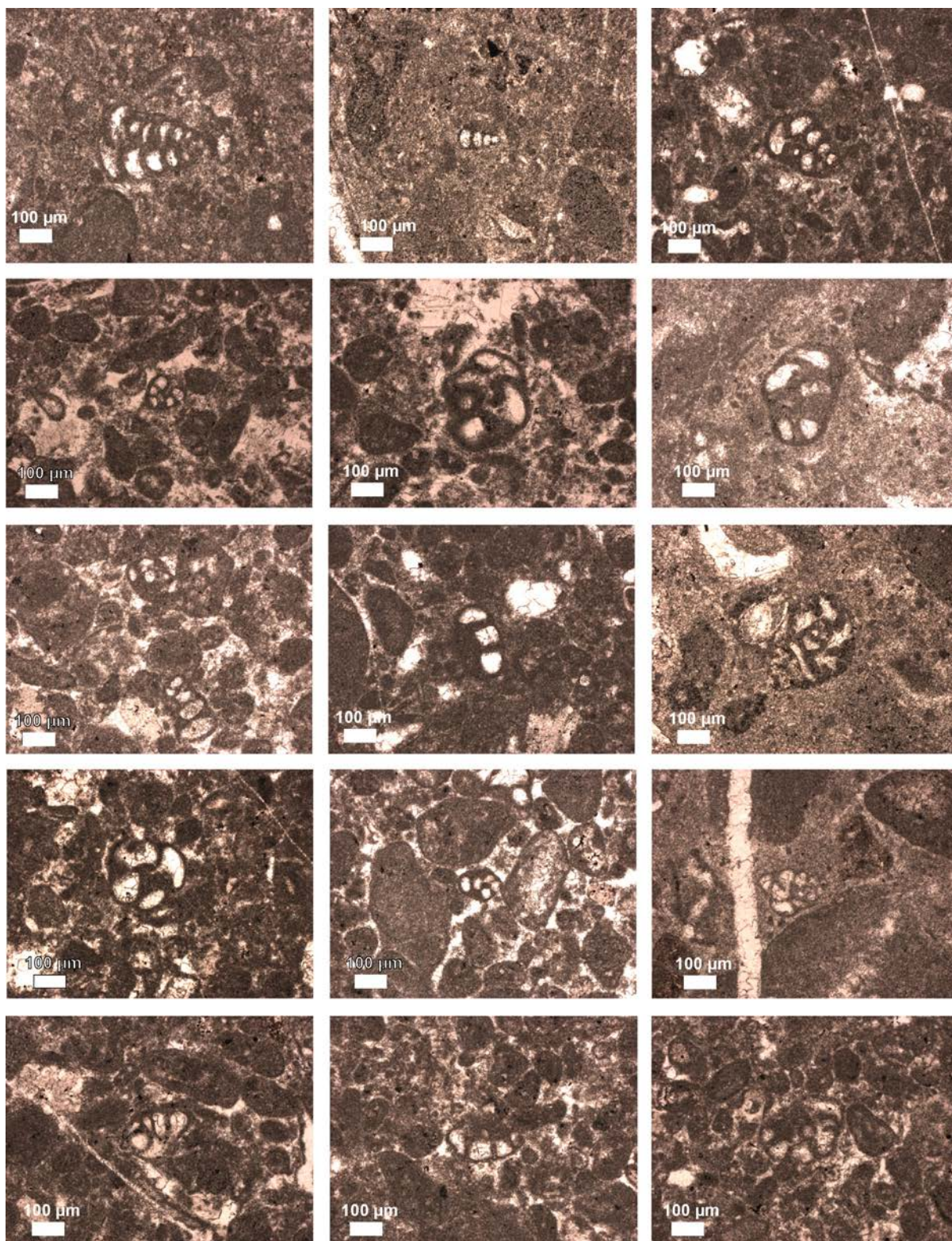
LATERAL Facies	width (m)
Biocl Lst	4.5 18.5 7 7
Biocl-Oo	7 1.25 11.5
Coral F	14 18.5
Cor-Biocl-Oo	18.5 7 7 11.5 11.5

Median	9.25
Mean	10.34
Stdev	5.48
Cv	0.53

VERTICAL Facies	(m) min	max
Biocl Lst	0.05	0.25
Biocl-Oo	0.1	0.3
Coral F	0.05	0.24
Cor-Biocl-Oo	0.2	0.9

Samankassou et al., 2003. Ecl. Geol. Helv. - Fig. 9

D.7 Foraminifera from the Middle Jurassic bedset.



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Dept. Earth Science and Engineering
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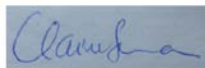
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