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Controls on reservoir quality in Early Cretaceous carbonate oil fields and implications for basin modelling

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ABSTRACT

Carbonate reservoirs hold more than 50 % of Earth’s remaining conventional hydrocarbon. However, recovery from these reservoirs is notoriously difficult due to the complex and multiple scales of porosity. This heterogeneity is a function of both the depositional environment and of subsequent diagenetic processes. This thesis examines the processes that have controlled the reservoir quality of three Early Cretaceous carbonate oil fields (A, B, and C), in particular the role of deposition, diagenesis and the timing of oil charge in controlling final properties. Results are then used to help provide a theoretical basis for the modelling and prediction of reservoir quality and to improve the calibration of basin models.

Field A and B are stacked and highly compartmentalised giant oil fields in the U.A.E. that are dominated by muddy fabrics and have a highly variable porosity (0-35 %) and permeability (0.01-830 mD). Although the depositional environment strongly determines the location of reservoirs extensive diagenesis, through cementation and dissolution, has greatly modified the porosity and permeability of the reservoirs. Bulk δ¹³C values obtained from the main pore occluding calcite and dolomite cements are similar to the δ¹³C values of bulk micrite for the reservoir interval in which they are now present. This suggests that the cements that are occluding the pore space in each stacked reservoir are locally sourced and implies that each reservoir behaves as a relatively closed system during cement precipitation.

In-situ (SIMS) δ¹⁸OVPDB values were obtained for the complete calcite cementation history of multiple reservoirs in Field A and B. The δ¹⁸OVPDB values for the first (oldest) calcite cement zone in each reservoir can be related to the global δ¹⁸OVPDB marine curve during the Hauterivian-Aptian and to million-year scale major climatic cooling events. The δ¹⁸OVPDB values for successive cement zones then progressively decrease, which is related to successive precipitation as a result of increasing temperature during burial in a relatively closed system.
In-situ (SIMS) \( \delta^{18}O_{VPDB} \) data together with oil inclusion occurrence suggest that initial oil charge (from the Dukhan Formation), at ~ 55-45 Million years ago (Mya) in Field A, reduced the cementation rate in the oil reservoir and preserved porosity. Whereas in the coeval aquifer a large volume of cement precipitated, after oil entered the oil reservoir, that greatly reduced porosity. Furthermore, the most reduced \( \delta^{18}O_{VPDB} \) and \( ^{m}\text{Mg}/^{m}\text{Ca} \) values are obtained from the cements in the shallowest (youngest) reservoirs, suggesting that cementation ceased in the deepest reservoirs first. This can be related to hydrocarbon stopping cementation or to the complete occlusion of effective porosity in the older reservoirs prior to the younger.

After calcite and dolomite cementation ceased in the reservoirs of Field A and B a large scale dissolution event has been identified which significantly enhanced porosity. This dissolution event is then followed by the precipitation of authigenic kaolinite. Basin modelling reveals that this dissolution event is likely to be related to the thermal maturation of sedimentary organic matter that is present within local intraformational seals and to the migration of organic acids prior to a second hydrocarbon charging event (at ~ 45 Mya). The aluminium, that is required for the formation of kaolinite, would then have been brought into the system by complexing with the organic compounds derived from this maturation event.

Field C is an oil field located in offshore Brazil. The field is dominated by high energy facies that have porosities which range from 5 % to 39 %, and permeabilities from 0.1 mD to 8.1 D. The depositional poro-perm properties of the oil reservoir have undergone little diagenetic alteration; however, the aquifer is extensively cemented and the porosity is much reduced. All the cements identified, by both petrography and stable isotopic analyses, in the oil reservoir are early and are thought to have formed from a pore fluid similar to, or slightly evolved from, Early Cretaceous seawater. Basin modelling suggests that oil may have entered the field slightly after deposition (at ~105 Mya) and led to the preservation of high porosities and permeabilities in the oil reservoir by stopping cementation.
DECLARATION

This thesis is entirely my own work, except where indicated otherwise. This work has not been submitted for any other professional qualification.

Dean Timothy Thorpe

Date
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CHAPTER 1
GENERAL INTRODUCTION

1.1 Introduction

Carbonate reservoirs hold > 50 % of the Earth’s remaining hydrocarbon (Sayers, 2008; Sloan and Sun, 2003), yet the recovery factors from these reservoirs are notably low, often < 35 % (Xie et al., 2005) compared with 62 % in some siliciclastic reservoirs (Kabir et al., 2008). The lower recovery factor in carbonate reservoirs is primarily due to the highly complex and heterogeneous pore systems found in carbonate rocks, where pores range in size over several orders of magnitude from microns to meters, and pore shapes are varied and irregular.

The complex pore systems in carbonates are typically thought to be a result of both the depositional environment and of subsequent diagenetic modifications (Ahr, 2008; Ehrenberg and Nadeau, 2005; Ehrenberg et al., 2009). The depositional environments of carbonates, that result in lithofacies, are typically well understood and can be used to estimate the initial porosity and permeability of carbonate sediment (Enos and Sawatsky, 1981). However, diagenetic modifications can act to greatly modify the initial depositional pore systems to create vastly different petrophysical properties compared to the original lithofacies. This results in highly tortuous and multi-scalar pore networks which commonly leads to there being no direct relationship between porosity and permeability, making the prediction of subsurface petrophysical properties difficult (Lucia, 2007).

Diagenesis can greatly affect the petrophysical properties of subsurface reservoirs by enhancing (via dissolution) or occluding (via cementation) the initial depositional porosity of sediment. However, the origins and effects of diagenetic processes on the evolution of porosity and permeability are typically difficult to
determine and have received little quantitative documentation (Ehrenberg and Nadeau, 2005; Moore and Druckman, 1981; Rosales and Perez-Garcia, 2010). This has led to numerous controversies as to the role of diagenetic processes in modifying the depositional pore networks, such as: the role of burial cementation in modifying porosity, the sources of burial cement (Lucia, 2007), whether diagenetic processes can be related to specific burial depths/temperatures; like in siliciclastics where chemical compaction is known to predominate at temperatures >90 °C (Bjørkum et al., 1998), the role of oil charge and whether diagenetic systems are ‘open’ or ‘closed’ (Heydari and Moore, 1993).

This thesis aims to advance our understanding of diagenesis in three Early Cretaceous carbonate oil fields by determining the origin, effects, timing and controls on diagenetic processes. Such an analysis can be used to understand the causes of pore heterogeneity, and so enable more accurate prediction of the distribution of subsurface porosity and permeability which can be used to increase the recovery efficiency from carbonate reservoirs. In particular, quantification of diagenetic effects can then be included in basin models.

Basin models have been extensively used to understand all the processes essential to determining the probability of a viable hydrocarbon accumulation, such as pressure and temperature distribution, hydrocarbon accumulation locations and volumes, and the evolution of fluid composition (Al-Hajeri et al., 2009; Baur et al., 2010; Loutfi and Sattar, 1987). Yet while these models enable examination of the dynamics of sedimentary basin evolution, they do not currently incorporate the key diagenetic processes of cementation and dissolution for carbonate systems, and in addition use generic compaction curves to predict the final porosity of any given reservoir. These shortcomings can lead to numerous inaccuracies such as when calculating the hydrocarbon reserves for a reservoir and when assessing the fluid flow pathways in a basin. Therefore, by including an understanding of diagenesis into basin models the models will be improved.
1.2 Aims and Objectives

Three Early Cretaceous carbonate oil fields are assessed in this thesis. Field A and B are located in the Rub Al Khali Basin in the United Arab Emirates (U.A.E.) (Figure 1.1) and are producing from the Thamama Group (134-124 Million years ago (Mya)); Field C is located in the Campos Basin, offshore Brazil (Figure 1.2) and is producing from the Quissamã Formation (110-107 Mya). The Thamama Group in Field A and B is typically composed of wackestone and packstone fabrics that have undergone extensive diagenesis through a long and complex burial history. By contrast, the Quissamã Formation in Field C is dominated by packstones and grainstones that have undergone little diagenetic alteration.

![Figure 1.1](image_url) Location map of the major oil fields in Abu Dhabi (black). Also shown is the location of the studied outcrop at Wadi Rahabah (modified from Alsharhan (1989) and Strohmenger et al (2006)).
Figure 1.2: Distribution of the oil and gas fields in the Campos Basin, Brazil. Modified from Guardado et al. (2000).

The overarching aim of this thesis is to understand the evolution of porosity and permeability from original sediment through all subsequent major diagenetic modification during burial for all three fields. This new understanding will then be incorporated into basin models so improving the accuracy of reservoir quality prediction in carbonate reservoirs. To achieve this, this thesis has:

1. Undertaken an assessment to determine the control that depositional environment has on current reservoir quality, by comparing the lithofacies identified to the current plug derived porosity and permeability data.

2. Obtained the bulk micrite $\delta^{13}$C and $\delta^{18}$O$_{VPDB}$ values through separate component analyses. The $\delta^{13}$C values can be used for chronostratigraphic correlation and the $\delta^{18}$O$_{VPDB}$ values can be used to determine the extent of rock-water interaction.

3. Constructed paragenetic sequences to help identify the diagenetic features present in the fields and their relative chronology.
4. Quantified the stratigraphic and lateral distribution of major cement phases and pore types. The cement volumes are considered in relation to depositional fabric to determine to what degree the initial depositional environment controls subsequent diagenetic processes.

5. Undertaken separate component analyses to obtain the $\delta^{13}$C and $\delta^{18}$O$_{VPDB}$ of the main pore occluding calcite and dolomite cement phases. This assessment can be used to help understand the pore fluid chemistry from which the cements precipitated and to determine the origin of the solutes required for cement formation. The $\delta^{18}$O$_{VPDB}$ of the cements can be used as a temperature proxy to constrain the temperature at which the cements formed.

6. Determined, via Electron Probe Microanalysis (EPM), the concentration of Ca, Na, Mg, Al, Si, K, P, Sr, Ti, Mn, Fe and O within calcite cement to aid in the assessment of the dynamics of cementation. The elemental concentrations derived from the cements can be used to help constrain how pore fluid chemistry has altered during the precipitation of calcite cement. Specific attention is paid to $^{m}$Mg$/^{m}$Ca ratios because they can be used to help predict the thermal history of the reservoir and to determine the effect of oil charge on the dynamics of cementation. This independent temperature proxy is integrated with in-situ $\delta^{18}$O$_{VPDB}$ data.

7. Identified, via cathodoluminescence assessment, the micro-cement stratigraphy of the main pore occluding cement phase (low-Mg calcite). High resolution Secondary Ion Mass Spectrometer (SIMS) in-situ $\delta^{18}$O$_{VPDB}$ data are then obtained, for the complete micro-cement stratigraphy identified in the reservoirs, to help understand the dynamics of cementation. The cause for any lateral or stratigraphic variability in the dynamics of cementation has then been discussed.

8. Incorporated this new insight of the main diagenetic processes into basin models using the facies piercing tool. These basin models have then been used to understand and reconstruct the geological and thermal evolution of the reservoirs in order to help understand the origin and controls on the
diagenetic features identified in the reservoirs. The models have also been used to assess the time of hydrocarbon generation, migration and accumulation to help gain a better understanding of the wider petroleum system.

1.3 Controls on Porosity and Permeability in Carbonates

This section will outline the origin and causes of porosity modification in carbonates, covering the controls on depositional (primary) porosity followed by the effects of diagenesis, how porosity can be classified, and the relationships identified between porosity and permeability in carbonate reservoirs.

1.3.1 Depositional (primary) porosity

The depositional environment in which carbonate sedimentation occurs determines the primary porosity and permeability of the carbonate rock and can greatly control the subsequent diagenetic processes affecting this rock (Lucia, 2007). Although the majority of carbonate production occurs in shallow, warm tropical marine waters (James, 1977), present day carbonate production occurs in a variety of different depositional environments and at all latitudes, and in tropical, warm and cold waters, this results in highly variable depositional (primary) porosities and permeabilities.

Primary porosity in carbonates is usually a function of sediment sorting and shape which are typically related to energy and the biological community. Typically in higher energy settings lime mud is winnowed away, leaving grain-dominated, porous carbonates (Figure 1.3). By contrast in lower energy settings, after the sediment has undergone dewatering, porosity is much reduced (Enos and Sawatsky, 1981; Lucia, 2007).

Primary permeability is a function of pore-size distribution and tortuosity, and typically increases as the volume of lime mud decreases (Figure 1.3). The permeability of mud-dominated sediment is typically between 1 and 200 mD whereas the permeability for grain-dominated sediment with some intergrain mud averages ~ 2 D (Lucia, 2007).
When porosity of carbonate rocks is considered in relation to the initial lithofacies and geologic time, porosity is notably reduced in increasingly older rocks, especially in those with mud-dominated fabrics (Figure 1.4). The cause for the lower porosity in the older sediments than in their more recent equivalents is typically because they have been buried deeper and are therefore more prone to burial diagenesis.

Figure 1.4: The relationship between porosity, depositional fabric and geologic time (Lucia, 2007).
1.3.2 Diagenetic (secondary) porosity

Diagenesis refers to all the physical, chemical and biological processes that affect a rock after initial deposition and lithification. Carbonates are far more chemically reactive than siliciclastics (Ehrenberg and Nadeau, 2005), such that most carbonates have undergone considerable diagenetic modification. Diagenetic processes, such as dissolution and cementation, are however difficult to predict and are dependent on: the rate and total volume of water movement, pH and chemical composition of the pore fluid, CO$_2$ activity, temperature and pressure, mineral stability, oil charge and its effect on wettability, and many others (Harris et al., 1985; Tucker et al., 1991).

Although diagenetic processes are difficult to predict, the diagenetic features that are a result of these processes are commonly placed within distinct diagenetic zones (Figure 1.5). The eogenetic, mesogenetic and telogenetic zones defined by Choquette and Pray (1970) are commonly used (Mehrabi and Rahimpour-Bonab, 2013; Mørk, 2013; Tucker et al., 1991). These terms will not be used in this thesis because they do not consider pore fluid chemistry. Instead the six diagenetic zones defined by Harris et al. (1985) are used because they are based on pore fluid chemistry variations:

1) Vadose zone: above the water table where pores are occupied by freshwater and air.

2) Meteoric phreatic zone: pores are occupied by meteoric water.

3) Mixing zone: both marine and meteoric water occupy the pore volume.

4) Marine phreatic zone: normal marine water fully saturates the pores.

5) Shallow marine burial zone: diagenesis occurs in pore water slightly more evolved than marine waters, at elevated temperatures and pressures away from near-surface processes.
6) Deep burial zone: where pore fluid may be connate or from other sources. The zone extends to the realm of low-grade metamorphism; it is in this zone where the majority of carbonates spend most of their geological existence.

![Diagram of diagenetic zones](image)

**Figure 1.5:** Typical calcite cements for each diagenetic zone (Moore, 1989).

The two main diagenetic processes that affect initial primary porosity and permeability are dissolution and cementation. Dissolution can enhance porosity in the subsurface and can occur at any point during the burial history of the rock (Figure 1.6a). This can lead to the generation of a large volume of secondary pore space. Pore fluid with a low Saturation Index (SI) and therefore undersaturated with respect to carbonate, is a primary cause for the development of secondary porosity. A low pore fluid SI is commonly observed in the vadose, meteoric phreatic and mixing zones where carbonates are dissolved by freshwater, leading to the formation of cave systems and karst surfaces (Saller et al., 1994). Dissolution can also occur in the burial environments by: mixing of water with dissimilar compositions, dissolution by acidic waters through decarboxylation, and dissolution by hydrogen sulphide due to the thermal maturity of organic matter (Mazzullo and Harris, 1991).
Pore-occluding processes can greatly reduce primary and secondary porosity and can also occur at any point during the burial history of the rock (Figure 1.6b). All cements propagate from the pore walls into pore space and therefore reduce pore space as they grow, with pore size reducing in proportion to the amount of cement precipitated. Some diagenetic cement types are characteristic of a particular diagenetic zone and therefore can be used to reconstruct the burial history of the rock. For example, meniscus calcite cements are thought to only develop in the vadose diagenetic zone. However, calcite cements with a similar fabric can develop in multiple diagenetic zones, an example being equant and blocky cements. This makes understanding the origin of the cement difficult in the absence of an accurate paragenetic sequence. Indeed, some authors suggest that burial cements lead to the greatest reduction in porosity (Choquette and James, 1987; Dutton, 2008; Feazel and Schatzinger, 1985; Scholle and Halley, 1985) while others suggest that cements precipitated in the near-surface/marine phreatic environment are the dominant pore occluding cements (Budd et al., 1993; Friedman, 1975; Halley and Schmoker, 1983; Longman, 1980). Therefore, one of the most important challenges to carbonate geologists is to unravel the diagenetic history, for each stage of dissolution and cementation and to understand its origin and effect on the porosity and permeability of the rock.
1.3.3 Pore types

The depositional environment and the subsequent diagenetic processes that affect the initial porosity and permeability of a carbonate rock can lead to a very varied and complex arrangement of pore types. Research on carbonate pore space has shown the importance of relating the present day porosity observed within the rock to depositional and diagenetic fabrics (Archie, 1952; Choquette and Pray, 1970; Lucia, 1983; Murray, 1960). In order to classify visible porosity a number of classification schemes can be used.

1) Archie Classification: This scheme was one of the first to classify carbonate porosity based on textural descriptions (Archie, 1952). Four classes (A-D) were identified for visible porosity. Class A has no porosity at 10 magnifications, Class B has microporosity (< 10 µm (Cantrell and Hagerty, 1999)), Class C has visible pores > 10 µm, and Class D includes large pores visible in cutting.

2) Choquette and Pray Classification: Choquette and Pray (1970) discussed the geologic concepts surrounding carbonate pore space and presented a classification scheme that is widely used (Figure 1.7). They emphasise the importance of pore space genesis and their classification scheme is genetic and not petrophysical, unlike Lucia (1983). 15 basic pore types were identified that can be organised into three classes based on whether they are fabric-selective, i.e. follow the initial depositional fabrics, non-fabric selective where porosity is independent of original fabric or fabric selective or not which is a combination of the aforementioned classes.

![Figure 1.7: Carbonate pore types separated into three classes: fabric selective, non fabric selective and fabric selective or not (Choquette and Pray, 1970).](image-url)
3) Lucia Classification: The Archie classification method was expanded on by Lucia (1983, 1995) who emphasised the petrophysical aspects of carbonate pore space. Lucia (1983) showed that by comparing pore types with laboratory measurements of porosity, permeability, capillarity and Archie $m$ values the most useful way of dividing pore space was between pore space located between grains, called interparticle porosity, and all other pore types called vuggy porosity. Vuggy porosity is subdivided further into two groups based on how the vugs are interconnected: 1) vugs that are interconnected only through interparticle pore network, termed separate vugs, and 2) vugs that form an interconnected pore system, termed touching vugs (Figure 1.8).

![Vuggy Pore Space Table](image)

**Figure 1.8: Lucia’s (2007) definition of vuggy porosity which is based on the interconnection of vuggy pores.**

4) Ahr Classification: Ahr (2005) suggests that present day carbonate porosity is a function of depositional porosity, diagenesis and mechanical fracture and defines these as three end members on a ternary plot (Figure 1.9). Hybrid porosity is porosity that has been created by more than one of these end member processes. This scheme uses the pore types identified by Choquette and Pray (1970) but tries to impart a stronger understanding as to their origin.
**1.3.4 Porosity-permeability relationships**

Porosity can be defined through both visual and laboratory methods. Unlike porosity, permeability is a more difficult parameter to measure effectively and is usually predicted from porosity-permeability cross-plots (Lucia, 2007). Where there is typically a good correlation between porosity and permeability in siliciclastic sediments due to the dominance of simple interparticle pore types, and spherical and well-sorted grains the heterogeneous pore space in carbonates creates a complex relationship between porosity and permeability.

While many studies have described the relationship between porosity and permeability in carbonates (Archie, 1952; Choquette and Pray, 1970; Hollis et al., 2010; Hulea and Nicholls, 2012; Kopaska-Merkel and Mann, 1993; Lønøy, 2006; Lucia, 1983; Mazzullo and Harris, 1992), two methods are most commonly used when discussing this relationship:

1) The Lucia method (Lucia, 1983, 1995): Here the porosity and permeability relationships for carbonate fabrics are separated into three classes that are based on the interparticle pore distribution (Figure 1.10). The volume of interparticle porosity is controlled by grain size and sorting and by the volume of interparticle cement.
2) The Lønøy method (Lønøy, 2006): Carbonate pore types are classified using the Lucia (1983) and Choquette and Pray (1970) classification schemes and are then subdivided based on whether there is a uniform pore distribution or a patchy pore distribution. The method then relates porosity to permeability for multiple pore types in order to observe statistical relationships. In this method 1 mD is considered the permeability threshold for all pore types which is then related to a critical porosity (Figure 1.11). Critical porosity is used to exclude porosity that is thought to not contribute significantly to flow and is usually defined on porosity-permeability cross-plots by defining a permeability threshold (typically 1 mD). The critical porosity for vuggy carbonates is 6.2 %, whereas for intraparticle porosity the critical porosity is 14.1 %, this suggests that a lower volume of porosity is required for flow in vuggy carbonate rocks. Further, whether the porosity has a uniform or patchy distribution also affects the critical porosity required for flow of 1 mD, with a lower volume of patchy porosity being required than for uniformly distributed porosity (Figure 1.11).
Figure 1.11: Porosity vs. Permeability cross plot for intergranular macropores (Lønøy, 2006). The trend lines that describe the relationship between patchy and uniform intergranular macroporosity distribution and permeability have $R^2$ values of 0.87 and 0.88 respectively. This would suggest that the trend lines can be used to provide a good estimate for permeability when the volume of intergranular porosity is known. The porosity cut-off for intergranular porosity is typically lower than for other pore types, suggesting that a lower volume of this pore type is required in-order to achieve flow $> 1 \text{ mD}$.

1.4 Early Cretaceous Climate

The dominant factors controlling carbonate reservoir quality and the ways porosity and permeability can be classified have now been discussed. Because all three fields assessed in this thesis were deposited during the Early Cretaceous it is now important to present background information regarding the climate during this time, because this will affect the depositional environments and the fauna and flora present.

The Early Cretaceous is commonly considered to be a time of sea-level highstand during a greenhouse period (Föllmi, 2012), although details remain controversial (Föllmi, 2012; Steuber et al., 2005). The highstand and greenhouse
conditions probably relate to the increased $pCO_2$ during the Early Cretaceous that is thought to be due to a 50-70 % increase (Larson, 1991) in ocean crust formation rate between 120 to 80 Million years ago (Mya). This increase in mid-ocean spreading rates could have led to the predicted 1.5-8 times (Royer et al., 2001) or 3-12 times (Berner and Kothavala, 2001) increase in $pCO_2$ compared with modern pre-industrial levels. This increase in greenhouse gases contributed to a predicted $100 \pm 50$ m (Miller et al., 2005) increase in sea-level compared with present day, which led to widespread continental flooding.

However, evidence is beginning to accumulate that suggests the climate during the Early Cretaceous could have been variable. The climate may have oscillated between arid regimes during “normal” greenhouse conditions and humid regimes during intensified greenhouse conditions. Humid regimes were particularly intense during shorter episodes of environmental change (Föllmi, 2012). These humid regimes have been related to numerous cooler periods during the Early Cretaceous, however the timing and significance of these events is typically not well constrained (Steuber et al., 2005).

Recent research has demonstrated the presence of glendonites (a calcite pseudomorph after ikaite which is characteristic of near freezing temperatures) in Early Valanginian sediments in high latitude areas of Canada, Svalbard and Russia. This has been combined with the identification of tillites and dropstones in the Eromanga Basin in Australia (Föllmi, 2012) and with the $\delta^{18}O_{VPDB}$ of high latitude belemnites to postulate the presence of ice caps at the Berriasian-Valanginian boundary (Föllmi, 2012; Price and Mutterlose, 2004; Price, 1999; Price and Nunn, 2010). Other major cooling events have also been identified by Pucéat et al (2003) during the earliest Late Valanginian, and during the earliest Aptian. These cooler periods are supported by the intrashell $\delta^{18}O_{VPDB}$ variability in Cretaceous rudists, that suggest large seasonal variability of up to 18°C near 25°N that is comparable to the range found today and with the existence of polar ice sheets (Steuber et al., 2005).

Early Cretaceous seawater is thought to have been low-Mg calcite producing, with a molar Mg:Ca ratio of 1:1 in comparison with 5:1 for today’s aragonite seas.
Consequently high-Mg calcite and aragonite was dissolved in these waters to form low-Mg calcite cements (Figure 1.12).

![Figure 1.12: The composition of marine water has varied through time and has favoured the precipitation of aragonite and MgSO\(_4\) salts at certain time intervals (areas highlighted in blue), and low-Mg calcite and KCl salts at different time intervals (areas highlighted in red). During the Cretaceous the concentrations of Mg\(^{2+}\) and Ca\(^{2+}\) in seawater are similar leading to the precipitation of low-Mg calcite (Lowenstein et al., 2003).](image)

1.5 Geological Setting of the Studied Fields

1.5.1 Rub Al Khali Basin, U.A.E.

Field A and B are in the Rub Al Khali Basin, U.A.E., this is one of the most prolific petroleum producing basins in the World and is located in the Middle East region (Figure 1.13) which holds ~ 42 % of the Earth’s conventional oil and has ~ 70 % still to produce (Bentley, 2002). Both fields are producing from the Early Cretaceous Thamama Group (Alsharhan, 1990, 1993), the geological setting and the Cretaceous stratigraphy of the fields is briefly described below.

The U.A.E. is located on the Arabian Peninsula, this peninsula has an approximate area of 3,003,200 km\(^2\) (Alsharhan et al., 2001) and is bound to the West and North-West by the Red Sea rift and Gulf of Aden rift, respectively, to the North by the thrust belt of the Alpine Orogeny, to the East and South East by the Zagros...
thrust belt and the Oman Mountains, and to the South by wrench faults associated with the Owen fracture zone (Powers et al., 1966).

There are two main geological provinces on the Arabian Peninsula (Figure 1.13) (Dull et al., 2006; Pinnington, 1981):

- The Arabian Shield: an area of 770,000 km$^2$ (Alsharhan et al., 2001) to the west of the Arabian Platform with a dominant dip direction towards the East. The shield is composed mostly of Precambrian igneous and metamorphic rocks.

- The Arabian Platform: a vast area to the east of the Arabian Shield that has undergone periodic subsidence leading to the accumulation of a thick succession of sedimentary rocks ranging in age from the Cambrian to Recent. The platform includes one of the most prolific petroleum producing basins in the World - the Rub Al Khali Basin (Pollastro, 2003). The Rub Al Khali Basin holds all the major oil fields in the U.A.E. (Figure 1.1, 1.13).
During the Cretaceous, the U.A.E. was located close to the equator and began to drift northwards in conjunction with the Afro-Arabian continent. This caused the climate to change from an arid to a progressively more humid climate. The depositional history of the Middle East during this time was dominated by carbonate sedimentation, on a very broad, stable and extensive platform, that was controlled by marine transgressive cycles (Dickson et al., 2008; Murris, 1980; Pinnington, 1981).

The Thamama Group is a regional term for the Early Cretaceous carbonate succession present within the U.A.E (Figure 1.14), and can be up to 1075 m thick (Scott, 1990). The Group is thought to have deposited on a carbonate ramp (Ahr, 1973; Alsharhan and Nairn, 1997; Murris, 1980), and is characterised by numerous shallowing upward cycles (Alsharhan, 1990; Oswald et al., 1995). Each cycle can be correlated over large distances, in the order of 500 kilometres (Alsharhan, 1989; Alsharhan and Nairn, 1997), parallel with the depositional strike as well as perpendicular to it (Murris, 1980). The individual cycles are constant in thickness.
and lithology with any changes occurring very gradually, this suggests that the cycles were deposited on a near horizontal platform (Alsharhan and Kendall, 1991; Strohmenger et al., 2006).

Figure 1.14: Chronostratigraphy, magneto-stratigraphy and eustatic sea-level during deposition of the Thamama Group. The four formations of the Thamama Group were deposited during two supersequences which correspond to changes in 2nd order sea-level. During the first supersequence the Habshan Formation was deposited and during the second supersequence the Lekhwair, Kharaiib and Shu’aiba’ Formations were deposited. The disconformable contact, between the two supersequences, is thought to be related to far field stresses induced by the opening of the South Atlantic (Al-Fares et al., 1998). Modified from Vahrenkamp et al., (1996) and Alsharhan (1990).
The Group is composed of two second-order supersequences. The older sequence corresponds to the Habshan Formation (Berriasian to Early Valanginian), while the younger sequence encompasses the formations from the top Habshan to the top Shu’aiba’ and includes the Lekhwair, Kharaib and Shu’aiba’ Formations. The formations are bound by unconformities and were identified by sub-surface data that can be correlated throughout the entire southeastern Arabian Gulf (Alsharhan, 1990; Granier, 2000; Haq and Al-Qahtani, 2005; Scott, 1990; Sharland et al., 2001; Strohmenger et al., 2006).

The Habshan Formation was deposited between 144-140 Mya (Davies et al., 2000; Simmons et al., 2007; Ziegler, 2001) and corresponds to the first Early Cretaceous 2nd order depositional sequence of Sharland et al (2001) and Strohmenger et al (2006). During deposition the Western Arabian plate was undergoing uplift while the Eastern Arabian plate was subsiding, this led to the titling of the Arabian Plate to the East and caused the development of an extensive shallow marine shelf (Figure 1.15a) (Ziegler, 2001). The Habshan Formation was deposited on this shelf in a restricted to semi-restricted environment.

The disconformable contact that separates the older and younger second order super-sequence has been dated based on calcareous algae and foraminifera (Sharland et al., 2001) and is located at the Habshan-Lekhwair contact (~140 Mya) (Simmons et al., 2007). This contact has been related to far field stresses induced by the opening of the South Atlantic (Al-Fares et al., 1998). Deposition of the Lekhwair and Kharaib Formations began due to a progressive rise in sea-level during the Hauterivian-Barremian (Figure 1.14) (Haq et al., 1988).

The Lekhwair and Kharaib Formations were deposited between 140-125 Mya and represent shoaling upward cycles (Lambert et al., 2006; Scott, 1990; Simmons et al., 2007; Vahrenkamp, 2010). Deposition occurred during a period of relatively low sea level, in normal unrestricted marine-shelf conditions, and on an extensive carbonate ramp behind the shelf rim of the Habshan (Figure 1.15b) (Granier et al., 2003; Scott, 1990). The formations are composed of a number of cycles, with each cycle starting with variably argillaceous limestone (mostly lime-mudstones and wackestones) that grade into wackestones/packstones with some grainstones...
(Dickson et al., 2008; Van Buchem et al., 2002a). The argillaceous limestone represents the transgressive phase of an open shelf sub-tidal environment whereas the more porous wackestones/packstone and sub-ordinate grainstones represent the highstand phase of the cycle. In general, the transgressive phase formed the dominant part of the cycles, however the Kharaib Formation exhibits a greater thickness of the highstand deposits than the Lekhwair that is most likely due to the progressively shallowing conditions of the Thamama Group (Alsharhan and Kendall, 1991; Haq et al., 1988). A sedimentary hiatus that is of minor significance terminated the deposition of the Lekhwair Formation (Granier et al., 2003).
Figure 1.15: The Arabian Peninsula during the deposition of a) the Habshan (Yamama), and b) Lekhwair and Kharai Formations. After Ziegler (2001).
The transgressive phase of the Early Aptian Shu’aiba’ Formation terminated the deposition of the Kharai Formation. The Shu’aiba’ Formation was deposited between 125-113 Mya during a rise in sea-level (Haq et al., 1988; Vahrenkamp, 2010). The formation shows a general shoaling from deepwater limestones to rudist-rimmed platforms, to back bank, and then lagoonal deposits (Hughes, 1998; Murris, 1980; Ziegler, 2001), and has a similar cyclical deposition pattern as the Lekhwair and Kharai Formations. A cycle is typically composed of underlying shales and wackestones overlain by shelf carbonates. The intrashelf basin mudstones and wackestones of the Bab Member (Aldabal and Al sharhan, 1989; Azzam and Taher, 1995), completes the deposition of the Shu’aiba’ Formation and the Thamama Group.

Each cycle of the Lekhwair, Kharai and Shu’aiba’ Formations has a highly porous reservoir interval, that is typically within the Highstand Systems Tract (HST) of each cycle and an underlying low porosity non-reservoir interval that is within the Transgressive Systems Tract (TST) of the cycle (Figure 1.16) (Alsharhan, 1990; Strohmenger et al., 2006). The intervals are laterally continuous and are observed in all subsurface fields and outcrop analogues in the U.A.E., allowing for regional correlation (Strohmenger et al., 2006).

The regional disconformity of the Early-Albian separates the Shu’aiba’ and Nahr Umr Formations and has been dated at 113 Mya (Murriss, 1980; Vahrenkamp, 2010). This break probably coincides with the worldwide low stand of sea-level (Haq et al., 1988) that terminated the deposition of the Thamama Group and is the result of far-field stress associated with the opening of the central Atlantic Ocean (Al-Fares et al., 1998).
Figure 1.16: Typical well showing porosity and permeability data in relation to a sequence stratigraphic framework for the Lower Kharaiib, Upper Kharaiib and the Lower Shu’aiba’ reservoir units. Figure from Strohmenger et al. (2006). Note: In the Bab Basin the top of the Shu’aiba’ Formation is not thought to have been exposed; however, elsewhere in the U.A.E. it is likely that exposure has occurred (see Figure 2.6).

This lowstand was then followed by a gradually rising sea level that culminated in maximum flooding by end Albian. This gradual rise in sea-level after the pre-Albian disconformity led to the deposition of the Nahr Umr Formation that is dominated by shale with minor lenses of glauconitic silt and carbonate. With the termination of Nahr Umr Formation, the influx of terrigeneous sediments ceased and a transgressive phase resulted in the deposition of normal marine, clear water carbonate sediments of the Mauddud Member (Salabikh Formation). Continued
subsidence resulted in the deposition of the Shilaif member (Salabikh Formation) that mostly consists of fine-grained limestones and the Mishrif Formation that consists of a complex sequence of carbonates. The Wasia Group comprises all the Middle Cretaceous deposits (Nahr Umr, Salabikh and Mishrif Formations). Middle Cretaceous deposition was terminated by a major unconformity in the Turonian that has led to the complete erosion of the Wasia Group over the south-east of Abu Dhabi (Pinnington, 1981).

During the Late Cretaceous renewed subsidence after the Turonian led to incursion of the Tethyan Ocean and to the deposition of the Aruma Group. This Group consists of two major transgressive-regressive cycles, with the first cycle constituting the Laffan and Halul Formations, while the second represents the Fiqa and Simsima Formations (Haq et al., 1988). The Laffan Formation is composed mainly of shales, whereas the conformably overlying Halul Formation was deposited during a gradual regression which led to the development of shallow clear-water carbonate sediments (Beydoun, 1991; Haq and Al-Qahtani, 2005; Pinnington, 1981). The Fiqa Formation was deposited during the transgressive phase overlying the Halul Formation and consists of soft marl, calcareous shales and argillaceous limestones. The Simsima Formation comprises packstones and wackestones deposited during the second regressive cycle.

1.5.2 Campos Basin, Brazil

Field C is from the Campos Basin that is located offshore from Rio de Janeiro on the East coast of Brazil (Figure 1.2), South America and covers an area of ~100,000 km², this basin is separated from the Espirito Santo Basin to the North by the Vitoria High, and from the Santos Basin in the South by the Cabo Frio High (Guardado et al., 2000; Guardado et al., 1990). The basin is a prolific oil producing basin and accounts for > 65 % of the total oil production in Brazil (Guardado et al., 1990). There are numerous oil fields discovered in this basin that produce from multiple stratigraphic intervals, the field studied in this thesis is producing from the Early Cretaceous Quissamã Formation (Macaé Group) (Figure 1.17).
Figure 1.17: Chrono-stratigraphic column for the Campos Basin. The tectono-stratigraphic framework of the Campos Basin can be divided into three phases: the rift, transitional and marine phases. The Quissamã Formation is highlighted by the arrow and underlined. Modified figure courtesy of Petrobras.

There is similarity in tectono-stratigraphic evolution of several basins along the South American coastline which points to a common mechanism of basin formation (Mohriak et al., 1990). These basins have evolved from the Early
Cretaceous to present day as rift basins that are part of the breakup of Gondwana. All these basins have several important structural elements, including 1) NE and NW trending horsts and grabens; 2) pre-Aptian structures related to the NE-SW basin bounding Campos fault; and 3) a salt dome province (Guardado et al., 1990; Mohriak et al., 1990).

A general depositional model for the Macaé Group is of a carbonate ramp that slopes towards the east and northeast (Figure 1.18) (Guardado et al., 1990). During deposition eastward basin tilting together with differential compaction triggered salt movement and led to the development of growth fault structures that sole out on the underlying Aptian salt. The growth faults that became active during the Albian persisted until the Holocene, and have played a decisive role in controlling sedimentary facies and trap formation for major hydrocarbon accumulations in the Campos Basin (Guardado et al., 1990) (Mohriak et al., 1990).

![Ramp type depositional model](image)

*Figure 1.18: Depositional environment of the Macaé Group. Modified from Guardado et al. (1990).*

The Quissamã Formation (Lower Macaé Group) represents the Albian shallow water carbonate platform system, with oncolitic calcarenites forming offshore shoals. The carbonate platforms of the Lower Macaé Group underwent vertical aggradation due to the continued subsidence of the basin, this was caused by crustal thermal contraction and continued sea-level rise (Guardado et al., 1990). Each cycle of the aggrading Lower Macaé Group is composed of oncolitic packstones at
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the base that grade upward into peloidal packstone and oolitic grainstones at the top. The packstones are poorly sorted, typically bioturbated and include a large number of pelletoids (Guardado et al., 1990). The grainstones are massive, locally displaying cross-stratification and are mainly composed of sub-spherical oncolites (Guardado et al., 1990).

The deposits of the Albian-Cenomanian Macaé Group formed in an anoxic restricted marine environment (Figure 1.19) (Beglinger et al., 2012b; Guardado et al., 1990). A palaeoecological study of the neritic carbonate sequence of the Lower Macaé Group suggests that the sediments were deposited under warm, dry climatic conditions. The sea was probably hypersaline and generally less than 200 m deep. During the Late Cretaceous water depths increased progressively in response to subsidence and eustatic sea-level rise (Chang et al., 1992; Guardado et al., 1990) this led to the drowning of the Lower Macaé carbonate platform (Chang and Kowsmann, 1987; Schlager, 1981).

![Figure 1.19: The depositional environment of the South Atlantic margin during the Late Mesozoic. Until the Campanian-Maastrichtian the depositional environment of the Campos Basin is characterised by an anoxic, hypersaline restricted depositional environment. After Beglinger et al. (2012a) original data from Brownfield and Charpentier (2006).](image)

The Cenomanian Upper Macaé Group is composed of deeper water calcilutites and marls (Moraes, 1989). The turbidite Namorado Sandstone sequence
of the Upper Macaé Group forms regionally extensive blanket sands on a flat sea floor that can be correlated over wide areas. The abundance of turbidites in the basin suggests that the sands entered the basin from several sources as opposed to a single major feeder (Guardado et al., 1990).

1.6 Samples and Datasets

1.6.1 Rub Al Khali Basin, U.A.E.

1.6.1.1 Field A

Field A is a giant anticline with estimated reserves of 10.5 billion barrels (Mallinson and Sharp, 1975) and an ultimate recovery factor of 34 % (Alsharhan, 1993). In 1974 460,000 Barrels of Oil Per Day (BOPD) were produced from 22 wells from the Thamama Group (Mallinson and Sharp, 1975). The field is 26 km long and 9 km wide (Alsharhan, 1993).

Well 1A and 3A were sampled by the author, with Well 2A being sampled by Prof R. Wood on a previous visit. The samples obtained for the three wells are presented in Table 1.1. Samples were obtained from every HST and TST for each cycle that most appropriately describes the reservoir and non-reservoir intervals identified (Figure 1.20). Thin section rounds with a width of 2.5 cm and a height of 2.5 cm were made for all plug samples obtained. Cycles v to i are oil bearing with each cycle thought to have a separate Oil Water Contact (OWC), cycle vii is gas bearing. Plug porosity and permeability data are only available for cycles iii-i of Well 1A. There are no wireline logs available for the field.
<table>
<thead>
<tr>
<th>Well</th>
<th>Well deviation (degrees)</th>
<th>Cored interval (m)</th>
<th>Sampled thickness (m)</th>
<th>Number of plug samples obtained from the cycle</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>S</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>i</td>
</tr>
<tr>
<td>1A</td>
<td>Near vertical</td>
<td>2501-2376</td>
<td>125</td>
<td>4</td>
</tr>
<tr>
<td>2A</td>
<td>Near vertical</td>
<td>2802-2414</td>
<td>388</td>
<td>n/a</td>
</tr>
<tr>
<td>3A</td>
<td>Near vertical</td>
<td>2628-2589</td>
<td>39</td>
<td>3</td>
</tr>
</tbody>
</table>

Table 1.1: Plug samples obtained for the three wells of Field A, all samples are taken parallel to bedding. The cored interval that was available for sampling and the sampled thickness of each well are also presented along with the well deviation. Key: S – Shu’aiba’ Formation, K – Kharaih Formation, L – Lekhwair Formation.
1.6.1.2 Field B

Field B has reserves of 12.25 billion barrels and an ultimate recovery of 35% (Hassan and Wada, 1981). In 1963 50,000 BOPD were produced from the Thamama Group, the field has a length of 40 km and is 24 km wide (Aldabal and Alsharhan, 1989).

All Field B samples were obtained by Prof R.Wood, Dr J.A.D. Dickson and Dr. P. Cox. 5 wells were sampled, with Wells 1B-4B in the crest of the field, and Well 5B in the flank (Figure 1.21). Samples were obtained from the HST and the

Figure 1.20: Field A well location map. The grey segment marks when the HST was sampled, a black internal band on each segment marks that the underlying TST was sampled. The TST of cycle vii in Well 2A has not been sampled.
TST for all cycles observed in the core (Figure 1.21). The samples obtained for the five wells are presented in Table 1.2. Thin sections with a height of 4.5 cm and a width of 7.5 cm were available for each plug sample. Plug derived porosity and permeability data are available for all 5 wells. There are no wireline logs available for the field. The location of the OWC in Field B is unknown however the field is known to be producing from the Lekhwair, Kharaiab and Shu’aiba’ Formations, with 69 % of production being from the Zakum Member in the lower Lekhwair Formation (Hassan and Wada, 1981); no samples were obtained from this member. Reservoir intervals v-i are oil stained, therefore these cycles are taken to be oil bearing and the older cycles (ix-vi) taken to be water bearing (aquifers).

<table>
<thead>
<tr>
<th>Well</th>
<th>Well deviation (degrees)</th>
<th>Depth range (m)</th>
<th>Sampled thickness (m)</th>
<th>Number of plug samples obtained from cycle</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>S</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>i</td>
</tr>
<tr>
<td>1B</td>
<td>42</td>
<td>3636-3465</td>
<td>171</td>
<td>3</td>
</tr>
<tr>
<td>2B</td>
<td>Near vertical</td>
<td>2556-2407</td>
<td>149</td>
<td>9</td>
</tr>
<tr>
<td>3B</td>
<td>46</td>
<td>3424-3342</td>
<td>82</td>
<td>n/a</td>
</tr>
<tr>
<td>4B</td>
<td>44</td>
<td>3416-3317</td>
<td>99</td>
<td>n/a</td>
</tr>
<tr>
<td>5B</td>
<td>35</td>
<td>3213-3189</td>
<td>24</td>
<td>10</td>
</tr>
</tbody>
</table>

Table 1.2: Plug samples obtained for the five wells in the field, all samples obtained are bedding parallel. The depth and thickness of the sampled interval are shown along with the well deviation. Key: S – Shu’aiba’ Formation, K – Kharaiab Formation, L – Lekhwair Formation.
1.6.1.3 Outcrop

The outcrop equivalent to the Kharaib and Shu’aiba’ Formation is exposed at Wadi Rahabah (Figure 1.1). The reservoir intervals and non-reservoir intervals were defined by Strohmenger et al., (2006) and related to a sequence stratigraphic framework (Figure 1.16). All the surfaces that are exposed at Wadi Rahabah were sampled, with two samples obtained from the surfaces of the K60_SB and K70_SB and four samples from the K50_MFS, K60_FS100, K60_MFS and K60_FS1000 flooding surfaces (Figure 1.16).

1.6.2 Campos Basin, Brazil

1.6.2.1 Field C

Field C is located in the Campos Basin at water depths of around 105 m, and has a length of 4 km and a width of 5 km (Tigre et al., 1983). The field is producing
from both the Quissamã Formation of the Macaé Group, and the Coqueiros Formation of the Lagoa Feia Group (Figure 1.17). This thesis only assesses the Quissamã Formation due to the marked lower recovery of 10.1%, compared with 32.1% in the Coqueiros Formation. This is because the thesis aims to understand the controls on reservoir quality and therefore the results from this project can be used to more accurately predict the distribution of subsurface porosity and permeability which will help to improve recovery. The reserves within the Quissamã reservoir are 385 million barrels. There is a single OWC within the Quissamã reservoir.

Eight cycles are present in crestal Well 1C (Table 1.3, Figure 1.22); sampling was undertaken by Dr C. Van der Land in UNESP. 17 thin sections were made from these plug samples that have a width of 2 cm and a height of 4 cm (Table 1.3). A further 8 thin sections with a width of 1.5 cm and a height of 3 cm were provided by Petrobras from crestal Well 2C (Table 1.3, Figure 1.22).

Plug porosity and permeability data are available for both wells, and three additional wells. Wireline log data are available for both wells and include: neutron porosity, density, resistivity and gamma ray logs. Mud logging data are also available for 17 wells.

<table>
<thead>
<tr>
<th>Well</th>
<th>Well deviation (degrees)</th>
<th>Sampled depth range (m)</th>
<th>Sampled thickness (m)</th>
<th>Number of plug samples obtained from cycle</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>1</td>
</tr>
<tr>
<td>1C</td>
<td>Near vertical</td>
<td>1855-1767</td>
<td>88</td>
<td>3</td>
</tr>
<tr>
<td>2C</td>
<td>Near vertical</td>
<td>1817-1769</td>
<td>48</td>
<td>2</td>
</tr>
</tbody>
</table>

Table 1.3: Plug samples obtained from Field C, all samples are taken parallel to bedding. The thickness of the sampled interval and the well deviation are also shown.
1.7 Methods

The methods used in this thesis are presented in this section.

1.7.1 Thin section staining

The thin sections used in this thesis were stained with potassium ferricyanide and alizarin red S to help identify ferroan and non-ferroan calcite and dolomite cement (Dickson 1965, 1966). This aids with determination of the chronology of diagenetic phases, and when assessing if ferrous iron is present.
1.7.2 Quantification of cement and pore volume

Cements and porosity were quantified based on thin section point count analyses (Dutton, 2008; Howarth, 1998; Van der Plas and Tobi, 1965), using the Prior ES11LB motorised stage. A random sample distribution was used for a specified Region Of Interest (ROI) and 350 counts made for each thin section. When designating the ROI the entire thin sectional area was used. The diagenetic cements present in all available thin sections were quantified to assess the impact of cementation on porosity occlusion.

A Python™ code was written to help determine the volume of blue epoxy resin in a thin section. However, due to the poor impregnation of the thin sections with epoxy resin the code could not be used to quantify total porosity, but has been included in Appendix 1 to aid future studies with quantifying total porosity.

1.7.3 Cathodoluminescence assessment

A Cathodoluminescence Cold cathode CITL 8200 MK3A was used to identify differences in luminescence between Cathodoluminescent (CL) zones. CL zones represent changes in the pore water chemistry during cement growth (Braithwaite, 1993; Machel, 1985), and can be used to establish cement fabric changes and micro-cement stratigraphies.

The luminescence of calcite and dolomite is predominantly controlled by the Mn$^{2+}$ concentration, although some REE ions may also be activators (Boggs, 2010), with 15-30 ppm and 30-35 ppm being the minimum concentration required in order for the cement to luminesce (ten Havet and Heijnen, 1985). Fe$^{2+}$ is thought to be the main quencher of luminescence (Boggs, 2010; Machel, 1985; Mason and Mariano, 1990).

1.7.4 Oil inclusions

Inclusion petrography is a key factor influencing the interpretation of pore fluid evolution. Primary inclusions are formed during cement growth, provided no alteration has taken place, and they can be used to help determine the pore fluid from which the cement formed. Secondary inclusions are formed after crystal growth, and
entrapped along microfractures healed during burial. Their characteristics can be used to document the evolution of reservoir conditions during burial, but are no longer relevant to the timing and conditions of cement growth.

Oil inclusions were observed with a Leitz Metallux Leica 3 microscope, with a 100w Hg HBO 103 W/2 bulb and a blue filter creating UV light with a waveband of 440-490 nm (Burruss, 1991). All counted inclusions were assessed to be primary and contained a non-luminescent portion reflecting the presence of a vapour bubble and fluoresced yellow-green to pale blue. Any string of aligned oil inclusions along a straight line may also have been formed along a fracture, fault and cleavage plane; these inclusions were disregarded.

1.7.5 Scanning Electron Microscope (SEM)

Rock chips were broken from the same core samples that the thin sections were made from, and cleaned in a sonic bath for 5 minutes and then gold coated. These stubs were placed in a SEM (Phillips XL30CP) in secondary electron mode to obtain high magnification images of five cutting samples, allowing for a detailed understanding of cement morphology and chronology. A 20Kv, 6 μm diameter electron beam was used to image at 120× magnification and reduced to 4 μm in diameter at magnifications above 1950×.

Electron BackScatter Diffraction (EBSD) is an SEM based technique used to assess the diffraction patterns for a crystalline sample. The technique is a powerful tool for assessing the orientation of individual grains, grain textures and phase identification (Schwartz et al., 2000). EBSD measurements were made every 5 μm with the diffraction patterns being imaged on a phosphor screen and analysed by the HKL Channel5 software, the resolution of the technique is ~ 1 μm.

1.7.6 Separate component, stable isotope assessment

Separate component analysis allows the largest, most accessible carbonate components to be sampled to provide an average δ¹⁸OVPDB and δ¹³C value for the cement. Separate component analyses of the main pore occluding cements were undertaken by removing individual carbonate fractions from thin sections with a steel
needle under a Leica 240 microscope. Micrite powder was obtained from hand specimens using a microdrill. The samples were prepared for isotopic analysis by dissolving the carbonate at 25°C with 100 % phosphoric acid this was followed by conventional mass spectrometry using a Thermo Delta+ Advantage. The $\delta^{18}\text{O}_{\text{VPDB}}$ and $\delta^{13}\text{C}$ of this powder are reported relative to a VPDB standard and precision was measured as better than 0.1 ‰ for $\delta^{13}\text{C}$ and $\delta^{18}\text{O}_{\text{VPDB}}$. The $\delta^{13}\text{C}$ and $\delta^{18}\text{O}_{\text{VPDB}}$ values can be used to assess the pore fluid chemistry and temperature of the fluid from which the cement precipitated.

1.7.7 In-situ $\delta^{18}\text{O}_{\text{VPDB}}$ assessment

Separate component analysis provides average $\delta^{18}\text{O}_{\text{VPDB}}$ and $\delta^{13}\text{C}$ value that can incorporate multiple cement zones. However detailed $\delta^{18}\text{O}_{\text{VPDB}}$ transects that assess individual cement zones can be made using the Secondary Ion Mass Spectrometer (SIMS), this procedure was discussed by Feazel and Schatzinger (1985) and applied by Heydari and Moore (1993), Lyon et al (2000) and Cox et al. (2010).

The samples were gold coated and placed within the Cameca 1270 ion microprobe at the University of Edinburgh to gain in-situ $\delta^{18}\text{O}_{\text{VPDB}}$ measurements for individual cement phases. A Cs ion beam was used to ablate 10-15 µm diameter spots from the sample. The internal precision for each spot ranges between 0.009 % and 0.015 % (Standard Error). The external precision was estimated to be 0.3 ‰ as determined by consecutive analysis of a UWC (University of Wisconsin Calcite) standard that was assessed to be homogenous.

The sample holder used for the ion microprobe can accommodate two samples at a time. The typical procedure is to include the round to be assessed along with another round that includes the standard. A sample change can take ~6 hours and therefore only 1 round can be assessed per day. This meant that in the initial week allocated on the ion microprobe only 5 samples could be assessed, this was not suitable for undertaking a detailed diagenetic assessment on multiple reservoir intervals. Therefore, for this assessment the UWC standard was included in the round to be assessed, allowing for twice as many samples to be assessed in the time given.
A new procedure was developed to include the standard after cathodoluminescence and oil inclusion analysis; this procedure is described here. The microdrill is used to drill through the round making sure that no analysis locations in the round are affected, the hole needs to be ± 0.5 cm from the centre. The surface of the round is then covered by one layer of adhesive tape and a knife is used to cut a hole into the adhesive tape to uncover the hole drilled by the microdrill. Another layer of tape was then added onto the previous layer. Making sure a large crystal plane rests on the tape, a standard was placed into the drilled hole. The hole is then filled with epoxy and left to dry. Both tape layers are then removed and the standard polished to the same plane as the thin section surface.

1.7.8 Electron Probe Microanalysis (EPM)

The EPM technique has been used to obtain the elemental concentrations of Ca, Na, Mg, Al, Si, K, P, Sr, Ti, Mn, Fe, O present in the calcite and dolomite cements, making sure that the complete micro-cement stratigraphy was evaluated. The trace element composition of carbonate minerals gives an indication of the type of fluid present at the time of crystallisation (Boles et al., 2004) and can aid analysis of the palaeo-fluid flow (Cao et al., 2010). Trace element analyses can also be used to determine the origin of the pore water, with high-Mg commonly being observed in marine water, whereas calcite forming from meteoric water typically has high Mn relative to Mg and Fe (Boles, 1998; Morad, 1998).

Samples were coated with a thin layer of carbon and placed in the electron microprobe (Cameca SX100) located at the University of Edinburgh. The EPM facility uses an incident electron beam that has a spatial resolution as high as 1 μm to excite electrons within the sample. Each element releases a distinct photon which can be used to determine the elements present and their concentration. Routine instrument calibration was carried out by analysing reference materials with known compositions. Routine detection limits are in the order of 0.009 to 0.05 wt %, results below the detection limit are given null values; the concentrations of P^{5+}, Ti^{4+} and Mn^{2+} are all below the detection limit and will not be discussed further.
In order to understand how the basin has developed, the PetroMod\textsuperscript{TM} basin and petroleum systems modelling software has been used. During this project multiple versions of Petromod\textsuperscript{TM} have been used due to the software undergoing continued improvement since 2010. The results presented in this thesis are solely from Petromod\textsuperscript{TM} version 12 (2012 edition). PetroMod\textsuperscript{TM} typically uses a variety of inputs (Figure 1.23) including the present day geometry with present and/or historical thermal states and assigned lithology parameters to calculate a 1D/2D/3D set of states including porosity, temperature and pore pressure with time. It is typically used to calculate hydrocarbon maturation, migration, and trapping and therefore can be used to reduce the risk of a prospective structure before drilling.

Once a basin model’s ability to adequately represent the geological system has been assessed against independent parameters such as vitrinite reflectance, maturity and temperature and pressure data (Figure 1.23), it can then be used to assist in our assessment of the basin. The model produces plots that can be assessed for geological viability, and migration pathways to be compared against accumulations. This makes the software a very powerful tool when trying to visualise how the reservoir has developed and can be used to relate diagenetic events to basin scale processes.
Figure 1.23: Diagram illustrating the typical forward modelling approach of basin modelling. A number of inputs are incorporated into the model (shown in blue), the model is then simulated and the output (shown in yellow) calibrated to observed data sets and, if required, the inputs are refined and the model re-simulated. After Al-Hajeri et al. (2009).
1.8 Thesis Outline

Chapter 2 outlines new observations on the depositional setting of the Lekhwair, Kharaib and Shu’aiba’ Formations from both Field A and Field B and from outcrop analyses, with the primary aim of understanding how the depositional environment affects current reservoir quality. The depositional cycles which best represent the reservoir intervals, are described and the depositional environments discussed. The cycles identified are then placed in a chronostratigraphic framework through the identification of 3rd order well constrained sequence boundaries and maximum flooding surfaces (Sharland et al., 2004) and through biostratigraphic analyses. Bulk micrite stable isotope assessments (δ^{13}C) on all plug samples are then used to determine the pore fluid chemistry of Early Cretaceous marine seawater, and are also used to aid with the chronostratigraphical interpretation by relating the δ^{13}C data to well constrained marine secular curves. The lithofacies and fabrics identified in these chronostratigraphically constrained depositional cycles, are then related to plug derived porosity and permeability data to determine if there is a relationship between depositional setting and final reservoir quality.

Chapter 3 identifies the diagenetic constituents present in Field A and B and determines their relative chronology. The main pore occluding cements and pore types present are then quantified to determine the effect of diagenetic processes on modifying porosity and are related to lithofacies to determine the control of lithofacies on subsequent diagenetic processes. Bulk cement stable isotope data (δ^{18}O_{VPDB} and δ^{13}C) for the main pore occluding cements are then used to help determine their origin, to constrain their relative time of formation and to help understand the evolution of pore fluid chemistry during cement formation.

Chapter 4 assesses the dynamics of cementation for the Field A and B. A detailed elemental assessment is undertaken on the main pore occluding cement type, low-Mg calcite, in Field A and B to help determine the origin of the cement phases and to identify any changes in pore fluid chemistry during cement formation; specific attention is paid to the nMg/mCa ratio because it is thought to be a good temperature proxy. The micro-cement stratigraphy of low-Mg calcite is then determined by a cathodoluminescence assessment. The distribution of oil inclusions is then related to
this micro-cement stratigraphy to help understand the effect that oil charge has on the
dynamics of cementation. The micro-cement stratigraphy is then assessed by ion
microprobe analyses to obtain in-situ $\delta^{18}O_{VPDB}$ values that can be used as a
temperature/burial depth proxy. This assessment is undertaken on coeval reservoir
intervals in the crest and flank of Field A to determine any lateral variability in the
dynamics of cementation and the causes for any variability. A stratigraphic
assessment is also undertaken on the stacked reservoir intervals in both Field A and
B to determine any stratigraphic variability in the dynamics of cementation. These
data will be related to the cement volumes and pore volumes quantified in the
previous chapter to determine how any lateral and stratigraphic differences in pore
and cement distribution relate to the dynamics of cementation.

Chapter 5 aims to understand the burial, structural and thermal history of the Rub Al
Khali Basin, and to input an understanding of diagenesis into the basin models for
Field A. Three models were run with different porosity-depth curves for the reservoir
and non-reservoir intervals of the nine cycles assessed in this thesis, the first uses the
porosity-depth curve of Schmoker and Halley (1982), the second used the porosity-
depth curves defined from the PetromodTM database and the third attempts to include
an estimation of diagenesis. The maturity of the source rocks in the basin does not
correlate with the observed maturity if the heat flux of the basin is described only by
the McKenzie (1978) stretching model, therefore two heat fluxes have been
simulated for each of the three models where periods of higher heat flux are related
to corresponding tectonic events. All the models are then calibrated and the outputs
used to help determine the origin of the diagenetic features observed within the
reservoirs. The predicted thermal histories for the reservoirs are also extracted from
the basin models and compared to the $\delta^{18}O_{VPDB}$ and $^{24}Mg/^{44}Ca$ temperature proxies to
help constrain the time of calcite cement precipitation. The known accumulation sites
of hydrocarbon are also compared to the predicted accumulation locations suggested
by the three models to determine which model most accurately predicts the location
of hydrocarbon.

Chapter 6 considers the development of reservoir quality in Field C. First, the
depositional setting was determined through seismic line interpretation and core and
petrographic analyses. The depositional fabrics identified were then related to plug derived porosity and permeability to determine if there is a relationship between depositional setting and reservoir quality. The porosity and permeability of the oil reservoir and aquifer are compared, in the same depositional fabric, to determine the effect of oil charge on reservoir quality. A petrographic assessment was then undertaken to identify the diagenetic processes that have affected reservoir quality and to determine their relative chronology. An elemental assessment was then undertaken on the calcite cements in the field to help identify how the pore fluid chemistry has changed during cement precipitation. The micro-cement stratigraphy of the main pore occluding cements was determined by cathodoluminescence and a fluid inclusion analysis undertaken, and related to this stratigraphy, to help understand the effect that oil charge has on cementation. Ion microprobe in-situ $\delta^{18}O_{VPDB}$ data for the complete micro-cement stratigraphy were then obtained, and combined with the elemental assessment to understand the dynamics of cementation for the field. To constrain the geological evolution of the basin and to help determine the time of oil charge into Field C two basin models were then developed. The first model was developed from a field scale seismic line whereas the second model uses a published interpretation of a basin scale seismic line. The time-temperature evolution for the reservoir interval is then extracted from the models, and compared to the $\delta^{18}O_{VPDB}$ and $^{26}Mg/^{44}Ca$ temperature/burial depth proxies to help constrain the time of cement precipitation.

Chapter 7 synthesises how the data presented in this thesis have improved our understanding of the controls on porosity evolution and the origin of diagenetic features in the three oil fields. The chapter also discusses future areas of research that will help to further improve our understanding of the dynamics of cementation and the ways in which basin models can be improved.
CHAPTER 2
DEPOSITIONAL CONTROLS ON RESERVOIR QUALITY IN THE LEKHWAIR, KHARAIB AND SHU’AIBA’ FORMATIONS

2.1 Introduction

The depositional environment controls the initial petrophysical properties (porosity and permeability) and the porosity network (diameter, shape, and tortuosity) of sedimentary rocks. These properties can be subsequently altered by diagenesis; this is especially true for carbonate sediments due to their highly reactive nature. This chapter aims to understand the control that depositional environment has on the present day petrophysical properties of the Lekhwair, Kharaib and Shu’aiba’ Formations.

Lithofacies are not deposited randomly but occur in a sequence stratigraphic framework. This framework relates lithological variability to changes in relative sea-level in order to identify depositional cycles. A depositional cycle is divided up by key surfaces: Sequence Boundaries (SBs), Maximum Flooding Surfaces (MFSs), which are used to internally divide a cycle into genetically related stratal packages (systems tracts).

Systems tracts can then be used for correlation purposes and can also be related to the location of reservoir and non-reservoir intervals to determine the control of depositional environment on reservoir location. If a good relationship is observed between depositional environment and reservoir location a predictive model, based on relative sea level changes, can be produced.
The established sequence stratigraphic framework can then be chronostratigraphically constrained by relating the 3rd order MFSs identified to chronostratigraphically constrained 3rd order MFSs (Sharland et al., 2004). Furthermore, biostratigraphic and stable isotope ($\delta^{13}C$) analyses, where the stable isotope data of interest is placed in context with established global trends, can be used to further constrain the time of deposition for each depositional cycle. The established time of deposition can then be included in basin models, which will allow the subsidence history of the basin to be more accurately modelled.

The first aim of this chapter is to place the reservoir intervals considered in this thesis into a sequence stratigraphic framework. A new sequence stratigraphic framework for Well 1A and 3A of Field A and for the outcrop is presented in this study. This framework is then related to the previously established framework of Field B (Cox, 2010) and of Well 2A (Prof. R. Wood; unpublished/confidential dataset) of Field A, so that the controls on reservoir quality can be compared. Biostratigraphic and chronostratigraphic constraints are then applied to this sequence stratigraphic framework to help constrain the time of deposition for the cycles. The location of the reservoir and non-reservoir intervals are then related to this chronostratigraphic framework to determine whether there is a good relationship between relative sea level change and reservoir quality.

2.2 Previous Work

2.2.1 Sequence stratigraphy and depositional setting

The Thamama Group is composed of four formations: the Habshan, Lekhwair, Kharaib and Shu’aiba’ Formations (Figure 1.14). The formations were deposited during two second order supersequences (Sharland et al., 2001) separated by the disconformable contact at the Habshan-Lekhwair contact; the younger supersequence is assessed in this thesis (Section 1.5.1).

The Lekhwair, Kharaib and Shu’aiba’ Formations are composed of a number of depositional cycles, with extensive research been previously undertaken to understand their depositional setting. This section presents a summary of this work for each formation.
2.2.1.1 Lekhwair Formation

Relative sea-level curves describing the period of deposition for the Lekhwair Formation are quite variable (Figure 2.1), this is because they are typically made from a relatively limited dataset. Unlike these previous studies, Sharland et al. (2001) have recently published a a 3rd order regionally correlatable Arabian plate sequence stratigraphic framework which has attempted to integrate a vast quantity of Phanerozoic data not only from the Arabian Platform but also from surrounding basins (Section 2.2.2); this framework has been well accepted (Haq and Al-Qahtani, 2005; Van Buchem et al., 2010; Ziegler, 2001). Haq and Al-Qahtani (2005) related this framework to eustatic sea-level patterns and found that during the Mesozoic and early Palaeogene the sedimentary cycles of the Arabian Platform show good agreement with eustatic patterns. This suggests that during the deposition of the Thamama Group eustasy was the dominant control over sedimentary sequences.

Typically the Lekhwair formation is thought to have deposited during an overall second order eustatic rise in sea-level (Figure 2.1) (Haq and Al-Qahtani, 2005; Haq et al., 1988). During this overall rise numerous minor perturbations in sea-level are observed (Figure 2.1), these perturbations can be related to the many cycles identified within the Lekhwair Formation (Alsharhan and Kendall, 1991; Haq et al., 1987; Haq et al., 1988). Alsharhan and Kendall (1991) identified six shoaling upward cycles in core and compared these cycles to the duration of the perturbations observed on the sea-level chart of Haq et al. (1988) (Figure 2.1), this would suggest that the cycles deposited over 1-1.5 Myr periods and are therefore likely to be related to 3rd-4th order sea-level variations (Goldhammer et al., 1993; Goldhammer et al., 1990).

The cycles have been interpreted to have been deposited in a variety of depositional settings from deep water subtidal open marine, to restricted inner shelf lagoons (Alsharhan and Kendall, 1991; Alsharhan, 1989; Cox, 2010; Granier et al., 2003). A typical cycle is on a meter to decameter-scale (Alsharhan and Kendall, 1991; Cox, 2010; Hillgärtner et al., 2003b) and contains dense argillaceous impermeable mud- and wackestones with aragonitic foraminifers and algae as well as
non-aragonitic foraminifers such as *Choffatella* and *Hensonella* (Granier et al., 2003) that shoal up into microporous wackestones and peloidal packstones/grainstones (Alsharhan and Kendall, 1991; Cox, 2010).

The depositional porosity for each of these cycles is thought to be enhanced by leaching with freshwater during relative falls in sea-level (Alsharhan and Kendall, 1991). Sub-aerial exposure is also suggested to have resulted in the leaching of bioclasts at the 3rd order SB that marks the transition from the Lekhwair to Kharai Formation (Granier et al., 2003), this 3rd order SB can be correlated over hundreds of kilometres on the platform (Hillgartner et al., 2003b; Van Buchem et al., 2002b).
Figure 2.1: Sea level fluctuations during the deposition of the Thamama Group. Modified from Haq and Al-Qahtani (2005) and Pratt and Smewing (1993).
2.2.1.2 Kharaib Formation

The Kharaib Formation is typically represented by two depositional cycles (Alsharhan and Kendall, 1991). To help constrain the time of deposition for these cycles they have been related to the two perturbations observed on the sea-level chart of Haq et al. (1988) and Haq and Al-Qahtani (2005), this suggests that the cycles were deposited over 1.5 Myr periods and are of a 3rd order (Goldhammer et al., 1990). The cycles are therefore likely to have been deposited over a greater duration to the cycles of the Lekhwair Formation (Section 2.2.1.1), but appear to have deposited under comparable conditions (Alsharhan and Kendall, 1991).

The Transgressive Systems Tracts (TSTs) of the two cycles are characterised by skeletal wackestones and packstones with a low biodiversity that can be correlated for tens to hundreds of kilometres across the platform (Strohmenger et al., 2006). Peloids, small echinoderm fragments, dasycladacean algae and red algae are frequently identified as well as abundant Thalassinoides burrows that are commonly dolomitised, this is taken to indicate deposition in a low energy subtidal setting (Figure 2.2) (Alsharhan and Kendall, 1991; Hillgärtner et al., 2003b; Strohmenger et al., 2006).
Figure 2.2: Depositional environment for the TSTs of the Kharaib Formation (Strohmenger et al., 2006).

The Highstand Systems Tracts (HSTs) of these cycles are composed of grain dominated lithofacies that are characterised by coated grain, algal, skeletal grainstones and rudstones (Figure 2.3). The depositional environment is thought of as a tidal-influenced, high-energy bioclastic shoal environment with a water depth not > 10 m, even during the highest rates of sea-level rise (Hillgärtner et al., 2003a). Normal marine conditions with good water circulation prevailed, as indicated by the diverse faunal content. Nutrients were sufficiently abundant to sustain the development of algal buildups and rudists, and water turbidity was low as indicated by the presence of Lithocodium/Bacinella (Strohmenger et al., 2006). The sequence boundaries identified by Strohmenger et al. (2006), both in outcrop and in core are thought to be marked by subaerial exposure (Figure 1.16).
In contrast to the two 3rd order depositional cycles identified by most authors (Alsharhan and Kendall, 1991; Strohmenger et al., 2006; Vahrenkamp, 2010; Van Buchem et al., 2010), Granier et al (2003) identified four 3rd order depositional cycles. All cycles start with a thin TST and display a trend of upward shallowing into more homogeneous deposits. The first three cycles are thought to deposit in the mid ramp setting, with the 4th thought to deposit in the inner or inner-most ramp setting. No evidence to suggest that the formation has experienced subaerial exposure was obtained by Granier et al. (2003), this is in contrast to the multiple exposure surfaces identified by Strohmenger et al. (2006).

2.2.1.3 Shu’aiba’ Formation

The Shu’aiba’ Formation is thought to have deposited over one depositional cycle by Alsharhan and Kendall (1991), by comparing this cycle to the Haq et al (1987) sea-level chart Alsharhan and Kendall (1991) suggest that the cycle was deposited over a 2.5 Myr period. This suggests that the cycle of the Shu’aiba’ Formation formed over a greater duration than the cycles of the Lekhwair and Kharaib Formations and is of a 3rd order (Goldhammer et al., 1990). However, recent
work has sub-divided the Shu’aiba’ Formation into a Lower and Upper Member (Figure 2.4) (Granier et al., 2003; Strohmenger et al., 2006; Vahrenkamp, 1996; Van Buchem et al., 2010).

![Figure 2.4: Sequence stratigraphy for the Kharaib and Shu’aiba’ Formations, modified from Van Buchem et al. (2010). The locations of the regionally correlatable 3rd order MFSs (Sharland et al., 2004) are also shown in relation to the Kharaib and Shu’aiba’ Formations.](image)

Prior to the formation of the Lower Shu’aiba’ Member (LSM) the Hawar Member was deposited (Figure 2.4). The Hawar Member is a deci-metre-scale interval that is argillaceous, dominated by *orbitolinids*, peloids, small echinoderm fragments and dasycladacean algae (Pittet et al., 2002). The Hawar Member is organic and siliciclastic rich, with mudcracks, palaeosols and blackened grains indicating frequent exposure (Strohmenger et al., 2006).

The LSM (Figure 2.4) is a shallow to deep water wackestone to packstone that consists of *Lithocodium/Bacinella, Palorbitolina lenticularis*, foraminifera, and
peloids (Granier et al., 2003; Immenhauser et al., 2005; Masse et al., 1998). The third order K80 Maximum Flooding Surface (MFS) (Figure 2.4) of the LSM marks the time at which rudist mounds began to thrive. After this MFS a general tendency towards shoaling, and the spread of lagoonal facies is indicated by the presence of dasyclad algae and encrusting Lithocodium sp. aggregates and implies a SB at the close of the LSM (Strohmenger et al., 2006; Ziegler, 2001). Strohmenger et al. (2006) suggests that the LSM was subaerially exposed, however this is contrary to the findings of Granier et al. (2003).

The Upper Shu’aiba’ Member (USM) (Figure 2.4) shows a subtle-shallowing upward trend as well as graining upwards and consists of packstones and wackestones with rudist fragments and rare Bacinella (Figure 2.5) (Strohmenger et al., 2006). The USM has been observed to prograde over deeper open marine beds for 10 to 15 km (Masse et al., 1998), with the elevated rudist reef feeding the slope, this agrees with the observations of Pratt and Smewing (1993). The USM culminates with exposure and the development of a 2nd order SB (Alsharhan and Kendall, 1991; Strohmenger et al., 2006).

![Figure 2.5: Depositional model for the Shu’aiba’ Formation in Oman (Masse et al., 1998).](image)

The Bab Member, which is coeval to the USM (Figure 2.4), deposited in an intrashelf basin and palaeoforeslope, this basin occupies a depression on the stable craton that was most likely formed by differential subsidence rates associated with relative sea-level rise (Alsharhan, 1995; Murris, 1980). The Bab Member consists of
well bedded grey-brown lime mudstones and wackestones that contain thick shaly,
bituminous, and stylolitic partings that become more argillaceous upwards. Burrowing is common and planktonic foraminifera and pelagic ammonites have been found (Alsharhan, 1995).

### 2.2.2 Chronostratigraphic framework for the Lekhwair, Kharaiib and Shu’aiba’ Formations

Recently numerous publications have aimed to place the aforementioned depositional cycles of the Lekhwair, Kharaiib and Shu’aiba’ Formations into a regional sequence stratigraphic framework (Alsharhan and Kendall, 1991; Droste, 2010; Sharland et al., 2001; Strohmenger et al., 2004; Suwaina et al., 2004; Vahrenkamp, 2010; Van Buchem et al., 2010; Van Buchem et al., 2002a); this is contrary to Pratt and Smewing (1993) where no clear correlation was found between interpreted relative sea-level curves (Figure 2.1). These publications have used regional MFSs to correlate 3rd order sea-level cycles across the Arabian platform and have shown that the depositional environments of the Thamama Group do not occur randomly in time and space; they occur systematically in specific 3rd order composite sequences that can be regionally correlated (Section 2.2.1). This regional 3rd order stratigraphic framework is chronostratigraphically well constrained, therefore by relating the 3rd order cycles identified in this thesis to these 3rd order cycles the time of deposition for each cycle can be determined. This section presents the previously established 3rd order chronostratigraphy for the Arabian plate.

A 3rd order sequence stratigraphic framework was developed by Sharland et al. (2001) for the Arabian plate and is based on the identification of Maximum Flooding Surfaces (MFSs). This work was subsequently expanded on (Figure 2.6) (Sharland et al., 2004), and the age of the 3rd order MFSs have been recently revised (Simmons et al., 2007; Van Buchem et al., 2010), the revised ages are used in this thesis.
The studies suggest that the Lekhwair, Kharaiib and Shu’aiba’ Formations are bound by two regionally correlatable 2nd order sequence boundaries (K40 and K90) that mark the beginning and end of deposition. The formations are then composed of five 3rd order MFSs: K40, K50, K60, K70 and K80 (Figure 2.6) that have all been chronologically constrained (Sharland et al., 2004; Simmons et al., 2007; Van Buchem et al., 2010).

The K40 MFS is taken to mark the top of the Zakum Member (Figure 2.6) (Davies et al., 2002; Sharland et al., 2004; Strohmenger et al., 2006), the age of which is thought to be 134.5 Mya (Simmons et al., 2007).

The K50 SB marks the transition from the Lekhwair to Kharaib Formation (Strohmenger et al., 2006; Van Buchem et al., 2010) and the K50 MFS (129 Mya) is thought to be located at the 3rd order MFS of the Lower Kharaiib Member (LKM) (Figure 2.4), the age of this surface is supported by a new evolutionary scheme for *Palaeoorbitolina lenticularis* (Schroeder et al., 2010), carbon and strontium isotope data and cyclostratigraphic modelling (Strohmenger et al., 2010; Vahrenkamp, 2010).

Typically the K60 SB is taken to separate the LKM and the Upper Kharaiib Member (UKM) (Figure 2.4) (Strohmenger et al., 2006; Van Buchem et al., 2010), this is based on the identification of a surface that shows significant emersion. However, at Wadi Mu’ Ayin Van Buchem et al. (2002a) placed the K60 SB within the LKM, between rudist grainstone/rudstone and miliolid wackestones/mudstone at surfaces lined with mud cracks that are suggestive of subaerial exposure.

The 3rd order MFS of the UKM (Figure 2.4) is typically thought to represent the K60 MFS (125.5 Mya) (Strohmenger et al., 2006; Van Buchem et al., 2010), this interpretation agrees with the findings of Schroeder et al. (2010) where the presence
of *Montseciella arabica* within the UKM is taken to suggest deposition in the Upper Barremian.

K70 SB is located at the base of the Hawar Member (Strohmenger et al., 2006; Van Buchem et al., 2010) and shows significant emersion of Kharaib platform carbonates. The Lower Aptian K70 MFS (124.5 Mya) is thought to be within the Hawar Member of the Shu’aiba’ Formation (Figure 2.6) (Matthews and Frohlich, 2002; Van Buchem et al., 2010) or within the LSM (Figure 2.6) (Sharland et al., 2004). Advanced forms of *Palaeoorbitolina lenticularis* are now observed to be typical of the K70 MFS (Schroeder et al., 2010).

The age of the K80 MFS has recently been updated from 116 Mya (Matthews and Frohlich, 2002; Sharland et al., 2004) or 119 Mya (Simmons et al., 2007) to 123 Mya (Van Buchem et al., 2010) (Figure 2.4). They modify the date of the this surface based on new evidence of a marine incursion in Yemen, Ethiopia and Saudi Arabia, where a late Early Aptian age is confirmed by Orbitolinid fauna (Bosellini et al., 1999; Schroeder et al., 2010) and by ammonites (Le Nindre et al., 2010).

The 2nd order K90 SB marks the end of deposition for the Thamama Group and is typically represented by a disconformity. However, Field A and B are located in the Bab Basin and so the surface is located within the Bab Member (Figure 2.4). The K90 MFS is located within the Nahr Umr Formation (Figure 2.4) (Van Buchem et al., 2010).

### 2.2.3 Established $\delta^{13}C$ trends

$\Delta^{13}C$ can be used as a chronostratigraphic tool to help confirm the time at which the cycles deposited. This will be achieved by comparing the $\delta^{13}C$ signal of the bulk micrite for each cycle to established marine curves. $\delta^{13}C$ values can also be used to correlate between multiple wells, provided that the $\delta^{13}C$ value has not been altered by diagenesis. This section presents the previously established secular $\delta^{13}C$ trends for the Early Cretaceous.
Sprovieri et al. (2006) published marine bulk δ\textsuperscript{13}C profiles for the Berriasian-Barremian, the δ\textsuperscript{13}C curve for the end Valanginian to the Barremian is shown in Figure 2.7a. This curve shows that marine δ\textsuperscript{13}C signal increases sharply from ~1.5‰ to ~3.1‰ during the late Valanginian this episode is known as the Weissert event. Δ\textsuperscript{13}C then progressively decreases to 1.6 ‰ by ~ 133 Mya and then begins to progressively increase to ~ 2.6 ‰ at ~ 126 Mya. The Early Aptian is marked by the Taxy Episode where δ\textsuperscript{13}C decreases to ~2 ‰ by 125 Mya and then increases to ~5 ‰ in the Upper Aptian (Figure 2.7b) (Föllmi, 2012; Vahrenkamp, 1996; Vahrenkamp, 2010).

Figure 2.7: Previously established δ\textsuperscript{13}C profile for the Late Valanginian-Early Aptian with the Weissert Event and Taxy Episode labelled (black dashed lines), a). The left panel is from the Umbria-Marche composite section in Italy (Sprovieri et al., 2006), whereas the right panel is from sections in Northern Italy and Switzerland (Weissert and Erba, 2004). The arrow marks the progressive increase in δ\textsuperscript{13}C from the Hauterivian until the Upper Barremian-Early Aptian. Modified from Sprovieri et al. (2006). The previously established Upper Barremian-Early Aptian composite δ\textsuperscript{13}C profile for three onshore wells in Abu Dhabi is shown in b), after the Taxy episode there is a progressive increase in δ\textsuperscript{13}C to ~5 ‰ in the Upper Aptian. H. – Hawar Member, N.U. - Nahr Umr Formation. Modified from Vahrenkamp (2010).
2.3 Chronostratigraphic Framework

Numerous reservoir intervals are present in the Thamama Group, these reservoir intervals are separated and encased by non-reservoirs that have very low porosities and permeabilities (Figure 1.16) (Strohmenger et al., 2006). The definition of a reservoir and its relative relationship to a depositional cycle is described in Section 2.4.1. This section assumes the layer-cake nature of the Lekhwair, Kharab and Shu’aiba’ Formations.

The reservoir intervals in Field B have already been placed in a sequence stratigraphic framework by Cox (2010) and the reservoir intervals identified in Well 2A of Field A have also been placed in a sequence stratigraphic framework by Prof. R. Wood (unpublished dataset). However, the reservoir intervals identified in Well 1A and 3A have not been placed in a sequence stratigraphic framework.

The first aim of this section is to relate the reservoir intervals identified in Well 1A and 3A and in outcrop to a sequence stratigraphic framework. This framework is then related, wherever possible, to the established 3rd order chronostratigraphic framework (Section 2.2.2) (Sharland et al., 2004) to constrain the time of deposition for the cycles. A biostratigraphic assessment is then undertaken on all available thin sections from all wells to help further constrain the time of deposition for the cycles and finally the stable carbon isotope signals obtained for the cycles are then used to aid with establishing a chronostratigraphic framework.

87Sr/86Sr isotope data were made available to the author during this project. This data was correlated to secular marine curves (e.g. McArthur et al. (2001)) to validate the chronostratigraphy presented here, however due to reasons of confidentiality the data have been omitted from this thesis. A chronostratigraphic summary figure for the nine cycles assessed in this thesis is presented in Section 2.3.4.

2.3.1 Sequence stratigraphy of the Kharab and Lower Shu’aiba’ Formations

The purpose of this assessment is to place the depositional cycles, that best describe the reservoir intervals of the Kharab Formation (Figure 2.4) (cycles iii and ii) and the Shu’aiba’ Formation (Figure 2.4) (cycle i) in Well 1A and 3A into a
sequence stratigraphic framework, these cycles are also assessed in an outcrop analogue, which has been previously studied by Strohmenger et al. (2006). The sequence stratigraphic framework identified is related, wherever possible, to the 3rd order MFSs of Sharland et al. (2004) (Section 2.2.2) to help determine the time at which the cycles deposited.

The MFS of cycle iii is taken to represent the location of the 3rd order K50 MFS and is shown in Figure 2.8. The surface is marked by encrusting microsolenid corals (Figure 2.9a), stromatoporoids (Figure 2.9b) and large gastropods (Figure 2.9c). The surface has a red colour, due to the presence of haematite, and marks the transition from mudstone and wackestone dominated TST deposits to rudist and orbitolinid dominated packstone and grainstone lithofacies of the HST.

![Figure 2.8: MFS of cycle iii, which is exposed on the Southeast face of Wadi Rahabaha hammer is shown for scale (red ellipsoid). The MFS is encrusted by microsolenid corals and stromatoporoids (Figure 2.9a, b, c).](image)

The overlying SB of cycle ii (K60 SB) is marked by an extensive hardground that was observed to be extensively bored with *Gastrochaenolites* (with in-situ *Lithophaga*) and *Trypanites* borings (Figure 2.9d, e), unlike Strohmenger et al. (2006) there is no evidence for sub-aerial exposure. Overlying the SB is a heavily stylolitised thin shale interval. The MFS of cycle ii (K60 MFS) is represented by the transition from orbitolinid rich wackestones of the TST to coated grain, rudist rich packstones of the HST.
The SB of cycle i is heavily stylolitised and so the features that are present on this surface are difficult to identify. Strohmenger et al. (2006) suggested the SB has a pedogenic overprint however, no clear evidence to suggest this was obtained in this assessment. The surface is overlain by shales of the Hawar member (Figure 2.9f).

Figure 2.9: SBs and MFSs identified on the Southeast wall of the Wadi Rahabah outcrop. a) Microsolenid corals, b) Stromatoporoids, and c) Gastropods, encrust the MFS of cycle iii. The SB of cycle ii is marked by a hardground that contains d) Trypanites, and e) Lithophaga borings with in-situ bivalves (arrow). The SB of cycle i (K70 SB) is overlain by a shale interval that is observed at the contact between cycle ii and i, camera bag for scale (red ellipsoid).
2.3.2 Biostratigraphy

Simmons (1994) showed that foraminifera can be used to date the Cretaceous deposits of the Arabian Peninsula. This work was expanded on by Schroeder et al. (2010) where they developed an orbitolinid biostratigraphic zonation scheme for the Barremian-Aptian of the Eastern Arabian Plate. A biostratigraphic assessment has therefore been undertaken to help constrain the time of deposition for all the cycles assessed in this thesis.

Low evolutionary rates during the deposition of the Lekhwair Formation make dating difficult. However, the oldest cycle (ix) contains the benthic foraminifera *Choffatella deciphens* (Figure 2.10) that is characteristic of the Hauterivian-Early Aptian (Granier et al., 2003; Robertson et al., 2012; Simmons, 1994) and may suggest that cycle ix deposited during the Hauterivian.

![Image](image.png)

*Figure 2.10: Choffatella deciphens that is characteristic of the Hauterivian-Early Aptian.*

*Palorbitolina lenticularis* is thought to be characteristic of the Lower Barremian (128.5 Mya; Schroeder et al. (2010)). *Orbitolinids* have only been observed within the HST of cycle iii (Figure 2.11a), implying that deposition commenced in the Lower Barremian. This further suggests that the MFS of cycle iii does correspond to the K50 MFS.
Montseciella arabica (Henson, 1948; Săsăran et al., 2004) is characteristic of the Upper Barremian (~ 127 Mya) and thought to represent the K60 MFS (Schroeder et al., 2010). Montseciella arabica is observed in the HST of cycle ii (Figure 2.11b), but not in the underlying TST. Therefore in this study the MFS of cycle ii is taken to represent the K60 MFS.

Figure 2.11: a) Early form of P. Lenticularis observed in the HST of cycle iii; b) M. Arabica thought to be characteristic of the K60 MFS.

2.3.3 Stable carbon isotope signatures

2.3.3.1 Results

To help further constrain the chronostratigraphy of all cycles a δ¹³C bulk micrite assessment was undertaken (Section 2.2.3). New δ¹³C results are presented in this section for Field A and the outcrop and new δ¹³C results are also presented for Well 1B and Well 4B of Field B, but for Field B the majority of the results have been obtained from the Ph.D. thesis of Cox (2010); see Appendix 2 for the data values and their origin. However, because similar trends are observed in both fields and in the outcrop the data will be discussed together.

Measured δ¹³C values for Field A (Figure 2.12) and B (Figure 2.13) vary between 1.7 ‰ to 4.8 ‰, and -0.1 to 4.4 ‰ respectively, the δ¹³C values for the outcrop are between -1.5 ‰ and 2.2 ‰ (Figure 2.14). The δ¹³C values obtained in this thesis are in good agreement with the δ¹³C values of 0 ‰ to 4 ‰ that are typical
of Early Cretaceous seawater (Föllmi, 2012; Sprovieri et al., 2006; Vahrenkamp, 1996; Veizer et al., 1999; Weisset et al., 1998).

The $\delta^{13}C$ values during the deposition of cycles ix to i progressively increase from $\sim2\%o$ in cycle ix to $\sim3.5\%o$ in cycle i. The progressive upward increase in $\delta^{13}C$ is observed in Field A (Figure 2.12) and B (Figure 2.13) and in outcrop (Figure 2.14).

There are episodes, during this overall increase in $\delta^{13}C$, where the $\delta^{13}C$ values become more negative, these episodes are typically observed in proximity to SBs and the decrease observed is of a varying magnitude (0.5–2 %o) (Figure 2.12 and 2.13).
Figure 2.12: Bulk micrite $\delta^{13}C$ for Field A. There is a general upward increase in $\delta^{13}C$ into cycle ii this is most clearly observed in Well 2A. There are two negative episodes observed in Well 1A and one episode in Well 3A; these episodes are located in proximity to the SB of each cycle. The sequence stratigraphy for Well 2A is provided by Prof. R. Wood.
Figure 2.13: Bulk δ^{13}C for Field B. The general upward increase of δ^{13}C values is most clearly observed in Well 1B and Well 2B. Numerous negative δ^{13}C episodes are observed in the wells; these episodes are typically observed at the SB of each cycle. The sequence stratigraphy for the nine cycles is taken from Cox (2010).
Figure 2.14: Bulk $\delta^{13}$C for the Wadi Rahabah outcrop section for all exposed surfaces. The mean $\delta^{13}$C value is presented along with standard error and with the number of samples assessed (N); terminology of the surfaces from Figure 1.15 (Strohmenger et al., 2006). The $\delta^{13}$C progressively increases upwards towards the HST of cycle ii. Four samples at the K60 FS100 surface have a negative $\delta^{13}$C that does not agree with the progressive upward increase in $\delta^{13}$C.
2.3.3.2 Discussion

Diagenesis can alter the primary δ\(^{13}\)C signal of bulk micrite, with sub-aerial exposure, sulphate reduction (Sass et al., 1991) and methanogenesis known to affect the bulk rock signature. However, typically for Field A and B and for the outcrop the profiles are smooth and are within the range of Early Cretaceous seawater (0–4 ‰) (Föllmi, 2012; Sprovieri et al., 2006; Vahrenkamp, 1996; Veizer et al., 1999; Weissert et al., 1998); with the exclusion of the δ\(^{13}\)C values obtained from the K60 FS100 surface in outcrop. Based on these observations it is concluded that the δ\(^{13}\)C profiles reflect changes in seawater composition over time and have not been greatly affected by the input of meteoric water, this agrees with the outcrop and core observations where no evidence for subaerial exposure was observed (Section 2.3.1). Therefore, the δ\(^{13}\)C values obtained from the outcrop and core can be related to regional trends (Section 2.2.3).

The extrusion of one of the World’s largest flood basalt provinces, the Paraña Etendeka Large Igneous Province (LIP), at 134.7 ± 1 Mya (Thiede and Vasconcelos, 2010) is thought to have resulted in the Weissert event (Coffin and Eldholm, 1993; Sprovieri et al., 2006); a δ\(^{13}\)C excursion from ~1.5 ‰ to ~3.1 ‰ (Figure 2.7) (Föllmi, 2012; Sprovieri et al., 2006). The emplacement of this LIP is thought to have increased sea-level and led to more humid conditions which increased erosion and caused a greater flux of C\(^{12}\) to the oceans (Follmi et al., 1994). This in turn resulted in the deposition of a greater amount of organic carbon which may have caused the δ\(^{13}\)C positive excursion. The δ\(^{13}\)C values obtained in this thesis do not record such a sharp positive excursion, implying that all the cycles are likely to have deposited after this excursion.

The progressive increase in δ\(^{13}\)C that is observed in Field A and B and the outcrop during the deposition of the Lekhwair Formation (cycles ix–iv) and Kharaib Formation (cycles iii, ii) is similar to that observed by Sprovieri et al. (2006) (Figure 2.7a) and van de Schootbrugge et al. (2000) for the Hauterivian to the Barremian. This excursion begins in the Hauterivian where values of ~1.9 ‰ are obtained and culminates at ~2.5 ‰ in the Barremian (Sprovieri et al., 2006). Progressive platform
drowning during the Hauterivian-Barremian (Figure 2.1) (Haq et al., 1988; Masse, 1993) can result in increased weathering and run-off (Follmi et al., 1994; Weissert et al., 1998) which can be related to the decrease in carbonate carbon to organic carbon burial rates (Scholle and Arthur, 1980) and result in the progressive increase in δ13C during the Hauterivian-Barremian.

The Hauterivian is thought to be a stage dominated by humid conditions and platform drowning, with progressive evolution, environmental and climate change that culminated with the Faraoni Episode at the end of the Hauterivian; an episode marked by a small decrease (< 1‰) in δ13C values (Föllmi, 2012). This episode can be related to a peak in submarine volcanism of the Tristan da Cunha plume along the Rio Grande Rise at around 127 Mya (Baudin, 2005) which is thought to have resulted in the release of light carbon in the form of volcanic CO2. Numerous small δ13C decreases are observed in the δ13C data (Figures 2.12, 2.13) and therefore this event cannot be used for chronostratigraphic purposes.

There are multiple intervals with a lower δ13C value during an overall progressive increase in δ13C during the Hauterivian-Barremian. These decreases are observed in proximity to the SBs of the cycles and into the overlying TST, the decreases may be a result of increased volcanic activity, or due to the release of methane clathrates (Föllmi, 2012). Volcanic activity and the release of methane clathrates may have resulted in the short negative excursions due to the increase in light carbon in the form of volcanic or methane related CO2. A similar process to this has been related to the negative δ13C episodes observed during the Cretaceous oceanic anoxic events (Baudin, 2005; Kuhnt et al., 2011).

Another possible cause for a more negative δ13C signal is due to the formation of hardgrounds. Sulphate reduction can cause supersaturation with respect to calcite (Dickson et al., 2008) and lead to the formation of hardgrounds and to the deposition of 13C-depleted carbonate. However, this is likely to be localised and therefore the general decrease in δ13C around the SBs is unlikely to be related to this process; this is discussed further in Section 3.4.1.
Furthermore, the decrease in \( \delta^{13}C \) may also be related to subaerial exposure and to the development of pedogenic surfaces. Isotopically light carbon from soil derived \( \text{CO}_2 \) may be incorporated into the carbonate lattice during recrystallization which will lead to a decrease in \( \delta^{13}C \) (Allan and Matthews, 1982; Dickson and Saller, 2006; James and Choquette, 1984) . However, no evidence to suggest subaerial exposure was obtained from either outcrop or core analyses and no vadose diagenetic features (e.g. meniscus cements) were observed in petrographic assessments. Therefore, subaerial exposure is unlikely to have caused the decrease in \( \delta^{13}C \) at the SBs of the cycles.

The \( \delta^{13}C \) decrease from 3.5 \( \% \) to 2 \( \% \) marks the Upper Barremian-Early Aptian (Figure 2.7a, b) (Sprovieri et al., 2006; Vahrenkamp, 1996; Vahrenkamp, 2010) and the transition from the Kharaib Formation to the Shu’aiba’ Formation (K70 SB). This event can be related to the Taxy episode that coincides with the onset of the Ontong-Java LIP in the Pacific that began at around 125 Mya (Tejada et al., 2009) and can be related to an increase in light carbon in the form of volcanic \( \text{CO}_2 \). The related decrease in \( \delta^{13}C \) is potentially observed at the SB of cycle i in Field A and B (Figure 2.12) however, due to the multiple decreases in \( \delta^{13}C \) at the SBs of most cycles, this excursion cannot be used for chronostratigraphic purposes.

### 2.3.4 Composite chronostratigraphic framework

All the sequence stratigraphic data and chronostratigraphic data previously presented (Sections 2.3.1-2.3.3) have been integrated for all the cycles assessed in this thesis so that a composite chronostratigraphic framework can be presented (Figure 2.15). This framework is presented along with that identified by other authors (Hardenbol et al., 1998; Sharland et al., 2004; Strohmenger et al., 2006). This framework is the most accurate possible for the data available and is also in agreement with \( ^{87}\text{Sr} / ^{86}\text{Sr} \) data; although these data have been omitted due to reasons of confidentiality.

The time of deposition for cycles iii-i is relatively well constrained by the identification of 3\(^{rd}\) order MFSs and by biostratigraphic assessment (Section 2.3.1-2.3.2). However, there are little chronostratigraphic constraints available for cycles.
ix-iv, with the only constraint being that the oldest cycle was deposited in the Hauterivian (Section 2.3.2) and after the δ^{13}C Weissert event that is dated at ~135 Mya (Section 2.3.3). Therefore the precise time of deposition for these cycles is uncertain and the cycles can deposit at any time between 135 Mya and 130 Mya.

The cycles identified here, which best describe the reservoir intervals assessed in this thesis, correspond reasonably well with the 4th and 3rd order cycles identified by Strohmenger et al. (2006) (Figure 2.15). However, unlike Strohmenger et al. (2006) a 3rd order SB is not identified at the base of the Kharai Formation (cycle iii) and instead the SB has been identified within the Lekhwair Formation. The author has only had access to core for cycles iii to i in Field A (not including the SB of cycle iii). Therefore no comment can be made on this discrepancy.
Figure 2.15: Composite sequence stratigraphy for the nine cycles assessed in the Lekhwair, Kharai and Shu’aiba’ Formations in relation to the sequence stratigraphic framework of other authors. Timescale from Ogg et al (2008); 3rd order sequence from Hardenbol et al. (1998) and 2nd, 3rd and 4th order sequences from Strohmenger et al. (2006). Age of the MFSs from Simmons et al. (2007). Note: the author has only undertaken sequence stratigraphic analyses on cycles iii-i from Well 1A and 3A of Field A; cycles ix-i of Field B were assessed by Cox (2010) and cycles vii-ii of Well 2A, for Field A, were assessed by Prof R Wood (unpublished dataset).
2.4 Results

A chronostratigraphic framework has now been established for the nine cycles assessed in this thesis (Section 2.3). The overriding aim for this section is to determine the control that depositional environment has on present day reservoir quality. This section will firstly present a typical depositional cycle for the Kharaih (cycles iii-ii) and Shu’aiba’ Formations (cycle i) as identified in Well 1A and 3A and in outcrop. Secondly, the depositional lithofacies and fabrics identified in the cycles are then presented and finally the plug derived porosity and permeability data are related, wherever possible, to these lithofacies and fabrics to determine the control that depositional environment has on present day reservoir quality.

2.4.1 Description of a typical cycle for the Kharaih and Shu’aiba’ Formations

A typical cycle starts with the deposition of a TST which is composed of shale and argillaceous micrite/mudstone (Figure 2.17a, b, c) that grade upwards into a pyritised (Figure 2.17d) mudstone to wackestone, this deposit continues to grade into an orbitolinid rich wackestone-packstone (Figure 2.17e). The HST is composed of more grain dominated fabrics that are rich in *Lithocodium/Bacinella* (Figure 2.17f) and rudists (Figure 2.17g) with interbeds of coated grains. These rudist rich deposits are overlain by stylolitised (Figure 2.17h), microdolomitised, miliolid and rudist rich wackestones and packstones (Figure 2.17i). The overlying SB marks the end of a cycle and is characterised by *Lithophaga*-bored hardgrounds.
Figure 2.16: Typical cycle for the Kharaih and Shu’aiba’ Formations as observed in the outcrop and core sections from Well 1A and 3A of Field A. The letters on the figure relate to the photographs shown in Figure 2.17. The locations of the non-reservoir and reservoir intervals in relation to the systems tracts of the cycle are also shown. The interpreted depositional environments are included in the blue box; they are discussed in Section 2.5.1.
Figure 2.17: a) Mudstone with the K60 SB interpreted below the shale; b) Stylolitised mudstone/wackestones; c) Haematite-rich burrowed surface; d) Pyritised mudstone/wackestones; e) Orbitolinid-rich wackestones; f) Lithocodium/Bacinella packstone; g) Grain-dominated packstone; h) Mud-dominated packstone overlain by mudstone – the change in fabric is marked by a rectangular stylolite; i) Lagoonal section that is dominated by stylolites and rudist fragments; hammer for scale (red ellipsoid).
All the cycles observed in Well 1A and 3A consist of a reservoir and non-reservoir interval (Figures 2.12, 2.16). The reservoir intervals are defined based on the presence of oil stain in core and plug samples and are commonly associated with the HST (Azer and Toland, 1993; Boichard et al., 1994; Van Buchem et al., 2002a) (Figure 2.18a, b). The reservoirs are encased and separated by non-reservoir intervals that have very low porosities and permeabilities; the non-reservoirs are commonly associated with TST deposits (Figure 2.18a, b). Oil stain is only observed within parts of the HST due to extensive cementation of some grainy HST lithofacies, the cause for this extensive cementation will be discussed in Chapter 3 and 4. Therefore, in this thesis, the non-reservoir intervals include the cemented lithofacies of the HST as well as deposits from the TST and always underlie the reservoir intervals (Figure 2.18a, b).

Figure 2.18: a) Definition of reservoir and non-reservoir intervals in relation to the sequence stratigraphy, b) Core photograph showing the oil stained reservoir interval and an overlying non-reservoir interval.

Minor non-reservoir intervals are observed within the reservoir intervals of Field A (Figure 2.16), these minor intervals correspond to a variety of features: 1) calcite cemented layers that are formed by a sharp textural change from wackestones to grainstones; 2) stylolitised horizons. Some of these horizons could be controlled by textural changes, however stylolites can also occur within homogeneous facies; 3)
low porosity horizons occurring near weak textural changes; 4) Dolomitised intervals.

2.4.2 Lithofacies and fabrics

To assess the depositional environments for the nine cycles, 22 lithofacies were identified for Field A and Field B (Appendix 3). These lithofacies are based on the distinctive fauna, floral and abiotic components in the petrographic sections, and are used to group rocks deposited in a similar environment.

The depositional fabric/texture for each lithofacies is also included because these fabrics can be used to group lithofacies that deposited in a similar environment. The fabrics used in this thesis are modified from the Dunham classification scheme (Dunham, 1962) by applying the modifications of Lucia (1995). The Dunham classification scheme uses the Mudstone (M), Wackestone (W), Packstone (P) and Grainstone (G) fabrics (Dunham, 1962). However, the packstone fabric will be subdivided into Grain Dominated (GDP) or Mud Dominated (MDP) Packstones (Lucia, 1995). This subdivision will allow for a more accurate representation of the depositional petrophysical properties of the fabric, because this division takes into account the presence (GDP) or absence (MDP) of intergrain pore space (Lucia, 1995).

2.4.3 Lithological trends in reservoir quality

The lithofacies that were identified through petrographic assessment have been related to their corresponding plug derived porosity and permeability (Figure 2.19). The results presented here are therefore from the five wells from Field B and from Well 1A of Field A because plug porosity and permeability data are not available for Well 2A and 3A of Field A. Figure 2.19 shows that there is a large variability in the petrophysical properties for all lithofacies, however the variability is more pronounced in the grainy facies of the HST (highlighted by the black dashed line) than in the more mud dominated TST deposits (highlighted by the red dashed line).
All lithofacies of the HST have a highly variable porosity and permeability, this is demonstrated by a weak linear relationship ($R^2 = 0.16$) between porosity and permeability for the lithofacies of the HST. A highly variable porosity and permeability is also present for a single lithofacies, examples include L2 and L13 that have a porosity of 6.1-35 % (n=21) and 1.9-28.19 % (n=16) respectively, and a permeability of 0.01-90 mD (n=21) and 0.01-83.03 mD (n=16) respectively. The average porosity for the HST is $11.9 \pm 0.9$ % (n=86), whereas the average permeability is $20.5 \pm 10.9$ mD (n=86).

A smaller variability and a lower average porosity and permeability are observed for the lithofacies of the TST (Figure 2.19). L11 is one of the most variable lithofacies observed in the TST and has a porosity and permeability of 0-6.9 % and 0-2.6 mD respectively (n=27). However, most lithofacies have a lower variability such as L8 that has a porosity of 0-5.9 % and permeability of 0.01 mD (n=4). The average porosity for the TST is $2.5 \pm 0.5$ % (n=51) and the average permeability is $0.1 \pm 0.5$ mD (n=51).
Figure 2.19: Porosity and permeability plots for the lithofacies observed in the Lekhwair, Kharai and Shu’aiba’ Formations of Fields A and B. The HST and TST lines circumscribe all lithofacies for those tracts. No porosity and permeability data were available for L15, L16 and L22. The colour and label for each lithofacies in the plot is related to the colour coded lithofacies identified in Appendix 3.

Five depositional fabrics have been identified in the nine cycles (Section 2.4.2). These fabrics are used here to group carbonate lithofacies that are deposited in a similar depositional environment in order to simplify the assessment.

The plug derived porosity vs. permeability data in relation to the depositional fabric is shown in Figure 2.20. This figure shows that there is a greater correlation between porosity and permeability for the muddier fabrics (mudstones, wackestones ($R^2 = 0.43$) and mud dominated packstones ($R^2 = 0.39$)) than for the grainy fabrics (grain dominated packstones ($R^2 = 0.17$) and grainstones ($R^2 = 0.21$)).
Figure 2.20: Porosity vs. permeability for the depositional fabrics observed in the Lekhwair, Kharaiib and Shu’aiba’ Formations of Fields A and B, with their corresponding $R^2$ (coefficient of determination) values.

Porosities for the grainstone and grain dominated packstones range between 2.3-30.4 % and 0.8-35 % respectively, with permeabilities ranging between 0.03-640 mD and 0.01-670 mD, respectively. Whereas the porosities and permeabilities for the mud dominated packstone and wackestone fabrics are typically lower, with the porosities for mud dominated packstone and wackestone fabrics being between 0.9-30 % and 0.3-15 %, respectively, and with permeabilities between 0.01-83.03 mD and 0.01-30.2 mD, respectively.
2.5 Discussion

2.5.1 Depositional environment

It needs to be determined whether the ramp is rimmed or not, because this will affect the depositional environments present on the shelf and can greatly determine the sites of exploration wells. This is because the shoals tend to be dominated by higher energy depositional facies and typically have higher porosities and permeabilities.

The lithofacies rich in algal components (Bacinella and Lithocodium) could sometimes constitute mounds (Van Buchem et al., 2002a) which should develop a discrete relief. But the more probable relief is associated with the rudist-rich lithofacies which characterise the deposits of the Kharaib and Shu’aiba’ Formations. These lithofacies are most often packstones than grainstones and so are interpreted as typical of a low to medium energy environment. Nevertheless, it is probable that these deposits compose a rim of rudist bioherms or biostromes which could act as a more or less developed shoal barrier.

The observations of this study support this hypothesis. The upper part of each cycle records packstone and wackestone fabrics that are composed of miliolids and rudists (Figure 2.16). These deposits are thought to form in a more proximal environment and may indicate the presence of a lagoonal environment typically behind a shoal. Therefore the depositional models presented here have been modified from a typical ramp model (Ahr, 1973) to include a shoal and lagoon.

Two depositional models have been developed (Figure 2.20, 2.22), one for TST deposits and another for the deposits of the HST. Both models assume an increasing grain size towards the shallowest parts of the ramp. Biological assemblages were used to indicate the location on the ramp. A distal open marine or a restricted inner ramp setting is indicated by Orbitolinids (flattened), ubiquitous Choffatellids and Lithocodium/Bacinella, with proximal high energy depositional locations being marked by rare planktics, rudists, coated grains, ooids, fragmented dasycladacean, and miliolids.
The lithofacies of the TST are more mud rich and are typically represented by mudstone and wackestone fabrics (Figure 2.21). The lithofacies of the TST have been interpreted to form in a variety of depositional environments from the basin to upper ramp, with mass flows thought to lead to the deposition of some lag deposits (e.g. L11).

**Figure 2.21: Depositional environments for the lithofacies in the TST. The numbers relate to the predefined lithofacies (Appendix 3).**

The HST deposits are interpreted to deposit above the fair weather-wave base in a proximal open marine and restricted marine (inner ramp) depositional environment (Figure 2.22). The deposits of the HST are commonly high energy deposits and are dominated by packstone and grainstone fabrics.

Therefore, a typical depositional cycle is interpreted as commencing in a deep, low energy, open marine environment that shallows upwards into a condensed section capped by bored hardgrounds that formed in a lagoonal environment (Figure 2.16, 2.21, 2.22).
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Figure 2.22: Depositional environments for the lithofacies of the HST. The numbers relate to the predefined lithofacies (Appendix 3).

2.5.2 Lithological trends in reservoir quality

There is typically a greater porosity and permeability in the more grain dominated fabrics (grainstones and grain dominated packstones) than in the mud dominated fabrics (mudstones, wackestones and mud dominated packstones) (Figure 2.20). This may suggest that the grain dominated fabrics still retain their higher depositional porosities and permeabilities when compared to the mud dominated fabrics (Section 1.3.1).

However, typically there is a poorer correlation between porosity and permeability for grainstones and grain dominated packstones (Figure 2.20) when compared to the more mud dominated fabrics. This may be due to the initially higher porosity and permeability in the grain dominated fabrics which led to a more complex diagenetic history, where grain rearrangement, extensive intergranular cementation and dissolution can all greatly affect the porosity and permeability of the fabric. However in the more mud dominated fabrics porosity and permeability is likely to be lower, after initial dewatering, which will make the effect of diagenesis
on the pore system less pronounced and lead to a better correlation between porosity and permeability.

The present day porosities and permeabilities for the lithofacies that compose the HST are also more variable than those of the TST (Figure 2.19); this is because the fabrics of the HST are more grain dominated in comparison to the mud dominated TST (see aforementioned reasons). Furthermore, on average the porosity and permeability of the lithofacies that compose the TST is lower than for the HST (Section 2.4.3), this has led to these deposits commonly being thought of as impermeable seals to the underlying HST reservoir interval (Alsharhan and Kendall, 1991).

Therefore, this assessment suggests that the products of diagenesis conform reasonably well to the depositional fabrics and have not completely altered the depositional control on porosity and permeability, with the more grain dominated deposits of the HST typically having better porosities and permeabilities than the TST. This would in turn imply that the location of potential reservoirs can be predicted by three dimensional modelling of the petrophysically important depositional fabrics.

2.6 Conclusions

Multiple authors have suggested that the SBs of the Lekhwair, Kharaib and Shu’aiba’ Formations have experienced sub-aerial exposure (Granier et al., 2003; Strohmenger et al., 2006); however, in this thesis no evidence for sub-aerial exposure was obtained either from the core or outcrop, or through petrographic assessment.

A chronostratigraphic framework has been established for the nine cycles assessed in this thesis based on new core, outcrop, biostratigraphic and δ^{13}C assessments. This framework has established that cycle ix of the Lekhwair Formation most likely deposited in the Hauterivian after the Weissert event which has been dated previously at ~135 Mya. Cycle iii of the Kharaib Formation began to deposit in the Lower Barremian with the MFS of cycle iii taken to correspond to the K50 MFS at ~129 Mya and the MFS of cycle ii is likely to mark the K60 MFS at ~125.5 Mya.
The $\delta^{13}C$ of cycle ix to ii shows an upward increase from $\sim 2 \text{‰}$ to $\sim 3.5 \text{‰}$. This is compatible with established trends (Sprovieri et al., 2006; Van de Schootbrugge et al., 2000) and can be related to progressive platform drowning during the Hauterivian and Barremian. Numerous negative $\delta^{13}C$ episodes are observed at the SBs of most cycles; these events are most likely a result of volcanic activity, methane clathrate release or sulphate reduction, however further work is required to confirm this.

Deposition of the cycles of the Kharaib and Shu’aiba’ Formations occurred on a carbonate ramp in a distal open marine to proximal restricted marine depositional environment. A typical cycle for the Kharaib and Shu’aiba’ Formations commences at the SB which is represented by a hardground where *Gastrochaeonolites* (with in-situ *Lithophaga*) and *Trypanites* borings are observed. The surface is overlain by shallowing upward deposits that start with a transgressive shale of the TST that is then succeeded by pyrite and microdolomite rich mudstones and wackestones that are then overlain by the rudist and coated grain dominated HST deposits. The upper section of each cycle is rich in miliolid and rudist wackestone and packstone deposits that are thought to deposit in a lagoon, these deposits are in turn capped by the SB of the following cycle.

The 22 lithofacies identified in the TSTs and HSTs of the cycles have distinct petrophysical properties that are a reflection of their depositional environment. The TST lithofacies typically have lower porosities and permeabilities than the HST leading to the TST commonly forming non-reservoir intervals and the HST reservoir intervals; this would suggest that the location of reservoir intervals can be predicted by developing a three dimensional model of the petrophysically important depositional fabrics.

The porosity and permeability ranges of the grain dominated fabrics that typically compose the HST are larger than those of the mud dominated fabrics that dominate the TST. The variability in the petrophysical properties is thought to be a result of the depositional porosity and permeability of the grain dominated fabrics being greater, this has likely led to a more complex diagenetic history and has resulted in the pore volume and tortuosity of the pore space being altered during
burial. This is evident in the core where highly porous lithofacies of the HST are fully cemented and now compose part of the non-reservoir. The diagenetic effect on the porosity and permeability of the depositional cycles is assessed in Chapter 3 and 4.
CHAPTER 3
DIAGENETIC CONSTITUENTS OF THE LEKHWAIR, KHARAIB AND SHU’AIBA’ FORMATIONS: THEIR CHRONOLOGY, DISTRIBUTION AND BULK ISOTOPE GEOCHEMISTRY

3.1 Introduction

Changes in relative sea-level during the deposition of the Lekhwair, Kbaraib and Shu’aisba’ Formations have led to the deposition of a number of cycles. These cycles are composed of a Transgressive Systems Tract (TST) and a Highstand Systems Tract (HST) where the HST typically forms reservoir intervals and the TST non-reservoir intervals, this has resulted in Field A and B being highly cyclic successions composed of numerous reservoir and non-reservoir intervals (Chapter 2).

Diagenesis through cementation and dissolution has, however, modified the primary (depositional) petrophysical properties of each cycle and has resulted in a different layering being imposed through diagenetic modifications. The current reservoir thickness for each cycle is therefore partly controlled by diagenesis where the extent of cementation and dissolution determines the position of the reservoir tops (Section 2.4.1). Although diagenesis can greatly control the present-day petrophysical properties of the cycles very little published work deals with the diagenesis of the entire Thamama Group: most publications consider the Kbaraib Formation with the most thorough being that of Alsharhan (1990), Neilson et al.

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1 The information presented in this chapter was submitted to Abu Dhabi Company for Onshore Oil Operations for permission to publish in April 2013, but as of September 2014 permission has not been granted.
This chapter presents a detailed diagenetic investigation of the Lekhwair, Kharaib and Shu’aiba’ Formations and aims to:

1) Identify the pore types and diagenetic constituents present in both Field A and B and determine their relative chronology through petrographic analyses. The diagenetic constituents identified are then divided into three diagenetic zones based on pore fluid chemistry variations (Section 1.3): 1) marine phreatic/syn-depositional, 2) shallow burial, and 3) deep burial.

2) Quantify the cements and pore types present in the cycles of Field A and B by point counting (350 points). This allows for the distribution of macroporosity and macrocement in the cycles to be estimated.

3) Calculate the total Minus Cement Porosity (MCP) for a specific pore type to allow an understanding as to the effect that cementation has had on occluding a single pore type. The Lønøy (2006) 1 mD permeability cut-off is then applied to each of these pore types to help identify the pore types that have most significantly contributed to permeability > 1 mD.

4) Discuss the distribution of cement and pore types, as well as the origin and timing of cementation in the cemented reservoir intervals. This includes a discussion on the stratigraphic distribution of porosity and cement and compares the distribution of porosity and cement in the oil reservoirs and coeval aquifers to determine the effect of oil charge on reservoir quality.

5) Present a separate component geochemical assessment on the main pore occluding cement phases to determine if the system is relatively closed during cement precipitation. This is achieved by relating the δ\textsuperscript{13}C obtained for the cements in a reservoir interval to the δ\textsuperscript{13}C of bulk micrite for that same reservoir interval; if there is little variability between the δ\textsuperscript{13}C of the cements precipitated in that reservoir interval and the δ\textsuperscript{13}C of bulk micrite in that same reservoir it is likely that the cements are locally sourced and therefore the reservoir is thought to behave as a relatively closed system. If the system is relatively closed and the cements precipitated in equilibrium with the pore fluid, the δ\textsuperscript{18}O\textsubscript{VPDB} of the
cements can be used as an approximate temperature proxy that will help to constrain the temperature of precipitation for the cement.

3.2 Diagenesis in Field A and B

The same diagenetic constituents are observed in both Field A and B. The diagenetic constituents have been identified and grouped, based on the interpreted pore fluid chemistry variations, into the Marine Phreatic Zone (MPZ), Shallow Burial Zone (SBZ), and Deep Burial Zone (DBZ) (Section 1.3). The marine phreatic zone incorporates all the early diagenetic features observed that formed syn-depositionally and includes microdolomite, glauconite and hardground formation and continues until aragonite and high-Mg calcite stabilisation. The deposition of saddle dolomite cement marks a change in pore fluid chemistry and the onset of the deep burial zone.

A chronological overview of the cross cutting relationships identified in Field A and B is provided in Figure 3.1. This paragenetic sequence is based on the cross-cutting relationships that were observed through petrographic analyses. The main diagenetic constituents present in Field A and B and their cross cutting relationships are discussed in the following subsections.
Figure 3.1: Paragenetic sequence for Field A and B. The relative time of pore formation is presented in the lower part of the figure. The approximate time of initial oil charge is identified by the presence of oil inclusions and is marked by the red arrow, oil inclusions are then observed in subsequent calcite and dolomite cements.

3.2.1 Pore types contributing to reservoir quality

This subsection identifies the pore types present in the Lekhwair, Kharaiib and Shu’aiba’ Formations to help determine the origin and types of porosity that have, and are currently, contributing to the reservoir quality of Field A and B. The pore classification scheme is modified from Choquette and Pray (1970).
The pore classification scheme of Choquette and Pray (1970) groups moldic and intragranular porosity because they are classified as fabric selective, these pore types are also grouped with intergranular and intercrystalline porosity. However, Lucia (1983) and Lønøy (2006) have demonstrated that moldic and intragranular porosity have a different effect on permeability than intergranular and intercrystalline porosity (a lower volume of the intergranular and intercrystalline porosity is required for 1 mD permeability when compared to intragranular and moldic porosity) and thus should be grouped separately. Therefore in this assessment the pore types are not grouped but instead are identified as separate pore types.

There are many types of primary and secondary pore types within the Thamama Group. Primary pore types are those that formed during initial deposition of the sediment, whereas secondary pore types formed after initial deposition and are a product of diagenesis (Section 1.3.1, 1.3.2). The primary pore types include: intragranular and intergranular, whereas the secondary pores include: moldic, vuggy, fracture, intercrystalline and microporosity.

### 3.2.1 Depositional (primary) porosity

Intergranular porosity is defined as the porosity that is between grains (Figure 3.2) (Choquette and Pray, 1970; Radke and Mathis, 1980). All the main pore occluding diagenetic phases are observed to occlude intergranular porosity.

![Figure 3.2: a), b) Intergranular porosity in a grainstone. Non-ferroan saddle dolomite cement occludes some of the intergranular pore space in a).](image)

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Intragranular porosity is observed within fauna where it is enclosed by skeletal walls; the classification scheme of Choquette and Pray (1970) has been slightly modified so that the pores that are related to dissolution are not included. This is so that pores that are of a primary/depositional origin can be separated from those of a secondary/diagenetic origin. The distribution of intragranular porosity is therefore fauna dependant and is typically observed in the chambers of orbitolinids, miliolids and *lithocodium/bacinella* (Figure 3.3).

*Figure 3.3: Occluded and un-occluded intragranular porosity of a) A benthic foraminiferan, and b) Lithocodium/Bacinella. c) The intragranular porosity of an encrusting Lithocodium/Bacinella is free of cement; also note the presence of vuggy porosity. d) SEM image of the intragranular porosity of a benthic foraminiferan; microporosity is common throughout the surrounding matrix.*
3.2.2 Diagenetic (secondary) porosity

Moldic pores formed by partial or complete dissolution of grains, this definition is similar to that of Choquette and Pray (1970) and Lønøy (2006). Moldic porosity is most common within the dissolved internal aragonitic valve of rudists and in the dissolved tests of foraminifera (Figure 3.4a, b). The moldic porosity created by the dissolution of rudists is normally partly occluded by equant, blocky and saddle dolomite cement (Figure 3.4a, b).

![Figure 3.4: a) Moldic porosity in the internal dissolved aragonitic valve of a rudist and within the dissolved test of a foraminiferan. b) Moldic porosity of a dissolved coated grain and of a rudist; note the presence of microporosity (hazy blue) in the coated grains.](image)

Vuggy porosity is related to dissolution and is defined as non-fabric selective porosity and is commonly irregular in shape (Choquette and Pray, 1970). Vuggy porosity is commonly seen within wackestone and packstone fabrics where it is seen within the micritic matrix (Figure 3.5a, b).
Fracture porosity is defined by Choquette and Pray (1970) as porosity formed by fracturing where there has been little mutual movement of opposing blocks. Fractures typically have symmetrical margins and have a large length to width ratio. Fracture porosity is rare and most commonly observed within the TST lithofacies, this type of porosity may be due to the decompaction processes that occur when a sample is extracted from depth and therefore has been disregarded in this thesis.

Intercrystalline porosity is the porosity typically associated with the dolomitisation process (Choquette and Pray, 1970). Intercrystalline porosity is normally observed between dolomite microrhombs within burrows and borings, and is of a very low volume.

Microporosity is defined by Cantrell and Hagerty (1999) as all pores that are < 10 µm in diameter, microporosity is therefore not clearly distinguishable for most standard petrographic microscopes. Here, microporosity is typically observed as a blue haze between micrite crystals which is due to the infiltration of blue-dyed epoxy (Figure 3.6a, b, 3.3d, 3.4b). Diagenetic kaolinite has a high volume of microporosity.
3.2.2 Diagenetic features and their cross cutting relationships

3.2.2.1 Marine Phreatic Zone (MPZ)

Marine phreatic features are subject to be formed under marine conditions, during the development of the carbonate ramp, from pore fluids similar to Early Cretaceous seawater. The cements that formed in the MPZ consist of fluid inclusion rich microdolomite, pyrite, glauconite, syntaxial and fringing calcite cements.

1) Microdolomite occurs as euhedral to subhedral microrhombs that are ~50 µm in size. These rhombs typically have a cloudy core and a clear outer rim that is associated with an increased abundance of fluid inclusions in the inner core (Figure 3.7a). Non-ferroan and ferroan cement zones are observed in the microdolomite cements; the ferroan zones cannot be correlated between samples. Under the cathodoluminoscope the inner core typically appears dull black and the outer rim is normally bright orange (Figure 3.7b).
Microdolomite is only observed within muddy lithofacies and is seen to primarily replace the soft sediment that fills the *Thalassinoides* burrows of the firmground and the *Lithophaga* borings of the hardground (Figure 3.7a). Microdolomite is also distributed throughout the matrix and in some samples seen to increase in abundance over 50 mm.

2) Pyrite commonly occurs as brassy coloured crystals ~60 µm in size and appears dull to non-luminescent under the cathodoluminoscope. Pyrite is commonly dispersed throughout the mudstone and wackestone lithofacies of the TST as frambooidal, anhedral and euhedral pyrite (Figure 3.8a). When borings are observed a pyrite “halo” is seen ~ 2 mm from the boring, this is composed of multiple clusters of anhedral pyrite (Figure 3.8b).
Figure 3.8: a) Pyrite is typically dispersed throughout the matrix and in grains. b) Pyrite “haloes” are commonly seen close to borings; in the photograph the micrite initially filling the boring is now replaced by microdolomite.

3) Syntaxial cements are in optical continuity with the host grain (typically an echinoderm fragment) and are composed of a single crystal that is elongate in the c-axis. The cements are most commonly observed in intergranular primary porosity and are observed to grow competitively with microdolomite (Figure 3.9a) and saddle dolomite (Figure 3.9b). Syntaxial calcite can be ferroan and non-ferroan with a single spar seen to alternate between non-ferroan and ferroan zones (Figure 3.9b). The ferroan zones cannot be correlated between samples. Oil inclusions are observed in these cements in both Field A and B.

Figure 3.9: a) Syntaxial cements growing competitively with microdolomite. b) A syntaxial cement with a non-ferroan core (stained pink) and ferroan rim (stained purple), that is growing competitively with saddle dolomite.
4) Bladed fringing calcite cement is only observed within primary porosity (inter- and intra-granular porosity), and is < 100 μm in length and most easily observed on large uniform grains (Figure 3.10). The fringing cement is not equidimensional and not fibrous, with individual crystals increasing in width along their length and exhibiting pyramid like terminations. The thin peri-granular calcite cements irregularly rim the grains leading to partial primary pore space occlusion.

![Figure 3.10: Fringing calcite cement forming a peri-granular rim.](image)

5) Dissolution of aragonite and high-Mg calcite is most commonly observed in the reservoir intervals where it has led to the formation of a large volume of moldic porosity. Dissolution of the internal aragonitic valve of rudists is relatively common, but the outer low-Mg calcite valve is largely unaffected by this dissolution event and still displays internal growth bands (Figure 3.11a). The moldic porosity that formed due to dissolution can be preserved until present day (Figure 3.11a), or can be occluded by equant, blocky and saddle dolomite cement (Figure 3.11b).
3.2.2.2 Shallow Marine Burial Zone (SMBZ)

6) Equant calcite cements are typically equidimensional, < 300 \( \mu \text{m} \) in size, pore lining and coarsen into the pore centre. This cement type has also been recognised by Alsharhan (1990) where it is described as a fine to medium equant cement that is frequently seen on the surface of allochems. Equant cements occur as isolated spar or as massive mosaics (Figure 3.12a) and are seen to post-date fringing cement (Figure 3.10), this is also observed by Alsharhan (1990). They are observed within both primary and secondary pore space and can lead to complete pore occlusion.

Potassium ferricyanide staining has revealed that equant calcite cement can either be ferroan or non-ferroan, with ferroan equant calcite cement being most commonly observed in more muddy facies. A single equant spar can alternate between non-ferroan and ferroan calcite, leading to zonation, these ferroan zones cannot be correlated between samples (Figure 3.12b).
Figure 3.12: a) Equant calcite mosaic fully occluding the intergranular pore space. b) In this figure the core of the equant calcite cement is ferroan and the external rim non-ferroan; however, the location of the ferroan and non-ferroan cement zones is highly variable and cannot be correlated between samples.

7) Stylolites form parallel to bedding, indicating dominant vertical lithostatic stress. They are commonly observed between the MFS and HST (Figure 2.16), where there is a change in rock texture and within the fine grained homogeneous muddy deposits of the TST, where they form swarms (Figure 3.13a). The porosity close to the seams is commonly filled with blocky calcite, saddle dolomite and anhydrite which results in a noticeable reduction in porosity. Grain sutures are observed at grain contacts when there has been no early protective cement rim precipitated (Figure 3.13b).

Figure 3.13: a) Wispy stylolite seams are typically observed within the TSTs of the cycles. b) Sutured grain contacts of coated grains.
8) Fractures are relatively uncommon in the samples examined, except when associated with stylolites (Figure 3.14). Fractures can be 30 µm to several centimetres wide and cross cut stabilised or lithified sediment, palaeo-porosity and pore occluding cement; they are not observed to cross cut skeletal grains. Typically fracture walls are matching, indicative of a lack of both dissolution and mechanical erosion along the fracture walls. The smaller fractures (< 0.5 mm in width) are usually filled with equant mosaic calcite cement or blocky calcite cement (Figure 3.14), with the large (> 0.5 mm in width) fractures typically being occluded by saddle dolomite. The fractures are most commonly seen within the muddier lithofacies of the TST.

![Figure 3.14: Cross cutting relationships between stylolites and veins (fracture fills) which suggest both stylolites and veins formed at a similar time.](image)

3.2.2.3 Deep Burial Zone (DBZ)

9) Blocky calcite cement is composed of large (> 300 µm) coarse grained crystals that have distinct crystal boundaries. This cement phase is equivalent to the coarse sparry calcite of Alsharhan (1990), where the cement phase is observed to overlie the finer equant calcite cement phase. The cements are observed to grow competitively with saddle dolomite in some moldic pores (Figure 3.15a) and with syntaxial calcite
cement (Figure 3.15b). Blocky calcite cements are observed within all pore types, and can fully occlude porosity (Figure 3.15b). The cements can be ferroan or non-ferroan and can have alternating ferroan and non-ferroan cement zones.

Figure 3.15: a) Non-ferroan blocky calcite cement and ferroan saddle dolomite cement occluding a moldic pore of a rudist. b) Blocky calcite cement competitively growing with syntaxial calcite cement and saddle dolomite cement.

10) Saddle dolomite crystals are typically > 200 µm in size and are of a euhedral to anhedral morphology (Figure 3.16a). Fluid inclusions (both water and oil) are observed within some saddle dolomite cements and have led to a cloudy appearance. Saddle dolomite is observed in every macropore type and occurs principally as cement but is observed to replace blocky calcite and equant calcite cements in some intergranular pore space (Figure 3.16b). However, in some pores saddle dolomite appears to grow competitively with blocky calcite cement (Figure 3.16c). The surfaces of some saddle dolomite crystals are etched and kaolinite is now seen to precipitate on these etched surfaces (Figure 3.16d).

Under transmitted light dolomite crystals are brown to grey, but exhibit a dull orange, red luminescence or are non-luminescent under the cathodoluminoscope (Figure 3.16d). The alternation between the luminescent and non-luminescent zones allows for the spear head form of saddle dolomite to be easily observed (Radke and Mathis, 1980) (Figure 3.16d). The orange to red luminescent zones and the non-luminescent zones are in agreement with non-ferroan and ferroan saddle dolomite respectively. Ferroan dolomite is typically observed around muddier facies that tend
to contain *Lithocodium/Bacinella* nodules. The ferroan bands are of a variable thickness, and cannot be correlated between samples.

Figure 3.16: a) SEM image of saddle dolomite. b) Saddle dolomite replacing calcite cement. c) Blocky calcite cement, saddle dolomite and kaolinite occluding an intragranular pore within lithocodium/bacinella. d) The alternating luminescence of spear head saddle dolomite; note that kaolinite now rests on the etched surface of saddle dolomite.

Electron Back Scatter Diffraction (EBSD) can be used to obtain the orientation of the axes for a crystal (Figure 3.17), in order help understand how the cement has occluded a pore space (Section 1.7.5). The EBSD technique shows that the crystal orientation of saddle dolomite is not constant during growth but changes by up to 12°; with the gradients in orientation being mainly perpendicular to the pore wall. The c-axis is typically parallel to the cavity wall and often fan-shaped into the pore centre.
Figure 3.17: a) Unprocessed EBSD map of saddle dolomite cement. The change in colour is related to the orientation of the \{0001\}/c axis, see d) for the key. b) Processed EBSD map with the interpreted crystal boundaries marked by a turquoise outline. c) Stereonet for the main growth axis of saddle dolomite.

11) Late stage, DBZ, dissolution of low-Mg calcite is widespread and typically observed in the reservoir and non-reservoir intervals of the succession. The dissolution event is related to the etching of blocky calcite and saddle dolomite cement (Figure 3.18a, b, 3.16d). This event has also greatly increased the porosity of both the reservoirs and non-reservoirs by dissolving the micritic matrix, leading to
the development of a large volume of vuggy porosity (Figure 3.18c, d). Kaolinite is then observed to precipitate in this porosity (Figure 3.18d) and on the etched surfaces of calcite (Figure 3.18a) and saddle dolomite cement (Figure 3.16d).

12) Kaolinite is randomly observed throughout both fields and only precipitated on etched surfaces or in the porosity generated by the previous dissolution event. It is white and soft in hand sample and is readily recognised through SEM analysis due to its hexagonal lattice (Figure 3.19a) and by using cathodoluminescence microscopy because it displays a bright blue luminescence that fades over time (Figure 3.19b). Kaolinite is observed to contain a significant proportion of microporosity; this can be observed by its blue hazy appearance (Figure 3.18a, d).

Figure 3.18: a) Etched surfaces of calcite cement, with kaolinite then precipitating on these surfaces. b) Dissolved micrite and etched calcite spar surfaces with a large volume of vuggy porosity. c), d) Vuggy porosity associated with the dissolution of the micritic matrix. Kaolinite then precipitates in this pore space, d).

Figure 3.19: a) Kaolinite in hand sample with its typical blue hazy appearance. b) Kaolinite displaying bright blue luminescence that fades over time.
Figure 3.19: a) SEM image showing the hexagonal appearance of kaolinite. b) Kaolinite is readily identifiable via cathodoluminescence because of its bright blue luminescence; this colour fades with continued exposure to the electron beam.

13) Anhydrite appears as needles with a high birefringence (Figure 3.20a, b) and is only associated with pressure solution seams within reservoir intervals ix-iii.

Figure 3.20: a) Anhydrite in proximity to a stylolite. b) Cladocoropsis is now partially displaced by stylolitisation, and Fe-saddle dolomite and anhydrite are observed in the porosity of the stylolite.

3.2.3 Interpretation and discussion of cross cutting relationships

Through petrographic observations and the identification of cross cutting relationships the diagenesis of both Field A and B can be discussed in terms of two main stages that formed over three diagenetic zones (Figure 3.1). The first stage begins at the sediment-water interface in the syn-depositional environment with the development of micritic envelopes, hardgrounds and borings. Borings are thought to
continue to develop in the sediment to depths of 4-5 cm (Dickson et al., 2008) and are then infilled with micrite, which is in turn replaced by microdolomite (Figure 3.7a).

The microdolomite crystals typically have an internal core that is non-luminescent and rich in fluid inclusions (Figure 3.7). Rapid growth of the cement, possibly due to high fluid flux rates, would lead to the incorporation of fluid inclusions in the core of microdolomite, therefore the internal core is likely have formed in the marine environment where high fluid flow rates are likely to have been present. This is also suggested by the non-luminescent cement zone because modern marine environments are oxidizing which would preclude the incorporation of Mn$^{2+}$ and Fe$^{2+}$ in the cement and would lead to the formation of a non-luminescent core (Moore, 1989). This may suggest that the non-luminescent core, that is rich in fluid inclusions, began to form in a marine phreatic environment.

A pyrite halo is typically observed around the borings (Figure 3.8b). Pyrite is thought to result from organic metabolism and anaerobic bacterial sulphate reduction (Berner, 1970; Wilkin and Arthur, 2001). Bacterial sulphate reduction will result in the formation of H$_2$S (Albarède, 2003), which in turn will react with iron minerals leading to the development of pyrite. Because pyrite is observed around borings it is thought to be a subsequent diagenetic phase, pyrite is therefore unlikely to develop at the sediment water interface but at a depth ≥ 4-5 cm (Dickson et al., 2008).

Syntaxial cements are observed to grow competitively with microdolomite, (Figure 3.9a) and also with blocky calcite and saddle dolomite (Figure 3.9b), implying that cementation began in the marine phreatic zone and continued into the deep burial zone, this is also suggested by Cox et al. (2010).

Fringing cements are observed within intergranular pores (Figure 3.10) and within the intragranular pores of _lithocodium/bacinella_ and of benthic foraminifera (Figure 3.3a). This would suggest that precipitation occurred in the marine phreatic zone before the pores became isolated during burial, this is also suggested by Flügel (2004). However, fringing cements are not observed within the inner dissolved
aragonitic valve of rudists, suggesting precipitation prior to the dissolution of aragonite.

The Early Cretaceous was a period with calcite seas (Section 1.4). However, because aragonitic and high-Mg calcite biota are common (rudists, dasycladaceans and gastropods) it is thought that mechanical erosion and bio-erosion led to the development of a highly heterogeneous mud (Deville de Periere et al., 2011). Therefore, because there is a group of different minerals with differing solubilities (aragonite, low-Mg calcite, high-Mg calcite) the main process will not be low-Mg calcite dissolution but instead that of more soluble aragonite and high-Mg calcite (Scholle et al., 1989; Tucker et al., 1991). The important point here is that in this process low-Mg calcite will not be dissolved because initially Early Cretaceous seawater is favourable to low-Mg calcite precipitation (Section 1.4) and oversaturation can be achieved and maintained by the dissolution of aragonite and high-Mg calcite (Lambert et al., 2006).

Currently growth bands are observed in the exterior (low-Mg calcite) valve of rudists, however either the internal aragonitic valve is dissolved to leave an un-occluded moldic pore or no growth bands are present in the cements, which suggests that the aragonitic valve has been dissolved and diagenetic cements are then precipitated in this pore space (Figure 3.11). The dissolution of aragonite and high-Mg calcite corresponds to the main stage of carbonate stabilisation and is thought to lead to the formation of moldic pores by the dissolution of biota and to vuggy pores in the matrix by the dissolution of bio-eroded aragonite and high-Mg calcite mud. Microporosity is also thought to be generated at this time by the re-precipitation of micrite as low-Mg calcite microrhombs (Lambert et al., 2006).

Pressure-solution seams can form during shallow burial (10’s metre) in muddy lithofacies (Mazzullo and Harris, 1992), therefore the onset of stylolitisation has been placed in the SBZ. Fractures are thought to be related to the formation of stylolites (Figure 3.14) (Alsharhan, 1990; Alsharhan and Sadd, 2000; Moshier, 1989) and therefore are also thought to start to develop in the SBZ. Pressure solution has been previously studied within the Lower Cretaceous Thamama Group and is
thought to provide the majority of solutes needed for burial cementation (Alsharhan and Sadd, 2000; Koepnick, 1987; Oswald et al., 1995).

Bitumen is primarily found along stylolites, but also within the matrix and on grain surfaces. Bitumen is observed within oil reservoirs and aquifers and likely formed by sulphate reduction of organic matter (Machel, 1987). This organic matter can be derived from the mudrier facies directly, or by degradation of hydrocarbons along stylolites. Bitumen can then be concentrated along stylolites by either compression of the matrix and concentration along the pressure solution seam or by in-situ degradation of organic matter. The concentration of bitumen along stylolites will aid in the development of impermeable layers and therefore with field compartmentalisation.

Equant and blocky calcite cements precipitated after the formation of fringing calcite cement (Figure 3.10, 3.15b) and can develop in the near-surface meteoric as well as in burial environments (Flügel, 2004). Blocky cement is thought to grow competitively with and postdate the precipitation of syntaxial cement (Figure 3.15b) and is also observed to be replaced by, and to grow competitively with, saddle dolomite suggesting formation prior to and synchronously with saddle dolomite (3.15a, 3.16b). Therefore it is interpreted here that calcite cements continue to form from the marine phreatic zone (fringing cement) into the deep burial zone (blocky calcite cement) because saddle dolomite cement is characteristic of the deep burial zone.

The precipitation of saddle dolomite is characteristic of high temperature burial diagenesis (Gregg and Sibley, 1984) and precipitation is thought to begin at ~60°C (Machel, 1987; Radke and Mathis, 1980). Using $\delta^{18}O_{VPDB}$ data, Alsharhan and Williams (1987) suggest that the saddle dolomite cements in the Shu’aiba’ Formation precipitated between 67°C and 112°C; 67°C is therefore taken to mark the onset of the DBZ.

In some cases non-ferroan saddle dolomite is observed to precipitate after ferroan dolomite and in other cases the reverse, this suggests that local changes in pore fluid chemistry caused the precipitation of ferroan and non-ferroan dolomite
cement (Figure 3.16d). This is similar to calcite cements where alternating ferroan and non-ferroan cement zones are observed that cannot be correlated between samples (Figure 3.9b). Oil inclusions are observed within saddle dolomite cements and within syntaxial, equant and blocky calcite cements, implying their formation during and after oil charge, therefore the Fe$^{2+}$ may be derived from hydrocarbon charge or may be a result of dewatering of the shales and mudstones in the TST.

The formation of fringing, syntaxial, equant and blocky calcite cement as well as saddle dolomite was not accompanied by massive dissolution of low-Mg calcite. Furthermore, core and outcrop descriptions present no evidence for strong meteoric diagenesis (no enlarged vugs or traces of karstification etc.). Therefore during Stage 1, cementation is likely to have occurred from porewater that was always supersaturated with respect to low-Mg calcite. The composition of this water is thought to be equal to, or slightly evolved from, Early Cretaceous low-Mg bearing seawater.

The entry of acidic fluid into the fields marks the end of Stage 1, where the formation water was supersaturated with respect to low-Mg calcite, to Stage 2 where the formation water was undersaturated with respect to low-Mg calcite (Figure 3.1). This event led to the etching of the surfaces of calcite cement and saddle dolomite and led to extensive dissolution and to the formation of vuggy porosity (Figure 3.5). The porosity generated by this dissolution event is typically located in the micrite matrix where large vuggy pores are formed (Figure 3.18).

A late stage generation of a significant volume of microporosity by the partial dissolution of low-Mg micrite has been suggested by Lambert et al. (2006). This is most likely related to the low-Mg dissolution event identified here. In the study by Lambert et al. (2006) late stage dissolution is identified by observing that euhedral spar crystals engulf rounded micrite spar. When these rounded crystals became detached from the spar the surface revealed is a mold with a plane face which suggests the spar grew around a euhedral micrite spar, however subsequent dissolution rounded the micrite spar and led to the development of microporosity. The samples where rounded micrite is observed typically have a better reservoir quality, suggesting that the rounding event is important in enhancing porosity. This
interpretation disagrees with Budd (1989) who attributes the generation of microporosity to meteoric diagenesis, but agrees with the observations of Moshier (1989).

After this dissolution event kaolinite is then precipitated, it is therefore unsurprising that the mineral rests on the etched surfaces of the earlier saddle dolomite and calcite cements and partially occludes the newly generated vuggy porosity. The precipitation of kaolinite after a dissolution event has been observed in several other carbonate reservoirs, with precipitation of kaolinite typically thought to occur at temperatures $> 90^\circ \text{C}$ (Neilson and Oxtoby, 2008). The precipitation of kaolinite will result in a net loss of protons and increase the pH of the pore fluid (Albarède, 2003), which will help to return the system to a near neutral pH.

Kaolinite requires a source of aluminium in order to form. The transport of aluminium ions in diagenetic water is very limited but aluminium can complex with organic compounds, it has therefore been suggested that the presence of kaolinite may be linked to oil charge (Maliva et al., 1999). The source of the low-Mg dissolution event may therefore be related to the migration of organic acids prior to hydrocarbon charge, with kaolinite precipitation being related to the oil charging event. However, this interpretation contradicts a previous observation where oil inclusions are observed within the cements prior to this dissolution event. Therefore this hydrocarbon charging event cannot have led to the late stage dissolution event. It is therefore suggested that dissolution occurred due to the migration of acidic acids prior to a second hydrocarbon charging event; this will be discussed in Chapter 5 (Section 5.6.3).

Anhydrite is only observed in proximity to stylolites and there were no cross cutting relationships observed (Figure 3.20), therefore the placement of anhydrite in the paragenetic sequence is uncertain but it is thought to be one of the final cements to form. Neilson and Oxtoby (2008) discussed the origin of anhydrite in carbonate reservoirs and suggested that precipitation typically occurs at temperatures of 85-160 $^\circ \text{C}$ and from fluids in the salinity range of 15-25 wt. % NaCl. Therefore, in Field A and B anhydrite may have formed in the deep burial zone (when the reservoirs were
at their highest temperature) with precipitation being related to the upward migration of saline waters derived from the Hith halite or from anhydrite present in the Habshan Formation (Alsharhan and Kendall, 1991). These waters are likely to have primarily migrated along stylolites, which will allow for the preferential precipitation of anhydrite along these seams.

3.3 Diagenetic Controls on Reservoir Quality: A Quantitative Study

3.3.1 Main pore occluding cements

The distribution of cement and porosity in the nine cycles of the Lekhwair, Kharaiib and Shu’aiba’ Formations was estimated by undertaking point count analyses (N = 350). This assessment was undertaken to determine the main pore occluding cement phase in Field A and Field B and to quantify the effect of cementation on porosity.

A plot showing the total volume of pore occluding cement (calcite, saddle dolomite and other cements) in a thin section, in relation to the total remaining macroporosity can be used to help identify the main pore occluding cement (Figure 3.21). It is possible to observe in these plots that for both Field A and B calcite cement is the main pore occluding cement type, with there typically being a greater volume of calcite cement present in Field B in comparison to Field A.
Figure 3.21: Main pore occluding cements of Field A and Field B. See Appendix 4 for the data values.
3.3.2 Minus Cement Porosity (MCP)

The effects of compaction and cementation on the petrophysical properties of a sample are difficult to separate, but they both reduce porosity. In order to determine the effect of cementation on occluding pore space the MCP was calculated. The MCP corresponds to the sum of the remaining macroporosity and the volume of cement that occludes that pore type. The MCP for intergranular porosity corresponds to the primary intergranular porosity (of depositional origin), minus the reduction of porosity due to compaction. The plot of the intergranular cements (%) vs. the intergranular MCP (%), allows the direct reading, from a single plot, of the main three parameters characterizing reservoir quality, namely the porosity loss due to compaction, the cementation porosity loss, and the present-day porosity. Although this plot can be used to understand the effect of mechanical compaction on intergranular porosity because the depositional porosity can be estimated (Enos and Sawatsky, 1981), the effect of mechanical compaction on other pore types is uncertain because either the pore type (moldic and vuggy) was not present at deposition, or the initial volume is unknown (intrgranular).

Lønøy (2006) determined the minimum macroporosity required for 1 mD permeability in multiple carbonate pore types, this value is known as the porosity cut-off (Section 1.3.4). The porosity cut-offs have $R^2$ values of between 0.79 and 0.96. These values are included on Figure 3.22 and on subsequent MCP plots to help identify the pore types which are most probably contributing to flow in the depositional fabric. This is an estimation as to which macropore types allow for flow and does not include the microporosity commonly present in the muddier fabrics. This microporosity may connect the different pore types and improve permeability.

The initial volume of intergranular macroporosity in a sample is determined by the initial depositional environment. This is evident in Figure 3.22a where the intergranular MCP of the wackestone and mud-supported packstone fabrics is low (< 10 %) whereas, in the grain supported packstones and the grainstones a higher MCP is present.
Although mechanical compaction has undoubtedly reduced the intergranular porosity in the fabrics, the remaining MCP (y-axis) of some of these samples is above the permeability cutoff of Lønøy (2006) (shown by the brown lines). This would imply that although mechanical compaction has reduced the MCP it has not reduced the volume remaining to below the porosity cut-off. However cementation (x-axis) has typically occluded some of this intergranular MCP and has reduced the porosity to below the permeability cut-off of Lønøy (2006).
Figure 3.22 (previous page): Cement volume vs. MCP for a) intergranular, b) intragranular, c) vuggy, and d) moldic pore types for all the reservoir intervals of Field A and B. The Lønøy (2006) porosity cut-off values are shown by the brown lines.

The burial depth at which total porosity occlusion occurred can be estimated by translating the intergranular MCP values from Figure 3.22a onto a typical porosity vs. depth curve (Figure 3.23) for carbonates (Schmoker and Halley, 1982). Assuming that original porosity was in the range between 45 and 50 % (Enos and Sawatsky, 1981), the most elevated MCP value of ~ 41 % is found in a sample in the non-reservoir of cycle iv and may suggest a maximum burial depth of < 200 m for the completion of compaction, and hence of cementation (black arrow; Figure 3.23). This depth estimation should not be considered as a precise value, but rather as an order of magnitude because the shape of compaction curves will depend on the nature of the grainstone or grain supported packstone as well as many other factors (e.g. type and shape of the grains). Assuming a lower initial porosity this would lead to an even lower depth for the completion of cementation. The intergranular MCP in some samples has followed the porosity vs. depth trend closely suggesting that compaction has continued during burial.
Figure 3.23: a) MCP vs. depth curve for the intergranular pore types present in the grain-supported packstones and grainstones of Field A and B. These volumes have been related to the shallow water carbonate compaction curve of Schmoker and Halley (1982), with their standard error of 5.1 % marked by the dashed lines. The thin section sample marked by the asterisk in a) is shown in b).
The intragranular MCP (Figure 3.22b) is determined by the initial fauna present (Section 3.2.1), with a larger intragranular MCP for the the muddier fabrics than for the grain-dominated fabrics. Lithofacies dominated by *Lithocodium/Bacinella* typically have a large volume of intragranular porosity, this porosity has likely allowed for flow >1 mD in some reservoirs. However, some intragranular pores have remained un-occluded by cement, this would suggest that the pores may not contribute to effective porosity and therefore do not contribute to flow (Figure 3.3c).

The remaining vuggy porosity is on occasion greater than the porosity threshold (Lønøy, 2006), suggesting that this pore type may contribute to flow (Figure 3.22c). There is no clear relationship between depositional fabric and the volume of vuggy pore space remaining, with some wackestone fabrics having a vuggy porosity > 6.2 % and therefore above the porosity threshold of Lønøy (2006).

Moldic porosity is commonly observed within most depositional fabrics, but there is typically a greater volume within the grain supported fabrics than in the mud supported fabrics (Figure 3.22d). Moldic porosity can allow for flow in the grain-supported packstones and the grainstone fabrics of Field B, however moldic porosity in Field A is not thought to contribute to permeability > 1mD, although through connections with other pore types moldic porosity may contribute to flow.

Therefore, in the reservoir intervals, the macropore types that are most likely to contribute to flow in the more mud rich depositional fabrics are the intragranular and vuggy pores. Whereas in the grain dominated fabrics the pore types that are most likely to contribute to flow are the intergranular, vuggy and moldic pores (Figure 3.22).

The porosity present in all the non-reservoir intervals is significantly below the minimum porosity required for flow in all pore types (Figure 3.24a-d); with the exception of one sample. Cementation in these non-reservoir intervals has led to significant porosity reduction for all pore types.
3.3.3 Distribution of pore and cement types

3.3.3.1 Stratigraphic distribution of cement and porosity

Calcite cement is the main pore occluding cement phase in both Field A and B (Figure 3.21). There are four calcite cement phases that have led to porosity occlusion these are: fringing, syntaxial, equant and blocky calcite cements. The stratigraphic distribution of porosity and calcite cement is shown for Field A and B in Figures 3.25-3.27 and Figures 3.28-3.32 respectively, in relation to the cycle assessed and to whether the sample is from a reservoir or non-reservoir interval (see Appendix 4 for the data values).

Fringing and syntaxial cements only occlude primary porosity and are relatively minor constituents of the total volume of calcite precipitated (Figures 3.25-3.32), with equant and blocky calcite cements being the main pore occluding cements identified.

In both fields there is a similar volume of primary MCP in all reservoir intervals, however currently there is a greater volume of intergranular porosity in the stratigraphically younger reservoir intervals than in the older reservoir intervals (Figures 3.26, 3.29, 3.30, and 3.31). This is coupled with a decline in total intragranular porosity into the younger reservoir intervals. There is a similar volume of secondary/diagenetic vuggy MCP in all reservoir intervals in both fields. However, there is a greater volume of moldic porosity in the younger reservoir intervals.

Therefore, the reservoir intervals are primarily associated with lithofacies that have high depositional porosities (Section 2.3.2) which have then been enhanced by...
the development of secondary porosity. This is in contrast to the non-reservoir intervals of both Field A and B which typically have a low primary and secondary MCP.

Wells 1B (Figure 3.28) and 2B (Figure 3.29) of Field B have the largest sample coverage of any well. In these wells it is possible to observe that the total volume of blocky calcite cement is typically similar in reservoir intervals viii and v-i, when samples with a similar MCP are compared. It is also possible to observe that reservoir intervals v and iv typically have a greater volume of pore occluding cement (Figure 3.30) than intervals iii, ii and i. Only one sample was obtained from reservoir vii and vi and so no conclusions can be drawn.

**Figure 3.25: The distribution of pore and cement types in relation to the total MCP in Well 1A, Field A. The primary MCP is the sum of all the cement occluding the primary porosity and the remaining primary porosity in the sample.**
Figure 3.26: The distribution of pore and cement types in relation to the total MCP in Well 2A, Field A.

Figure 3.27: The distribution of pore and cement types in relation to the total MCP in Well 3A, Field A.
Figure 3.28: The distribution of pore and cement types in relation to the total MCP in Well 1B, Field B.
Figure 3.29: The distribution of pore and cement types in relation to the total MCP in Well 2B, Field B.

Figure 3.30: The distribution of pore and cement types in relation to the total MCP in Well 3B, Field B.
Figure 3.31: The distribution of pore and cement types in relation to the total MCP in oil Well 4B, Field B.
3.3.3.2 The effect of oil charge on preserving reservoir quality

Oil charge is thought to stop or retard cementation in carbonate reservoirs, thereby preserving porosity (Cox et al., 2010; Feazel and Schatzinger, 1985). The following subsection presents data for coeval oil bearing reservoirs and aquifers in Field A and B to help determine whether oil charge has preserved porosity in these fields.

Generally, when similar depositional fabrics are compared, there is a lower intergranular porosity in the aquifers of both Field A and B, due to a greater volume of cement precipitating in the intergranular pores of the aquifers than in the coeval oil bearing reservoirs (Figure 3.33). This has resulted in a higher remaining porosity in the oil bearing reservoir intervals (Figure 3.34). This relationship is observed for all depositional fabrics but is most clearly seen in the grainstone and grain-supported packstone fabrics, where a higher intergranular MCP is present (Figure 3.33a, b).
Figure 3.33: Cement vs Intergranular MCP for the coeval reservoir intervals of a) reservoir i and ii in Field A and b) reservoir i in Field B. The depth difference between coeval reservoir intervals in both Field A and Field B is < 200 m, so the difference in porosity is unlikely because of changes to lithostatic pressure.
Secondary porosity is significant in both Field A and Field B and is likely to contribute to flow $> 1 \text{ mD}$ in the reservoirs of both fields (Figure 3.35). There is a similar total MCP and remaining volume of secondary porosity (vuggy and moldic) in comparable depositional fabrics in the coeval aquifers and oil reservoirs of both Field A and B (Figure 3.35).
The calcite cements that have led to the loss in porosity in the oil reservoirs and coeval aquifers include fringing, syntaxial, equant and blocky calcite cements (Section 3.3.3.1) (see Appendix 4 for the data values). Fringing and syntaxial cements are of a relatively minor volume (< 3 %) in all samples, however equant and blocky calcite cements have led to significant porosity occlusion in both Field A and B. In Field A there is typically a lower volume of equant cement in the oil bearing reservoirs of Well 1A (< 13 %) than in the coeval aquifers of Well 3A (< 27 %), and also a lower volume of blocky calcite cement in the oil bearing reservoir intervals (< 15 %) than in the coeval aquifer (< 55 %). In Field B there is also typically a lower volume of equant calcite cement (< 14 %; although one sample has 31 %) in oil bearing reservoir interval i (Wells 1B, 2B) than in the coeval aquifer (< 23 %) (Well 5B). Blocky cements are also of a lower volume in the oil bearing reservoir intervals of Field B (< 8%) than in the aquifers (< 30 %).

3.4 Separate Component Analysis

Now that the control of cementation on porosity has been established, a separate component stable isotope analysis ($\delta^{13}$C and $\delta^{18}$O$_{VPDB}$) of the main pore-occluding cements will help to constrain the origin of the cement and to help determine if the reservoir intervals are relatively closed or open systems. The $\delta^{18}$O$_{VPDB}$ values for the cements, assuming a decreasing $\delta^{18}$O$_{VPDB}$ with increasing burial depth and temperature, are also used to help validate the paragenetic sequence (Figure 3.1).

3.4.1 Micrite

3.4.1.1 Results

The stable isotope values for reservoir and non-reservoir micrites are shown on Figure 3.36a and Figure 3.36b respectively. The data obtained for Field A is new
to this thesis as well as some data for Well 1B and 4B of Field B, however the data for Field B has been integrated with that of Cox (2010); see Appendix 2 for the data values and their origin.

The $\delta^{13}$C values for the reservoirs are between 1.0‰ to 4.5‰, the $\delta^{13}$C of the non-reservoir intervals are from 0‰ to 4.5‰. Marine $\delta^{13}$C values for the Hauterivian-Aptian are between 0‰ and 4‰; these values are in good agreement (Section 2.3.3).

$\delta^{18}$O$_{VPDB}$ values for micrite range between ~-8‰ and ~-2‰ for the reservoirs and non-reservoirs; those samples with $\delta^{18}$O$_{VPDB}$ values below -4‰ are of diagenetic origin. The youngest reservoir and non-reservoir intervals (iii-i) have on average more reduced $\delta^{18}$O$_{VPDB}$ values than the deeper intervals (ix-iv).

Sulphate reduction has been suggested as a possible cause for the $\delta^{13}$C negative episodes that are observed around the SBs of each cycle (Dickson et al., 2008) (Section 2.3.3). Pyrite is thought to be a by-product of sulphate reduction (Berner, 1970, 1984; Dickson et al., 2008) therefore the samples where pyrite is observed have been marked on Figure 3.36. These samples are not significantly different to the $\delta^{13}$C of other bulk micrite samples for that same interval.
Figure 3.36: $\delta^{18}O_{VPDB}$ vs. $\delta^{13}C$ for a) the reservoir intervals and b) the non-reservoir intervals of Field A and B. The average $\delta^{18}O_{VPDB}$ and $\delta^{13}C$ composition for each reservoir and non-reservoir is also presented with standard error. The box represents the postulated ranges for Lower Cretaceous carbonates and is in good agreement with Neilson et al. (1998), Alsharhan et al. (2000) and Vahrenkamp (2010).
3.4.1.2 Interpretation

The $\delta^{13}$C values for micrite within the reservoir and non-reservoir intervals are within the range for that of Early Cretaceous seawater (Section 2.3.3). However, in Section 2.3.3 it was observed that negative $\delta^{13}$C excursions are observed at the SB of each cycle, the cause for these excursions is equivocal. To determine whether these excursions are related to the effect of sulphate reduction (as suggested in Section 2.3.3) the samples where pyrite, a by-product of sulphate reduction, is observed are marked in Figure 3.36.

The micrite samples that were taken in proximity to pyrite are not significantly different to that of other micrite samples from that same reservoir interval. This would suggest that sulphate reduction is unlikely to have caused the reduced $\delta^{13}$C values observed in proximity to the SB of each cycle.

Sub-aerial exposure can cause reduced $\delta^{13}$C values in proximity to the SBs through the incorporation of soil derived CO$_2$ during recrystallization (Allan and Matthews, 1982; Dickson and Saller, 2006; James and Choquette, 1984). This process is unlikely to have affected the cycles assessed in this thesis, because no further evidence to suggest subaerial exposure was obtained; such as the presence of karst or pedogenic surfaces in the core or outcrop, and no petrographic evidence was identified (e.g. meniscus cements). Therefore it is likely that the $\delta^{13}$C of bulk micrite does accurately record the composition of Early Cretaceous seawater (Section 2.3.3). The reduced $\delta^{13}$C episodes observed in Section 2.3.3 may therefore be related to either volcanic activity (possibly related to the opening of the South Atlantic (Al-Fares et al., 1998)) or methane clathrate release which will reduce the $\delta^{13}$C of Early Cretaceous seawater.

The more negative $\delta^{18}$O$_{VPDB}$ values in the younger reservoir and non-reservoir intervals are the opposite of what is expected. This is because if the $\delta^{18}$O$_{VPDB}$ values are a result of rock-water interaction at increasing temperatures we would expect the most negative $\delta^{18}$O$_{VPDB}$ values to be at the greatest depth and therefore highest temperature. This trend is most likely a result of either rock-water interaction ceasing in the older reservoirs prior to the shallowest reservoir intervals
or the micropores of the shallowest reservoirs being cemented at a higher temperature. Another possibility is that meteoric diagenesis after the deposition of the Shu’aiba’ Formation has altered the $\delta^{18}$O$_{VPDB}$ of the shallowest reservoirs intervals by introducing pore fluid of a different isotopic composition, but with a similar $\delta^{13}$C to Early Cretaceous seawater. However, this latter explanation is unlikely because no evidence for sub-aerial exposure has been observed during the deposition of the Lekhwair, Kharai or Shu’aiba’ Formations in outcrop, core or through petrographic analyses. Also, for the fields located in the Bab Basin (Fields A and B) no evidence has been obtained for significant sub-aerial exposure after the deposition of the Shu’aiba’ Formation (Figure 2.6).

3.4.2 Calcite cement

3.4.2.1 Results

The data presented in this section are from Field B and obtained by Cox (2010). However, new conclusions will be reached in this assessment with respect to determining the origin of the diagenetic cements and therefore these data are re-discussed. The $\delta^{13}$C data for the blocky calcite macromcements and for calcite cements in fractures range from $\sim$ 0.7 ‰ to $\sim$ 3.5 ‰ (Figure 3.37), these values are similar to those of Cretaceous seawater (Figure 3.36). However, all $\delta^{18}$O$_{VPDB}$ data for the calcite macromcements range from $\sim$ -10 ‰ to $\sim$ -7.5 ‰, with the exception of one sample that has a $\delta^{18}$O$_{VPDB}$ of $\sim$ -5.8 ‰. All the $\delta^{18}$O$_{VPDB}$ values are too low for the cements to have been precipitated under normal marine conditions.
Figure 3.37: $\delta^{18}O_{VPDB}$ vs. $\delta^{13}C$ cross-plot for blocky calcite cement and for the calcite cement in fractures. The box represents the postulated ranges for Lower Cretaceous carbonates and is in good agreement with Neilson et al. (1998), Alsharhan et al. (2000) and Vahrenkamp (2010).

### 3.4.2.2 Interpretation

The $\delta^{13}C$ values are compatible with Lower Cretaceous seawater (Section 2.3.3). This indicates that the main source of carbon for both the blocky and fracture calcite cements is likely to have been derived from marine carbonate; most likely by dissolution and re-precipitation and diffusive supply of solutes, as opposed to mass flow, and has not been significantly affected by temperature fractionation (Mook et al., 1974). However, the $\delta^{18}O_{VPDB}$ data do not indicate precipitation from Lower Cretaceous seawater and suggest precipitation from evolved warm water where the $\delta^{18}O_{VPDB}$ of the cement is primarily controlled by the temperature-dependant fractionation between pore water and calcite (Kim and O'Neil, 1997). This suggests that the cement may be derived from dissolution and re-precipitation of Early Cretaceous marine carbonate but at an increased temperature.
The $\delta^{18}\text{O}_{\text{VPDB}}$ is plotted against $\delta^{13}\text{C}$ with specific emphasis placed on stratigraphic position (Figure 3.38). It was expected that the data would show no stratigraphic differences as the entire sequence has undergone a similar burial history and a similar diagenetic history. However the data are distinctively grouped, some units have high $\delta^{13}\text{C}$ and others low $\delta^{13}\text{C}$. The plot shows that the $\delta^{13}\text{C}$ within each cycle is progressively more positive into the younger reservoir intervals. The simplest explanation would be that the calcite cements are sourced locally and from marine carbonate belonging to the reservoir interval in which the cements are now present, and therefore the stratigraphic trend is related to that of Early Cretaceous seawater (Section 2.3.3).

When the average $\delta^{13}\text{C}$ and $\delta^{18}\text{O}_{\text{VPDB}}$ of calcite cement in a reservoir is related to the average bulk micrite $\delta^{13}\text{C}$ and $\delta^{18}\text{O}_{\text{VPDB}}$ for the same reservoir interval, the cements appear to follow a “thermal alteration pathway” (Figure 3.38). The thermal alteration pathways for the reservoirs are shown in Figure 3.38 by the red dashed lines. There is some variability in $\delta^{13}\text{C}$ for the cements of each reservoir, but this is consistent with the variability in Cretaceous seawater during reservoir precipitation (Figure 3.36). Therefore, the thermal alteration pathways suggest that the $\delta^{13}\text{C}$ and $\delta^{18}\text{O}_{\text{VPDB}}$ for the blocky calcite cements are derived from dissolution and re-precipitation of marine carbonate from the reservoir interval in which they are now present, but at elevated temperatures. This implies that during precipitation of the blocky calcite cements the reservoirs behaved as relatively closed systems, where there was no significant external source of oxygen or carbon and the carbon composition of the cements is buffered by the rock.
Figure 3.38: $\delta^{18}O_{VPDB}$ vs. $\delta^{13}C$ cross-plot for the calcite cement in five reservoir intervals (viii, vii, v, iv, i). Thermal alteration pathways (red dashed lines) are drawn from the average $\delta^{18}O_{VPDB}$ and $\delta^{13}C$ composition of the micrite within each reservoir interval (shown by the blue points), to the average composition of the calcite cements within that same reservoir interval (red points). There is an outlier for reservoir viii this has led to a different gradient for the thermal alteration pathway when compared to other reservoirs. However, two points are in good agreement with the other thermal alteration pathways and would suggest that with further sampling the thermal alteration pathway of reservoir viii will be in good agreement with that for the other reservoir intervals; this is also suggested by the data from the calcite in the fracture cement of reservoir interval viii. The box represents the postulated ranges for Lower Cretaceous carbonates and is in good agreement with Neilson et al. (1998), Alsharhan et al. (2000) and Vahrenkamp (2010).
The calcite cements precipitated within the fractures of reservoir interval viii have δ¹³C values similar to Cretaceous seawater and are also compatible with cementation along a thermal alteration pathway. This suggests that the calcite precipitated in the fractures is also locally sourced.

### 3.4.3 Dolomite cement

#### 3.4.3.1 Results

The data presented here are from Field B and obtained by Cox (2010), however new observations are also made in this subsection. The δ¹³C of all dolomite cements with the exception of two values are compatible with Early Cretaceous seawater (Figure 3.39) (Section 2.3.3); the exceptions of ~ 4 ‰ were obtained from patchy dolomite that is distributed throughout the matrix.

The δ¹⁸OVPDB of microdolomite has a wide range of δ¹⁸OVPDB from ~ -2 ‰ to ~ -6.5 ‰. The values of ~ -4 ‰ to ~ -2 ‰ are compatible with marine values however the δ¹⁸OVPDB of ~ -6 ‰ to ~ -4 ‰ would require more evolved pore water. The saddle dolomite cements form a coherent group with δ¹⁸OVPDB of ~ 7 ‰. However, one exception exists with a δ¹⁸OVPDB of ~ -10.5 ‰.
Figure 3.39: $\delta^{18}O_{VPDB}$ vs. $\delta^{13}C$ cross-plot for microdolomite and saddle dolomite cement. The box represents the postulated ranges for Lower Cretaceous carbonates and is in good agreement with Neilson et al. (1998), Alsharhan et al. (2000) and Vahrenkamp (2010).

3.4.3.2 Interpretation

The $\delta^{18}O_{VPDB}$ of microdolomite is always more positive than that of saddle dolomite, this is most likely related to saddle dolomite precipitating at high temperatures and after the formation of microdolomite. Furthermore, the $\delta^{18}O_{VPDB}$ of both micro- and saddle- dolomite cements are more positive than that for blocky calcite cement (Section 3.4.2). However, when comparing calcite with dolomite $\delta^{18}O_{VPDB}$ values their differing fractionation must be taken into account.

The isotopic fractionation of calcite has been investigated numerous times and resulted in minor changes to the fractionation equation; the equation of Kim and O’Neil (1997) will be used in this thesis. The fractionation equation for dolomite is problematic because dolomite is difficult to synthesis at sedimentary temperatures. However, recently this has been achieved with the aid of bacteria (Vasconcelos et al.,
This fractionation equation gives a calcite/dolomite difference of \( \sim 2.5 \% \) which means that the saddle dolomite precipitated from \(^{18}\text{O}\) enriched waters that were similar to the waters from which calcite precipitated. This would also agree with microdolomite precipitating prior to the formation of blocky calcite and saddle dolomite cement, and supports the petrographic sequence presented in Figure 3.1.

The \( \delta^{13}\text{C} \) of the dolomite cements are similar to those of calcite cements; except for the two more positive \( \delta^{13}\text{C} \) values observed in microdolomite that could be a result of methanogenesis (Maliva et al., 1991). The decreasing \( \delta^{13}\text{C} \) trend into the older reservoir intervals for the microdolomite and saddle dolomite is similar to that observed for blocky calcite cements (Figure 3.40). This is most likely a result of the source of the \( \delta^{13}\text{C} \) being similar i.e. from dissolution and re-precipitation of local marine carbonate.

The “thermal alteration pathways” observed in the blocky calcite cement also exist for the dolomite cements. However, the difference in fractionation factor for \( \delta^{18}\text{O}_{\text{VPDB}} \) between micrite (calcite) and dolomite cement makes the pathways less pronounced. This would suggest that the source for the dolomite cements is local and likely from the same reservoir interval in which the cements are now present, this would imply a relatively closed system for each reservoir interval during the precipitation of the dolomite cements.
Figure 3.40: $\delta^{18}O_{VPDB}$ vs. $\delta^{13}C$ cross-plot for saddle- and micro-dolomite. The average bulk micrite composition for each reservoir (shown in blue) is presented with the data for the dolomite cements (shown in green and black). Thermal alteration pathways (shown by the red dashed lines) are made from the average micrite $\delta^{18}O_{VPDB}$ and $\delta^{13}C$ for a reservoir interval to the average composition of the dolomite cement within that same reservoir interval (shown in red). The box represents the postulated ranges for Lower Cretaceous carbonates and is in good agreement with Neilson et al. (1998), Alsharhan et al. (2000) and Vahrenkamp (2010).

3.5 Conclusions

The Lekhwair, Kharib and Shu’aila’ Formations have a very complex diagenetic history; however, the same diagenetic constituents are observed in both Field A and B. The diagenetic features identified appear in the same chronological order which allowed for a single paragenetic sequence to describe the diagenetic history of both fields. This paragenetic sequence is in agreement with $\delta^{18}O_{VPDB}$.
values obtained for the main pore occluding cements; with the diagenetically later cements having more reduced $\delta^{18}O_{VPDB}$ values.

The paragenetic sequence suggests that calcite cementation began in the marine phreatic zone and continued until the deep burial zone; this has resulted in significant porosity loss in both fields and reduced the thickness of the reservoir intervals. During calcite cementation no episodes of extensive low-Mg calcite dissolution were observed (although oil inclusions suggest oil charge did occur) implying that the pore fluids were always sufficiently saturated with respect to low-Mg calcite to prevent dissolution. At this stage the composition of formation water was equal to, or slightly evolved from, that of seawater.

Extensive dissolution after the development of blocky calcite and saddle dolomite cement has significantly enhanced porosity. The extensive dissolution event has etched the surfaces of calcite and dolomite cements and led to the development of vuggy porosity. The fluid causing this dissolution is thought to be related to a second hydrocarbon charging event, but the origin of this event is currently unclear and is discussed in more detail in Chapter 5.

The main pore occluding cement type in both Field A and B is low-Mg calcite cement. This cement can completely occlude intergranular, intragranular, vuggy and moldic pores, and significantly controls the flow properties of reservoir intervals. Using the porosity cut off of Lønøy (2006), flow >1 mD in the macropores of grain dominated fabrics is likely to be primarily from intergranular, moldic and vuggy pores, whereas flow in the muddier fabrics is typically through vuggy and intragranular pores. Hence, because the distribution of intragranular porosity is controlled by the fauna present (Section 3.2.1), if flow is through intragranular porosity alone the permeability of the sample will be dependent on the abundance of certain fauna (specifically the distribution of Lithocodium/Bacinella). No significant remaining macroporosity is present in the non-reservoir intervals which has most likely caused them to behave as seals to the underlying reservoir.

There is a greater volume of intergranular and moldic porosity in the younger reservoir intervals than in the older reservoirs. This is coupled with there being a
greater volume of intragranular porosity in the older reservoirs than in the younger reservoirs. A similar volume of vuggy porosity is present in all reservoir intervals; this is a result of late stage pervasive dissolution of low-Mg calcite.

The oil bearing reservoir intervals of both Field A and B typically have a lower volume of equant and blocky calcite cement precipitated in their intergranular pores than in the coeval aquifer, this results in a higher porosity in the oil reservoirs than in the aquifers. The reason for this is most likely related to the effect of oil charge on preserving porosity in the oil bearing reservoir intervals by reducing cementation rate. However a similar volume of vuggy porosity is present in both the coeval oil reservoirs and aquifers of Field A and B, this is most likely a result of late stage dissolution after calcite and dolomite cementation stopped.

The δ¹³C and δ¹⁸OVPDB data for calcite and dolomite cements can be directly related to the δ¹³C and δ¹⁸OVPDB of bulk micrite for the reservoir interval in which they are now present. This suggests that the δ¹³C and δ¹⁸OVPDB of calcite cement and for saddle dolomite follow thermal alteration pathways and implies that the source of the solutes is likely from pressure solution of marine carbonate that is in the same reservoir interval in which the cements are now present. This implies that during the formation of calcite and saddle dolomite cement, precipitation took place in a relatively confined aquifer system.

The likely cause for the reservoir intervals behaving as relatively confined aquifer systems is that the non-reservoir intervals acted as seals from an early burial stage. The non-reservoir intervals are primarily composed of the TST deposits for each cycle; however, the early cementation (< 200 m burial depth) of porous grainstones has created additional non-reservoir zones which have increased the thickness of the non-reservoir intervals. These non-reservoir intervals are therefore likely to behave as barriers to the effective migration of water during burial.
CHAPTER 4
DYNAMICS OF CALCITE CEMENTATION IN RESPONSE TO OIL CHARGE: EVIDENCE FROM THE LEKHWAIR, KHARAIB AND SHU’AIBA’ FORMATIONS

4.1 Introduction

Porosity in sedimentary rocks is known to decrease regularly with increasing burial depth but subsurface oil reservoirs are a notable exception (Emery et al., 1993; Robinson and Gluyas, 1992). Burial processes such as chemical compaction are thought to provide the majority of solutes available for cementation (Meyers, 1978) and are typically more dominant in the flanks of the fields than the coeval oil bearing reservoir intervals (Oswald et al., 1995). This is thought to be due to the inhibiting effect of oil charge due to partial water displacement, which is needed for the effective transport of solutes during chemical compaction (Heasley et al., 2000; Worden et al., 1998) and has likely led to the preservation of porosity in oil reservoirs. This is thought to be the case in the Thamama Group (Cox et al., 2010; Neilson et al., 1998). Neilson et al. (1998) showed that where petroleum emplacement occurred relatively early a lower volume of burial cement formed, leading to improved reservoir quality; however, in the aquifer a significant volume of porosity was occluded by burial cement.

Cementation and oil migration are thought to occur synchronously (Gluyas et al., 1993), but the oil-filling of reservoirs is a gradual process over millions of years.
where pores in the crest are filled first (Marchand et al., 2001). Cement growth will therefore cease or slow initially in the crest and continue down structure, in the older stratigraphic intervals, until the final oil/water contact is reached. Below the oil/water contact, i.e. within the aquifer, cementation can continue in the water-saturated pores as long as porosity is present. In many studies, this has been concluded to mean that water-leg diagenesis continues for longer (Heasley et al., 2000; Worden et al., 1998) in both quartz-cemented sandstone reservoirs (Bjorkum et al., 1993; Emery et al., 1993; Walderhaug, 1990) and calcite-cemented carbonate reservoirs (Kirkham et al., 1996).

The notion that cement growth and oil charge occur synchronously over millions of years during burial means that the conditions under which the cements grow are likely to change: temperature will increase and formation waters evolve through reactions with the rocks during burial and time. Calcite precipitated during burial will experience rising temperature with increasing depth, leading to progressively lower $\delta^{18}O_{VPDB}$ values, and indeed many analysed late calcites show decreasing $\delta^{18}O_{VPDB}$ values (Cox et al., 2010; Heydari and Moore, 1993; Hudson, 1977; Meyers and Lohmann, 1985). Neilson et al. (1998) showed that in the Kharaib Formation the $\delta^{18}O_{VPDB}$ of sparry calcite cements lie between -5 and -15‰ with the conclusion being that the cements formed from a variety of waters and/or up to relatively high temperatures.

Neilson et al. (1998) has previously undertaken bulk $\delta^{18}O$ analyses on the calcite cements present within the Thamama Group to help constrain the conditions in which the cements formed. However, this assessment did not undertake a geochemical assessment on the cements within a coeval oil reservoir and aquifer and therefore no comment was made, through geochemical techniques, as to the effect of oil charge on the timing of cementation. Cox et al. (2010) undertook a detailed in-situ ion microprobe assessment of the cements in oil reservoir i and its coeval aquifer i in Field B and concluded that the cements continued to precipitate to a similar temperature in both the oil reservoir and aquifer, but a larger volume of cement formed in the aquifer when compared to the oil reservoir.
This study expands on the study of Cox et al. (2010); which assessed oil reservoir i and its coeval aquifer for Field B, by determining the approximate time at which cementation ceased in nine vertically stacked reservoirs in Field B. This study also assesses a new field (Field A), where no previous assessment has been undertaken to assess the controls on cementation. Two coeval reservoirs (i, ii) in Field A are analysed to determine the effect that oil charge has had on the dynamics of cementation. This assessment also assesses the controls on cementation and the approximate time at which cementation ceased in nine vertically stacked reservoirs from Field A. Further this study is also the first to undertake a detailed elemental assessment of the calcite cements in the Thamama Group and is the first to undertake in-situ geochemical assessments on the cements in a non-reservoir. This study involved:

1) Undertaking elemental assessments on the calcite cements in multiple reservoir and non-reservoir intervals in Field A and B to help identify any changes to pore fluid chemistry during cement precipitation and to help understand whether individual reservoirs behave as relatively closed systems. The samples assessed are from two coeval oil bearing reservoirs and aquifers in Field A, so that the effect of oil charge on the dynamics of cementation can be assessed, and are also from different stratigraphic intervals to help understand how pore fluid chemistry has evolved in different reservoir intervals. The assessment involved using Electron Probe Microanalysis (EPM) to obtain the in-situ Ca, Na, Mg, Al, Si, K, P, Sr, Ti, Mn, Fe and O concentration for 10 μm sample sites along multiple transects. During this assessment specific attention was paid to the \(^{\text{mMg/mCa}}\) composition of the cements, because the ratio is thought to be a good geothermometer (Carpenter, 1980; Heydari and Moore, 1993). This \(^{\text{mMg/mCa}}\) ratio will be compared to in-situ δ\(^{18}\)O\(_{\text{VPDB}}\) data to help constrain the temperature of calcite cement precipitation.

2) The Secondary Ion Mass Spectrometer (SIMS) has been used to obtain in-situ δ\(^{18}\)O\(_{\text{VPDB}}\) values for 10-15 μm sample sites along single transects, from the oldest to the youngest cement zone. Calcite cements were selected in the oil
bearing reservoir intervals and coeval aquifers of Field A, with the position of oil inclusions relatively placed in context with the δ\(^{18}\)O\(_{VPDB}\) cementation sequence so that the effect of oil charge on cementation can be determined. Samples were also chosen in different stratigraphic reservoir intervals in Field A and B to understand how the δ\(^{18}\)O\(_{VPDB}\) has evolved during progressive cementation and to help determine the approximate temperature at which cementation ceased in successive reservoirs. The calcite cements in a fully cemented grainstone from a non-reservoir interval were also sampled to help understand the approximate temperature at which the pores were occluded.

### 4.2 Samples and Representivity

There can be significant variability in the distribution of porosity and cement volume in a single plug sample. Therefore when new highly polished thin sections were made from the plug samples chosen for SIMS and EPM assessment the cement volumes and porosity present in the newly created sections were estimated by point counting (350 points). These data are presented in Table 4.1 and Table 4.2 along with the bulk micrite stable isotope and, where available, commercially derived plug porosity and permeability data.

Each of the reservoir samples chosen have high plug derived porosities. This porosity is distributed in multiple pore types, with the macroporosity of the Kharaiib (cycles iii and ii) and Shu’aiba’ (cycle i) Formations being dominated by intergranular porosity and the macropores of the Lekhwair Formation (cycles ix-iv) being dominated by intragranular porosity. The remaining intragranular porosity in the Lekhwair Formation is typically un-occluded by any form of cement. A similar volume of secondary porosity (moldic and vuggy) is present in all reservoirs. These conclusions are similar to those reached in Chapter 3, suggesting that the samples chosen are representative.
Table 4.1: The remaining un-occluded pore volume for each pore type present in Field A, along with the Minus Cement Porosity (MCP) (shown in brackets). These data are presented in relation to the well and cycle the sample was taken from, along with the lithofacies and fabric of the sample and is shown in relation to bulk micrite $\delta^{13}C$ and $\delta^{18}O_{VPDB}$ and plug porosity and permeability. The rows marked in grey have been used in the SIMS assessment. Key: L. – Lithofacies, Fa. – Fabric, G – Grainstone, GSP – Grain Supported Packstone, W – Wackestone.
<table>
<thead>
<tr>
<th>Cycle</th>
<th>Reservoir/Non-reservoir</th>
<th>Well</th>
<th>L.</th>
<th>Fa.</th>
<th>PORE TYPE (%)</th>
<th>Inter-granular</th>
<th>Intra-granular</th>
<th>Moldic</th>
<th>Vuggy</th>
<th>Micro-porosity</th>
<th>$\Delta^{13}$C</th>
<th>$\Delta^{18}$O</th>
<th>Plug porosity (%)</th>
<th>Plug permeability (mD)</th>
</tr>
</thead>
<tbody>
<tr>
<td>i</td>
<td>Non-reservoir</td>
<td>5B</td>
<td>6</td>
<td>W</td>
<td></td>
<td>0 (3.7)</td>
<td>0 (0)</td>
<td>0 (0.9)</td>
<td>0 (7.1)</td>
<td>20.3</td>
<td>3.3</td>
<td>-6.1</td>
<td>20.30</td>
<td>0.000</td>
</tr>
<tr>
<td>i</td>
<td>Reservoir</td>
<td>1B</td>
<td>18</td>
<td>GSP</td>
<td></td>
<td>6 (15.7)</td>
<td>2 (2.9)</td>
<td>0 (0.9)</td>
<td>4.3 (10.3)</td>
<td>7.6</td>
<td>3.1</td>
<td>-6.6</td>
<td>19.90</td>
<td>5.500</td>
</tr>
<tr>
<td>i</td>
<td>Reservoir (Aquifer)</td>
<td>5B</td>
<td>3</td>
<td>GSP</td>
<td></td>
<td>0 (20.9)</td>
<td>0 (3.1)</td>
<td>2 (20)</td>
<td>4.9 (4.9)</td>
<td>14.8</td>
<td>3.3</td>
<td>-7.5</td>
<td>21.70</td>
<td>0.400</td>
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<td>Reservoir</td>
<td>1B</td>
<td>2</td>
<td>MSP</td>
<td></td>
<td>4.6 (8.3)</td>
<td>0 (0)</td>
<td>3.1 (14.3)</td>
<td>4.9 (8)</td>
<td>2.9</td>
<td>2.8</td>
<td>-6.2</td>
<td>15.50</td>
<td>1.100</td>
</tr>
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<td>iv</td>
<td>Reservoir</td>
<td>4B</td>
<td>2</td>
<td>MSP</td>
<td></td>
<td>0 (0.6)</td>
<td>7.1 (12.3)</td>
<td>1 (2.1)</td>
<td>4 (28.9)</td>
<td>-0.1</td>
<td>x</td>
<td>x</td>
<td>12.00</td>
<td>3.800</td>
</tr>
<tr>
<td>iv</td>
<td>Reservoir</td>
<td>3B</td>
<td>12</td>
<td>W</td>
<td></td>
<td>1.1 (3.7)</td>
<td>3.1 (5.1)</td>
<td>2 (3.1)</td>
<td>5.1 (22.6)</td>
<td>x</td>
<td>2.7</td>
<td>-4.0</td>
<td>x</td>
<td>x</td>
</tr>
<tr>
<td>iv</td>
<td>Reservoir</td>
<td>3B</td>
<td>18</td>
<td>GSP</td>
<td></td>
<td>2.3 (11.1)</td>
<td>0 (1.1)</td>
<td>0 (0)</td>
<td>9.1 (9.1)</td>
<td>x</td>
<td>2.2</td>
<td>-3.7</td>
<td>x</td>
<td>x</td>
</tr>
<tr>
<td>iv</td>
<td>Non-reservoir</td>
<td>4B</td>
<td>18</td>
<td>G</td>
<td></td>
<td>0 (33.1)</td>
<td>0 (0)</td>
<td>0 (0)</td>
<td>0 (0)</td>
<td>1</td>
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<td>0.010</td>
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<tr>
<td>v</td>
<td>Reservoir</td>
<td>1B</td>
<td>2</td>
<td>GSP</td>
<td></td>
<td>0 (6.6)</td>
<td>12 (16)</td>
<td>0 (8)</td>
<td>5.1 (14)</td>
<td>-3.9</td>
<td>2.3</td>
<td>-3.9</td>
<td>13.20</td>
<td>0.520</td>
</tr>
<tr>
<td>vii</td>
<td>Non-reservoir</td>
<td>4B</td>
<td>11</td>
<td>W</td>
<td></td>
<td>0 (1.7)</td>
<td>0 (2)</td>
<td>0 (6.9)</td>
<td>1.1 (10)</td>
<td>5.8</td>
<td>1.3</td>
<td>-4.3</td>
<td>6.90</td>
<td>0.010</td>
</tr>
<tr>
<td>viii</td>
<td>Reservoir (Aquifer)</td>
<td>3B</td>
<td>12</td>
<td>GSP</td>
<td></td>
<td>0 (0.6)</td>
<td>9.1 (18.3)</td>
<td>0 (5.1)</td>
<td>8.6 (18.9)</td>
<td>-3.85</td>
<td>2.1</td>
<td>-4.2</td>
<td>13.85</td>
<td>0.134</td>
</tr>
<tr>
<td>ix</td>
<td>Reservoir (Aquifer)</td>
<td>4B</td>
<td>13</td>
<td>MSP</td>
<td></td>
<td>0 (0.6)</td>
<td>10 (18)</td>
<td>0 (7.1)</td>
<td>9.1 (19.1)</td>
<td>-5.1</td>
<td>1.9</td>
<td>-3.5</td>
<td>14.00</td>
<td>5.300</td>
</tr>
</tbody>
</table>

Table 4.2: The remaining un-occluded pore volume for each pore type present in Field B, along with the Minus Cement Porosity (MCP) (shown in brackets). These data are presented in relation to the well and cycle the sample was taken from, along with the lithofacies and fabric of the sample and is shown in relation to bulk micrite $\delta^{13}$C and $\delta^{18}$OVPDB and plug porosity and permeability. The rows marked in grey have been used in the SIMS assessment. Key: L. – Lithofacies, Fa. – Fabric, G – Grainstone, GSP – Grain Supported Packstone, MSP – Mud Supported Packstone, W - Wackestone.
4.3 In-situ Elemental Assessment

In-situ elemental data obtained from calcite cements can be used to help understand how to pore fluid chemistry has varied during calcite cement precipitation. This assessment can be used to identify if any exotic fluid has entered the reservoir during cement precipitation and is typically undertaken prior to SIMS assessment.

4.3.1 Results

The results of the EPM assessment for Field A and B are presented in Tables 4.3 and 4.4, respectively (see Appendix 5 for the complete dataset). This assessment will help to determine the geochemical evolution of the reservoir and non-reservoir intervals and will be used to help evaluate whether the reservoirs are relatively closed systems.

Calcium and silicon are above the detection limit for all sample sites, with the elemental concentration being similar for all sampled calcite cements.

Sodium is typically below the detection limit in most reservoir and non-reservoir intervals. However, two reservoir intervals in Field B (Table 4.4), aquifer i and oil reservoir iv, have detectable levels of sodium in all calcite cements. Potassium is also distributed in a similar manner to sodium, with the concentration of potassium typically being below the detection limit for most reservoir and non-reservoir intervals. However, for the same reservoir intervals where a high sodium concentration is observed (aquifer i and oil reservoir iv), a high potassium concentration is also present.

The magnesium and calcium concentrations for calcite cement in all reservoir and non-reservoir intervals were converted to the molar $^{\text{m}}\text{Mg}/^{\text{m}}\text{Ca}$ ratio of pore water using a $K_d$ of 0.021 (Rimstidt et al., 1998). In all samples the $^{\text{m}}\text{Mg}/^{\text{m}}\text{Ca}$ ratio typically decreases from the cements closest to the pore wall into the youngest cements that are in the pore centre or adjacent to an open pore space (Figure 4.1).
Figure 4.1: Typical $^{26}\text{Mg}/^{44}\text{Ca}$ transects for calcite cements in Field A and B. The $^{26}\text{Mg}/^{44}\text{Ca}$ ratios for the cements that precipitate closest to a pore wall are typically greater than those for the cements that precipitate in the pore centre or adjacent to an open pore space.
The concentration of strontium and aluminium in the cements are typically below the detection limit and when the concentration is above the detection limit no clear trends are observed.

Iron is typically below the detection limit for all reservoir and non-reservoir intervals in Field A. This is also true for the reservoirs and non-reservoirs of the Kharai and Shu’aiba’ Formations in Field B. However, iron is of a detectable concentration for most sample sites in all reservoir and non-reservoir intervals of the Lekhwair Formation in Field B. When the concentration of iron for a sample site is above the detection limit the sample site can be related to a ferroan zone within the cement, this zone can clearly be seen through staining (Section 3.2.2).
<table>
<thead>
<tr>
<th>Cycle</th>
<th>Reservoir/Non-reservoir</th>
<th>Well</th>
<th>Ca</th>
<th>Na</th>
<th>Mg</th>
<th>Al</th>
<th>Si</th>
<th>K</th>
<th>Sr</th>
<th>Fe</th>
<th>$^\text{mMg}/^\text{mCa}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>i</td>
<td>Reservoir</td>
<td>1A</td>
<td>392832 ± 1576</td>
<td>401 ± 113</td>
<td>3704 ± 320</td>
<td>208 ± 45</td>
<td>252 ± 36</td>
<td>262 ± 64</td>
<td>406 ± 40</td>
<td>378 ± 162</td>
<td>0.74 ± 0.06</td>
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<td></td>
<td></td>
<td>(n = 23)</td>
<td>(n = 8)</td>
<td>(n = 23)</td>
<td>(n = 2)</td>
<td>(n = 22)</td>
<td>(n = 5)</td>
<td>(n = 3)</td>
<td>(n = 4)</td>
<td>(n = 23)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>3A</td>
<td>391144 ± 909</td>
<td>7300 ± 6639</td>
<td>3160 ± 157</td>
<td>612 ± 292</td>
<td>210 ± 43</td>
<td>3206 ± 2946</td>
<td>412 ± 32</td>
<td>1038 ± 240</td>
<td>0.64 ± 0.03</td>
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<tr>
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<td>(n = 43)</td>
<td>(n = 5)</td>
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<td>(n = 3)</td>
<td>(n = 8)</td>
<td>(n = 9)</td>
<td>(n = 43)</td>
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<tr>
<td>i</td>
<td>Non-reservoir</td>
<td>3A</td>
<td>389556 ± 987</td>
<td>146 ± 15</td>
<td>3346 ± 160</td>
<td>178 ± 58</td>
<td>171 ± 5</td>
<td>91 ± 17</td>
<td>570 ± 92</td>
<td>1166 ± 186</td>
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<td>394386 ± 1425</td>
<td>345 ± 112</td>
<td>2456 ± 268</td>
<td>2233 ± 1398</td>
<td>169 ± 11</td>
<td>187 ± 60</td>
<td>508 ± 68</td>
<td>492 ± 215</td>
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<td>(n = 11)</td>
<td>(n = 5)</td>
<td>(n = 38)</td>
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<td>Reservoir (Aquifer)</td>
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<td>394590 ± 3433</td>
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<td>5033 ± 3544</td>
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<td>166 ± 69</td>
<td>412 ± 12</td>
<td>974 ± 146</td>
<td>0.80 ± 0.05</td>
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<td>(n = 11)</td>
<td>(n = 25)</td>
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<td>397112 ± 4074</td>
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<td>3217 ± 183</td>
<td>254 ± 134</td>
<td>213 ± 48</td>
<td>104 ± 13</td>
<td>609 ± 127</td>
<td>472 ± 94</td>
<td>0.64 ± 0.04</td>
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<td>(n = 19)</td>
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<td>(n = 18)</td>
<td>(n = 1)</td>
<td>(n = 6)</td>
<td>(n = 12)</td>
<td>(n = 19)</td>
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<td>v</td>
<td>Reservoir</td>
<td>2A</td>
<td>390266 ± 1608</td>
<td>x</td>
<td>3553 ± 212</td>
<td>x</td>
<td>145 ± 8</td>
<td>x</td>
<td>381 ± 15</td>
<td>1154 ± 443</td>
<td>0.71 ± 0.04</td>
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<td>384806 ± 1489</td>
<td>211 ± 1</td>
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<td>573 ± 341</td>
<td>166 ± 7</td>
<td>102 ± 1</td>
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<td>2452 ± 910</td>
<td>0.99 ± 0.07</td>
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<td>(n = 7)</td>
<td>(n = 7)</td>
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Table 4.3: Average elemental composition (in ppm) for the calcite cements present in Field A. x = below detection limit. N is the number of assessments that are above the detection limit. Porewater $^\text{mMg}/^\text{mCa}$ concentration calculated using a partition coefficient of 0.021 (Rimstidt et al., 1998). The rows marked in grey have been used in the SIMS assessment.
<table>
<thead>
<tr>
<th>Cycle</th>
<th>Reservoir/Non-reservoir</th>
<th>Well</th>
<th>Ca</th>
<th>Na</th>
<th>Mg</th>
<th>Al</th>
<th>Si</th>
<th>K</th>
<th>Sr</th>
<th>Fe</th>
<th>$^{26}$Mg/$^{40}$Ca</th>
</tr>
</thead>
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<tr>
<td>i</td>
<td>Non-reservoir</td>
<td>5B</td>
<td>397406 ± 1203 (n = 26)</td>
<td>415 (n = 1)</td>
<td>1212 ± 115 (n = 26)</td>
<td>7011 ± 5179 (n = 5)</td>
<td>247 ± 102 (n = 24)</td>
<td>648 (n = 1)</td>
<td>461 ± 40 (n = 2)</td>
<td>472 (n = 1)</td>
<td>0.24 ± 0.02 (n = 26)</td>
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<tr>
<td>i</td>
<td>Reservoir</td>
<td>1B</td>
<td>399179 ± 591 (n = 16)</td>
<td>119 (n = 1)</td>
<td>1221 ± 167 (n = 16)</td>
<td>680 (n = 1)</td>
<td>150 ± 7 (n = 15)</td>
<td>x</td>
<td>384 ± 11 (n = 4)</td>
<td>x</td>
<td>0.24 ± 0.03 (n = 16)</td>
</tr>
<tr>
<td>i</td>
<td>Reservoir (Aquifer)</td>
<td>5B</td>
<td>393266 ± 774 (n = 31)</td>
<td>975 ± 78 (n = 31)</td>
<td>1860 ± 171 (n = 31)</td>
<td>1024 ± 623 (n = 11)</td>
<td>299 ± 14 (n = 31)</td>
<td>209 ± 19 (n = 29)</td>
<td>530 ± 197 (n = 2)</td>
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<td>398432 ± 1947 (n = 4)</td>
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<td>1592 ± 160 (n = 4)</td>
<td>x</td>
<td>162 ± 19 (n = 4)</td>
<td>73 (n = 1)</td>
<td>x</td>
<td>x</td>
<td>0.31 ± 0.03 (n = 4)</td>
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<tr>
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<td>Reservoir</td>
<td>4B</td>
<td>387639 ± 3614 (n = 36)</td>
<td>12982 (n = 1)</td>
<td>3374 ± 145 (n = 36)</td>
<td>860 ± 451 (n = 14)</td>
<td>260 ± 92 (n = 33)</td>
<td>8023 (n = 1)</td>
<td>521 ± 50 (n = 13)</td>
<td>1893 ± 198 (n = 34)</td>
<td>0.68 ± 0.03 (n = 36)</td>
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<tr>
<td>iv</td>
<td>Reservoir</td>
<td>3B</td>
<td>388224 ± 610 (n = 44)</td>
<td>279 ± 95 (n = 6)</td>
<td>3509 ± 150 (n = 44)</td>
<td>179 ± 67 (n = 2)</td>
<td>197 ± 4 (n = 43)</td>
<td>147 ± 61 (n = 4)</td>
<td>670 ± 83 (n = 12)</td>
<td>2048 ± 226 (n = 44)</td>
<td>0.71 ± 0.03 (n = 44)</td>
</tr>
<tr>
<td>iv</td>
<td>Reservoir</td>
<td>3B</td>
<td>387452 ± 824 (n = 62)</td>
<td>1069 ± 101 (n = 62)</td>
<td>4278 ± 225 (n = 62)</td>
<td>3748 ± 1173 (n = 16)</td>
<td>379 ± 65 (n = 62)</td>
<td>235 ± 30 (n = 40)</td>
<td>419 ± 15 (n = 14)</td>
<td>1680 ± 271 (n = 20)</td>
<td>0.87 ± 0.05 (n = 62)</td>
</tr>
<tr>
<td>iv</td>
<td>Non-reservoir</td>
<td>4B</td>
<td>387118 ± 1029 (n = 57)</td>
<td>274 ± 137 (n = 6)</td>
<td>4421 ± 141 (n = 57)</td>
<td>461 ± 156 (n = 30)</td>
<td>199 ± 6 (n = 56)</td>
<td>x</td>
<td>424 ± 34 (n = 16)</td>
<td>2585 ± 187 (n = 45)</td>
<td>0.90 ± 0.03 (n = 57)</td>
</tr>
<tr>
<td>v</td>
<td>Reservoir</td>
<td>1B</td>
<td>389577 ± 838 (n = 33)</td>
<td>162 ± 5 (n = 5)</td>
<td>3704 ± 115 (n = 33)</td>
<td>792 ± 279 (n = 5)</td>
<td>223 ± 9 (n = 32)</td>
<td>70 (n = 1)</td>
<td>535 ± 79 (n = 11)</td>
<td>2356 ± 263 (n = 32)</td>
<td>0.75 ± 0.02 (n = 33)</td>
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<tr>
<td>vii</td>
<td>Non-reservoir</td>
<td>4B</td>
<td>384885 ± 912 (n = 36)</td>
<td>182 ± 42 (n = 3)</td>
<td>3011 ± 140 (n = 36)</td>
<td>527 ± 217 (n = 5)</td>
<td>275 ± 70 (n = 36)</td>
<td>156 ± 41 (n = 2)</td>
<td>607 ± 47 (n = 20)</td>
<td>3521 ± 142 (n = 36)</td>
<td>0.62 ± 0.03 (n = 36)</td>
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<tr>
<td>viii</td>
<td>Reservoir (Aquifer)</td>
<td>3B</td>
<td>384648 ± 1023 (n = 20)</td>
<td>182 ± 27 (n = 3)</td>
<td>3237 ± 230 (n = 20)</td>
<td>95 (n = 1)</td>
<td>152 ± 7 (n = 20)</td>
<td>81 (n = 1)</td>
<td>639 ± 100 (n = 7)</td>
<td>3971 ± 283 (n = 20)</td>
<td>0.66 ± 0.05 (n = 20)</td>
</tr>
<tr>
<td>ix</td>
<td>Reservoir (Aquifer)</td>
<td>4B</td>
<td>379297 ± 795 (n = 29)</td>
<td>205 (n = 1)</td>
<td>4139 ± 207 (n = 29)</td>
<td>x</td>
<td>204 ± 5 (n = 29)</td>
<td>x</td>
<td>531 ± 65 (n = 13)</td>
<td>5373 ± 298 (n = 29)</td>
<td>0.86 ± 0.04 (n = 29)</td>
</tr>
</tbody>
</table>
Table 4.4 (on previous page): Average elemental composition (in ppm) of the calcite cements present in Field B. \( x = \) below detection limit. \( N \) is the number of assessments that are above the detection limit. The porewater \( "\frac{Mg}{Ca}\) ratio is calculated using a partition coefficient of 0.021 (Rimstidt et al., 1998). The rows marked in grey have been used in the SIMS assessment.

4.3.2 Discussion

To help determine how pore fluid chemistry changed during the formation of calcite cement in Field A and B and to identify any evidence for fluid mixing, the results of the elemental analysis will now be discussed.

4.3.2.1 Sodium and Potassium

The distribution coefficient (\( K_d \)) for Na is thought to be between \( 10^{-4} \) to \( 10^{-5} \) (Brand and Veizer, 1980) which would suggest that sodium is not preferentially incorporated into the growing cement. Two samples in Field B have high Na and K concentrations (Table 4.4), these samples are in different reservoir intervals and in different wells (reservoir interval i in Well 5B and reservoir iv in Well 3B). The Na and K concentration is above the detection limit for all sample sites in these reservoirs, suggesting that the initial pore fluid had a high concentration of these elements and the system remained relatively closed with respect to Na and K during cement precipitation. This is because if significant fluid flow occurred between reservoirs it is likely, because of the low \( K_d \), that the Na and K concentrations of the pore water would have mixed with other reservoirs, therefore the system is likely to have remained relatively closed during cement precipitation.

4.3.2.2 Aluminium

Aluminium is randomly distributed throughout Field A and Field B. Al has a \( K_d > 1 \) which suggests that the transportation of aluminium ions in diagenetic waters is very limited. However, it has been proposed that aluminium can complex with organic matter and therefore the presence of aluminium may be related to the migration of oil in many reservoirs (Maliva et al., 1999). The origin of Al and the
cause for its random distribution throughout both Field A and Field B is currently uncertain and will be discussed in Chapter 5.

4.3.2.3 Silicon

Silicon is a much larger element and of a greater valency than calcium and so does not readily substitute into the carbonate lattice. Similar concentrations of silicon are observed in all sampled calcite cements, this may suggest that there was a low concentration of Si in the pore fluid during cement formation and that there was not a significant input of Si into the system from weathering or from diagenetic alteration of silicate minerals.

4.3.2.4 Strontium

The $K_d$ of Sr is $\sim 0.1$ (Bjørlykke and Avseth, 2010), this would suggest that the concentration of Sr in calcite cements will be low. It was thought that due to the increasing abundance of aragonitic biota in the Kharai and Shu’aiba’ Formations, when compared to the Lekhwa Formation, there would be a higher Sr concentration in the calcite cements; if the reservoirs are relatively closed. However, no systematic trends in Sr are observed. This may be explained by 1) A progressive decrease in $K_d$ due to rising temperatures (Katz et al., 1972); 2) Sr in calcite is strongly-rate dependant and therefore extremely slow precipitation of the cement during burial may override the effects of temperature and Sr/Ca ratio (Heydari and Moore, 1993); 3) That aragonite stabilisation occurred in a system still open to seawater, this is supported by the paragenetic sequence (Figure 3.1); 4) Influx of exotic fluid into the system has led to the preferential removal of Sr – this is unlikely because Mg has a similar $K_d$ to Sr and no evidence for the removal of Mg from the system is observed (Section 4.3.2.6).
4.3.2.5 Iron

The $K_d$ of iron is thought to be 3.7 in seawater and increases with depth to ~7.7 (Dromgoole and Walter, 1990), this suggests that iron preferentially partitions into the cement from the pore fluid. Therefore the high distribution coefficient may be the cause for the random increase in Fe within the calcite cements. The distribution of iron rich calcite is in good agreement with the iron rich cement zones identified through staining (Section 3.2) (Dickson, 1965) and cannot be correlated between samples. Possible sources for the iron include dewatering of the underlying dense interval and fluid migration prior to hydrocarbon charge (Section 3.2.3).

The cements in the Lekhwair Formation of Field B are typically more ferroan than the cements in the Kharaib and Shu’aiba’ Formations (Table 4.4). This would suggest that the initial pore fluid of the Lekhwair formation in Field B had a higher iron content than the Kharaib and Shu’aiba’ Formations and would imply that the pore water for these cycles remained reducing from deposition and during subsequent cement precipitation and is unlikely to have mixed with the pore water in the Kharaib and Shu’aiba’ Formations.

4.3.2.6 Magnesium/Calcium ratio

The magnesium concentration in calcite cements is primarily controlled by the $\text{Mg}/\text{Ca}$ ratio of pore fluids, the distribution coefficient – which is thought to increase due to decreasing Mg/Ca (Mucci and Morse, 1983) and increasing temperature (Burton and Walter, 1991) - and temperature (Heydari, 1997a; Tucker et al., 1991). Although the concentration of $pCO_2$ and $SO_4^{2-}$ are known to affect the distribution coefficient of Mg they are likely to be buffered in a carbonate system (Heydari and Moore, 1993).

Thermodynamic calculations suggest that the $\text{Mg}/\text{Ca}$ of pore water decreases with increasing temperature (Johnson et al., 1992). This is observed to be the case in the Interior Oman Sedimentary Basin (Hartmann et al., 2000) and in multiple basins in North America (Heydari and Moore, 1993). The $\text{Mg}/\text{Ca}$ concentration of pore fluids are typically close to 1 at 25°C, and then decrease in a
near-linear manner to 0.05 at 150°C (Figure 4.2), this suggests that the $^{\text{m}}\text{Mg}/^{\text{m}}\text{Ca}$ ratio of subsurface waters can be used as an approximate geothermometer (Carpenter, 1980; Heydari and Moore, 1993).

![Figure 4.2: Temperature vs. $^{\text{m}}\text{Mg}/^{\text{m}}\text{Ca}$ of pore water (Heydari and Moore, 1993). The near linear decrease in $^{\text{m}}\text{Mg}/^{\text{m}}\text{Ca}$ would suggest that the $^{\text{m}}\text{Mg}/^{\text{m}}\text{Ca}$ ratio can be used as an approximate geothermometer.](image)

The $^{\text{m}}\text{Mg}/^{\text{m}}\text{Ca}$ ratio of the cements decreases from the pore wall into the pore centre (Figure 4.1). This would suggest that the distribution coefficient of Mg did not increase sufficiently to outweigh the decreasing $^{\text{m}}\text{Mg}/^{\text{m}}\text{Ca}$ ratios of pore water, implying that the $^{\text{m}}\text{Mg}/^{\text{m}}\text{Ca}$ ratio of the cements can be used to help understand the approximate temperature of precipitation for the cements.

The average $^{\text{m}}\text{Mg}/^{\text{m}}\text{Ca}$ is lower in the Shu’aiba’ (reservoir i) and Kharaib (reservoirs ii and iii) Formations than in the Lekhwaib Formation (reservoirs iv-ix) of Field B (Figure 4.3). The cause of this trend is equivocal but can be explained by: 1) The majority of cements precipitated from pore water with a similar $^{\text{m}}\text{Mg}/^{\text{m}}\text{Ca}$ ratio, but the cements in the younger reservoir intervals precipitated at a higher temperature. 2) The calcite cements precipitated from pore fluid where the $^{\text{m}}\text{Mg}/^{\text{m}}\text{Ca}$ ratio varied over time. 3) Influx of meteoric water after deposition, has preferentially removed the Mg$^{2+}$ from the youngest reservoir intervals.
Figure 4.3: Average \(^{\text{Mg}/}\text{Ca}\) of pore water for each cycle (ix-i), with standard error. The \(^{\text{Mg}/}\text{Ca}\) of porewater was calculated using a \(K_d\) of 0.021 (Rimstidt et al., 1998) and the time of deposition for the cycles is taken from Figure 2.15.
During the Hauterivian the $^{\text{m}}\text{Mg}/^{\text{m}}\text{Ca}$ ratio of seawater is thought to be 1.7-1.8 (Dickson, 2004) and decreases into the Upper Barremian (0.9-1.4; Steuber and Rauch (2005) and Aptian (1.1-1.3; Timofeeff et al. (2006)), this agrees with the modelled data of Hardie (1996) (Figure 4.4). Therefore, it is possible that the progressive decrease in average $^{\text{m}}\text{Mg}/^{\text{m}}\text{Ca}$ from the cements in the Lekhwair Formation to the cements in the Khaib and Shu’aiba’ Formations (Figure 4.3) is a result of changes to the composition of Early Cretaceous seawater. However, it has also been suggested that the $^{\text{m}}\text{Mg}/^{\text{m}}\text{Ca}$ ratio of Early Cretaceous seawater was relatively constant (Horita et al., 2002; Wilkinson and Algeo, 1989) (Figure 4.4).

**Figure 4.4**: $^{\text{m}}\text{Mg}/^{\text{m}}\text{Ca}$ composition of Phanerozoic seawater; taken from Steuber and Rauch (2005). The black dashed box marks the time of deposition for the nine cycles.

The $^{\text{m}}\text{Mg}/^{\text{m}}\text{Ca}$ of seawater during the deposition of the Lekhwair, Khaib and Shu’aiba’ Formations is thought to decrease (bold line on figure) (Hardie, 1996), however data presented in Steuber and Rauch (2005) (symbols on the figure) suggest that the $^{\text{m}}\text{Mg}/^{\text{m}}\text{Ca}$ of seawater may have been relatively constant during the deposition of the cycles.
It is possible to observe that in coeval reservoir intervals (assuming a similar initial $\text{Mg}/\text{Ca}$ ratio of the fluid and assuming a constant $K_d$) the $\text{Mg}/\text{Ca}$ ratio is similar for the oil bearing reservoir interval, aquifer and non-reservoir interval in reservoir interval i for both Field A and Field B (Figure 4.3). This would suggest that the majority of cements in each of the intervals precipitated at a similar temperature. However, the cements in the oil bearing reservoir interval ii of Field A have a lower $\text{Mg}/\text{Ca}$ ratio than the coeval aquifer (Figure 4.3). This may suggest that the majority of the cements in the coeval aquifer formed at a lower temperature than in the oil bearing reservoir interval.

The lowest $\text{Mg}/\text{Ca}$ ratio obtained for reservoir i in Field A and Field B is similar for the reservoir, aquifer and non-reservoir (Figure 4.5). This would suggest that cementation continued to a similar temperature in all intervals. Oil reservoir ii in Field A has a lower $\text{Mg}/\text{Ca}$ ratio than in the coeval aquifer this may suggest that cementation continued to a higher temperature in oil reservoir ii.

The $\text{Mg}/\text{Ca}$ ratios in Field A and Field B for the Lekhwair Formation (Figure 4.5) are typically higher than for the Kharaib and Shu’aiba’ Formations. Assuming a similar depositional $\text{Mg}/\text{Ca}$ ratio (Horita et al., 2002; Wilkinson and Algeo, 1989) and a constant $K_d$, cementation may have ceased in the Lekhwair Formation prior to that in the Kharaib and Shu’aiba’ Formations.

Minus cement porosity calculations suggest that the porosity in non-reservoir iv of Field B is thought to have been preserved due to early cementation at burial depths $<200$ m (Figure 3.23, Section 3.3.2); this is also supported by the $\text{Mg}/\text{Ca}$ ratio. The lowest $\text{Mg}/\text{Ca}$ ratio for the cements in this sample is higher than in the overlying reservoir interval (Figure 4.5), this is taken to indicate that cementation ceased at a lower temperature in the non-reservoir.
Figure 4.5: Minimum $\text{Mg/}^{\text{Ca}}$ ratios for the burial cements in cycles ix-i for Fields A and B. The $\text{Mg/}^{\text{Ca}}$ of porewater was calculated using a $K_d$ of 0.021 (Rimstidt et al., 1998) and the time of deposition is taken from Figure 2.15.
4.4 In-situ $\delta^{18}O_{VPDB}$ Assessment

The elemental assessment has shown that calcite cementation most likely occurred in a relatively closed system, this also agrees with the petrographic and separate component assessments (Chapter 3). The $^{\text{m}}$Mg/$^{\text{m}}$Ca ratio decreases from the pore wall into the pore centre which, in a relatively closed system, is taken to indicate cementation continuing at progressively higher temperatures. Therefore, it is probable that in the reservoir intervals of Field A and B, the $\delta^{18}O_{VPDB}$ geothermometer will most likely record cementation in a closed system and at progressively higher temperatures.

The $\delta^{18}O_{VPDB}$ composition of calcite cement has been used to help determine the pore fluid chemistry evolution of the system and to help constrain the temperature of cement precipitation. The $\delta^{18}O_{VPDB}$ was obtained for the individual calcite cement zones identified through cathodoluminescence in the reservoir and non-reservoir intervals and the occurrence of primary oil inclusions have been placed in this cement zone sequence. This assessment will be compared with the elemental data presented in Section 4.3 to help further understand the dynamics of cementation for Field A and B.

4.4.1 Results

4.4.1.1 In-situ $\delta^{18}O_{VPDB}$ data

Whenever possible syntaxial cements (Figure 4.6a, b) were analysed because these cements are thought to contain the most complete record of cementation (Cox et al., 2010). However, in some samples syntaxial cements were not observed, or contain fewer cement zones than that observed in the sample as a whole. Therefore, the following $\delta^{18}O_{VPDB}$ data comes from multiple transects that sample fringing, equant, blocky and syntaxial cements. The transects always start at the pore wall (oldest cement zone) and terminate at an open pore space or when all the cement zones observed for that transect are sampled. In all the $\delta^{18}O_{VPDB}$ transects it is always observed that the higher $\delta^{18}O_{VPDB}$ values are found in cements precipitated against
pore walls, while lower values are observed within cements precipitated in pore centres.

Figure 4.6: a) Syntaxial cements in reservoir interval iv of Field B and b) in aquifer i of Field A. The majority of the cement zones identified within the sample are present in the cement.

4.4.1.1 Field A

The $\delta^{18}$OVPDB in-situ data are shown for Field A in the following five tables (Tables 4.5-4.9), in relation to the cement zone and to the cement phase analysed and to the location of oil inclusions.
<table>
<thead>
<tr>
<th>Zone</th>
<th>Zone colour</th>
<th>Zone luminescence</th>
<th>Oil inclusions</th>
<th>Transsect 1 ($\delta^{18}O$)</th>
<th>Transsect 2 ($\delta^{18}O$)</th>
<th>Transsect 3 ($\delta^{18}O$)</th>
<th>Transsect 4 ($\delta^{18}O$)</th>
<th>Zone average ($\delta^{18}O$)</th>
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<tr>
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Table 4.5: In-situ $\delta^{18}O_{VPDB}$ data (%) for reservoir i in Well 1A (oil bearing reservoir).

Key: $^\dagger$ - syntaxial; $^\ddagger$ - equant
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<th>Transect 2 (δ¹⁸O)</th>
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Table 4.6: In-situ $\delta^{18}O_{VPDB}$ data (‰) for reservoir ii in Well 2A (oil bearing reservoir). Key: * - blocky
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<th>Transsect 2 ($\delta^{18}$O)</th>
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<td>-3.0±0.2</td>
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<tr>
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<td>-3.2$^\dagger$</td>
<td>-3.0$^\dagger$</td>
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<td>-3.5±0.3</td>
</tr>
<tr>
<td>4</td>
<td>Brown</td>
<td>Dull</td>
<td>-6.0$^\dagger$</td>
<td>-4.7$^\dagger$</td>
<td>-4.9$^\dagger$</td>
<td>-5.3$^\dagger$</td>
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<td>-5.1±0.2</td>
</tr>
<tr>
<td>5</td>
<td>Brown</td>
<td>Moderate</td>
<td>-6.1$^\dagger$</td>
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<td>-6.1</td>
</tr>
<tr>
<td>6</td>
<td>Brown</td>
<td>Dull</td>
<td>-6.7$^\dagger$</td>
<td></td>
<td></td>
<td>-6.9$^*$</td>
<td>-6.9±0.1</td>
<td></td>
</tr>
<tr>
<td>7</td>
<td>Orange</td>
<td>Dull</td>
<td>-8.2$^\dagger$</td>
<td>-8.5$^\dagger$</td>
<td>-8.0$^*$</td>
<td>-8.0$^*$</td>
<td>-8.2±0.1</td>
<td></td>
</tr>
<tr>
<td>8</td>
<td>Orange</td>
<td>Moderate</td>
<td>-9.0$^\dagger$</td>
<td>-9.9$^\dagger$</td>
<td>-8.7$^*$</td>
<td>-8.5$^*$</td>
<td>-9.0±0.3</td>
<td></td>
</tr>
</tbody>
</table>

Table 4.7: In-situ $\delta^{18}$OVPDB data (%o) for reservoir i in Well 3A (aquifer). Key: $^\dagger$ - syntaxial; $^\ddagger$ - equant; * - blocky
Table 4.8: In-situ $\delta^{18}O_{VPDB}$ data (‰) for reservoir ii in Well 3A (aquifer). Key: ‡ - equant; * - blocky

<table>
<thead>
<tr>
<th>Zone</th>
<th>Zone colour</th>
<th>Zone luminescence</th>
<th>Transect 1 ($\delta^{18}$O)</th>
<th>Transect 2 ($\delta^{18}$O)</th>
<th>Transect 3 ($\delta^{18}$O)</th>
<th>Zone average ($\delta^{18}$O)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
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<td>Dull</td>
<td>-2.7‡</td>
<td>-2.7</td>
<td></td>
<td></td>
</tr>
<tr>
<td>2</td>
<td>Brown</td>
<td>Moderate</td>
<td>-3.2‡</td>
<td>-3.2</td>
<td></td>
<td></td>
</tr>
<tr>
<td>3</td>
<td>Orange</td>
<td>Moderate</td>
<td></td>
<td></td>
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</tr>
<tr>
<td>4</td>
<td>Brown</td>
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<td>-5.2*</td>
<td>-6.9‡</td>
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</tr>
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<td>Orange</td>
<td>Moderate</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>6</td>
<td>Red-Brown</td>
<td>Moderate</td>
<td>-6.9*</td>
<td></td>
<td>-6.9</td>
<td></td>
</tr>
<tr>
<td>7</td>
<td>Brown</td>
<td>Dull</td>
<td>-8.3*</td>
<td>-8.1‡</td>
<td>-8.3±0.1</td>
<td></td>
</tr>
<tr>
<td>8</td>
<td>Brown</td>
<td>Moderate</td>
<td>-8.8*</td>
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<td>-8.8</td>
<td></td>
</tr>
<tr>
<td>9</td>
<td>Orange</td>
<td>Moderate</td>
<td>-8.9*</td>
<td>-9.0*</td>
<td>-8.9±0.1</td>
<td></td>
</tr>
</tbody>
</table>

Table 4.9: In-situ $\delta^{18}O_{VPDB}$ data (‰) for reservoir vii in Well 2A (gas bearing reservoir). Key: * - fringing; ‡ - equant; * - blocky

<table>
<thead>
<tr>
<th>Zone</th>
<th>Zone colour</th>
<th>Zone luminescence</th>
<th>Transect 1 ($\delta^{18}$O)</th>
<th>Transect 2 ($\delta^{18}$O)</th>
<th>Transect 3 ($\delta^{18}$O)</th>
<th>Zone average ($\delta^{18}$O)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Black</td>
<td>Dull</td>
<td>-2.9‡</td>
<td>-4.0‡</td>
<td>-3.21*</td>
<td>-4.01*</td>
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<td>Yellow</td>
<td>Bright</td>
<td>-3.4‡</td>
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<td></td>
</tr>
<tr>
<td>3</td>
<td>Brown</td>
<td>Dull</td>
<td>-4.7‡</td>
<td>-4.2‡</td>
<td>-5.8*</td>
<td>-6.4*</td>
</tr>
<tr>
<td>4</td>
<td>Orange</td>
<td>Moderate</td>
<td>-6.7*</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>5</td>
<td>Brown</td>
<td>Dull</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>6</td>
<td>Brown</td>
<td>Moderate</td>
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<td>-6.2‡</td>
<td>-7.1‡</td>
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<tr>
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<td>Orange</td>
<td>Moderate</td>
<td>-7.8‡</td>
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</tbody>
</table>
4.4.1.1.2 Field B

The $\delta^{18}O_{VPDB}$ in-situ data for Field B are shown in the following six tables (Tables 4.10-4.15), in relation to the cement zone and to the cement phase analysed and to the location of oil inclusions.

<table>
<thead>
<tr>
<th>Zone</th>
<th>Zone colour</th>
<th>Zone luminescence</th>
<th>Transect 1 $(\delta^{18}O)$</th>
<th>Transect 2 $(\delta^{18}O)$</th>
<th>Transect 3 $(\delta^{18}O)$</th>
<th>Transect 4 $(\delta^{18}O)$</th>
<th>Transect 5 $(\delta^{18}O)$</th>
<th>Transect 5 average $(\delta^{18}O)$</th>
<th>Zone average $(\delta^{18}O)$</th>
</tr>
</thead>
<tbody>
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<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>2</td>
<td>Orange</td>
<td>Moderate</td>
<td></td>
<td></td>
<td></td>
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<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>3</td>
<td>Yellow</td>
<td>Bright</td>
<td>-4.5*</td>
<td></td>
<td>-4.3*</td>
<td></td>
<td></td>
<td></td>
<td>-4.4±0.1</td>
</tr>
<tr>
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<td>Brown</td>
<td>Dull</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>5</td>
<td>Yellow</td>
<td>Moderate</td>
<td>-6.1*</td>
<td>-5.2*</td>
<td>-5.1*</td>
<td>-5.3*</td>
<td>-5.6*</td>
<td>-5.4±0.2</td>
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</tr>
<tr>
<td>6</td>
<td>Orange</td>
<td>Moderate</td>
<td></td>
<td>-7.3*</td>
<td>-6.6*</td>
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<td>-6.2*</td>
<td>-6.7±0.3</td>
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</tr>
<tr>
<td>7</td>
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<td>-8.3*</td>
<td>-8.2*</td>
<td>-8.6*</td>
<td>-7.4*</td>
<td>-7.9*</td>
<td>-8.0±0.1</td>
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</tr>
<tr>
<td>8</td>
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<td>Dull</td>
<td>-9.5*</td>
<td>-8.7*</td>
<td>-9.1*</td>
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<td>-8.8±0.2</td>
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Table 4.10: In-situ $\delta^{18}O_{VPDB}$ data (‰) for reservoir $i$ in Well 5B (aquifer). † - syntaxial; ‡ - equant; * - blocky
<table>
<thead>
<tr>
<th>Zone</th>
<th>Zone colour</th>
<th>Zone luminescence</th>
<th>Oil inclusions</th>
<th>Transect 1 ($\delta^{18}$O)</th>
<th>Transect 2 ($\delta^{18}$O)</th>
<th>Zone average ($\delta^{18}$O)</th>
</tr>
</thead>
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<td>1</td>
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<td></td>
<td></td>
</tr>
<tr>
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<td>Orange</td>
<td>Moderate</td>
<td></td>
<td></td>
<td></td>
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</tr>
<tr>
<td>3</td>
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<td>Bright</td>
<td>-4.8\textsuperscript{f}</td>
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<td>-4.8</td>
<td></td>
</tr>
<tr>
<td>4</td>
<td>Brown</td>
<td>Moderate</td>
<td>-6.1\textsuperscript{f}</td>
<td>-6.0\textsuperscript{f}</td>
<td>-6.3\textsuperscript{f}</td>
<td>-6.7\textsuperscript{f}</td>
</tr>
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<td>Dull</td>
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</tr>
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<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>7</td>
<td>Brown</td>
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<td>-7.6*</td>
<td>-7.3*</td>
<td>-7.5*</td>
<td>-7.4*</td>
</tr>
<tr>
<td>8</td>
<td>Orange</td>
<td>Moderate</td>
<td>4</td>
<td></td>
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<td></td>
</tr>
<tr>
<td>9</td>
<td>Orange</td>
<td>Bright</td>
<td>0</td>
<td>-10.5*</td>
<td>-9.9*</td>
<td>-8.2*</td>
</tr>
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<td>Moderate</td>
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<td>-10.9*</td>
<td>-10.9*</td>
<td>-10.2*</td>
</tr>
<tr>
<td>11</td>
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<td>Moderate</td>
<td>2</td>
<td>-7.8*</td>
<td>-7.3*</td>
<td>-7.6±0.4</td>
</tr>
</tbody>
</table>

*Table 4.11: In-situ $\delta^{18}$O\textsubscript{VPDB} data (‰) for reservoir ii in Well 1B (oil bearing reservoir). Key: \textsuperscript{f} - equant; * - blocky*
<table>
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<tr>
<th>Zone</th>
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<th>Zone luminescence</th>
<th>Transect 1 $({\delta^{18}}O)$</th>
<th>Transect 2 $({\delta^{18}}O)$</th>
<th>Transect 3 $({\delta^{18}}O)$</th>
<th>Zone average $({\delta^{18}}O)$</th>
</tr>
</thead>
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<td>Dull</td>
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<td></td>
<td></td>
</tr>
<tr>
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<td>Orange</td>
<td>Moderate</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>3</td>
<td>Brown</td>
<td>Moderate</td>
<td></td>
<td></td>
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<td></td>
</tr>
<tr>
<td>4</td>
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<td>Dull</td>
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<td>-5.8*§</td>
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</tr>
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<td>Orange</td>
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<td>-7.1*</td>
<td>-7.1*</td>
<td>-7.1±0.1</td>
</tr>
<tr>
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<td>Dull</td>
<td>-6.1*</td>
<td>-6.1*</td>
<td>-6.1</td>
<td></td>
</tr>
<tr>
<td>7</td>
<td>Orange</td>
<td>Moderate</td>
<td>-6.1*</td>
<td>-6.0*</td>
<td></td>
<td>-6.5±0.1</td>
</tr>
<tr>
<td>8</td>
<td>Brown</td>
<td>Moderate</td>
<td>-7.3*</td>
<td></td>
<td></td>
<td>-7.6±0.1</td>
</tr>
</tbody>
</table>
### Table 4.12: In-situ $\delta^{18}O_{VPDB}$ data (%) for reservoir iv in Well 3B (oil bearing reservoir). Key: ‡ - equant; * - blocky

<table>
<thead>
<tr>
<th>Zone</th>
<th>Zone colour</th>
<th>Zone luminescence</th>
<th>Transect 1 ($\delta^{18}$O)</th>
<th>Transect 2 ($\delta^{18}$O)</th>
<th>Transect 3 ($\delta^{18}$O)</th>
<th>Zone average ($\delta^{18}$O)</th>
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<td>-2.5‡</td>
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</tr>
<tr>
<td>3</td>
<td>Yellow</td>
<td>Bright</td>
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</tr>
<tr>
<td>4</td>
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<td>Dull</td>
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<td>-6.9‡</td>
<td>-6.9*</td>
<td>-6.5±0.4</td>
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<td>-6.6*</td>
<td>-7.0*</td>
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<td>-6.9*</td>
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</tr>
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<td></td>
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<td>-6.5*</td>
<td></td>
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<td>-6.9±0.1</td>
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<tr>
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<td>Orange</td>
<td>Moderate</td>
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<td>-7.4*</td>
<td>-7.6*</td>
<td>-7.5*</td>
</tr>
<tr>
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<td></td>
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</tr>
</tbody>
</table>

### Table 4.13: In-situ $\delta^{18}O_{VPDB}$ data (%) for reservoir v in Well 1B (oil bearing reservoir). Key: † - fringing; ‡ - equant; * - blocky

<table>
<thead>
<tr>
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<th>Zone colour</th>
<th>Zone luminescence</th>
<th>Transect 1 ($\delta^{18}$O)</th>
<th>Transect 2 ($\delta^{18}$O)</th>
<th>Transect 3 ($\delta^{18}$O)</th>
<th>Zone average ($\delta^{18}$O)</th>
</tr>
</thead>
<tbody>
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<td>-7.6*</td>
<td>-7.4*</td>
<td>-7.6*</td>
<td>-7.5*</td>
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<td>-7.5*</td>
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<td>-6.9*</td>
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<td>-8.4*</td>
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<td>Zone colour</td>
<td>Zone luminescence</td>
<td>Transect 1 ($\delta^{18}O$)</td>
<td>Transect 2 ($\delta^{18}O$)</td>
<td>Zone average ($\delta^{18}O$)</td>
<td></td>
</tr>
<tr>
<td>------</td>
<td>-------------</td>
<td>------------------</td>
<td>-----------------------------</td>
<td>-----------------------------</td>
<td>-----------------------------</td>
<td></td>
</tr>
<tr>
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<td>Black</td>
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<td>-3.8$^*$</td>
<td>-3.8$^*$</td>
<td>-3.8$^*$</td>
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</tr>
<tr>
<td>2</td>
<td>Orange</td>
<td>Moderate</td>
<td>-5.1$^f$</td>
<td>-5.6$^f$</td>
<td>-5.0±0.4</td>
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</tr>
<tr>
<td>3</td>
<td>Yellow</td>
<td>Moderate</td>
<td>-6.6$^f$</td>
<td>-6.0$^f$</td>
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<tr>
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<td>Dull</td>
<td>-6.1$^*$</td>
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</tr>
<tr>
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<td>Moderate</td>
<td>-7.5$^*$</td>
<td>-7.7$^*$</td>
<td>-7.6±0.1</td>
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</tr>
<tr>
<td>6</td>
<td>Brown</td>
<td>Moderate</td>
<td>-8.5$^*$</td>
<td>-8.3$^*$</td>
<td>-8.3±0.1</td>
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</tr>
</tbody>
</table>

*Table 4.14: In-situ $\delta^{18}O_{VPDB}$ data (‰) for reservoir viii in Well 3B (aquifer). Key: $^*$ - fringing; $^f$ - equant; $^*$ - blocky*
Oil inclusions

A fluid inclusion assessment has been undertaken with the goal of understanding the time at which oil entered the reservoirs. Most of the inclusions – both hydrocarbon and aqueous inclusions are of secondary origin i.e. not trapped during cement growth, but rather in healed microfractures and/or deformed cleavages (Section 1.7.4). Therefore these inclusions have been discounted. Primary oil inclusions are observed in late syntaxial, equant and blocky calcite cements, they are also commonly observed within saddle dolomite.

Oil inclusions within both Field A and Field B are typically between 5-50 μm in diameter and fluoresce yellow-green to pale blue. Oil inclusions are more commonly observed in the cements of Field B than the cements of Field A, with the amount of oil inclusions typically increasing into the youngest cement zone for both fields (Tables 4.5, 4.6, 4.11).

In Field A 8 oil inclusions are observed. 1 oil inclusion is identified in the penultimate cement zone (average δ¹⁸OVPDB of -8.8 ‰), with 2 oil inclusions being observed in the final cement zone of reservoir i in Well 1A, this zone has an average δ¹⁸OVPDB composition of -9.3 ‰ (Table 4.5). 1 oil inclusion is observed within the penultimate zone, and 3 in the final cement zone of reservoir ii in Well 2A, these

<table>
<thead>
<tr>
<th>Zone</th>
<th>Zone colour</th>
<th>Zone luminescence</th>
<th>Transect 1 (δ¹⁸O)</th>
<th>Zone average (δ¹⁸O)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Black</td>
<td>Dull</td>
<td>-0.6 †</td>
<td>-0.6</td>
</tr>
<tr>
<td>2</td>
<td>Brown</td>
<td>Moderate</td>
<td>-0.3 †</td>
<td>-0.3</td>
</tr>
<tr>
<td>3</td>
<td>Orange</td>
<td>Moderate</td>
<td>0.0 †</td>
<td>0.0</td>
</tr>
<tr>
<td>4</td>
<td>Brown</td>
<td>Moderate</td>
<td></td>
<td></td>
</tr>
<tr>
<td>5</td>
<td>Brown</td>
<td>Dull</td>
<td>-0.6 †</td>
<td>-0.8±0.2</td>
</tr>
</tbody>
</table>

*Table 4.15: In-situ δ¹⁸OVPDB data (‰) for non-reservoir iv in Well 4B. Key: † - syntaxial; ‡ - equant.*
cement zones have an average composition of -11.2 ‰ and -11.8 ‰ respectively (Table 4.6).

14 oil inclusions are observed in oil reservoir ii of Field B, these inclusions are initially observed within cement zone 7 which has an average $\delta^{18}O_{VPDB}$ composition of -7.4 ‰ (Table 4.11). Although reservoir v and iv are thought to be oil bearing, no oil inclusions were observed.

### 4.4.1.3 Cement stratigraphy

#### 4.4.1.3.1 Field A

Oil bearing reservoir interval i in Well 1A revealed 8 cement zones (Figure 4.7a) that are all volumetrically small. The first cement zone has a $\delta^{18}O_{VPDB}$ composition of -1.2 ‰ (n=1) and then progressively decreases to the final cement zone which has a $\delta^{18}O_{VPDB}$ of $-9.4 \pm 0.2$ ‰ (n=2) (Figure 4.8).
Figure 4.7: Cathodoluminescence photographs for a) oil reservoir i, b) aquifer i, c) oil reservoir ii and d) aquifer ii. The final cement zone in both aquifer intervals has greatly reduced porosity. A significant volume of pore occluding cement is occluding the intergranular porosity of interval vii (e,f), these cements have been etched by a subsequent dissolution event.
Figure 4.8: $\delta^{18}O_{VPDB}$ values for each cement zone (numbered) in a) oil bearing reservoir i and in b) the coeval aquifer. $\delta^{18}O_{VPDB}$ values for c) the oil bearing reservoir ii and d) its coeval aquifer and e) the $\delta^{18}O_{VPDB}$ values for reservoir vii. All transects show a progressive decrease in $\delta^{18}O_{VPDB}$ from the oldest to the youngest cement zone.

The $\delta^{18}O_{VPDB}$ composition of aquifer i is similar to the oil bearing reservoir interval, with a progressive decrease in $\delta^{18}O_{VPDB}$ during cement precipitation from $-2.2 \pm 0.4 \%$ (n=2) in cement zone 1 to $-9.0 \pm 0.3 \%$ (n=4) in cement zone 8 (Figure 4.8b). The main pore occluding cement zone in aquifer interval i is the youngest cement zone, cement zone 8 (Figure 4.7b), this is in contrast to the youngest cement zone in the coeval oil reservoir that has not significantly reduced porosity.

Only the main pore occluding cement zones were assessed in reservoir ii of Well 2A because the other cement zones were missed (Figure 4.7c). However, both cement zones show very negative values and have similar average $\delta^{18}O_{VPDB}$ compositions of $-11.2 \pm 0.3 \%$ (n=2) and $-11.8 \pm 0.1 \%$ (n=23) respectively (Figure 4.8c). These $\delta^{18}O_{VPDB}$ values are ~ 2 \% lower than in the overlying oil bearing reservoir i.

The coeval aquifer to oil bearing reservoir ii has a similar $\delta^{18}O_{VPDB}$ evolution to aquifer interval i, with a progressive decline in $\delta^{18}O_{VPDB}$ from $-2.7 \%$ (n=1) to $-8.9$
± 0.1 ‰ (n=2) (Figure 4.8d). The main pore occluding cement zones in this aquifer are cement zones 7 and 9 (Figure 4.7d). The δ¹⁸O_VPDB values of the cement zones in the oil bearing reservoir ii are lower than those obtained in this aquifer.

Reservoir interval vii has seven cement zones, the average δ¹⁸O_VPDB of the first cement zone is -3.4 ± 0.3 ‰ (n=5) with the δ¹⁸O_VPDB of the subsequent cement zones then progressively decreasing to -7.8 ‰ (n=1) by cement zone 7 (Figure 4.8e). Cement zones 6 and 7 are the greatest pore occluding cement zones identified (Figure 4.7c,f).

4.4.1.3.2 Field B

8 cement zones were identified in aquifer interval i, with the final cement zone being pore filling (Figure 4.9a, b). The δ¹⁸O_VPDB of these cements decreases from -4.4 ± 0.1 ‰ (n=2) in the third cement zone to -8.8 ± 0.2 ‰ (n=14) in the youngest cement zone (Figure 4.10a).
Figure 4.9: Cementation has greatly occluded the porosity of: a), b) aquifer i, c) oil reservoir iv, d) oil reservoir v, e) aquifer viii and f) non-reservoir iv. The main pore occluding cement zone in all these samples is the final cement zone. Dissolution has then enhanced the porosity in these reservoirs, this can be observed in oil reservoir iv (c).
Figure 4.10: The $\delta^{18}O_{VPDB}$ compositions for each of the reservoir intervals in relation to the cement zone analysed (numbered). The $\delta^{18}O_{VPDB}$ values progressively decrease into the youngest cement zones. Reservoir interval ii shows a higher $\delta^{18}O_{VPDB}$ composition in the final zone from -11 ‰ in the penultimate to -8 ‰ in the final zone.

Oil reservoir interval ii has 11 cement zones, these cement zones decrease in $\delta^{18}O_{VPDB}$ from -4.8 ‰ (n=1) in the third cement zone to -10.6 ± 0.2 ‰ (n=3) in the penultimate cement zone, this $\delta^{18}O_{VPDB}$ than increases into the final cement zone where a $\delta^{18}O_{VPDB}$ of -7.6 ± 0.4 ‰ (n=2) is observed (Figure 4.10b).

9 cement zones were identified in oil reservoir interval iv, the $\delta^{18}O_{VPDB}$ composition of these cement zones decreases from -5.9 ± 0.1 ‰ in the fourth cement zone to -9.0 ‰ (n=1) in the youngest cement zone (Figure 4.10c). Cement zones 7 and 8 have nearly led to the complete occlusion of porosity (Figure 4.9c).

Oil reservoir interval v has 6 cement zones, the first cement zone has a $\delta^{18}O_{VPDB}$ of -2.5 ‰ (n=1) this decreases into the youngest cement zone where the average $\delta^{18}O_{VPDB}$ obtained is -7.8 ± 0.2 ‰ (n=12) (Figure 4.10d). Cement zones 5 and 6 are typically pore filling (Figure 4.9d).
6 cement zones were identified in aquifer viii the $\delta^{18}O_{VPDB}$ of the first cement zone is -3.8 ‰ (n=1) this progressively decreases into the youngest cement zone where a $\delta^{18}O_{VPDB}$ of -8.3 ± 0.1 ‰ (n=5) is observed (Figure 4.10e). The final cement zone is the main pore occluding cement zone (Figure 4.9e).

5 cement zones were identified in non-reservoir interval iv (Figure 4.10f). The oldest cement zone has a $\delta^{18}O_{VPDB}$ of -0.6 ‰ (n=2), the following two cement zones increase in $\delta^{18}O_{VPDB}$, with the $\delta^{18}O_{VPDB}$ decreasing into the final main pore occluding cement zone where a $\delta^{18}O_{VPDB}$ of -0.8 ± 0.2 ‰ (n=2) is observed (Figure 4.9f).

4.4.1.4 $\Delta^{18}O_{VPDB}$ in relation to cement type

Not surprisingly, cement types show distinctive isotope signatures (Tables 4.16 and 4.17). In Field A the fringing cements hold the most positive $\delta^{18}O_{VPDB}$ value of -1.2 ‰, with the most negative $\delta^{18}O_{VPDB}$ being -4.0 ‰. The most positive $\delta^{18}O_{VPDB}$ for equant calcite cements is -2.7 ‰, with the most reduced $\delta^{18}O_{VPDB}$ value being -8.5 ‰, the average $\delta^{18}O_{VPDB}$ value for equant calcite cements is -5.0 ± 0.4 ‰ (n=22). The most positive $\delta^{18}O_{VPDB}$ value for syntaxial cements is -1.8 ‰, with the most reduced being -9.9 ‰; the average $\delta^{18}O_{VPDB}$ value for syntaxial cements is -5.6 ± 0.4 ‰ (n=32). The syntaxial cements usually contain the most complete record of cementation however when large pore filling cements are present the pore space into which the cement was growing is occluded and cementation ceases, implying that the pore filling blocky cements can post date the syntaxial cements. Blocky calcite cements are the youngest cement phase to precipitate and have a $\delta^{18}O_{VPDB}$ range of -5.2 to -13.5 ‰, the average $\delta^{18}O_{VPDB}$ for blocky calcite cement is -9.9 ± 0.3 ‰ (n=44).
Table 4.16: $\delta^{18}O_{VPDB}$ (‰) compositions for the calcite cements in the reservoir intervals of Field A.

The $\delta^{18}O_{VPDB}$ of the calcite cements in Field B follow a similar succession to that in Field A. The range over which fringing calcite cements precipitate is -2.5 ‰ to -3.8 ‰, the average composition is -3.1 ± 0.6 ‰ (n=2). The most positive $\delta^{18}O_{VPDB}$ for syntaxial cements is -4.3 ‰, with the most negative $\delta^{18}O_{VPDB}$ composition being -9.1 ‰, the average $\delta^{18}O_{VPDB}$ for syntaxial cements is -6.1 ± 0.5 ‰ (n=8). Equant calcite cements began to precipitate at a similar $\delta^{18}O_{VPDB}$ to syntaxial cements with cementation continuing to $\delta^{18}O_{VPDB}$ values of -4.3 ‰, the average $\delta^{18}O_{VPDB}$ value is -6.0 ± 0.2 ‰ (n=25). Blocky calcite cements began to precipitate at a similar $\delta^{18}O_{VPDB}$ to equant and syntaxial cements, but continued to precipitate to a $\delta^{18}O_{VPDB}$ of -10.9 ‰, the average $\delta^{18}O_{VPDB}$ composition for blocky calcite cement is -7.7 ± 0.1 ‰ (n=109).

Table 4.17: $\delta^{18}O_{VPDB}$ (‰) compositions for the calcite cements in the reservoir intervals of Field B.
4.4.2 Discussion

4.4.2.1 $\Delta^{18}O_{VPDB}$ of Early Cretaceous seawater

The first non-luminescent cement zone in each of the cycles is indicative of shallow marine cementation, where increased fluid circulation and oxidizing environments preclude the incorporation of Fe$^{2+}$ and Mn$^{2+}$ in the carbonate lattice (Moore, 1989). This cement zone is observed in all reservoir intervals, but is primarily observed in the grain-supported packstone and grainstone depositional fabrics. The $\delta^{18}O_{VPDB}$ of this first cement zone, if the cement zone formed in equilibrium with Early Cretaceous seawater, should have a $\delta^{18}O_{VPDB}$ similar to seawater.

There is a progressively more negative $\delta^{18}O_{VPDB}$ composition for this same, non-luminescent cement zone into the older reservoir intervals of both Field A and Field B (Figure 4.11). In Field A, interval vii has an average $\delta^{18}O_{VPDB}$ of $-3.4\pm0.3$‰ (n=5); interval ii a $\delta^{18}O_{VPDB}$ of $-2.7$‰ (n=1) and interval i a $\delta^{18}O_{VPDB}$ of $-1.9\pm0.4$‰ (n=3). In Field B the first non-luminescent cement zone in reservoir interval viii has a $\delta^{18}O_{VPDB}$ of $-3.8$‰ (n=1), reservoir interval v has a $\delta^{18}O_{VPDB}$ of $-2.5$‰ (n=1) and in the cemented non-reservoir interval iv a $\delta^{18}O_{VPDB}$ of $-0.6$‰ (n=2). The first non-luminescent zone was not successfully sampled in younger reservoir intervals of Field B. However, the data from Cox et al. (2010) for oil reservoir interval i of Field B also supports the progressive increase in $\delta^{18}O_{VPDB}$ with a value of $-1.2$‰ obtained for the first cement zone in the oil bearing reservoir interval. However, in this same assessment the non-luminescent cement zone of the aquifer, has a $\delta^{18}O_{VPDB}$ of $-6.6$‰, this value is too negative for Cretaceous marine values at shallow depths, suggesting precipitation from more evolved or meteoric water. The average $\delta^{18}O_{VPDB}$ composition for the first cement zone is in good agreement with that of Early Cretaceous seawater (Pucéat et al., 2003) (Figure 4.11).

The first cement is therefore likely to have precipitated from pore fluids similar to Early Cretaceous seawater and is likely recording the changing $\delta^{18}O_{SMOW}$ in the Early Cretaceous. The most positive $\delta^{18}O_{VPDB}$ in non-reservoir iv and the youngest reservoir intervals are thought to be related to the development of ice in the
Hauterivian and Early Aptian (Pucéat et al., 2003). These episodes can be related to progressively shallowing upward cycles in the Hauterivian (Scott et al., 1988) and Early Aptian (Alsharhan and Kendall, 1991; Haq et al., 1988; Scott et al., 1988; Strohmenger et al., 2006; Vail et al., 1977).

Figure 4.11: $\delta^{18}O_{SMOW}$ and $\delta^{18}O_{VPDB}$ for Cretaceous carbonates at different latitudes. The figure has been adapted from Pucéat et al. (2003) with a revised time-scale (Ogg et al., 2008), and the y-axis changed to display the $\delta^{18}O_{VPDB}$. The average $\delta^{18}O_{VPDB}$ composition (with standard error) for the first non-luminescent cement zone in Fields A and B (cross and triangle respectively) have then been included on the figure for each reservoir or non-reservoir interval sampled. Note the presence of glendonites that are thought to be indicative of glacial episodes (Pucéat et al., 2003).
4.4.2.2 Cementation of a non-reservoir

The calcite cements in a cemented grainstone from non-reservoir interval iv of Field B have a lower number of cement zones and have higher $\delta^{18}$O$_{VPDB}$ values (Figure 4.10f) than that observed in the reservoir intervals (Figure 4.10c). The cement phases observed within the non-reservoirs are typically syntaxial and equant calcite cements that are composed of 5 cement zones, do not contain any oil inclusions and have $\delta^{18}$O$_{VPDB}$ values between 0 ‰ and -1.0 ‰.

The positive $\delta^{18}$O$_{VPDB}$ values can be associated with climatic cooling and the accumulation of polar ice (Immenhauser et al., 2003) and are in good correspondence with that of Pucéat et al. (2003) (Figure 4.11; cycle iv, Field B). Therefore, the simple explanation is that the cements present within non-reservoir interval iv were precipitated from water that was similar to Early Cretaceous seawater. This would suggest that the intergranular pore space was fully occluded by cements that formed from pore fluid that had a similar composition to Early Cretaceous seawater. This interpretation agrees with the $\frac{n^mMg}{m^mCa}$ ratios obtained for this sample (Section 4.3.2.6) and is also suggested by the Minus Cement Porosity (MCP) assessment, where the intergranular MCP of the sample is ~39 %, which suggests the complete occlusion of porosity at < 200 m burial depth (Figure 3.23, Section 3.3.2).

The cause for extensive intergranular cementation penecontemporaneous with the deposition of carbonate sediment is equivocal, but typically requires several prerequisites: very low sedimentation and erosion rates, high fluid flow rates, appropriate water temperatures and salinities, biogenic activities which stabilise substrates and prevent reworking of the grains (Flügel, 2004). Once the grainstone was fully cemented it is likely to have behaved as a barrier to flow. These cemented reservoir intervals, in combination with the underlying shales and mudstones of the TST would have probably been effective seals and would have limited flow between successive reservoir intervals, causing the reservoir intervals to behave as relatively closed systems.
4.4.2.3 Evolution of pore fluid $\delta^{18}O_{SMOW}$ during burial

The $\delta^{18}O_{VPDB}$ compositions for each calcite cement type present in the reservoirs of Field A and Field B are shown in Figure 4.12. This figure shows that the average $\delta^{18}O_{VPDB}$ for fringing, equant and blocky calcite cements follows the relative order shown by the paragenetic sequence (Figure 3.1), with the oldest cements containing a $\delta^{18}O_{VPDB}$ in equilibrium with Cretaceous marine seawater, and the youngest cements having a $\delta^{18}O_{VPDB}$ that suggests precipitation from warmer water. Curtis (1978) noted that the progressive depletion of heavier isotopes from early to late cements can be expected if cement precipitation continues during burial to progressively increasing temperatures.

\[
1000 \ln \alpha_{(\text{calcite-water})} = 18.03 \left(10^3 T^{-1}\right) - 32.42 \quad \text{(equation 1)}
\]

Fringing cements are thought to precipitate in the Marine Phreatic Zone (MPZ) from pore water at temperatures similar to Cretaceous marine water (15-25
°C: Pucéat et al. (2003) and Wygrala (1989)) (Figure 3.1). Equant calcite cements are thought to primarily precipitate in the Shallow Burial Zone (SBZ) which has an upper temperature limit of ~ 67 °C (onset of saddle dolomite formation; Section 3.2.3), with the blocky cements thought to primarily precipitate in the Deep Burial Zone (DBZ). The maximum burial temperature reached was obtained by Neilson et al. (1998) via fluid inclusions, and is thought to be ~140°C. The average temperature for each of these diagenetic zones and the average $\delta^{18}O_{VPDB}$ composition for the cement phases can be plotted on an equilibrium diagram to estimate the evolution of $\delta^{18}O_{SMOW}$ during burial (Figure 4.13).

Figure 4.13: Temperature vs. $\delta^{18}O_{SMOW}$ plot for the $\delta^{18}O_{VPDB}$ of the calcite cements present in Field A and B (Section 4.4.1.4). The trend lines are plotted using equation 1, which relates temperature, pore fluid $\delta^{18}O_{SMOW}$ and the $\delta^{18}O_{VPDB}$ of the calcite cements. Key: MPZ – Marine Phreatic Zone, SBZ – Shallow Burial Zone, DBZ – Deep Burial Zone.
The plot shows a continuous evolution of pore fluid, during successive cement precipitation, towards higher $\delta^{18}\text{O}_{\text{SMOW}}$ values. The $\delta^{18}\text{O}_{\text{SMOW}}$ of the pore fluid is likely to become higher with burial as more cement is precipitated (Heydari and Moore, 1993; Hudson, 1977). Therefore, the simple explanation implies that formation waters evolved in-situ in a relatively closed system during the precipitation of calcite cement and were not significantly affected by the introduction of waters with isotopically different compositions.

4.4.2.4 The effect of oil charge on calcite cementation

There is typically a greater volume of pore occluding cement in coeval aquifers of both Field A and B, than in the oil bearing reservoir intervals (Section 3.3.3.2), this is also observed in the samples assessed here (Table 4.1 and 4.2). The intergranular porosity in the aquifers is mostly occluded by the final pore occluding cement zone (Section 4.4.1.3) and the remaining porosity is now in moldic, vuggy or microporosity (Table 4.1 and 4.2) that are interpreted to have formed relatively recently (Section 3.2.3). Whereas in the oil reservoirs a lower volume of cement is present and no final pore filling cement zone has formed, this has resulted in a greater volume of primary intergranular porosity remaining (Section 3.3.3.2).

The cause for porosity preservation in carbonate reservoirs can be due to several reasons, including: rapid burial so that the sediments pass quickly through the near-surface phreatic zone and undergo little early cementation (Harris et al., 1985); limited fluid migration due to bounding seals that limit the transport of solutes causing cementation rate to decrease (Feazel and Schatzinger, 1985); and early oil charge which displaces the pore water, causing cementation to slow or cease (Feazel and Schatzinger, 1985; Gluyas et al., 1993; Neilson et al., 1998).

Oil charge is thought to be the primary reason for the difference in cement volume between the oil bearing reservoirs and coeval aquifers in both Field A and Field B, this supports the observations of Neilson et al. (1996) and Neilson et al. (1998). Oil charge controversially prevents or significantly retards the precipitation of cement as most minerals are insoluble in hydrocarbon fluids (Heasley et al., 2000). When oil replaces pore water it therefore replaces the medium of solute transport
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(pore water) needed for the effective transport of solutes during chemical compaction. Burial processes (e.g. chemical compaction) provide the majority of solutes available for cementation (Meyers, 1978) and are typically observed to be more abundant in the flanks of the fields than the crests (Grotsch et al., 1998; Oswald et al., 1995). Oil charge can therefore reduce the amount of solute available for cementation and reduce cementation rate thereby preserving porosity.

The δ\(^{18}\)O\textsubscript{VPDB} values for the cements can be used to help understand the effect that oil charge has had on preserving porosity. The most negative δ\(^{18}\)O\textsubscript{VPDB} observed for aquifer interval i of Field A is -9.9 ‰ this is similar to that in the coeval oil bearing reservoir interval where a δ\(^{18}\)O\textsubscript{VPDB} of -9.6 ‰ is obtained. The temperature of precipitation can be estimated using equation 1 and a δ\(^{18}\)O\textsubscript{SMOW} of ~+6 ‰ for the pore fluid in the deep burial zone (Figure 4.13), this suggests that cementation continued to a similar temperature of ~ 111 °C and ~ 113 °C in the oil reservoir and aquifer respectively. This agrees with the \(^{m}\)Mg/\(^{m}\)Ca data where cementation in the oil reservoir and coeval aquifer was interpreted to continue to a similar temperature (Section 4.3.2.6).

This temperature estimate is only an approximation because although it has been shown that the cements formed in a relatively closed system, where external fluid has not greatly affected the composition of the pore waters during calcite precipitation, the cements are most likely derived from a local source (Section 3.4). Therefore the δ\(^{18}\)O\textsubscript{SMOW} of the pore water is likely to have been partially buffered by local rock-water interaction.

A similar conclusion to that reached in reservoir i of Field A is also reached for reservoir i in Field B. The δ\(^{18}\)O\textsubscript{VPDB} of the calcite cements present in aquifer interval i of Field B were compared to the data for the coeval oil bearing reservoir interval in Field B which is presented in Cox et al. (2010). These results correspond with that of Cox et al (2010), with similar negative δ\(^{18}\)O\textsubscript{VPDB} compositions being obtained in the youngest cement zones in the coeval oil bearing reservoir interval (-10.3 ‰: Cox et al. (2010)) and in the aquifer (-9.8 ‰). Using a δ\(^{18}\)O\textsubscript{SMOW} of ~+6 ‰ for the final burial cement (Figure 4.13), precipitation in the oil reservoir continued to a temperature of 117 °C, whereas in the aquifers cementation continued to 112 °C.
This suggests that cementation continued to a similar temperature in both the oil reservoir and aquifer and is in agreement with the $^{\text{m}}\text{Mg}/^{\text{m}}\text{Ca}$ data (Section 4.3.2.6).

Oil reservoir interval $i$ is at a similar depth in both Field A and Field B ~2.2-2.4 km, and therefore it should be possible to compare the data sets. However, oil is thought to enter the reservoir interval at a much earlier stage in Field B (~5 ‰: Cox et al. (2010)) than in Field A (~8.8 ‰). Therefore, it would appear that cementation continued in the presence of oil over a much greater pore water $\delta^{18}\text{O}_{\text{VPDB}}$ composition in Field B than in Field A, where oil entered the reservoir much later.

The cause for continued cementation in the oil reservoirs after oil charge may be related to the wettability of the system. Marzouk et al. (1995) and Cuiec and Yahya (1991) suggest that macropores are typically oil wet, mesopores are mixed-wet and micropores are water-wet. The micropore systems in the oil bearing reservoirs remain water-wet due to high capillary entry pressures which typically produce high irreducible water saturations (Heasley et al., 2000; Marzouk et al., 1995). These water-saturated micropores may be ineffective for the advective mass transfer of solutes; however, in the burial diagenetic environment where calcite is sourced mainly by local pressure dissolution they may provide a pathway for the diffusion of solutes even when macropores are partially oil-filled. Within carbonate reservoirs from the U.A.E., it has been shown that the aquifer possesses a microporosity-dominated system due to the occlusion of macroporosity by burial cements (Cantrell and Hagerty, 1999; Neilson et al., 1996), whereas in the oil leg a dual porosity system is retained in the oil leg reflecting limited macropore cementation (Cox et al., 2010; Heasley et al., 2000; Neilson et al., 1996).

During oil charge in carbonate rocks, a change in porewater movement from one dominated by macro- and mesoporosity to a mixed-wet mesopore system and a solely water-wet micropore system will increase the tortuosity of fluid flow so reducing permeability and limiting the diffusional supply of carbonate. Solutes will take longer to reach potential sites of precipitation, causing cementation rate to progressively decline leading to the cement zones becoming thinner and coated with oil. Ultimately, cementation in these areas within the crest ceases leaving most interparticle pores open and cement free. By contrast in the aquifers, the increasing
volume of solutes provided by pressure solution promotes precipitation of the thickest cements zones, so eventually occluding most of the remaining macropores.

Contrary to a previous study for Field B (Cox et al., 2010) and to the observations in reservoir interval i of Field A, the δ18OVPDB of oil reservoir and aquifer interval ii of Field A do not reach similar negative values: the most negative δ18OVPDB of the calcite cements within the aquifer yield a value of -9.0‰ but in the coeval oil reservoir the most reduced δ18OVPDB value for the cements is -13.5‰; similar negative values are also observed in the oil reservoir of Field B (-10.9‰) however no coeval aquifer sample was available. Using +6‰ as the δ18OVPDB composition of pore waters (Figure 4.13), precipitation of the blocky calcite cements in oil reservoir interval ii in Field A ceased at ~145°C compared with ~106°C within the aquifer. This suggests that cementation has not continued within the aquifer and oil reservoir to the same burial depths and temperatures, but that further cements formed selectively within oil reservoir ii with more negative δ18OVPDB values. This conclusion is also supported by the \(^{26}\)Mg/\(^{40}\)Ca ratios that suggest cementation continued in the oil reservoir after it stopped in the aquifer (Section 4.3.2.6).

The maximum temperature to which cementation continued in oil bearing reservoir interval ii is within the temperature range suggested by fluid inclusion and vitrinite reflectance data. Fluid inclusion studies provide a maximum burial temperature of ~140°C (Neilson et al, 1998) for the Thamama Group. The overlying Bab member has a vitrinite reflectance value of 0.75-0.8 (Lijmach et al, 1992), which corresponds to a temperature of ~100-140°C (Baker et al, 1990). This may suggest that δ18OVPDB accurately records continued cementation at the maximum temperature reached by the reservoir.

However, the cause for the lower δ18OVPDB values in oil reservoir ii of Field A is equivocal and can also be explained by: 1) Hydrothermal activity increasing the temperature of precipitation for oil reservoir ii, but this requires hydrothermal activity to only affect the oil reservoir and not its coeval aquifer – this is unlikely. 2) More saline fluid entering this reservoir that increased the δ18O_SMOW of the pore fluid. 3) Oil charge entered the pore space later than in oil reservoir i, as suggested by the
oil inclusions, and allowed for continued cementation to higher burial depths and temperatures.

4.4.2.5 Stratigraphic trends in cement dynamics

Section 4.4.2.1 has shown that the $\delta^{18}O_{SMOW}$ of Early Cretaceous seawater likely varied quite considerably during the deposition of the successive depositional cycles. Therefore, in order to remove the effect of changes to the initial pore fluid $\delta^{18}O_{SMOW}$, the range of $\delta^{18}O_{VPDB}$ over which cementation has occurred has been calculated for each sample (Table 4.18).

Table 4.18 suggests that the cements in the Lekhwair Formation formed over a much smaller $\delta^{18}O_{VPDB}$ range and would suggest that cementation ceased in the Lekhwair Formation (cycles ix–iv) prior to the Kharaib (cycles iii–ii) and Shu’aiba’ Formations (cycle i). This interpretation also agrees with $^{26}Mg/^{44}Ca$ ratios (Section 4.3.2.6) and the $\delta^{18}O_{VPDB}$ of bulk micrite in the reservoir and non-reservoir intervals (Tables 4.1 and 4.2 and Figure 3.36) which suggest cementation and rock-water interaction continued to higher temperatures in the shallowest reservoirs (Section 3.4.1). There are also fewer cement zones in the Lekhwair Formation than in the Kharaib and Shu’aiba’ Formations which may also indicate that cementation ceased in the older reservoirs prior to the youngest reservoirs. However, there is more iron in the cement phases of the Lekhwair Formation in Field B (Section 4.3.1) which causes the cements zones to be duller than the cement zones in the younger reservoirs and less clearly visible (Figure 4.9e) so this may be an artefact.
<table>
<thead>
<tr>
<th>Field A</th>
<th>Cycle</th>
<th>Interval</th>
<th>Average $\delta^{18}$O of cement zone 1 (‰)</th>
<th>Most negative $\delta^{18}$O (‰)</th>
<th>Range (‰)</th>
<th>Number of points</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>i</td>
<td>Oil res</td>
<td>-1.2</td>
<td>-9.6</td>
<td>8.4</td>
<td>15</td>
</tr>
<tr>
<td></td>
<td>i</td>
<td>Aquifer</td>
<td>-2.2</td>
<td>-9.9</td>
<td>7.7</td>
<td>31</td>
</tr>
<tr>
<td></td>
<td>ii</td>
<td>Oil res</td>
<td>? (-2.7)</td>
<td>-13.5</td>
<td>10.8</td>
<td>25</td>
</tr>
<tr>
<td></td>
<td>ii</td>
<td>Aquifer</td>
<td>-2.7</td>
<td>-9.0</td>
<td>6.3</td>
<td>12</td>
</tr>
<tr>
<td></td>
<td>vii</td>
<td>Aquifer</td>
<td>-3.4</td>
<td>-7.8</td>
<td>4.4</td>
<td>19</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Field B</th>
<th>Cycle</th>
<th>Interval</th>
<th>Average $\delta^{18}$O of cement zone 1 (‰)</th>
<th>Most negative $\delta^{18}$O (‰)</th>
<th>Range (‰)</th>
<th>Number of points</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>i</td>
<td>Aquifer</td>
<td>? (-1.2)</td>
<td>-9.8</td>
<td>8.6</td>
<td>37</td>
</tr>
<tr>
<td></td>
<td>ii</td>
<td>Oil res</td>
<td>? (-2.7)</td>
<td>-10.9</td>
<td>8.2</td>
<td>22</td>
</tr>
<tr>
<td></td>
<td>iv</td>
<td>Oil res</td>
<td>? (-2.5 or -0.6)</td>
<td>-9.0</td>
<td>6.5 or 8.4</td>
<td>37</td>
</tr>
<tr>
<td></td>
<td>v</td>
<td>Oil res</td>
<td>-2.5</td>
<td>-8.7</td>
<td>6.2</td>
<td>23</td>
</tr>
<tr>
<td></td>
<td>viii</td>
<td>Aquifer</td>
<td>-3.8</td>
<td>-8.6</td>
<td>4.8</td>
<td>27</td>
</tr>
</tbody>
</table>

Table 4.18: The $\delta^{18}$OVPDB range over which the cements in an interval formed. The cements in the Lekhwair Formation typically precipitate over a smaller range than the cements in the Kharaib and Shu’aiba’ Formations, suggesting that cementation stopped in the Lekhwair Formation first. Note that when the first cement zone was not sampled the $\delta^{18}$OVPDB value for the coeval or closest reservoir interval is used; this value is shown in brackets.

Therefore the final, main pore occluding cement zone (Section 4.4.1.3), which led to a substantial reduction of porosity in the reservoirs of the Lekhwair Formation and the coeval aquifers of the Kharaib and Shu’aiba’ Formations is
thought to have precipitated at a lower temperature in the Lekhwair Formation than in the Kharaib and Shu’aiba’ Formations. The cause for this may be related to the increased abundance of iron in the cements of the Lekhwair Formation (Section 4.3.1).

The concentration of iron in a sample is thought to be proportional to the size of the stylolite, with a greater concentration of iron typically leading to the development of rectangular stylolites and lower concentrations leading to the development of wave-like stylolites (Alsharhan and Sadd, 2000). Therefore, because there is a greater abundance of iron in the Lekhwair Formation, there may be a greater number of stylolites with large amplitudes, which have contributed a greater volume of solutes to the pore fluid and allowed for the precipitation of a large volume of cement. Although stylolites are common throughout both fields and thought to be more dominant in aquifers (Oswald et al., 1995), the abundance and type of stylolites observed was not quantified in this assessment and requires further core assessment.

Another possible contributing factor to cementation ceasing in the oldest reservoirs prior to the younger reservoirs is a change in dominant primary pore type (Section 3.3.3.1). The porosity of the Lekhwair Formation is dominated by unoccluded intragranular porosity, this pore type is unlikely to contribute to effective porosity and therefore there is a lower initial effective porosity present in the Lekhwair Formation than in the Kharaib and Shu’aiba’ Formations, where a larger volume of intergranular porosity is present. Therefore for the same volume of cement, the effective porosity in the Lekhwair Formation will be occluded with cement prior to the Kharaib and Shu’aiba’ Formations.

4.4.2.6 Late stage dissolution

Late stage burial cements are known to have greatly reduced the porosity of the intervals that have not experienced an early oil charging event (Section 4.4.1.3). This late stage cementation event is thought to have occurred at a higher temperature in the stratigraphically younger aquifers. However, although cementation has
significantly reduced the initial porosity of the aquifers a significant porosity currently remains (Tables 4.1 and 4.2).

After the significant occlusion of porosity in the aquifers, the porosity has been enhanced by late stage dissolution (Figure 4.14). This event has led to the development of vuggy porosity and microporosity throughout Field A and B and has led to the etching of previously deposited cements (Section 3.2.3). This has undoubtedly increased the reservoir quality of the reservoir intervals and has likely led to the good reservoir quality that is currently observed in the Lekhwair, Kharaisb and Shu’aiba’ Formations (Table 4.1, 4.2). The origin of this dissolution event will be discussed in Chapter 5.

*Figure 4.14: Occluded pore space in a) oil reservoir and b) non-reservoir in these same samples dissolution has enhanced the porosity and led to the development of microporous kaolinite(c, d)) and vuggy and moldic porosity. The same pore occluding cement zone seen in b), is now observed to be etched in d), with microporous kaolinite now present in the secondary pore space.*
4.5 Conclusions

Fringing, equant and blocky calcite cements have distinct δ<sup>18</sup>O<sub>VPDB</sub> compositions that are most likely a result of cementation at increasing temperature. This suggests that fringing cement typically precipitates from pore fluid similar to Early Cretaceous seawater, that equant calcite cements precipitate prior to and during the deposition of blocky calcite cement and that with increasing temperature and burial, blocky calcite cements dominate. This agrees with the paragenetic sequence presented in Chapter 3 and suggests that the δ<sup>18</sup>O<sub>SMOW</sub> of the pore fluid progressively evolved in-situ to ~ 6 ‰ during successive cementation.

The initial non-luminescent cement zone observed in the reservoir and non-reservoir intervals has δ<sup>18</sup>O<sub>VPDB</sub> values close to equilibrium with Early Cretaceous seawater at surface temperatures. This suggests that the Lekhwair, Kharaiib and Shu’aiba’ Formations were deposited during 2 glacial episodes, with the Kharaiib Formation depositing during the interglacial episode.

The m<sup>24</sup>Mg/m<sup>40</sup>Ca ratio in Field A and Field B is a good geothermometer and agrees with the δ<sup>18</sup>O<sub>VPDB</sub> obtained for the cements, with both temperature proxies suggesting cementation at continually increasing temperatures, in a relatively closed system.

The calcite spars present in the non-reservoirs are early marine phases, as shown by the paragenetic sequence, the presence of only 5 cement zones, by their very positive δ<sup>18</sup>O<sub>VPDB</sub> composition (0 to -1.1‰) and high m<sup>24</sup>Mg/m<sup>40</sup>Ca ratios. This agrees with the minus cement porosity calculations undertaken in Chapter 3 which suggests early cementation. Therefore some non-reservoir samples are likely to behave as seals at an early stage, which may have contributed to field compartmentalisation.

In-situ δ<sup>18</sup>O<sub>VPDB</sub> and m<sup>24</sup>Mg/m<sup>40</sup>Ca data demonstrate that the cements in oil reservoir and coeval aquifer i of Field A likely precipitated to a similar temperature; ~ 111 °C and ~ 113 °C in the oil reservoir and aquifer respectively. Here, it is suggested that calcite cementation and oil charge occurred synchronously.
and that earlier saturation with oil retarded cementation in oil reservoir i and preserved reservoir quality. Reservoirs located down dip in a more distal position relative to the crest, underwent continuous compaction and cementation as the burial depth increased, until the effective porosity was occluded by the youngest main pore occluding cement zone. Whereas, in the oil bearing intervals cementation continued but at a much lower rate and no large pore occluding cement formed.

In-situ $\delta^{18}O_{VPDB}$ and $^{m}\text{Mg}/^{m}\text{Ca}$ data suggest that cementation continued to a higher temperature in oil reservoir interval ii of Field A (\(\sim 145^\circ\text{C}\)), after cementation ceased in the coeval aquifer (\(\sim 106^\circ\text{C}\)) and in reservoir i. This may be due to oil charge filling the pore space later, as suggested by oil inclusions or that a sufficient volume of oil was not emplaced to cause the complete cessation of cementation, as suggested in reservoir i.

$\delta^{18}O_{VPDB}$ and $^{m}\text{Mg}/^{m}\text{Ca}$ data suggest that the most complete calcite cementation history is preserved within the younger reservoir intervals. This is a result of a large volume of relatively early calcite cement precipitating in the older aquifers (viii) and oil and gas bearing reservoirs (vii, v, iv) of Field A and B, which has led to the near complete occlusion of pore space and to the cessation of cementation. This main pore occluding event is likely to have occurred in the Lekhwair Formation prior to the Kharaib and Shu’aiba’ Formations.

The samples where a large pore occluding cement zone is present currently have a high porosity this porosity is primarily in intragranular and secondary porosity. The remaining intragranular pores currently in the samples are un-occluded by any form of cement, this would suggest that during the formation of the main pore occluding cement zone the pores were isolated and so were not occluded by cement. The secondary porosity in the samples is typically in un-occluded vuggy porosity; this is associated with the dissolution event identified in Chapter 3.

The early cementation of the oldest reservoirs most likely stopped continued rock-water interaction in the oldest reservoirs prior to the youngest. This has resulted in the most reduced $\delta^{18}O_{VPDB}$ for bulk micrite being obtained in the Kharaib and
Shu’aiba’ Formations. This would in turn suggest that the $\delta^{18}O_{VPDB}$ of micrite can be used to provide an approximate time at which cementation ceased, however further work is required to confirm this.
CHAPTER 5
IMPROVING BASIN MODEL PREDICTIONS BY INTEGRATING DIAGENESIS: A CASE STUDY OF FIELD A

5.1 Introduction

Basin modelling is dynamic modelling of geological processes in basins over geological time. Basin modelling reduces the investment risk in oil and gas exploration by reliably predicting the presence, type and volume of hydrocarbon present in a trap (Al-Hajeri et al., 2009). The models can be used to predict, amongst other things: burial history, temperature evolution and the timing of source rock maturation and hydrocarbon migration.

Petromod™ is one of the most frequently used basin modelling packages (Baur et al., 2010; Frielingsdorf et al., 2008; Naeth et al., 2005; Skeie et al., 2004), and is one of the most sophisticated programs currently available to industry and academia. However like all widely used basin modelling software the predefined model does not, amongst other things, incorporate a representation of carbonate diagenesis. This is important to include because in a carbonate system diagenetic modifications can further, if not completely, alter the initial petrophysical properties of the depositional sediment (Chapter 3, 4). Here, the aims are to:

1) Investigate and illustrate the geological evolution of Field A through 2D basin modelling.

2) Determine the effect that including of an estimate of the diagenetic evolution of the Lekhwair, Kharaiib and Shu’aiba’ Formations has on the location of hydrocarbon accumulation. This will be achieved by developing three
models: the first model uses the porosity-depth/porosity-loss trend of Schmoker and Halley (1982) which is obtained from the limestones and dolomites in the South Florida Basin where porosity loss can be described by the exponential function \( \Phi = 41.73e^{-2.8197z} \). This porosity-loss curve takes into account chemical and mechanical compaction. The second model uses the mechanical compaction curves obtained from the Petromod™ database for a specified lithofacies. In the third model diagenesis is represented by switching the porosity-depth curves used during burial. The switched porosity depth curves will simulate the effect of oil charge in reducing cementation rate and preserving porosity in the oil reservoirs and in the aquifers the porosity-depth curve is switched to simulate the effect of rapid cementation on porosity occlusion. All three models will be run twice with each using two separate basal heat fluxes.

3) Evaluate the effect that changing the heat flux has on the time-temperature history of the basin and on the time of oil generation, migration and petroleum accumulation. This will be achieved by using two different heat fluxes that are both equally valid: the first assumes that the greatest heat flux is during subduction and ophiolite emplacement in the Late Cretaceous, whereas the second assumes that continental collision during the Tertiary is the cause of greatest heat flux.

4) Identify, through basin modelling alone, the time at which oil entered the reservoirs and determine the temperature of the reservoirs during this oil charging event. This will be compared to the temperature, as derived from the \( \delta^{18}O_{VPDB} \) geothermometer, for the first cement zone in which primary oil inclusions are observed (Chapter 4). This allows the two predictions as to the temperature of the reservoir at initial oil charge to be validated.

5) Use the models to predict the temperature evolution for the reservoir intervals through geological time and relate this to the \( \delta^{18}O_{VPDB} \) geothermometer (Chapter 4). The temperature of precipitation, as derived from the \( \delta^{18}O_{VPDB} \) geothermometer, for the youngest cement zone within each reservoir will be compared to the temperature evolution predicted by the model. This will be
used to help constrain the approximate time at which cementation ceased in the reservoir intervals.

6) Determine the basin wide flow patterns to help constrain the possible origin of the acidic fluid that caused widespread dissolution and led to the precipitation of kaolinite after cementation stopped (Chapter 3).

5.2 Tectono-stratigraphic Evolution of the U.A.E.

The determination of the basin type and tectono-stratigraphic framework of the basin precedes the construction of a basin model and encompasses information about plate tectonics, depositional environments, global climate and tectonic events. A summary of the tectono-stratigraphic framework of the U.A.E. is presented in this section.

5.2.1 Cambrian-Carboniferous

During the Cambrian the evaporites of the Hormuz Formation were deposited on the Arabian platform (Figure 5.1). These deposits were followed in the Cambro-Ordovician by siliciclastic sediments and argillaceous deposits that were accompanied by minor carbonates (Alsharhan and Nairn, 1997; Alsharhan et al., 2001).

Deposition during the Ordovician to the Early Carboniferous is dominated by very shallow marine sandstones with sub-ordinate shales and silts. Towards the South unconformities become more pronounced which may represent wide time gaps (Murris, 1980).
In the Late Carboniferous the Hercynian Orogeny resulted in uplift and erosion (Pinnington, 1981), and clastic (glacial) sedimentation dominated (Murris, 1980). At the end of the Hercynian Orogeny the climate became gradually warmer and more arid, and the sedimentary conditions changed to predominantly shallow-marine carbonates until the Holocene (Alsharhan and Scott, 2000).

5.2.2 Permian

Rifting events began in the Mid-Permian and caused the separation of central Iran, northwest Iran and part of Turkey from the Gondwanan margin, this led to the opening of the Neo-Tethys Ocean (Ali and Watts, 2009; Heydari, 1997b; Ziegler, 2001).
In the Neo-Tethys Ocean deposition from the Late Permian to the Middle Cretaceous (Turonian) was on a very stable platform (Alsharhan, 1993; Oswald et al., 1995) that was dominated by three types of positive (relative high) elements: 1) broad regional highs; 2) horsts and tilted fault blocks with bounding faults trending NNE-SSW; and 3) salt domes that are thought to be due to diapirism of the intra-Cambrian Hormuz salt during the Jurassic (Alsharhan et al., 2001; Alsharhan and Salah, 1997; Murris, 1980). Hormuz salt is not present in the vicinity of Field A, however diapirism of this salt is thought to have led to the formation of the trap in Field B (Alsharhan and Salah, 1997).

The first deposits to form in the Neo-Tethys Ocean are the carbonates of the Khuff Formation that are a result of a marine transgression during the Late Permian. Deposition was accompanied by short term sea-level oscillations that led to the development of evaporates (Figure 5.2) (Ziegler, 2001).
Figure 5.2: Chronostratigraphic chart for the Arabian Peninsula, modified from Ziegler (2001). Potential source and reservoir rocks are marked by the green and blue circles (Alsharhan, 1993; Lijmbach et al., 1992). The Late Jurassic Tuwaïq Mountain Formation is equivalent to the Dukhan Formation (Alsharhan and Magara, 1994) and is marked as a source rock.
5.2.3 Triassic

During the Early Triassic the climate became hotter and more arid and the dolomites, limestones, anhydrites and shales of the Sudair Formation were deposited (Figure 5.2). The restricted platform conditions present during the deposition of the Sudair Formation were maintained and the Jilh Formation was deposited during the Middle Triassic (Pinnington, 1981).

The Jilh Formation consists of microcrystalline dolomite interbedded with subordinate anhydrite, minor shales and argillaceous limestone. In the Rub Al Khali Basin the shallow marine shales and carbonates grade distally into platform dolomites (Ziegler, 2001).

The Late Triassic was a period when the deltaic clastics and interbedded shales of the Minjur Formation were deposited on the Arabian Platform, this is thought to indicate a widespread regressive phase (Pinnington, 1981). In the Rub Al Khali Basin this formation is represented by shallow water carbonates that show intermittent subaerial exposure surfaces (Ziegler, 2001).

Following from the early stages of rifting in the Mid-Permian (Section 5.2.2) late stage rifting and sea-floor spreading occurred in the Late Triassic-Early Jurassic (Figure 5.3). Late Triassic-Early Jurassic rifting on the Arabian craton is characterised by volcanism, half grabens and growth on the main border faults that led to break up (Robertson and Searle, 1990).
Figure 5.3: Tectonic history of the Arabian plate, modified from Alsharhan and Nairn (1997) after Marzouk and Sattar (1993). The late stage of rifting and sea-floor spreading occurred in the Late Triassic-Early Jurassic, subduction then begins in the Turonian which leads to folding on the Arabian plate. Collision then commences in the Eocene which leads to the closure of the Neo-Tethyan Ocean and to the formation of the Zagros fold belt.
Chapter 5

5.2.4 Jurassic

At the onset of the Jurassic a major marine transgression led to the deposition of clastics and carbonates of the Marrat Formation in onshore Abu Dhabi (the Marrat Formation is not found offshore) (Alsharhan, 1989). Succeeding this is the Hamlah Formation that consists of a lower dolomite interval overlain by argillaceous limestone and subordinate shales (Alsharhan, 1989). The Izhara Formation follows conformably and consists of argillaceous limestone, mudstone and subordinate shales.

The Middle Jurassic spans the deposition of the Araej Formation in the U.A.E. which was deposited in an open marine, very shallow and warm environment while sea level was rising (Haq et al., 1988). The Araej Formation consists solely of carbonate facies that formed in a very stable environment (Figure 5.4).

The deposits of the Late Jurassic were deposited on the unconformity surface that marks the end of the Middle Jurassic. The deposits of the Late Jurassic are composed of a thick sequence of carbonates and evaporates that mark the transition from a deep water environment (Tuwaq Mountain/Dukhan Formation) to a shallow water carbonate to supratidal depositional environment (Arab Formation) (Pinnington, 1981; Taher, 1996; Ziegler, 2001). The Hith anhydrite caps the Arab Formation and represents the final regressive supratidal phase of the Late Jurassic (Pinnington, 1981; Ziegler, 2001).
Figure 5.4: Depositional environments on the Arabian peninsula during the deposition of the Izhara and Araej Formations in the Middle Jurassic and the Tuwaiq Mountain/Dukhan Formation in the Late Middle Jurassic (Ziegler, 2001).
5.2.5 Cretaceous

Until the Middle Cretaceous deposition was on a passive margin (Figure 5.2) (Alsharhan and Scott, 2000). However, the tectonic regime changed during the Turonian when regional uplift and the onset of folding occurred in the U.A.E. (Figure 5.2, 5.3). This is thought to be a result of subduction of the Arabian Platform in the Turonian-Maastrichtian (Ziegler, 2001).

Subduction during the Campanian to Maastrichtian on the eastern edge of the Arabian craton caused the development of a foredeep along the Tethyan margin (Figure 5.3) and led to the emplacement of the Semail ophiolite (Alsharhan and Scott, 2000; Beydoun, 1991; Hessami et al., 2001). The cessation of ophiolite obduction is thought to have been in the Eocene (~ 63 Mya) (Ziegler, 2001). The deposits that formed during the Cretaceous have been discussed in Section 1.5.1.

5.2.6 Palaeogene-Quaternary

Active compression ceased and stable platform conditions continued on the Arabian platform during the Palaeocene-Eocene (Mouthereau et al., 2007; Murris, 1980), when the subducting slab of the Arabian plate breaks off (Agard et al., 2011).

Rapid erosion and subsidence of the emergent Semail ophiolite (Ziegler, 2001) and a general 2nd order transgression caused the deposition of a thick sequence (400-750 m) of sediments in the remnant foredeep (Alsharhan and Scott, 2000). The sediments are composed of shelly and bioclastic limestones and dolomites with thin shales that represent deposition in shallow water conditions, collectively these deposits comprise the Umm er Radhuma Formation (Alsharhan, 1993; Alsharhan and Scott, 2000; Ziegler, 2001). In some areas the shale cycles of the Umm er Radhuma Formation are terminated by anhydrite that formed in a sabkha (Figure 5.5).
Figure 5.5: The Arabian peninsula during deposition of the Umm er Radhuma Formation in the Late Palaeocene-Early Eocene and the DibDibba Formation in the Pliocene to Quaternary (Ziegler, 2001).
In the Middle Eocene a wide evaporitic platform existed on the Arabian Peninsula, that gave rise to the carbonate/evaporate sequence of the Rus Formation (Ziegler, 2001). After the deposition of the Rus Formation a transgressive phase resulted in the deposition of the nummulitic carbonates of the Damman Formation. The Umm er Radhuma, Rus and Dammam Formations constitute the Hasa Group.

During the Late Eocene/Early Oligocene Arabia and Eurasia converge (Figure 5.3) and part of the oceanic lithosphere that is still attached to Arabia is subducted beneath Eurasia (Agard et al., 2011). This collision progressively migrates to the SW, leading to deformation and topographic build-up, which results in the development of the Zagros fold belt (Agard et al., 2011; Hessami et al., 2001; Mouthereau et al., 2007; Oswald et al., 1995) and to the closure of the Neo-Tethys Ocean (Alsharhan and Nairn, 1997).

The Asmari Formation (Oligocene-Miocene) was deposited during convergence when a transgression resulted in the deposition of fossiliferous limestone. The overlying Fars Formation was deposited in a restricted depositional environment and is composed of anhydrite, with minor quantities of dolomite and limestone (Najafzadeh et al., 2010; Pinnington, 1981).

During the Upper Miocene/Pliocene slab break-off occurs at ~10 Mya to present, this is thought to have resulted in uplift and erosion over the Middle East (Agard et al., 2011; Mouthereau et al., 2012). The DibDibba Formation is the final deposit of the Rub Al Khali Basin and consists of Miocene and Pliocene carbonate, fluvial sandstone and conglomerate (Alsharhan et al., 2001; Pinnington, 1981).

5.3 Petroleum Systems Evaluation

A petroleum system is defined as a natural system that encompasses a pod of active source rock and all related oil and gas and which includes all the geologic elements and processes that are essential if a hydrocarbon accumulation is to exist (Magoon and Dow, 1994). This section identifies the petroleum system of Field A so that the basin model can accurately predict source rock maturation and hydrocarbon
migration and accumulation. This involves determining the source rock potential, the migration pathways, the plays and reservoirs and the traps and seals in the system.

5.3.1 Source rocks and maturation

There are a number of source rocks on the Arabian Peninsula (Figure 5.2). The three main source rocks that are thought to have produced the oil that has charged Field A are the Tuwaiq Mountain/Dukhan Formation (Alsharhan and Scott, 2000; Lijmbach et al., 1992; Taher, 1996), the organic rich Bab member of the Shu’aiba’ Formation (Lijmbach et al. (1992) and Abu Dhabi Company for Onshore Oil Operations (ADCO); personal communication) and the dense intervals of the Thamama Group (Lijmbach et al. (1992) and ADCO; personal communication).

5.3.1.1 Dukhan Formation

The Dukhan Formation was deposited in the Late Jurassic and is predominantly composed of argillaceous lime mudstones/wackestones and shales that formed in an anoxic environment (Section 5.2.4) (Alsharhan et al., 2000; Ziegler, 2001). Therefore type 2 (B) source rock kinetics have been applied in the model (Pepper and Corvi, 1995). The anoxic conditions allowed for the preservation of a relatively high Total Organic Carbon (TOC) (Ziegler, 2001) with TOCs as high as 5.5 % been recorded (Alsharhan, 1993), however an average value of 1-2 % has been used in the model (Alsharhan, 1993). The Hydrogen Index (HI) of the Formation is unknown and therefore a typical value of 600 mgHC/gTOC is used in the model.

The Dukhan Formation is thought to be the first source rock to mature. Maturation is thought to have initiated in the Late Cretaceous (Lijmbach et al., 1992; Milner, 1998; Murris, 1980; Nairn and Alsharhan, 1997; Oswald et al., 1995; Taher, 1996), and optimum maturity is thought to have occurred in the Eocene (Alsharhan, 1993; Alsharhan and Nairn, 1997; Taher, 1996). The gas generation window was reached in the Early Eocene (Alsharhan and Nairn, 1997), and it has remained in this window until present (Al-Suwaidi et al., 2000; Lijmbach et al., 1992).
The major expulsion event is thought to have been in the early Tertiary (50-40 Mya) (Al-Suwaidi et al., 2000; Oswald et al., 1995), with 75% of the total oils thought to have been generated during that time (Al-Suwaidi et al., 2000).

Although the Dukhan Formation is the main source rock in the region, volumetric calculations suggest that the Dukhan Formation is insufficiently rich to account for all the oil attributed to it (Taher, 1996). The Thamama dense intervals and the Bab Member are thought to have contributed to some of the oil in the Thamama Group (Lijmbach et al., 1992).

### 5.3.1.2 Thamama Group dense intervals

The dense intervals of the Thamama Group are composed of shales and mudstones (Chapter 2) that are thought to have been deposited in an open marine (Transgressive Systems Tract) to lagoonal environment (upper Highstand Systems Tract). The source rock kinetics most fitting to this depositional environment are type 2S (A) kinetics (Pepper and Corvi, 1995). The dense intervals have a TOC of 1.2% (Taher, 1996), the HI is unknown so a typical value of 600 mgHC/gTOC is used in the model.

### 5.3.1.3 Bab Member

The Bab Member is composed of pelagic lime mudstones and wackestones (Taher, 1996) that deposited in an intrashelf basin (Chapter 2). Restricted water circulation in this basin resulted in anoxic conditions (Aldabal and Alsharhan, 1989; Azzam and Taher, 1995) and led to the preservation of a relatively high TOC content of 1.5% (Taher, 1996) and a HI between 150-1000 mgHC/gTOC (Lijmbach et al., 1992); with a value of 600 mgHC/gTOC being used for the models. These characteristics have led to the Bab Member being viewed as an excellent type 1 or 2 source rock (Lijmbach et al., 1992); type 2 S(A) kinetics have been used in the model (Pepper and Corvi, 1995). Thermal modelling of the Bab Member by using the Lopatin/Waples algorithm (Waples, 1980) and the Arrhenius equation (Hunt et al., 1991) suggests that it entered the oil window in the early Tertiary between ~ 50-40
Mya (Azzam and Taher, 1995; Taher, 1996), and is presently in the oil window over most of the area (Lijmbach et al., 1992).

### 5.3.1.4 Other source rocks

The other source rocks present in onshore Abu Dhabi are summarised in Table 5.1. These source rocks are not thought to have contributed significantly to the hydrocarbon present in Field A.

<table>
<thead>
<tr>
<th>Source rock</th>
<th>Lithology</th>
<th>Depositional environment</th>
<th>TOC (%)</th>
<th>HI value (mgHC/gTOC)</th>
<th>Kinetics used</th>
</tr>
</thead>
<tbody>
<tr>
<td>Jilh</td>
<td>Shale</td>
<td>Continental – shoreline complex (Sharief, 1986)</td>
<td>2</td>
<td>100</td>
<td>Types 2 (B)</td>
</tr>
<tr>
<td>Minjur</td>
<td>Cross bedded sandstones with intercalated shales</td>
<td>Alluvial plain (Ziegler, 2001)</td>
<td>2</td>
<td>100</td>
<td>Type 2 (B)</td>
</tr>
<tr>
<td>Izhara</td>
<td>Shales</td>
<td>Shallow open marine (Ziegler, 2001)</td>
<td>2</td>
<td>100</td>
<td>Type 2 (B)</td>
</tr>
<tr>
<td>Shilaif</td>
<td>Pelagic carbonates</td>
<td>Anoxic intrashelf basin (Loutfi et al., 1987)</td>
<td>4</td>
<td>800</td>
<td>Type 2 (B)</td>
</tr>
</tbody>
</table>

*Table 5.1: Table showing the other source rocks present in onshore Abu Dhabi, with the TOC, HI and the kinetics used in the model. Data from Lijmbach et al. (1992) and Pepper and Corvi (1995), unless otherwise stated.*

### 5.3.2 Migration pathways

The principal process of oil migration for Field A is thought to be by lateral migration from the Falaha syncline (area 3 on Figure 5.6) (Al-Suwaidi et al., 2000; Alsharhan, 1993; Taher, 1996); lateral migration is defined here as flow within one unit that is sealed by an impermeable layer. The diagenetic analysis presented here
suggests that there are numerous effective intraformational seals (Chapter 3, 4) that are unlikely to allow for the migration of water between reservoir intervals; this is also suggested by Taher (1996) and by Alsharhan (1993) however, no direct evidence to support this is presented in these publications. If oil enters the reservoirs through lateral migration, but the non-reservoirs are effective seals, a question arises as to how hydrocarbon migrates into the reservoir intervals through the non-reservoirs.

Figure 5.6: Source areas, field locations and migration pathways of the hydrocarbon, modified from Alsharhan (1993). The location of the Hith anhydrite is also shown by the black dashed line and is labelled; Hith anhydrite is only present to the West of this line. The Hith anhydrite is thought to greatly affect the migration of hydrocarbon but is not present in the vicinity of Field A. The location of the cross section chosen for the basin model is shown by the red dashed line, with the location of the Falaha syncline and Field A labelled.
The buoyant pressure created by the hydrocarbons in a trap can exceed the resisting capillary pressure of the cap rock (Schowalter, 1979). If this is the case in Field A it will allow for the vertical migration of hydrocarbons through the non-reservoir zones. The Inter-Facial Tension (IFT) of the oil in Field A was determined to be 23 mN/m (Cuiec and Yahya, 1991), in contrast to an IFT of 42 mN/m that is used as standard for Petromod™; an IFT of 23 mN/m is used in all the models presented here. When the IFT of oil is lower the flow rates increase thus creating conditions for improved hydrocarbon migration (Haniff and Ali, 1990). Therefore, although the non-reservoirs restrict the migration of water between reservoir intervals, the migration of hydrocarbons through the same intraformational seals is possible.

Another possibility for the communication between the reservoir intervals is by migration along faults or open fractures. Field A contains numerous normal faults and fracture zones in the south, southeast and central part of the field (Edwards et al., 2006; Tamura et al., 2004); this is a common occurrence in cyclic carbonate successions (Harris et al., 1985). All the faults are normal and are thought to have been active until the Late Cretaceous (Alsharhan, 1993).

These faults may have allowed communication between successive reservoir intervals. However, if this is the case a single Oil Water Contact (OWC) may be expected in the field, but this is not observed (Section 1.6.1.1) (Alsharhan, 1993). It is therefore likely the faults in the field did not allow for significant communication between reservoir intervals, however migration along faults off the structure may have aided with migration between reservoir intervals.

5.3.3 Plays and reservoirs

Hydrocarbons have been identified in numerous formations (Figure 5.2), these formations include the: Simsima, Mishrif, Arab, Araej and Khuff Formations (Alsharhan, 1993). However, the reservoir intervals of the Lekhwair, Kharaid and Shu’aiba’ Formations (Cycles ix-i) are the focus of this study.
5.3.4 Trap and seal potential

The trap of Field A is an anticline which has a general dip of < 5° and is 26 km long and 9 km wide (Alsharhan, 1993). The structural trap is thought to have begun to form slightly after deposition in the Albian with significant growth of the field occurring during the Late Cretaceous, and minor growth continuing until present day (Alsharhan, 1993; Gumati, 1993). Extensive growth during the Late Cretaceous is thought to be related to the obduction of the Semail Ophiolite (Section 5.2.5), which generated a series of domes with a low structural dip that trend north to south (Oswald et al., 1995).

The Nahr Umr shale is a regional seal in the U.A.E. (Taher, 1996). The non-reservoirs of the Thamama Group also behave as intraformational seals that are thought to have led to compartmentalisation of the field (Chapter 4) (Alsharhan, 1993).

5.4 Basin Modelling Methodology

The first step to creating a 2D basin model is to input, as an image, a structural cross section that is appropriate for the area; this section is then digitised in Petromod™. Typically seismic sections are used to develop a cross section for an area. However, no seismic sections were available for Field A and its surroundings therefore an interpreted cross section from the literature (Ayoub and Nadi, 2000) is used that incorporates the entire petroleum system of Field A.

Three models will be created from this digitised section, with each model having different porosity-depth curves included for the reservoir and non-reservoir intervals of the Lekhwair, Kharaiib and Shu’aiba’ Formations. Model 1 will include a typical porosity loss curve for carbonates (Schmoker and Halley, 1982); Model 2 will include the porosity loss curves from the existing database within Petromod™ and the third model will attempt to include an understanding of diagenesis.

Two different Heat Flux (HF) profiles will also be simulated in all three models to determine the effect that changing the HF has on the time of petroleum generation and charge. The first HF profile relates increased HF to subduction and
ophiolite emplacement in the Late Cretaceous (Section 5.2.5). Whereas, continental collision has increased HF during the Oligocene to Miocene for the second HF profile and the obduction of the Semail ophiolite has led to a relatively minor increase in HF.

5.4.1 Cross section

Multiple cross sections transect Field A and the Falaha syncline (Al-Suwaidi et al., 2000; Alsharhan and Scott, 2000; Loutfi and Sattar, 1987), with all cross sections illustrating a similar general structure of a gentle, simple anticlinal fold (Field A) and the corresponding regional Falaha syncline trend. A typical cross section was chosen that is ~30 km in length and crosses the Falaha syncline and the width of Field A, this section is shown in Figure 5.7 and its location marked in relation to the U.A.E. on Figure 5.6.

![Cross section used for the basin model](image)

**Figure 5.7**: Cross section used for the basin model (Ayoub and Nadi, 2000), with the area of interest (petroleum system) labelled. Note the vertical exaggeration and the location of the Falaha syncline, and the anticlinal ridge that represents Field A.
The published cross sections typically group multiple formations into stratigraphic packages and so multiple formations can be assigned a single lithofacies, this will oversimplify the model. The contacts for other Formations such as the Umm Er Radhuma, base of the Damman Formation and for the Neogene sediments were interpreted on other cross sections (Alsharhan and Scott, 2000). Therefore in order to mediate the effects of over-simplification these contacts have been digitised and included on the cross section. This method allows the lithological variability to be more adequately modelled, thereby improving the model.

5.4.2 Data input

Horizons that were deposited from the Early Triassic to the present day have been identified in the foreland basin (Figure 5.8) (Ali et al., 2008). All these horizons were assigned the following from the public domain (Appendix 6):

1) A depositional start and end age.

2) Erosion start and end dates where appropriate; primarily from Alsharhan (1993).

3) Facies/Lithologies; individual lithofacies were mixed (Section 5.4.6). The mixing tool in Petromod™ allows a formation to be assigned a mixture of lithofacies as opposed to a single lithofacies. Therefore, when a formation is predominantly composed of two facies that are intermixed, the properties of the formation will be a mixture of both facies.

4) Petroleum systems components (source, reservoir, seal; see Section 5.3).

5) Source rock characteristics (Section 5.3.1) with at least two sub-divisions per layer.
Figure 5.8: Pre-gridded cross section, with the location of Field A and the Falaha syncline. The Dukhan Formation (black layer) is known to thicken into the Falaha syncline and so this has been represented in the models (Al-Suwaidi et al., 2000).

5.4.3 Boundary conditions

Boundary conditions need to be included in the model so that the heat, pressure and fluid flow through the entire simulated geologic history of the basin can be determined. The boundary conditions that are required by Petromod™ are the palaeo-water depth, sediment water interface temperature and heat flux of the basin through time.

5.4.3.1 Palaeo-Water Depth (PWD)

The Petromod™ default assumes that all the sediments in the system were deposited at sea-level. As eustatic and relative sea-level varies through time a Palaeo-Water Depth (PWD) consideration is required. The PWD interpretation (Figure 5.9) was developed based on the interpretation of the depositional environment at that time (Section 5.2), and is related to the global sea-level curves of Haq et al. (1988)
and to Google Earth’s palaeo-geometry tool. Simulations were run with variations to the PWD, but no observable differences between these model runs were identified.

![Figure 5.9: Palaeo-Water Depth through time for Field A.](image)

**5.4.3.2 Sediment-Water Interface Temperature (SWIT)**

Petromod™ contains a tool that calculates the Sediment Water Interface Temperature (SWIT) through time, this tool is based on the Ph.D. dissertation results of Wygrala (1989). The current latitude and geographic location are input into this tool and Petromod™ calculates the mean SWIT (Figure 5.10), taking into account climate change, continental position and the previously input PWD (Figure 5.9).

![Figure 5.10: Sediment Water Interface Temperature for Field A.](image)

**5.4.3.3 Heat Flux (HF)**

Time and temperature controls the maturity of source rocks (Waples, 1980). Therefore, the Heat Flux (HF) within the basin is an important consideration for determining when the source rocks matured, and when both primary and secondary migration commenced. Petromod™ is an advanced heat flux simulator, and for the first time in a numerical simulator, is using a new and more advanced approach to define heat flux that includes a finite stretching period and radioactive heat sources (Baur et al., 2010).
Ali and Watts (2009) found that the tectonic subsidence, uplift history and temperature evolution of the Rub Al Khali foredeep can be explained by the implementation of the McKenzie (1978) uniform extension model. The McKenzie (1978) uniform stretching model describes an initial stage of block faulting and subsidence due to rapid stretching of the continental lithosphere. The lithosphere then thickens by heat conduction to the surface and further subsidence occurs that is not associated with faulting. Ali and Watts (2009) suggest that initial rifting commenced 210 Mya in the Late Triassic (Section 5.2.3) and ceased at 180 Mya. They also suggest that the subsidence and thermal history of the Rub Al Khali foredeep can be described by applying a stretching factor, $\beta$, of 2.5 to the uniform stretching model (Ali and Watts, 2009). Petromod™ uses this information to create a HF profile that has a peak in heat flux at 180 Mya which is followed by relaxation and thermal subsidence.

This stretching model adequately accounts for deposition on the Arabian passive margin until the Late Cretaceous. However, to get the simulated maturity of the source rocks to correlate with the observed maturity it is necessary to modify the HF profile that is output from Petromod™ and insert two intervals of increased HF during the Late Cretaceous and at $\sim$ 10 Mya. The intervals of increased HF are attributed to contemporary tectonic events: subduction and ophiolite emplacement during the Late Cretaceous (Section 5.2.5) and collision and topographic build up from the Late Eocene/Early Oligocene until slab break off at $\sim$ 10 Mya (Section 5.2.6) (Agard et al., 2011; Alsharhan and Nairn, 1997; Mouthereau et al., 2012; Okay et al., 2010).

Although there is no direct evidence for changes in HF in the U.A.E. (Alsharhan and Nairn, 1997), evidence does exist to suggest that subduction and continental collisions can cause an increased HF (Collins, 2002; Thompson and Connolly, 1995; Thompson et al., 1997; Vanderhaeghe et al., 2003). The HF values assigned to these events are kept within the limits for an active margin and for continental collision by using the typical HF values for a sedimentary basin in this setting (Figure 5.11).
To determine the effect of changing the HF in the basin, and the effect that this has on the timing of maturation and migration of the hydrocarbons, two different HFs are applied to the models. In the first HF profile (HF 1) subduction and ophiolite emplacement led to an increased HF in the Late Cretaceous (Figure 5.12a). Whereas, continental collision has increased HF during the Oligocene to Miocene for the second HF profile (HF 2) (Figure 5.12b). Because the HF of the basin through time is unknown, both HFs are valid and are therefore simulated in the models to help determine the effect that changing the HF has on the time of source rock maturation.

Figure 5.11: Typical heat fluxes for the different types of sedimentary basins (Allen and Allen, 2005).
5.4.4 Porosity depth curves for Models 1 and 2

PetroMod uses generic porosity-depth trends for carbonates and siliciclastics. These are based on the observation that porosity commonly decreases smoothly with depth in continually subsiding basins (Figure 5.13). However, this is almost always an oversimplification and the incorporation of porosity-depth trends (porosity loss curves) in basin modelling is one of the key uncertainties (Giles et al., 1998).

Figure 5.13: Porosity loss curves for carbonates, curve 3 is that of Schmoker and Halley (1982). Figure from Giles et al. (1998).
The generic porosity depth trends typically used in the model do not adequately represent the porosity loss of Field A because the reservoirs have undergone extensive diagenesis (Chapter 3, 4). Therefore, three models will be simulated: Model 1 uses a typical porosity loss curve for carbonates (Schmoker and Halley, 1982) (Figure 5.13, 5.14a); Model 2 uses the predefined porosity loss curves for a given carbonate lithofacies from the pre-existing Petromod™ lithology database (Figure 5.15a, b), these models will then be compared to Model 3 where an attempt has been made to incorporate diagenesis into the model.

Figure 5.14: a) The porosity loss curve of Schmoker and Halley (1982) was used for all the reservoir intervals in Model 1. b) The porosity loss for the non-reservoirs is taken from the pre-existing Petromod™ database.

Figure 5.15: The porosity loss curves used in Model 2 are taken from the pre-existing Petromod™ lithology database for a) the reservoir intervals, and b) the non-reservoir intervals. These curves are based on the lithofacies of each reservoir interval, with a higher mud content being included in the older reservoirs.
5.4.5 Incorporating diagenesis into Model 3

Oil charge in Field A is thought to preserve porosity in the oil bearing reservoir intervals, whereas in the coeval aquifers the precipitation of calcite cement has greatly occluded pore space (Chapter 3). Using typical compaction curves (Models 1 and 2) (Section 5.4.4) this leads to the porosity in the oil bearing reservoirs being underestimated and the porosity in the aquifers being overestimated. Therefore, when creating the basin models an attempt has been made to incorporate the effects of diagenesis (Model 3).

In the absence of kinetic models, like that for siliciclastics (Lander et al., 2008; Lander and Walderhaug, 1999; Walderhaug et al., 2009) a new method to include diagenesis into the model was developed. This involves modifying the porosity loss curve of Schmoker and Halley (1982) for the oil bearing reservoir intervals, the coeval aquifers, the older aquifers, and the non-reservoir intervals to include a representation of diagenesis. The way that diagenesis has been included in the model is by switching the porosity loss curves used by the model during burial.

All models suggest that the accumulation of hydrocarbons in Field A can occur in the Eocene at ~ 55 Mya (Section 5.5.3). Therefore, at this point the curve of Schmoker and Halley (1982) has been switched in the hydrocarbon bearing reservoirs (vii, v-i) with another porosity-loss curve that more adequately represents porosity loss in an oil or gas bearing reservoir (Figure 5.16a).

Porosity loss in the coeval aquifers (cycles vii, v-i) is interpreted to have been in a fairly rapid manner due to extensive cementation after initial oil charge into the coeval oil bearing reservoir intervals (Chapter 4). Therefore the porosity loss curve of the aquifers has been pierced at ~ 45 Mya to represent the approximate time at which the final pore occluding cement zone formed (Figure 5.16b).

In the aquifers of the Lekhwair Formation (cycles ix, viii, vi) the effective porosity is thought to have been occluded by cement prior to that in the aquifers of the Kharaib and Shu’aiba’ Formations (Chapter 4). This rapid cementation event is
thought to have occurred prior to initial oil charge (~ 55 Mya) and has therefore been placed at ~ 60 Mya (Figure 5.16c).

The non-reservoir intervals are likely to have been rapidly cemented during shallow burial, and are therefore early barriers to flow (Chapter 3, 4). The porosity loss curve for the non-reservoir intervals therefore shows a greater reduction in porosity at a shallower burial depth (Figure 5.16d).

Figure 5.16: Porosity loss curves that were used to include an understanding of diagenesis into Model 3. The porosity loss curve shown in blue is that of Schmoker and Halley (1982), this curve is pierced at a predefined time (see labels) by another porosity loss curve to include a representation of diagenesis (shown in red) for: a) reservoir intervals vii, v-i, b) the coeval aquifers and c) aquifers ix, viii and vi. The black dashed lines show the porosity loss profiles used in the model for the reservoir intervals. d) The porosity loss curve for the non-reservoirs is modified from Models 1 and 2 to include the rapid cementation of the non-reservoirs (Section 4.2.2.5) by increasing the Athy factor of the porosity loss curve (Athy, 1930). The large scale dissolution event that postdates calcite cementation cannot be included in the model (Section 5.4.6).
5.4.6 Model limitations

The cross sections used for the model have grouped several units together and do not include the thickness for every formation. Therefore depth values were taken from other cross sections, and extended across the model with the thickness kept relatively constant across the basin, this is likely to be unrealistic.

The way diagenesis has been included into Petromod™ is by switching the porosity loss curves during burial. However, because the carbonate-dominated lithologies available use porosity loss curves with mechanical compaction only (loss of the pore volume and loss of total volume by grain rearrangement), and do not represent carbonate cementation of the pore space (preservation of the pore volume). The rapid reduction of porosity by switching the mechanical compaction curves will cause a rapid reduction in layer thickness and so does not represent the effect of cementation on preserving layer thickness.

The mechanical compaction curves can also be switched only once, therefore multiple episodes of cementation and dissolution cannot be included in the model. The curves have been switched during model simulation to include the extensive cementation event of the pore space; therefore the later enhancement of porosity by dissolution cannot be included.

All the reservoirs that are known to be oil bearing have the same porosity loss curves implemented which preserves porosity due to the effect of oil charge (Figure 5.16a). However, the porosity of hydrocarbon reservoirs vii, v and iv has initially been greatly occluded by cement and then enhanced by a later dissolution event, this is the same as what is observed in the aquifers of the Lekhwair Formation (Chapters 3, 4). However, this diagenetic history cannot be properly included because the curves can only be switched once.

An understanding of the dynamics of cementation for the reservoir and non-reservoir intervals is typically obtained from one sample via SIMS and EPM analyses (Chapter 4). Although the sample analysed is thought to be representative of the interval, the use of relatively few samples to characterise the entire interval may
not be appropriate. Also the early cementation of one non-reservoir sample has been taken as the standard for all non-reservoir intervals (Section 4.4.2.2) in Model 3, this may not be appropriate.

Field A is known to contain multiple faults (Alsharhan, 1993) however, no faults have been defined in this model because the location of the faults in the field and basin are not known. Off structure these faults may aid the migration of hydrocarbons between reservoir intervals. Therefore, the migration pathways may not be properly represented.

Unconformities were added to the model by estimating the approximate thickness of the eroded sediment and duration of the erosive event. Although there is some confidence in the unconformities’ stratigraphic location, both the thickness of the eroded strata and the exact time over which erosion occurred is estimated. Both unknowns can have a significant influence on porosity loss during burial.

The facies chosen for all the layers in all three models are mixtures of multiple facies that best represent the layer (Appendix 6). Mixing of facies results in a homogeneous layer that is composed of a mixture of the properties of the facies. This will oversimplify the sometimes complex relationships between the facies present in the layer.

5.4.7 Simulation

Once the 2D models were built and populated with all the necessary data they were each simulated. This allows the geological evolution of the basin to be predicted from its initiation to its present day configuration, and the evolution of the hydrocarbon system to be tracked. The simulation input parameters used for all runs are presented in Table 5.2.
Table 5.2: Input parameters used for all simulations. A brief description of the parameter is also given.

When the simulation finishes the output summary report gives an optimisation percentage (percentage of geometrical difference between model runs and the modelled present day cross section). This value should be $< 1\%$ and in all realisations the optimisation percentage was $\sim 2 \times 10^{-6}$. This would suggest that the model is an excellent fit with the present day cross section, and to all previous model runs.

<table>
<thead>
<tr>
<th>Simulation Parameter</th>
<th>Value</th>
<th>Reason for choice</th>
</tr>
</thead>
<tbody>
<tr>
<td>Number of runs</td>
<td>5</td>
<td>Increasing the number of runs improves the optimisation of the final model that is output. Typically a simulation should have a $&lt; 1%$ difference from the other model runs to ensure solution convergence.</td>
</tr>
<tr>
<td>Sampling stepsize</td>
<td>1</td>
<td>This determines how many cells are used in the simulation. The higher the number the faster the simulation, but the coarser the result. A stepsize of 1 was used to provide the greatest resolution, this is equivalent to 102 m.</td>
</tr>
<tr>
<td>Cell thickness</td>
<td>10-300 (m)</td>
<td>The maximum extent in the Z direction (depth), the higher the number the coarser the result. The homogeneous units have a higher value with the reservoir intervals having a resolution of 10 m.</td>
</tr>
<tr>
<td>Petroflow conditions</td>
<td>Hybrid (Darcy + Flowpath)</td>
<td>Less porous layers ($&lt; 30%$) are modelled using Darcy flow and more porous layers ($&gt; 30%$) are modelled using buoyancy.</td>
</tr>
<tr>
<td>Max time step duration</td>
<td>1</td>
<td>The time step that is used by the simulator. The value chosen provides the highest resolution possible for the model.</td>
</tr>
<tr>
<td>Minimum migration steps per time step</td>
<td>100</td>
<td>The number of migration steps used for each time step.</td>
</tr>
</tbody>
</table>
5.4.8 Calibration

The models have been calibrated to all available data. This includes: Vitrinite reflectance (Vr), fluid inclusions within the calcite and dolomite cements and to the present-day geothermal gradient.

For maturity modelling and calibration the Easy%Ro kinetic model of Sweeney and Burnham (1990) was adopted. The present day Vr of the source rocks agrees with that output from the simulation, where the Dukhan Formation is overmature in the Falaha syncline, and the Bab Member and dense intervals of the Thamama Group are currently in the oil window for all models (Table 5.3). The HF profile used in the models does not appear to lead to significant variations to the simulated Vr, with both HF 1 and HF 2 predicting a similar Vr for all source rocks (Table 5.3).

<table>
<thead>
<tr>
<th>Source rock</th>
<th>Depth (m)</th>
<th>Observed vitrinite reflectance (%)</th>
<th>Model 1</th>
<th>Model 2</th>
<th>Model 3</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td>HF 1</td>
<td>HF 2</td>
<td>HF 1</td>
</tr>
<tr>
<td>Dukhan Formation</td>
<td>3438</td>
<td>&gt;1.2</td>
<td>1.27</td>
<td>1.27</td>
<td>1.32</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>1.30</td>
<td>1.30</td>
<td></td>
</tr>
<tr>
<td>Dukhan Formation*</td>
<td>-</td>
<td>1.7</td>
<td>0.81</td>
<td>0.81</td>
<td>0.80</td>
</tr>
<tr>
<td>Bab Member</td>
<td>2532</td>
<td>0.75-0.8</td>
<td>0.75</td>
<td>0.75</td>
<td>0.76</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td>0.76</td>
<td>0.76</td>
<td>0.76</td>
</tr>
<tr>
<td>Bab Member</td>
<td>2450</td>
<td>0.70-0.75</td>
<td>0.70-0.75</td>
<td>0.70-0.75</td>
<td>0.70-0.75</td>
</tr>
</tbody>
</table>

Table 5.3: Vitrinite reflectance of the source rocks from wells drilled in Field A, data from Lijmbach et al. (1992) in relation to modelled Vr data. * Vr data from Alsharhan (1993), these samples are from the Falaha syncline but are from an unknown depth. Models 1-3 have been defined in Sections 5.4.4 and 5.4.5 and are based on changes to the porosity loss curves of the reservoir and non-reservoir intervals. Each model has been simulated twice with two different HF profiles (Section 5.4.3.3).
The maximum burial temperature constrained from fluid inclusions from the calcite and dolomite cements in reservoir interval ii is ~140°C (Neilson et al., 1998). This is in agreement with the maximum temperature suggested by all simulations (Section 5.5.4).

The present day geothermal gradient simulated by the model can be calibrated to the present day geothermal gradient of Abu Dhabi. The geothermal gradient for Abu Dhabi is ~25°C/km (Alsharhan, 1993; Neilson et al., 1998), this gradient is in good agreement with the simulated present day gradient for all models (Figure 5.17).

![Temperature gradient diagram](image)

*Figure 5.17: Comparison of simulated geothermal gradient (blue line), with the present day geothermal gradient (black crosses).*
5.5 Results

The results presented in this section are from calibrated models (Section 5.4.8). The burial history plot, timing of source rock maturation, the migration pathways for the hydrocarbon, the time of oil accumulation and the temperature evolution profiles for the reservoirs have all been extracted from these models.

5.5.1 Burial history plot

The subsidence history for all three models is identical and shows gradual subsidence from the Lower Triassic to the Lower Cretaceous (Figure 5.18). In the Late Cretaceous to the Early Tertiary subsidence rates increase due to subduction and the obduction of the Semail ophiolite. The Middle Tertiary was a time of relative tectonic quiescence and low subsidence rates however, by the Late Tertiary continental collision results in greater burial rates until slab break off at ~ 10 Mya which results in uplift. The uplift events shown in the model are related to documented erosive events (Section 5.4.6).
Figure 5.18: Burial history plot for Field A. The bold black lines on the plot are 50°C isolines. The stratigraphic column is given to the right of the figure, with the location of cycles ix-i shown in bold.
5.5.2 Generation

When HF 1 or HF 2 is applied to all three models a characteristic Transformation Ratio (TR) is obtained for the source rocks in all three simulations. Therefore, in the following sections when a TR profile is presented for a particular HF in a model, the same profile is also obtained in all other models.

All three source rocks (Dukhan Formation, dense intervals and Bab Member) are thought to have matured at a similar time (Figure 5.19) in the Falaha syncline (~ 75 Mya) (Figure 5.19a, 5.20a) and in the anticline (~ 65 Mya) (Figure 5.19b, 5.20b) of Field A when either HF 1 or HF 2 is used. This is similar to that suggested by other authors where the onset of maturation is thought to be in the Late Cretaceous (Section 5.3.1) (Azzam and Taher, 1995; Taher, 1996).

Using HF 1, peak generation of hydrocarbon from the Dukhan formation and dense interval ix in the Falaha syncline is simulated to have occurred 60-50 Mya when 80 % of the organic matter was cracked (Figure 5.19a). This is similar to other authors where optimum maturation is thought to be in the Eocene (Section 5.3.1) (Alsharhan, 1993; Alsharhan and Nairn, 1997; Taher, 1996). Secondary cracking commenced 20-10 Mya where a further 15 % of the organic matter was cracked. For the Bab Member 60 % of the total organic matter matured 60-50 Mya and a further 30 % 20-10 Mya.

HF 1 also suggests that for the source rocks located in the anticline of Field A, 50 % of the organic matter in the Dukhan Formation and dense interval ix had expelled hydrocarbon by 45 Mya (Figure 5.19b). At this time 30 % of the hydrocarbon in the Bab Member was also mature. No significant increase in maturation occurred for the Dukhan Formation, dense intervals or Bab Member between 45 and 20 Mya. However, a further 40 % of the total organic matter in these source rocks generated hydrocarbon from 20-10 Mya.

HF 2 suggests that in the Falaha syncline 50 % of the organic matter for the Dukhan Formation and dense interval ix was cracked by 50 Mya (Figure 5.20a). A gradual increase in hydrocarbon generation occurs until 20 Mya, by this point 60 %
of the total organic matter has been transformed with a further 40% of the organic matter being cracked 20-10 Mya. 40% of the organic matter in the Bab Member was cracked 20 Mya, with a further 50% mature 20-10 Mya.

Using HF 2, 20% of the organic matter in the Dukhan Formation and dense interval ix had cracked by ~ 45 Mya in the anticline of Field A (Figure 5.20b). A further 10% matured by ~ 30 Mya, with major hydrocarbon generation occurring 20-10 Mya where a further 60% of the organic matter expelled hydrocarbon. 10% of the organic matter in the Bab Member was mature ~ 45 Mya, with a further 10% mature by 20 Mya, major expulsion occurred between 20 and 10 Mya when 50% of the total organic matter in the Bab Member matured.
Figure 5.19: Transformation Ratio (TR) profiles for the three main source rocks of Field A when HF 1 is applied to the three models. a) The TR profiles for the source rocks in the Falaha syncline and b) for the source rocks in the anticline of Field A have been extracted.
Figure 5.20: Transformation Ratio (TR) profiles for the three main source rocks of Field A when HF 2 is applied to the three models. a) The TR profiles for the source rocks in the Falaha syncline and b) for the source rocks in the anticline of Field A have been extracted.
5.5.3 Migration and accumulation

Hydrocarbon migration was initially from the Falaha syncline when the Dukhan Formation, dense intervals and Bab Member matured, this led to vertical migration from these source rocks. The hydrocarbon then either vertically migrated through the overlying non-reservoir interval, or migrated laterally below the overlying non-reservoir interval and began to accumulate beneath the intraformational seals in Field A (Figure 5.21). Hydrocarbon was then produced from the source rocks in the anticline of Field A and migrated vertically into the overlying reservoir intervals.

The time at which oil initially entered Field A is different when HF 1 or HF 2 is used. HF 1 causes all the reservoir intervals in Field A to be charged by oil at an earlier stage (55 Mya), whereas HF 2 leads to charging of the older reservoirs first (51 Mya) and the younger reservoirs are then charged at a later stage (45 Mya). All models have experienced a gas charging event, however this gas has typically migrated through the overlying cap rock prior to present day and only oil remains in the reservoirs (Table 5.4).
Figure 5.21: Flow pathways for oil (green arrows), and gas (red arrows) at 25 Mya for Model 3, HF 2; all the other models show identical migration pathways. The overlay is the transformation ratio, which shows that hydrocarbons are generated from the Dukhan Formation, dense intervals and the Bab Member.
Model 1 has an initial oil charge into the older reservoir intervals when HF 1 is used, whereas when HF 2 is applied reservoir intervals ix and viii are not charged (Table 5.4). The oil in Model 1, regardless of whether HF1 or HF 2 is used, then progressive migrates from the oldest reservoirs, when porosity is < 16%, into successively younger reservoirs during burial; oil is currently within reservoir intervals i and ii (Table 5.4; Model 1) (Figure 5.22). The porosity-depth curves used in Model 2 (Figure 5.15) allow for oil accumulation in all reservoirs (Table 5.4) (Figure 5.22). The porosity-loss curves used in Model 3 allow for oil to accumulate in all the reservoir intervals that are currently oil bearing (Table 5.4) (Figure 5.22).

<table>
<thead>
<tr>
<th>Reservoir interval</th>
<th>Model 1</th>
<th>Model 2</th>
<th>Model 3</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>HF 1</td>
<td>HF 2</td>
<td>HF 1</td>
</tr>
<tr>
<td>i</td>
<td>55; 52-12</td>
<td>45; 21-14</td>
<td>55; 52-51, 45-11</td>
</tr>
<tr>
<td>ii</td>
<td>55; 52-11</td>
<td>45; 21-14</td>
<td>55; 52-11</td>
</tr>
<tr>
<td>iii</td>
<td>55-25; 21-15; 52-15</td>
<td>45-15; 21-15</td>
<td>55; 55-54.5, 51-11</td>
</tr>
<tr>
<td>iv</td>
<td>55-35; 52-25</td>
<td>45-25</td>
<td>55; 52-11</td>
</tr>
<tr>
<td>v</td>
<td>55-35; 52-35</td>
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<td>55; 55-11</td>
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<tr>
<td>vi</td>
<td>55-45; 52-45</td>
<td>50-45</td>
<td>55; 55-11</td>
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<td>vii</td>
<td>55-45; 52-45</td>
<td>50-45</td>
<td>55; 55-53, 52-11</td>
</tr>
<tr>
<td>viii</td>
<td>55-50; 52-50</td>
<td>x</td>
<td>55; 55-53, 52-10</td>
</tr>
<tr>
<td>ix</td>
<td>55-51; 52-51</td>
<td>x</td>
<td>55; 54.5-53, 52-10</td>
</tr>
</tbody>
</table>

Table 5.4: Modelled time (Mya) of oil (regular font style) and/or gas (italics) accumulation in a reservoir interval. The values in bold identify when hydrocarbon is still currently present in the reservoir from the charging event. x – not charged with hydrocarbon. Models 1-3 have been defined in Section 5.4.4.4 and are based on changes to the porosity loss curves of the reservoir and non-reservoir intervals. HF 1 and HF 2 are defined in Section 5.4.3.3.
Figure 5.22: Oil accumulation sites (shown in green) for Models 1, 2 and 3 at present day. The models suggest that oil is currently present in reservoir intervals i and ii in Model 1, in Model 2 oil is currently present in all reservoir intervals (ix-i) and in Model 3 oil is currently present in reservoir intervals vii, vi, iv, iii, ii, and i.

The white arrow in Model 3 marks the location of reservoir vi, where no hydrocarbon is present.
5.5.4 Temperature modelling

The temperature evolutions for the reservoir intervals of the Lekhwair, Kharaib and Shu’aiba’ Formations are different for HF 1 (Figure 5.23a) and HF 2 (Figure 5.23b). HF 1 shows a steady increase in temperature from deposition to ~40°C at 80 Mya, the temperatures of the reservoirs then increase from 40°C to 120°C at 40 Mya, temperature then decreases to 100°C 30 Mya and then increases to 135°C at 10 Mya, a period of cooling then reduces the temperature to 120°C, which agrees with the present day reservoir temperature (Section 5.4.8). Filling of the reservoirs is thought to have commenced at 55 Mya (Section 5.5.3), at this time the reservoirs were at a temperature of ~85-95°C (Figure 5.23a).

Using HF 2 the temperatures of the reservoir intervals gradually increase to 40°C by 80 Mya and then gradually increase to 100°C 50 Mya (Figure 5.23b). The temperatures then sharply increase to 130°C 10 Mya and then gradually decrease until the present reservoir temperature (120°C). Oil is thought to enter the older reservoirs 51 Mya (Section 5.5.3), when at a temperature of ~87°C and is thought to enter the younger reservoirs at 45 Mya (Section 5.5.3), when they are at a temperature of ~100°C (Figure 5.23b).
Figure 5.23: Temperature evolution profiles through time for HF 1 and HF 2, the profiles are extracted from the crest of Field A. The temperature range for the reservoir intervals at the time of oil charge (Table 5.4) is shown by the red lines (Section 5.5.3). This would suggest that at the time of oil charge the reservoirs were at a temperature of between 85°C and 100°C.
5.6 Discussion

5.6.1 The origin of hydrocarbon and the time of charge into Field A

The time of major maturation of the source rocks is different for HF 1 and HF 2, with maturation of the source rocks being in the Early Eocene for HF 1, whereas for HF 2 maturation was mainly in the Miocene (Section 5.5.2). This is to be expected from the HF profiles used because HF 1 has higher HF values in the Late Cretaceous-Early Eocene due to ophiolite obduction, whereas HF 2 has higher HF values in the Miocene due to continental collision (Figure 5.12).

Although changing the HF profiles for the models results in a different time for major source rock maturation (Section 5.5.2), the time at which generation commenced and the time of initial charge into Field A remain similar. Initial generation from the Dukhan Formation, dense intervals and Bab member commenced at ~ 75 Mya in the Falaha syncline, with charge from this maturation event into Field A being between 55 and 45 Mya. The source rocks in the vicinity of Field A also contribute to the hydrocarbon present within the field, with migration from these source rocks into the overlying reservoir occurring at ~ 45 Mya (Section 5.5.2).

Although gas is generated from the source rocks in the Falaha syncline gas is not simulated to be presently in the reservoir intervals, however gas is known to be in reservoir vii at present day (Alsharhan, 1993). This would suggest that the gas present in reservoir interval vii may be derived from a secondary source to the North East (Alsharhan, 1993) or that the gas was expelled from an area of the Falaha syncline that is not currently represented by the model. The source region is likely to have expelled the gas relatively recently so that the gas does not migrate through the intraformational seals.
5.6.2 The effect of changing porosity loss curves in basin models

Model 1 suggests that when the porosity of a reservoir interval falls below ~16% the hydrocarbon in the reservoir migrates from that reservoir and enters the overlying reservoir interval (the hydrocarbon is expelled) (Table 5.4; Model 1); this is not observed in any other model. This has led to the successive upward loss of hydrocarbon from the older reservoir intervals into the younger reservoirs. The most likely cause for the expulsion of oil is due to the porosity of the reservoir intervals typically being lower than for Models 2 and 3 (Section 5.4.4 and 5.4.5); the properties of the intraformational seals are the same as those for Model 2 where there is no noticeable upward loss of hydrocarbon (Table 5.4; Model 2). The lower porosity will increase the capillary pressure of the reservoir intervals (equation 2)

\[ P_c = 2\gamma \cos \theta / r \] (equation 2; Washburn (1921))

Where: \( P_c \) = capillary pressure; \( \gamma \) = Inter-Facial Tension (IFT); \( \theta \) = contact angle; \( r \) = pore throat radius

The IFT has been discussed previously (Section 5.3.2) and assumed to be constant. The contact angle may change due to alterations to the wettability state of the pore space but is assumed to be constant in the model. The pore throat radius is the only parameter that will change during burial due to mechanical compaction.

The seal has an initially lower porosity and higher capillary entry pressure than the reservoirs and so will behave as a barrier to flow. However, due to continued burial the pore throat radius will decrease in the reservoirs, assuming all other variables remain constant, this will result in an increase of capillary pressure and is likely to reduce the pressure difference between the seal and the reservoir. So during burial, as the capillary pressure of the reservoir approaches that of the seal, the capillary barrier becomes smaller and oil may breach the seal. The porosity at which the capillary pressure of the reservoir surpasses that of the non-reservoirs is predicted to be ~16% by the model. The effect of continued oil charge into the reservoir
interval and the resulting increase in buoyancy pressure is also likely to contribute to migration through the seal by reducing the pressure difference between the seal and reservoir. Another possibility is that at the critical porosity for the reservoir, the continued increase in bound water with depth will apparently cause the expulsion of hydrocarbon (Xiongqi et al., 2012).

The porosity loss curves for the reservoir intervals used for Model 2 typically overestimate the porosity in the older aquifers, with all reservoirs currently having porosities > 20 %. This higher porosity for the reservoirs allows hydrocarbon to accumulate in all reservoir intervals (Table 5.4).

Model 3 more accurately represents present day accumulations, with hydrocarbon currently in all producing reservoir intervals. The inclusion of diagenesis into the models has led to the porosity in some of the reservoir intervals (ix, viii, vi), at the time of hydrocarbon migration to be below the critical porosity that is required for hydrocarbon accumulation.

Changing the porosity loss curve for the non-reservoir intervals in Model 3, so that there is greater porosity loss in the initial 500 m of burial by early cementation, does not appear to significantly limit hydrocarbon migration. This is because when oil entered the reservoirs at 55-45 Mya (Section 5.6.1) the reservoirs were at a burial depth of ~ 1.7 km (Figure 5.18) and the porosity of the non-reservoirs in all the models at this burial depth is < 4 % (Figure 5.14b, 5.15b, 5.16d). These low porosities for the non-reservoir intervals will make them effective barriers to flow prior to and during hydrocarbon filling.

5.6.3 Constraints on diagenesis

The temperature profiles extracted from the models for HF 1 and HF 2 suggest that the temperature of the field progressively increased during burial (Section 5.5.4). This is in agreement with the progressive decline in $\delta^{18}O_{\text{VPDB}}$ and $^{\text{a}}\text{Mg}^{\text{a}}\text{Ca}$ for all reservoir intervals (Chapter 4) and would suggest that cementation continued during the course of burial and that the $\delta^{18}O_{\text{VPDB}}$ and $^{\text{a}}\text{Mg}^{\text{a}}\text{Ca}$ temperature proxies are most likely recording the progressive burial of the field.
The accumulation of hydrocarbon in all the basin models suggests that emplacement occurred at 55-45 Mya. Assuming that the cement zones in reservoir i in Field A record this charge, the first cement zone that contains an oil inclusion has an average \( \delta^{18}O_{VPDB} \) of \( -8.8 \% \) (~104 °C, assuming a \( \delta^{18}O_{SMOW} \) of +6 ‰; Chapter 4). This is in good agreement with the modelled temperature of the reservoir during initial hydrocarbon charge (85-95 °C) (Section 5.5.4).

The maximum temperature at which cementation continued to in oil reservoir and coeval aquifer i is thought to be ~ 111 °C and ~ 113 °C respectively (Chapter 4); this agrees with the majority of fluid inclusion data presented by Neilson et al. (1998). This suggests that cementation continued to ~ 50 Mya using HF 1 and to ~ 25 Mya using HF 2 (Figure 5.23). Cementation in oil reservoir ii is thought to have continued to a temperature of ~ 145 °C, the maximum temperature reached by both models is ~ 140 °C at ~ 10 Mya suggesting that cementation in oil reservoir ii may have continued during continental collision in the Miocene. Cementation in aquifer ii is thought to continue to a similar temperature to that in aquifer i (Chapter 4) and therefore is thought to continue to ~ 50 Mya (HF 1) or ~ 25 Mya (HF 2).

The \( \delta^{18}O_{VPDB} \) range over which the cements precipitated in all reservoir intervals both in Field A and B, excluding reservoir ii, is less than or equal to that in oil reservoir i of Field A (Section 4.3.2.5). Using HF 1, cementation is therefore thought to have ceased in all reservoir intervals by ~50 Mya. This is the approximate time (~45 Mya) at which the source rocks in the vicinity of Field A began to expel hydrocarbon (Figure 5.19b, 5.20b) (Section 5.5.2). During source rock maturation organic acids such as CH\(_4\), NH\(_3\), CO\(_2\) are produced (Hantschel and Kauerauf, 2009). The amount of organic acids produced by a source rock is thought to be dependent on the TOC of the source rock, with a lower TOC typically leading to a greater production of organic acids (Barth and Bjørlykke, 1993). Therefore because the TOC of the dense intervals is low (~1.2 %) there should be a greater volume of organic acids produced. However, the effect of these acids on the development of secondary porosity is currently controversial.

Although the initial expulsion of the organic acids from the source rock may be from the petroleum phase, the acids will rapidly partition into the water phase,
leading to the majority of reactions with the mineral phase occurring close to the source rock (Barth and Bjørlykke, 1993). The overlying or underlying reservoir interval will be the primary location into which the organic acids migrate and therefore are likely be the sites of any dissolution (Barth and Bjørlykke, 1993). Therefore it is probable that the generation of the late stage secondary porosity in both Field A and B and the cause for cementation to cease in some reservoirs may be due to the production of organic acids from local source rocks.

Kaolinite is then thought to precipitate after this dissolution event with its random distribution being observed in both Field A and B (Chapters 3 and 4 and Appendix 4). The source of kaolinite in the cycles is indeterminate but can be from the reaction between organic acids (e.g. CO$_2$) and depositionally concentrated aluminosilicate minerals in the TST intervals (reaction 1) or from the thermal degradation of organic matter in the dense intervals (Maliva et al., 1999). Another possibility is that aluminium was brought into the system by the acidic fluids generated from the maturation of the Dukhan Formation, these fluids would migrate through Palaeozoic siliciclastic rock which would lead to reaction 1 and to the formation of kaolinite.

\[
\text{M}^{(\text{II})}\text{Al}_x\text{Si}_y\text{O}_z + \text{CO}_2^{(\text{aq})} + \text{H}_2\text{O} \rightarrow \text{Al-silicate mineral} + \text{M}^{(\text{II})}\text{CO}_3
\]

(reaction 1: from Baines and Worden (2004))

5.7 Conclusions

Including different porosity loss curves into the model changes the location of hydrocarbon in the field. Model 1 suggests hydrocarbon accumulates primarily in reservoir interval i and ii, thereby incorrectly identifying the location of hydrocarbon in Field A. Model 2 suggests that hydrocarbon is accumulated beneath the intraformational seal for each reservoir, thereby overestimating the accumulations in Field A. The only model that correctly identifies the presence of hydrocarbon in all
producing reservoir intervals includes an understanding of diagenesis (Model 3). This demonstrates how crucial the inclusion of diagenesis is into any basin model.

A critical porosity for oil accumulation in a reservoir interval has been identified as ~ 16 % in Model 1, with migration from the reservoir intervals occurring when the porosity is reduced below this value. This is thought to be related to the capillary pressure of the reservoir increasing during burial to a similar pressure as in the overlying seal which allows for the upward migration of hydrocarbon.

The models simulated with HF 1 suggest that oil migrated into the reservoirs from the Falaha syncline at ~ 55 Mya, whereas HF 2 suggests hydrocarbon migrated into the reservoirs 51-45 Mya. These charging events occurred when the reservoirs were at a temperature of between 85-95°C and 87-100°C, respectively. This is in good agreement with the δ¹⁸O_VPDB value for the first cement zone in reservoir i that contains an oil inclusion (~ 8.8 ‰), which is thought to precipitate at a temperature of ~ 104 °C (Chapter 4).

The temperature at which cementation ceased in oil reservoir and aquifer i is suggested by δ¹⁸O_VPDB data to be ~ 111 °C and ~ 113 °C respectively. This suggests that cementation ceased in reservoir i by ~ 50 Mya (HF 1) or ~ 25 Mya (HF 2). The range in δ¹⁸O_VPDB in all other reservoir intervals, with the exclusion of oil reservoir ii of Field A, is less than or equal to that in oil reservoir i. This may suggest that cementation in all reservoir intervals, both in Field A and B ceased by 50 Mya (HF1). After cementation ceased in the reservoirs a large scale dissolution event occurred and kaolinite was then precipitated in the newly generated secondary pore space (Chapter 3, 4).

The origin of the final dissolution event is most likely due to the maturation of the dense intervals of the Thamama Group in the anticline of Field A. The models suggest that the dense intervals within the anticline of Field A began to expel hydrocarbon by ~ 45 Mya, prior to this expulsion event the organic acids generated due to source rock maturation most likely caused dissolution in the underlying and overlying reservoir intervals. This may suggest that the acidic charge during initial
source rock maturation can lead to significant porosity enhancement; when the reservoirs are in proximity to the source rock.

Kaolinite is found in the secondary pore space generated by this late dissolution event. The aluminium that is required for the formation of kaolinite may be transported into the field by complexing with the organic compounds derived from the maturation of the dense intervals of each cycle, allowing for the random distribution of kaolinite throughout the field.
CHAPTER 6
CONTROLS ON RESERVOIR QUALITY FOR THE QUISSAMÃ FORMATION, CAMPOS BASIN, BRAZIL

6.1 Aims and Objectives

The overall aim for this study is to determine the controls on reservoir quality for the Albian Quiassamã Formation of Field C (Section 1.5.2). This has been achieved by undertaking the following:

1) Seismic lines were interpreted to help determine the depositional system and the controls on platform formation and termination. The depositional environment is then assessed through core and thin section analyses to constrain the depositional environment and to better determine the depositional fabrics that compose the reservoir. The depositional fabrics identified are then related to plug derived porosity and permeability measurements to determine the control of depositional environment on current reservoir quality.

2) The porosity vs. depth trends for the grainstones in Well 1C and 2C are then related to a typical compaction curve for grainstones to determine whether oil charge has greatly affected the porosity within the field (Schmoker and Halley, 1982). This is because oil charge is thought to retard or stop cementation and therefore preserves reservoir quality. Thus if oil charge has preserved reservoir quality in Wells 1C and 2C, the porosity should be greater than that typically expected at a given depth. The effect of oil charge on preserving reservoir quality will also be assessed by comparing the porosity and permeability of similar depositional fabrics (mudstones, wackestones,
packstones and grainstones) in the oil reservoir and the aquifer. This is because similar depositional fabrics should have, initially, comparable porosities and permeabilities (Section 1.3.1) therefore by comparing similar depositional fabrics in the oil reservoir and aquifer the effect of oil charge can be assessed.

3) A petrographic assessment of thin sections from the core of Wells 1C and 2C is then undertaken to determine the chronology of diagenetic constituents and to understand their distribution in the oil reservoir and aquifer. This will help to assess the effect that diagenesis has had on modifying the initial depositional petrophysical properties of the reservoir rocks.

4) A separate component analysis (δ13C and δ18OVPDB) and in-situ elemental and δ18OVPDB analyses are then undertaken on the main pore occluding calcite cements to help determine the chemistry of the pore fluid from which the cements formed. This assessment was undertaken on cements from the oil reservoir and the aquifer to help understand if oil emplacement has had an impact on cementation.

5) 2D basin models are then developed to investigate and illustrate the geological evolution of Field C, determine the source area, timing of oil generation, migration and accumulation and evaluate the hypothesis that early oil charge in Field C has led to the preservation of porosity.

6.2 Previous Work

The following section presents the current understanding of the depositional environments and diagenetic constituents of the Quissamã Formation.

6.2.1 Depositional environment

The Quissamã Formation is composed of a number of High Frequency Cycles (HFCs). The transgressive portion of each HFC contains mudstones and wackestones
that are sometimes dolomitised, whereas the highstand portion is composed of bioclastic/oncolitic/oolitic grainstones and packstones (Silva et al., 2012).

The burrowed lime wackestones/packstones, terrigeneous siltstones and very-fine-grained sandstones are thought to have deposited in a lagoonal environment in the protected areas of the basin (Carozzi et al., 1983). Whereas the oncolitic carbonates are thought to develop on an open marine shelf in shallow subtidal turbulent waters (Carozzi et al., 1983). The most common components of these rocks are: oolites, fragments of bivalves, gastropods, echinoderms and red algae, but quartz and feldspar grains locally compose as much as 20 % (Guardado et al., 1990).

**6.2.2 Diagenesis**

There is little published information on the diagenesis of the Quissamã Formation, with the only known assessment being undertaken by Carozzi et al. (1983). Through petrographic analyses and bulk isotope geochemistry this study identified six diagenetic environments, each is summarised below:

1) Low energy intertidal environment: deposition of unconsolidated sediment that consists of abraded piso-oncoids and interstitial sand-size bioclasts that are set in a finer matrix. In this environment a thin isopachous rim of fibrous cement was deposited.

2) High energy environment: the interstitial constituents were removed and reorganised into sediment with geopetal features and horizontal surfaces. High energy and the downward circulation of seawater dominate this environment.

3) Beachrock environment: where the diagenetic lithification of internal sediment by interparticle sparite is followed by solution and the generation of secondary vuggy porosity. A rim of cement is then precipitated along the boundaries of all previously generated open pore spaces.
4) Beachrock vadose environment: the rock is sub-aerially exposed and intense vadose dissolution occurs. The removal of a large volume of bladed rim cement and the corrosion of piso-oncoids are typical of this environment.

5) Freshwater meteoric phreatic environment: in this environment extensive cementation by calcite cement occurs, this cement typically forms an interlocking drusy calcite mosaic.

6) Burial environment: identified by the occurrence of intense pressure solution and stylolitisation. The pores generated during burial diagenesis are uncemented or show a very thin rim of calcite cement. Late cementation is widespread but is not an important porosity reducer.

6.3 Depositional System

This section aims to determine the depositional environment of the Quissamã Formation in Field C by undertaking seismic, core and petrographic analyses.

6.3.1 Seismic scale geobodies

Numerous seismic transects have been made available for Field C by Petrobras. The seismic transects show that the Quissamã Formation developed on a basement high that is related to basement faulting and the development of a horst (Figure 6.1). This has led to the characteristic steep sides and the relatively gently sloping crest typical of an isolated platform (Figure 6.1).
Figure 6.1: Vertically exaggerated seismic line across Field C, from a Petrel™ model provided by Petrobras. The horizon of the top Quissamã Formation (shown in blue) is strongly related to that of the top of basement (shown in red). The orange arrow marks the location of Well 1C and the red arrow the location of Well 2C, both wells are near vertical.

Time slices of a seismic cube allow the morphology and the depositional system of the Quissamã Formation to be assessed. These horizontal slices have been interpreted to show the initial commencement of an isolated platform (Figure 6.2a) and the subsequent, vertically aggrading stacking patterns of the isolated platform (Figure 6.2b). The platform then retrogrades (Figure 6.2c, d) towards the coastline due to sea-level rise and turbidite channels of the Namorado sandstone (Section 1.5.2) then develop over the isolated platform (Figure 6.2e).

The Albian Namorado turbidites are thought to form regionally and to deposit extensive blankets on a flat sea floor (Guardado et al., 2000); the meandering nature of the channel would support their development on a relatively flat sea floor (Figure
6.2e). The blanket sands deposited from the turbidite flows can be correlated over wide areas, suggesting that the sands entered the basin from several sources as opposed to a single major feeder (Guardado et al., 1990). The turbidite sands are intercalated with and overlain by marine mudstones and shales (Figure 6.2f).

Figure 6.2: Seismic time slices during the deposition of the Quissamã Formation on the isolated platform; the locations of the studied wells are shown on all figures. The interpreted location of the platform is marked by the red dashed line. The deepest time slice is shown in a), with the shallowest presented in f). Initial aggradation (a, b) on the isolated platform is followed by retrogradation (c, d) and finally flooding (e,f). The red arrow on e) marks the location of the meandering turbidite channel. The location of the current oil bearing reservoir is shown in a) and b) by the yellow contoured surface.
6.3.2 Depositional environment

A typical depositional cycle for the Quissamã Formation is represented by an upward shallowing succession from low energy mudstones and wackestones (Figure 6.3a) to packstones (Figure 6.3b) that are overlain by coarse oncoidal, ooidal, peloidal grainstones (Figure 6.3c). These fabrics are dominated by oncoids, ooids, peloids, gastropods, echinoderms and bivalves, some of the grains are also bored.

Figure 6.3: Core photographs of a typical depositional cycle for the Quissamã Formation. a) The mudstones and wackestones identified in the core section are not oil bearing, whereas the, b) packstones and, c) grainstones typically show oil stain.

The field scale depositional model provided by Filho (2011) (Figure 6.4) demonstrates that the crest of the structure is dominated by grainstones, whereas the flanks are predominantly composed of packstones. The overlying seal (shown in green in Figure 6.4) is a transgressive wackestone deposit that is dominated by calcispheres (Figure 6.5a), glauconite (Figure 6.5b) and planktonic foraminifera.
Figure 6.4: Schematic geological section showing the distribution of depositional fabrics in Field C. Grainstones typically dominate the crest of the isolated platform, whereas packstones are most common on the flanks. Modified figure courtesy of Filho (2011). Note the location of Well 2C; Well 1C was also drilled within the crest of the field.

Figure 6.5: Transgressive mudstones and wackestones of the seal. a) Calcispheres and b) glauconite are commonly observed.
6.4 Controls on Reservoir Quality

This section aims to understand the distribution of porosity and permeability in Field C. The first subsection presents an assessment to determine the control that depositional environment has on present day porosity and permeability. The second subsection then aims to understand the effect that oil charge has on porosity and permeability.

6.4.1 Depositional control on reservoir quality

The porosity vs. permeability plot for Wells 1C and 2C show that typically the mud dominated fabrics have the poorest reservoir quality, and the more grainy fabrics have the best reservoir quality (Figure 6.6). This would be expected if no significant diagenetic alteration has occurred to the fabrics (Section 1.3.1) (Enos and Sawatsky, 1981). However, the porosity and permeability for some grainstones and packstones is markedly lower in some samples (black dashed area on Figure 6.6) when compared to the typical values for that fabric.
6.4.2 The effect of oil charge on reservoir quality

To help assess whether the variation in porosity and permeability within the same depositional fabric (Figure 6.6) is a result of oil charge preserving porosity in the oil reservoir, the porosity of the grainstones in Wells 1C and 2C are plotted with that of a typical porosity loss curve for grainstones (plotted as a solid line on Figure 6.7). This is because if oil charge has preserved porosity within the oil reservoir the porosity should be greater than that typically expected at a given depth.

Figure 6.7 shows that the plug porosity for grainstones in Well 1C is adequately described by the typical porosity loss curve (Figure 6.7a). This may
suggest that for the sampled section in Well 1C oil charge has not caused any significant deviation from the porosity loss expected due to burial. However, for some samples the porosity is greater than that expected by the typical porosity loss curve and in other samples the porosity is reduced by > 10 % compared to the predicted porosity at that depth.

Although the majority of data for Well 2C fall within the ± 5.1 % standard error of a typical compaction curve (Figure 6.7b), there is a significant contrast between the porosity in the oil reservoir and aquifer; this is marked by the OWC (Oil Water Contact). The porosity in the oil reservoir is currently markedly higher than in the aquifer which may be due to the effect that oil charge has on preserving porosity in the oil reservoir.

![Figure 6.7: Porosity vs. depth for the grainstones in a) Well 1C and b) Well 2C. The Oil Water Contact (OWC) is marked by the dashed line. The distribution of porosity for Well 1C is in good agreement with the typical porosity loss curve (5.1 % standard error marked by the dashed lines). However, the porosity of the oil reservoir in Well 2C is significantly greater than for the aquifer.](image)

The data previously presented in Figure 6.7 are solely from plug derived porosity and permeability measurements from grainstones. Therefore because there are fewer data points from Well 1C and these data points are proximal to the OWC, there is likely to be no observable difference in porosity between the oil reservoir and aquifer. To determine whether oil charge may have preserved porosity in Well 1C,
like in Well 2C (Figure 6.7b), neutron log porosity data are presented along with the plug porosity data for all the depositional fabrics in Well 1C (Figure 6.8a).

Figure 6.8: Porosity and permeability data for all depositional fabrics in a), b) Well 1C and c), d) Well 2C; the permeability data are presented in the panels on the right. The porosity data in a) is a combination of neutron porosity log data (hollow circles) and plug porosity data (filled circles). The two zones highlighted in a) and c) are discussed in the text.

The neutron porosity log for Well 1C suggests that the porosity distribution in Well 1C is similar to that in Well 2C for the same parts of the formation (Figure 6.8a, c). Therefore in both Well 1C and Well 2C the oil reservoir typically has a greater porosity than the aquifer (Figure 6.8a, c). The permeability for Well 2C also increases into the oil reservoir (Figure 6.8d), with the aquifer typically having a permeability of 168 ± 70 mD (N=139) and the oil reservoir having a permeability of
1155 ± 123 mD (N=173). This trend is not observed in Well 1C most likely because plug derived permeability measurements are not available for parts of the well (Figure 6.8b).

Although the porosity in the oil reservoir is greater than in the aquifer for both Well 1C and 2C, there are fluctuations in porosity within the oil reservoir of both wells. In both wells two high porosity zones have been identified that are separated by a relatively lower porosity interval (Figure 6.8a, c). This change in remaining porosity can be related to a change in depositional fabric from Zone 1 which is dominated by grainstones, to a low porosity zone composed of packstones and to a high porosity zone (Zone 2) that is dominated once more by grainstones.

To help determine the effect that oil charge has had on reservoir quality for individual depositional fabrics, the data presented in Figure 6.6 is separated into whether the sample is from the oil reservoir or aquifer (Figure 6.9).

**Figure 6.9**: Plug porosity vs. plug permeability for the grainstone and packstone fabrics present in the oil reservoir (black circles) and aquifer (hollow circles) of (a, b) Well 1C and (c, d) Well 2C. The permeability data are shown on a log scale.
The plug derived porosities and permeabilities in Well 1C are similar in the oil reservoir and aquifer, with the grainstones in the oil reservoir having an average porosity and permeability of $19 \pm 0.4 \% \ (n = 71)$ and $370 \pm 110 \ \text{mD} \ (n = 71)$ respectively (Figure 6.9a), compared to $17 \pm 1 \% \ (n = 25)$ and $460 \pm 200 \ \text{mD} \ (n = 25)$ in the aquifer (Figure 6.9a). Porosities and permeabilities are also similar for the packstones in the oil reservoir and aquifer with an average porosity and permeability for the packstones in the oil reservoir being $20 \pm 0.4 \% \ (n = 140)$ and $43 \pm 10 \ \text{mD} \ (n = 140)$ respectively, whereas in the aquifer the average porosity and permeability is $17 \pm 0.6 \% \ (n = 34)$ and $7 \pm 2 \ \text{mD} \ (n = 34)$ respectively (Figure 6.9b). As discussed previously this is most likely due to the samples being obtained from a smaller depth interval around the OWC.

The plug derived porosities of the grainstone and packstone fabrics in the oil reservoir are greater than in the aquifer for Well 2C (Figure 6.9c, d). The average porosity for the grainstones in the oil reservoirs is $22 \pm 0.3 \% \ (n = 173)$ in comparison to $16 \pm 0.3 \% \ (n = 143)$ in the aquifers. The packstones have an average porosity of $23 \pm 0.5 \% \ (n = 88)$ and $16 \pm 0.5 \% \ (n = 49)$ in the oil reservoirs and aquifers respectively.

The plug derived permeabilities for the grainstone and packstone fabrics in the oil reservoir are also significantly greater than in the aquifers (Figure 6.9c, d), with the average permeability of the grainstones in Well 2C being $1155 \pm 123 \ \text{mD} \ (n = 173)$ in the oil reservoirs in comparison to $168 \pm 70 \ \text{mD} \ (n = 139)$ in the aquifers. The packstones have a permeability of $95 \pm 59 \ \text{mD} \ (n = 88)$ in the oil reservoirs in contrast to $0.6 \pm 0.1 \ \text{mD} \ (n = 49)$ in the aquifers; this permeability is below the 1 mD permeability cut-off that is typically required for a reservoir to be economic (Lønøy, 2006).

Although permeability is greater in the oil reservoir than for the aquifer of Well 2C, the grainstones present within the oil reservoir have on average a greater permeability than the packstones. This would be expected if the reservoir quality is primarily determined by depositional fabric (Section 1.3.1) and most likely indicates that the petrophysical properties of the oil reservoir have not been greatly modified.
by diagenesis and that the present day reservoir quality is primarily determined by depositional heterogeneity.

Therefore in summary the plug porosity and permeability data of Well 2C suggest that the porosities and permeabilities in the oil reservoir are much greater than for the aquifer. This is also likely to be the case in Well 1C (as suggested by the neutron porosity log), however because the plug samples are all from a similar depth the effect of oil charge on preserving porosity and permeability is not clear. The cause for the higher porosities and permeabilities in the oil reservoir when compared to the aquifer is most likely a result of diagenesis and the effect of oil charge on retarding or stopping cementation. This will now be investigated through a petrographic and geochemical assessment.

6.5 Diagenesis

This section presents the results of the diagenetic assessment undertaken for Field C and is separated into three subsections. The first subsection involves petrographic analyses where the main pore occluding cements and any pore enhancing events are identified and placed in chronological order. The second subsection presents the results of a separate component (bulk) and in-situ geochemical assessment for the main pore occluding cement phase. The final subsection then discusses the results of the geochemical assessment with the aim to understand the origin and relative time of cement formation and to assess the effect that oil charge has on the dynamics of cementation.

6.5.1 Petrography

Petrography reveals that, for similar fabrics, the more porous and permeable samples of the oil reservoir have a lower volume of pore occluding cement than the aquifer (Figure 6.10). This has led to the marked difference in porosity and permeability between the oil reservoir and aquifer that was identified in Section 6.4.
Figure 6.10: a) In the oil reservoir there is a low volume of pore occluding cement present, with fringing and meniscus cements leading to a slight reduction in porosity.

b) In some oil reservoir samples there is a lack of cement altogether. This is in contrast to the aquifer (c, d) where the formation of drusy calcite mosaics has greatly reduced porosity.

The main pore occluding cement phase identified in Field C is non-ferroan calcite cement. Four calcite cement types have been identified:

1) Peri-granular fringing calcite rims that are 50-100 μm in size. The cements are present around former grains and can be partly dissolved (Figure 6.10a).

2) Syntaxial calcite cements (< 1mm in size) are commonly observed and typically form on fragments of echinoderms (Figure 6.10c).

3) Meniscus cements are rarely observed, and are more common in the oil reservoirs than in the aquifers (Figure 6.10a).
4) Mosaic drusy calcite cement (50–200 μm in size) can lead to near complete pore occlusion. This cement phase is most commonly observed within the samples of the aquifer (Figure 6.10c, d).

Cross cutting relationships have been used to determine the chronology of diagenetic events. However, the cements identified in Field C are all relatively early with a limited number of cross cutting relationships identified; this makes the discernment of a paragenetic sequence difficult.

There are no cross cutting relationships observed for meniscus cements, but this cement phase is thought to precipitate in the Vadose Zone (VZ) (Figure 6.10a) (Flügel, 2004). Fringing cements form a peri-granular rim around grains, this rim is typically thought to develop in a phreatic environment however, whether the cement formed in a marine or meteoric phreatic environment is difficult to determine through petrography alone (Figure 6.10a). It can be observed that fringing and syntaxial calcite cements grow competitively suggesting precipitation at a similar time (Figure 6.11a). Drusy calcite cement mosaics precipitate after or synchronously with syntaxial cements (Figure 6.10c) and appear to precipitate after the formation of fringing calcite cement (Figure 6.10d). Dissolution is then observed to etch the surfaces of fringing calcite cement and the surfaces of grains that do not have a protective cement rim (Figure 6.11b, c, d). No hydrocarbon inclusions are observed in the cements, therefore it is thought that hydrocarbon charge occurred after cementation stopped and so the dissolution event may be related to an acidic front prior to hydrocarbon migration. The aforementioned cross cutting relationships have been used to develop a paragenetic sequence for Field C (Figure 6.12).
Figure 6.11: a) Unlike other grains, the echinoderm grain on which the syntaxial cement formed is not coated by fringing cement, this suggests that the syntaxial cement developed prior to or during the formation of fringing calcite cement. Dissolution has etched the surfaces of (b, c) fringing cements and of (d) peloids.

Figure 6.12: Paragenetic sequence for Field C. All the cements identified in the Quissamã Formation are early cements, making the paragenetic sequence difficult to discern. Meniscus cements are interpreted to form in the Vadose Zone (VZ). The fringing, syntaxial and drusy cements are interpreted to form in a Marine Phreatic Zone (MPZ) but may have formed in a meteoric environment, however based on petrography alone it is not possible to tell. Hydrocarbon charge is thought to occur in the Shallow Burial Zone (SBZ), with the onset being marked by the red dashed line.
6.5.2 Geochemistry

The results for the separate component and in-situ elemental and $\delta^{18}O_{VPDB}$ analyses are presented in the following two subsections and are discussed in Section 6.5.3. The primary purpose of this study is to determine the chemistry of the pore fluid from which the main pore occluding cements formed and to determine the relative temperature at which cementation ceased in the oil reservoir and aquifer.

6.5.2.1 Separate component assessment

6.5.2.1.1 Micrite

The $\delta^{18}O_{VPDB}$ vs. $\delta^{13}C$ results for micrite from Well 2C are shown on Figure 6.13. The range in $\delta^{13}C$ values are from 1.7 ‰ to 4.1 ‰. The $\delta^{18}O_{VPDB}$ values for micrite range between -2.9 ‰ and -1.0 ‰.

Figure 6.13: Cross plot for $\delta^{18}O_{VPDB}$ vs. $\delta^{13}C$ for micrite in Well 2C. Data courtesy of Petrobras. See Appendix 7 for the data values. The box represents the postulated ranges for Albian carbonates and is in good agreement with Veizer et al. (1999), Huber et al. (2011) and Price and Harwood (2012).
6.5.2.1.2 Calcite cements

A separate component assessment was undertaken on the cements present in Field C to determine their bulk δ¹³C and δ¹⁸OVPDB values prior to a more detailed in-situ analysis. The range in δ¹³C values for peloids is between 2.9 ‰ and 3.8 ‰. The δ¹³C for syntaxial cement is 3.4 ‰ and the δ¹³C values for drusy cements range between 2.6 ‰ and 2.8 ‰ (Figure 6.14).

The δ¹⁸OVPDB values for peloids range between -3.1 ‰ and -1.5 ‰. δ¹⁸OVPDB for syntaxial cement is -1.9 ‰ and for drusy cements between -3.6 ‰ and -2.2 ‰ (Figure 6.14).

Figure 6.14: δ¹⁸OVPDB vs. δ¹³C cross plot for the separate component analysis of peloids and calcite cements. The data are from oil reservoir samples taken from Well 2C. The geochemical data presented are bulk signatures because the cements analysed are small, making sampling difficult. See Appendix 7 for the data values.

The box represents the postulated ranges for Albian carbonates and is in good agreement with Veizer et al. (1999), Huber et al. (2011) and Price and Harwood (2012).
6.5.2.2 In-situ assessment

In-situ geochemical analysis can help to further constrain the origin of the calcite cements in Field C. Isotopic and elemental assessments have been undertaken for the complete cementation history of Field C.

Two samples were chosen from Well 1C for in-situ assessment; one of these samples is from the oil reservoir and the other from the aquifer. These samples were chosen because they contain all the cements identified in Field C and they also have > 5 % remaining porosity, which would suggest that cementation could continue to the present day because a sufficient volume of pore space remains to be occluded (Table 6.1). The in-situ geochemistry of the cements in the samples is presented in the following subsections.

<table>
<thead>
<tr>
<th>Poroperm data</th>
<th>Porosity (%)</th>
<th>Permeability (mD)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Oil reservoir</td>
<td>17.3</td>
<td>61</td>
</tr>
<tr>
<td>Aquifer</td>
<td>4.5</td>
<td>0.1</td>
</tr>
</tbody>
</table>

Table 6.1: Plug derived porosity and permeability data. The samples used for the in-situ elemental and δ¹⁸O VPDB assessment have not had their plug porosity and permeability measured. Therefore, the data presented here is taken from plugs that are closest to the chosen sample, these plugs are from a similar depositional fabric and are taken as representative of the chosen sample.
6.5.2.2.1 Elemental data

Multiple Electron Probe Microanalysis (EPM) transects were made across the cements in both samples. These analyses are made from the pore wall into the pore centre, making sure that all the cement zones were sampled. This assessment allowed the elemental concentrations of the cements in both samples to be obtained.

There are similar concentrations of calcium, silica and strontium in the cements of the aquifer and oil reservoir (Table 6.2). However, aluminium is only observed within the oil reservoir. Also on average the concentration of magnesium is higher in the oil reservoir than in the aquifer. The lowest $^{m}$Mg/$^{m}$Ca ratio for the cements is obtained in the cement adjacent to an un-occluded pore or from cement at the centre of an occluded pore; the lowest $^{m}$Mg/$^{m}$Ca ratio obtained in the oil reservoir is 0.15 in comparison to 0.13 for the aquifer.

Table 6.2: Elemental composition (in ppm) for the calcite cements present in Field C (see Appendix 5 for the data values). $x$ = below detection limit. $N$ is the number of assessments that are above the detection limit in relation to standard error. The porewater $^{m}$Mg/$^{m}$Ca concentration was calculated using a partition coefficient of 0.21 (Rimstidt et al., 1998).

<table>
<thead>
<tr>
<th></th>
<th>Ca</th>
<th>Na</th>
<th>Mg</th>
<th>Al</th>
<th>Si</th>
<th>K</th>
<th>Sr</th>
<th>$^{m}$Mg/$^{m}$Ca</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Oil reservoir</strong></td>
<td>395899 ± 745</td>
<td>219 ± 39</td>
<td>2473 ± 341</td>
<td>384 ± 214</td>
<td>200 ± 15</td>
<td>432 (n = 1)</td>
<td>480 ± 72 (n = 3)</td>
<td>0.49 ± 0.07 (n = 17)</td>
</tr>
<tr>
<td>(n = 17)</td>
<td>(n = 4)</td>
<td>(n = 17)</td>
<td>(n = 7)</td>
<td>(n = 17)</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td><strong>Aquifer</strong></td>
<td>399943 ± 900</td>
<td>134 ± 1</td>
<td>1099 ± 184</td>
<td>x</td>
<td>197 ± 7</td>
<td>x</td>
<td>384 ± 25 (n = 2)</td>
<td>0.22 ± 0.04 (n = 13)</td>
</tr>
<tr>
<td>(n = 13)</td>
<td>(n = 1)</td>
<td>(n = 13)</td>
<td></td>
<td></td>
<td>(n = 13)</td>
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</tr>
</tbody>
</table>

6.5.2.2 Ion-microprobe $^{18}$O$_{VPDB}$ data

In-situ $^{18}$O$_{VPDB}$ values were obtained from a single sample in the oil reservoir of Well 1C (Table 6.1, 6.2). All transects made are from the pore wall into the pore centre making sure that all cement zones are analysed. The $^{18}$O$_{VPDB}$ values are then placed in terms of the cement phase analysed and in terms of the cement
zone assessed (Table 6.3). The $\delta^{18}O_{VPDB}$ values for fringing calcite cement are between -3.4‰ to -2.3‰, for syntaxial calcite cement -2.8‰ to -0.2‰ and for drusy cement between -3.9‰ and -2.4‰.

Table 6.3: In-situ $\delta^{18}O_{VPDB}$ results for the cements in the oil reservoir. Key: † - syntaxial; • - fringing; ‡ - drusy. No oil inclusions were observed in the cement zones.

Three cement zones were identified in the sample, an initial dull black cement zone, a moderate orange zone and then the youngest dull brown cement zone (Figure 6.15). The $\delta^{18}O_{VPDB}$ progressively decreases from the oldest cement zone into the youngest cement zone (Figure 6.16).
Figure 6.15: Cathodoluminescence images for (a, b) syntaxial and (c, d) fringing calcite cements the moderate orange cement zone (cement zone 2) is marked by the white arrow. The same cement zones are visible in syntaxial and fringing calcite cements.

Figure 6.16: Composite plot for all the in-situ $\delta^{18}O_{VPDB}$ data obtained from the cements in the oil reservoir of Field C. The numbers on the figure correspond to the cement zone sampled (Table 6.3).
6.5.3 Discussion

The cements present in the Macaé Group were thought to have precipitated in freshwater phreatic conditions (Carozzi et al., 1983). This is based on very light $\delta^{13}$C values and the lack of luminescence of the cement that indicates an iron-rich, manganese poor calcite. This conclusion is not supported by this study.

Influx of meteoric water is thought to have affected the Macaé Group (Carozzi et al., 1983; Meyers, 1978). If meteoric water has affected the $\delta^{13}$C of bulk micrite, it may be expected that the $\delta^{13}$C concentration would become more negative due to the addition of organic derived carbon (James and Choquette, 1984). However the $\delta^{13}$C of micrite is between 1.7 ‰ to 4.1 ‰ which is similar to Albian seawater where $\delta^{13}$C values of between 0.3 ‰ and 3.7 ‰ are observed (Price and Harwood, 2012; Veizer et al., 1999). Furthermore, the $\delta^{18}$OVPDB of modern meteoric groundwater in tropical areas with an average monthly temperature of ~20°C is -4 ‰ (± 2) (Rosales and Perez-Garcia, 2010). The majority of $\delta^{18}$OVPDB data for bulk micrite (Figure 6.13) are greater than -2 ‰. Therefore the $\delta^{18}$OVPDB values obtained from micrite are indicative of precipitation from Albian seawater where $\delta^{18}$OVPDB values between -2 ‰ and 0 ‰ are typical (Price and Harwood, 2012; Veizer et al., 1999). The $\delta^{13}$C and $\delta^{18}$OVPDB data presented here for micrite suggest that it is unlikely that micrite has been altered by meteoric water and indicate that the cements precipitated from pore water that has a similar composition to Albian seawater.

However, the $\delta^{13}$C and $\delta^{18}$OVPDB of bulk micrite also becomes progressively more positive at a shallower depth, suggesting that although the $\delta^{13}$C and $\delta^{18}$OVPDB of micrite do correspond with Albian seawater, micrite is likely to have been altered at increased depths and temperatures (Figure 6.17). The linear trend for $\delta^{18}$OVPDB vs. depth can be described by: $y=-101.76x+1629$, this line has $R^2 = 0.8$. Using this trend, the Kim and O’Neil (1997) equation and an initial pore fluid $\delta^{18}$OSMOW of -1.2 ‰ (Carvalho et al., 1995), a palaeo-geothermal gradient of 33°C/km can be calculated for the time at which rock-water interaction ceased.
Figure 6.17: Bulk micrite a) $\delta^{13}C$ and b) $\delta^{18}O_{VPDB}$ vs. depth for Well 2C. There is a strong relationship between $\delta^{18}O_{VPDB}$ and depth ($R^2 = 0.8$).

Mohriak et al. (1990) suggests that after rifting the geothermal gradient of the Campos Basin is ~30°C/km and that this decays linearly until the current geothermal gradient (20°C/km). The geothermal gradient of 33°C/km from bulk micrite therefore suggests that rock-water interaction ceased slightly after rifting, when the geothermal gradient was high.

Therefore, it is likely that no significant input of external fluid with a different $\delta^{18}O_{SMOW}$ to Albian seawater has occurred and that the $\delta^{18}O_{VPDB}$ of micrite is recording the palaeo-geothermal gradient and rock-water interaction at increasing temperature. This data would also suggest that rock-water interaction ceased in the aquifer and oil reservoir at a similar time; otherwise the $\delta^{18}O_{VPDB}$ of either the aquifer or oil reservoir would not lie on a linear gradient which can be related to the geothermal gradient.

The $\delta^{13}C$ values of the calcite cements are between 2.6 ‰ and 3.4 ‰. This is similar to that of bulk micrite (Figure 6.17a) and to that of peloids (Section 6.5.2.1). This suggests that the cements that precipitated in Field C formed from pore fluid that has a similar $\delta^{13}C$ composition to Albian seawater. This is contrary to the findings of Carozzi et al. (1983) where $\delta^{13}C$ values of -8 ‰ to -20 ‰ are taken to indicate that the cements formed from meteoric pore water.

The $\delta^{18}O_{VPDB}$ value of calcite cements is -1.9 ‰ for syntaxial cements and -3.6 ‰ to -2.2 ‰ for drusy cements. Assuming a $\delta^{18}O_{SMOW}$ of -1.2 ‰ for preglacial
oceans (Carvalho et al., 1995) and using the Kim and O'Neil (1997) equation, the temperature of precipitation for syntaxial cement is 18°C and between 20°C and 26°C for drusy cement. This data suggests that the cements formed from pore water that was at a similar temperature to Albian seawater (20-30°C; Pucéat et al. (2007)). A similar conclusion is also supported by the in-situ δ¹⁸OVPDB data.

The average in-situ δ¹⁸OVPDB value for the three cement zones identified in the oil reservoir is -0.5 ± 0.1 ‰ (n=4), -1.8 ± 0.2 ‰ (n=8) and -3.2 ± 0.2 ‰ (n=12). Assuming a δ¹⁸OSMOW of -1.2 ‰ for preglacial oceans (Carvalho et al., 1995) and using the Kim and O'Neil (1997) equation, the average temperature of precipitation for the cement zones is 12°C, 18°C and 24°C respectively. The highest and lowest δ¹⁸OVPDB value obtained by this study for the oldest and youngest cement zone is -0.2 ‰ and -3.9 ‰ respectively, this would suggest that all the cements present in the sample were formed at temperatures between 10°C and 28°C; assuming a pore water δ¹⁸OSMOW of -1.2 ‰ (Carvalho et al., 1995) and using the Kim and O'Neil (1997) equation.

The youngest cement zone is non-luminescent under the cathodoluminoscope, this was taken by Carozzi et al. (1983) to indicate precipitation from meteoric water. However, a non-luminescent cement zone does not indicate a meteoric deposit, but indicates an environment that is oxidising (Moore, 1989). Therefore it is possible that the first cement zone identified in Field C formed in a marine depositional setting, this is also suggested by the temperatures derived from the δ¹⁸OVPD temperature proxy which suggest the first cement zone precipitated at a temperature of 12°C. This temperature is cooler than that typically predicted for Albian seawater (Pucéat et al., 2007), but is in agreement with some studies (Huber et al., 2011; Price and Harwood, 2012). This would therefore imply that the first cement zone may accurately record the δ¹⁸OVPD of calcite cement precipitated from marine water. Cementation is then thought to continue into the shallow burial zone until a temperature of ~ 28°C, this temperature is lower than the current reservoir temperature (70°C) and therefore suggests that cementation ceased prior to present day.
Assuming that the initial $^{\text{m}}\text{Mg}/^{\text{m}}\text{Ca}$ of pore water is similar for the samples assessed from the oil reservoir and aquifer; this is valid because the samples are < 20 m apart, the $^{\text{m}}\text{Mg}/^{\text{m}}\text{Ca}$ of the cements can be used as an approximate geothermometer. The average $^{\text{m}}\text{Mg}/^{\text{m}}\text{Ca}$ ratio for the cements in the oil reservoir is higher than for the aquifer (Table 6.2), it is therefore suggested that the majority of cement precipitated at a lower temperature in the oil reservoir (~15°C; Figure 4.2) than in the aquifer (~ 60°C; Figure 4.2). However, the lowest $^{\text{m}}\text{Mg}/^{\text{m}}\text{Ca}$ ratio for the cements in the oil reservoir and aquifer is similar which may suggest that cementation continued to a similar temperature, and so to a similar burial depth in the oil reservoir and in the aquifer (Section 6.5.2.2); but this cannot be confirmed by this study.

Aluminium is only present in the youngest cements of the oil reservoirs. It has been proposed that aluminium can complex with organic matter and therefore may be related to the migration of oil in many reservoirs (Maliva et al., 1999). However, in Field C no oil inclusions are observed in the cements and therefore the increase in aluminium cannot be shown to be directly related to the migration of hydrocarbons. Oil inclusions may not be present in the cements either due to the slow precipitation rates of calcite cement, or due to cementation ceasing prior to, or because of, oil charge.

There are no burial cements observed in Field C, with no blocky calcite cement or saddle dolomite being present. Minor stylolitisation is observed in the field however stylolitisation can begin in the early burial diagenetic zone at depths of ~ 300 m (Ebner et al., 2009). This would suggest that the calcite cement in Field C formed in the marine phreatic zone or shallow burial zone at temperatures $\leq$ 28°C and that cementation ceased, in the oil reservoir, prior to the current reservoir temperature (70°C), this may be a result of early oil charge.

**6.6 Basin Modelling**

Basin modelling will help to determine the origin of the petroleum that is within Field C and can be used to understand the approximate time and temperature at which hydrocarbon entered the field. This data can be used with the
aforementioned depositional and diagenetic assessment to evaluate the hypothesis that early oil charge in Field C has preserved reservoir quality by stopping cementation, which in turn has led to the preservation of anomalously high porosities in the oil reservoir.

A variety of information is required to construct a basin model this includes: the tectono-stratigraphic history of the basin, the petroleum system (including source rock geochemistry), a present day structural cross section and the palaeo-water depth, sediment water interface temperature and heat flow of the basin through time (Figure 1.23). The required inputs are summarised in the following subsections. This thesis utilises literature and well data to construct suitable basin models for Field C.

6.6.1 Tectono-stratigraphic setting

The tectono-stratigraphic framework of the Campos Basin has been discussed briefly in Section 1.5.2 and is summarised in Figure 6.18. This section presents the detailed tectono-stratigraphic framework of the Campos Basin that is needed for the construction of a basin model. The tectono-stratigraphic framework of the Campos Basin can be divided into three phases: the first phase is known as the rift phase and commences with the breakup of Gondwana in the Early Cretaceous and is characterised by mafic volcanism (Mohriak et al., 1990; Tigre et al., 1983), the transitional phase succeeds the rift phase and is characterised by sedimentary deposition during rifting (Guardado et al., 2000), the final phase is the marine phase where deposition occurs on a thermally subsiding margin during sea-floor spreading (Guardado et al., 2000).
Figure 6.18 (previous page): Chrono-stratigraphic column for the Campos Basin. The tectono-stratigraphic framework of the Campos Basin can be divided into three phases: the rift, transitional and marine phases. Modified from Contreras et al. (2010). Key: Outei. – Outeiro.

6.6.1.1 Rift phase

Hot-spot volcanism preceded the rifting of the continental crust (Beglinger et al., 2012b; Furlong and Fountain, 1986) and is thought to have led to the deposition of the Paraná-Etendeka igneous province (Renne et al., 1992). Rifting of Gondwana commenced in the Hauterivian/Aptian (Guardado et al., 1990; Tigre et al., 1983) and is associated with: widespread mafic volcanism that led to the deposition of tholeitic, vesicular basalts (Mohriak et al., 1990; Tigre et al., 1983) (Figure 6.18), and basement involved block rotated faulting in a rapidly subsiding crust (Figure 6.19).

Figure 6.19: Faults associated with the break-up of Gondwana. Numerous antithetic and synthetic normal faults are currently present in the Campos Basin. Note the extent and orientation of the faults (Guardado et al., 1990).
Faulting caused the development of a series of horsts, grabens and half grabens that are bound by synthetic and antithetic normal faults. The faults can have throws of as much as 2500 m (Guardado et al., 1990) and are in an orientation that is coincident with structural lineaments evident to the West in the Precambrian shield; this may suggest that extensional faulting reactivated pre-existing crustal weaknesses (Ponte and Asmus, 1978).

### 6.6.1.2 Transitional phase

Rift tectonics influenced the distribution of the 90-600 m (Guardado et al., 2000; Tigre et al., 1983) thick Barremian-Aptian Lagoa Feia Group by fault block movement, which changed the position and accumulation pattern (depocentre) of the Lagoa Feia Group. The deposits of the Lagoa Feia Group formed in a lacustrine environment that ranged from fresh to hypersaline (Beglinger et al., 2012b; Rangel and Carminatti, 2000; Tigre et al., 1983), with organic rich shales depositing in the basin depocentre (Atafona Formation) and carbonate sedimentation on the basin rim (Guardado et al., 2000; Ojeda, 1982; Rangel and Carminatti, 2000).

The $\delta^{13}$C of Lagoa Feia carbonates carry signatures that are characteristic of anoxic conditions and bacterial degradation via sulphate reduction (Figure 6.20). The more negative $\delta^{13}$C in the younger deposits of the Lagoa Feia Group may be related to deposition in a progressively more saline environment. This is thought to be related to an early marine incursion into the basin and progressive stagnation (Guardado et al., 1990).
Most of the faults that were active during rifting and during the deposition of the lower sequence of conglomerates and carbonates became inactive prior to the deposition of the Retiro Formation (Guardado et al., 1990).

The São Paulo and Walvis ridges acted as major barriers to general Ocean flooding into the Campos Basin and led to its relative isolation. During an arid climate, seawater incursions from the South into this basin allowed for the development of a thick sequence of evaporite known as the Retiro Formation (Gamboa and Rabinowitz, 1981; Guardado et al., 1990). The Retiro Formation is composed of halite and anhydrite (Scotchman et al., 2010) and known to thicken to the East (Guardado et al., 1990). The progressive opening of the South Atlantic resulted in the breakdown of the São Paulo and Walvis ridge topographic barriers and allowed for the deposition of Albian Macaé Group.

**Figure 6.20: $\delta^{13}C$ for the lacustrine carbonates of the Lagoa Feia Group in Well 1C. Data courtesy of Petrobras.**
6.6.1.3 Marine phase

The onset of sea-floor spreading marks the transition from the transitional lagoonal depositional regime to the beginning of the marine phase (Figure 6.18), and is a time of continued thermal subsidence on the margins of the South Atlantic (Guardado et al., 2000). The first deposits of the marine phase are the carbonates of the Quissamã Formation of the Macaé Group; these deposits are discussed in Sections 1.5.2 and 6.3.

A regional transgression in the Middle-Upper Albian led to the deposition of the Outeiro Formation of the Macaé Group that is characterised by sedimentation consisting of carbonates and interbedded shale/mudstone. The final deposit of the Macaé Group is the Upper Albian-Cenomanian Imbetiba Formation that is characterised by marls and shales in the distal portions of the Campos Basin, but on the basin margins the formation is marked by shallow water carbonate (oolitic and bioclastic constituents) (Silva et al., 2012).

Deposition after the formation of the Macaé Group is characterised by: recurrent magmatic events from the Late Cretaceous to the Eocene that are indicated by the presence of basic intrusions and volcanic tuff (Ponte and Asmus, 1978), and by the deposition of a very thick marine sedimentary sequence. This sedimentary sequence was deposited over a long period of sea-level rise and is thought to be sourced by the Serra do Mar coastal mountain chain (Guardado et al., 1990). This marine sequence is known as the Campos Group and is composed of the Carapebus, Ubatuba and Embore Formations (Figure 1.16).

During the Cenomanian to the Maastrichtian the basin was starved of sediment and the prodelta black to grey marine shales of the Carapebus Formation were deposited. This is thought to be a consequence of tectonic subsidence, eustatic sea-level rise and a relatively low influx of terrigeneous sediments (Guardado et al., 1990).
The sandstone turbidites of the Ubatuba Formation were deposited in the Turonian and Santonian/Maastrichtian during short-term sea-level falls (Beglinger et al., 2012b; Guardado et al., 2000; Tigre et al., 1983).

Eustatic sea-level fall during the deposition of the Embore Formation led to the progradation of a clastic wedge that is composed of material derived from the Serra do Mar uplift (Siri Member) (Figueiredo and Mohriak, 1984) (Figure 6.18). Sea level fall also resulted in the formation of several canyon and turbidite systems (Figueiredo and Mohriak, 1984), the locations of which are strongly determined by salt tectonics (Moraes, 1989). These deposits are interbedded with deep marine calcilutites (Grussai Member) (Beglinger et al., 2012b; Guardado et al., 2000). The average thickness of the Embore Formation as a whole is 1600 m (Tigre et al., 1983).

6.6.2 Petroleum systems evaluation

This section identifies the petroleum system of Field C; see Section 5.3 for the definition of a petroleum system. This assessment will involve the identification of the source rocks present in the basin and their hydrocarbon generating potential, the migration pathways, the plays and reservoirs and the traps and seals in the system. These data are obtained from the literature and from Petrobras.

6.6.2.1 Source rocks and maturation

Potential source rocks in the basin include: the Atafona Formation and the marine shales of the Albian, Cenomanian-Turonian and Tertiary (Figure 6.18). The Albian shales are characterised by low organic carbon content, thus these carbonates lack source potential (Guardado et al., 1990). The shales of the Late Cenomanian/Early Turonian Stage were deposited in anoxic conditions as indicated by the high organic carbon, and therefore are composed of type 1 and type 2 kerogen. However, the vitrinite reflectance of these shales suggest that the rocks are thermally immature, and have not contributed to the hydrocarbon potential in the basin (Guardado et al., 1990). Therefore, all the oil found in the Campos Basin is thought to be derived from the organic rich shales of the Atafona Formation (Section 6.6.1.2) (Beglinger et al., 2012b; Guardado et al., 1990; Halbouty, 2003).
The transitional phase, lacustrine shales of the Atafona Formation are rich in algal and bacterial derived kerogen. This causes the source rocks to have a Total Organic Carbon (TOC) of < 9 % and an average of 2-6 %; an average value of 4 % was used in the model, the Hydrogen Index (HI) can be as high as 900 mg HC/g TOC, an average value of 600 mg HC/g TOC is used in the model (Figure 6.21). The kinetics used for these source rocks are for type 1 kerogen (Guardado et al., 1990; Pepper and Corvi, 1995).

Figure 6.21: Geochemical well log from the Campos Basin showing the source potential of the Lagoa Feia Group (Guardado et al., 2000; Mohriak et al., 1990). The V, data would suggest that the Lagoa Feia source rock does not enter the early oil generation window (0.55-0.7 %; Sweeney and Burnham (1990)) until ~ 3000 m.

The time at which the Lagoa Feia Group reached maturity is controversial. Guardado et al. (1990) suggests that the organic matter became thermally mature relatively late in the basins history (Eocene) and spent a long time in the oil generation window, this is also suggested by Nascimento et al. (1999) and Guardado et al. (2000) where peak maturity was identified in the Late Miocene. However, Halbouty (2003) suggests the source rocks entered the generation window in the
Albian to Coniacian-Santonian and reached peak generation in the Turonian-Santonian to the Miocene, this agrees with the findings of Beglinger et al. (2012b) and Mohriak et al. (1990).

6.6.2.2 Migration

The Atafona Formation (transitional phase) is separated from the Quissamã Formation (marine phase) in the Campos Basin by a salt layer (Retiro Formation) (Figure 6.18). Hydrocarbon migration from source rock (Atafona Formation) to reservoir (Quissamã Formation) requires that the oil migrates through salt windows in the evaporites, these windows are thought to be a result of the salt layer thinning and moving due to differential loading and increased overburden (Meister, 1984; Soldán et al.) (Figure 6.22).

The time of hydrocarbon migration into the reservoir is thought to be controlled by the time at which salt windows develop (Guardado et al., 2000; Soldán et al., 1995). However, the time at which halokinesis began and therefore at which salt windows may be present is controversial. Halokinesis is thought to commence between 100 to 80 Mya (Cobbold et al., 2001) and in the Late Cretaceous (Carozzi and Falkenhein, 1985; Soldán et al., 1995) in the Campos Basin. This is in contrast to halokinesis on the analogous African margin where halokinesis is thought to have begun in the Albian (Eichenseer et al., 1999), which may suggest that salt windows are present from the Early Cretaceous.
6.6.2.3 Reservoirs

Four reservoirs are present in the Campos Basin, these are: the turbiditic sandstones of the Carapebus Formation (Campos Group), the oncolitic–oolitic and calcilutites of the Quissamã Formation, the coquinas of the Coqueiros Formation and the Neocomian basalts when fractured (Tigre et al., 1983) (Figure 6.18). Stacked oil accumulations are common in most fields because of the wide distribution of reservoir and non-reservoirs both in time and space (Guardado et al., 1990). However, the most volumetrically important hydrocarbon accumulations are in the deep water fans that are distributed in the stratigraphic column from the Late Cretaceous to Late Tertiary, and in Albian carbonates of the Quissamã Formation (Mohriak et al., 1990).

6.6.2.4 Trap and seal potential

The oil in the Quissamã Formation of Field C appears to be in a stratigraphic trap (isolated platform) that developed on a basement high (Section 6.3). The lateral and top seal for the Quissamã Formation in Field C is the overlying argillaceous
pelagic lime mudstones (Figure 6.4) (Carozzi and Falkenhein, 1985). It has also been suggested that biodegraded oil in some of the fields in the Campos Basin could have sealed the reservoirs (Soldán et al., 1995), however this is unlikely to be the case in Field C.

The identification of biodegraded oil within the oil fields of the Campos Basin can provide a useful tool to assess whether a field has experienced an influx of meteoric water. This is because the biodegradation of hydrocarbon is a bioxidation process, and therefore requires the entry of freshwater into the reservoir. The biodegradation of oil will reduce the American Petroleum Institute (API) gravity of the oil (to ~10º for extremely biodegraded oil in the Campos Basin; Soldán et al. (1995)) and therefore by comparing the API of the oil in the reservoir to the API of unaltered oil any potential influx of meteoric water, after oil charge into the reservoir, can be identified. The API of the oil in the Quissamã Formation is 21º (Petrobras), this is close to the 25ºAPI of unaltered oil in the Campos Basin (Guardado et al., 1990; Soldán et al., 1995), suggesting no significant influx of meteoric water into the reservoir after hydrocarbon charge.

6.6.3 Data

2D seismic data were used to interpret and build a cross section that forms the base of the 2D field scale model (Figure 6.23), this was done in collaboration with Andrew Berrow (Berrow et al., 2013). The seismic lines provided by Petrobras are of a small length (< 10 km) and therefore unlikely to incorporate the entire petroleum system of the basin. Therefore, a second cross section was used to understand the wider petroleum system of the basin (Figure 6.24), the cross section used to build this model is taken from the literature (Contreras et al., 2010). The horizons from this study were digitised by Berrow et al. (2013); a Master’s project co-supervised by the author. Each of the horizons identified for the field- and basin- scale models were given a time of deposition and the layer given a lithology (see Appendix 8).

Mud logging data are available for 17 wells in Field C. These data have been used to help determine the stratigraphic column for each well, therefore in the field scale model the modelled lithofacies have been validated with this data.
Figure 6.23: Field-scale seismic line. The interpreted horizons identified on the seismic line are labelled and are calibrated to the formation tops identified on well logs; these horizons were interpreted in collaboration with Berrow et al. (2013).

Figure 6.24: Interpreted seismic line used for the basin scale model, from Contreras et al. (2010). The horizons identified on the cross section by Contreras et al. (2010) were digitised by Berrow et al. (2013).
6.6.4 Boundary conditions

6.6.4.1 Palaeo-Water Depth (PWD)

The Palaeo-Water Depth (PWD) profile used for Field C is based on depositional environment changes along with eustatic sea-level curves (Haq et al., 1988) and Google Earth’s palaeo-geometry tool. A PWD trend for the shelf (Figure 6.25a), slope (Figure 6.25b) and basin (Figure 6.25c) have been applied in the basin scale model. The isolated platform is located on the continental shelf, therefore the PWD trend used for the field scale model is that for the shelf (Figure 6.25a).

![Figure 6.25: Palaeo-Water Depth (PWD) profiles that have been used in the basin scale model, for a) the shelf, b) the slope, and c) the basin.](image)

6.6.4.2 Sediment-Water Interface Temperature (SWIT)

The Sediment-Water Interface Temperature (SWIT) has been obtained by using Petromod’s SWIT tool that is based on the Ph.D. dissertation of Wygrala (1989). Latitude and continent information is input and SWIT is estimated by the software. This tool takes into account continental drift, climatic changes and the previously input PWD. SWIT trends for the shelf (Figure 6.26a), slope (Figure 6.26b) and basin (Figure 6.26c) have been applied in the basin scale model. The trend for the shelf has been used in the field scale model (Figure 6.26a).
6.6.4.3 Heat Flux (HF)

Mohriak et al. (1990) showed that the subsidence history of the Campos Basin can be attributed to an initial rifting and subsequent thermal recovery of the lithosphere. However, there are some important deviations from the homogeneous stretching model (Hardenbol et al., 1998), because that model does not account for the observed tectonic subsidence (Beglinger et al., 2012b). Therefore numerous β factors for the Campos basin have been published (Beglinger et al., 2012b; Mohriak et al., 1990).

Most published β factors for the basin lie between 1.5 and 2. The high β factor of 2 suggests greater thinning and a larger heat flux than when a β factor of 1.5 is used. Through applying the uniform stretching model of McKenzie (1978) and by assuming rifting commenced at 137 Mya and continued until the transition zone (127 Mya) (Figure 6.18), the following Heat Flux (HF) profiles are produced when β = 2 (Figure 6.27a) and β = 1.5 (Figure 6.27b).
To determine the effect that changing the $\beta$ factor has on the timing of source rock maturation and the timing of petroleum emplacement, both HF profiles shown in Figure 6.27 will be applied to the basin and field scale model. The HF profiles included in the model therefore describe the maximum and minimum HF that is likely to have been present in the Campos Basin and are therefore likely to describe the earliest and latest time at which the source rocks in the basin matured.

6.6.5 Simulation parameters

The same simulation parameters used in the basin model for Field A have been used for Field C (Section 5.4.7). The optimisation percentage (percentage of geometrical difference between model runs and the modelled present day cross section) between individual model runs that used a $\beta$ factor of 2 is $< 0.019 \%$ for the field scale model and $< 0.024 \%$ for the basin scale model. When a beta factor of 1.5 is applied the optimisation for the field scale model is $<0.021 \%$ and for the basin scale model $<0.0013 \%$. This would suggest that the models are an excellent fit to the present day cross section.

6.6.6 Calibration

The models were calibrated to all available data this includes the present day geothermal gradient and the $V_r$ maturity index for the source rocks. The output from the models presented here is in good agreement with these present day values,
suggesting that the models are a good representation of the basin evolution for the Campos Basin.

The palaeo-geothermal gradient during the Cretaceous is ~30°C/km (Section 6.5.3) this gradient is then thought to decline until the present day geothermal gradient (~20°C/km). The modelled geothermal gradients agree with this decline and are in good agreement with the present day borehole temperatures obtained for 17 wells in Field C (Figure 6.28) and with the current reservoir temperature (70°C).

**Temperature calibration when β = 2**

![Figure 6.28](image)

*Figure 6.28*: Calibration of the modelled present day geothermal gradient (blue line) with borehole temperature data (crosses) when a β factor of a) 2 and b) 1.5 is used.

For maturity modelling and calibration the Easy%Ro kinetic model of Sweeney and Burnham (1990) was adopted. It is generally assumed that the Atafona Formation is immature at depth ~ < 3000 m (Figure 6.21). In the field scale and basin scale model the source rocks are immature until buried to depths ~ > 3000 m, when a β factor of 2 (Figure 6.29a) and 1.5 (Figure 6.29b) is used.
Figure 6.29: Present day maturity of the source rocks in the field scale model when a $\beta$ factor of a) 2 and b) 1.5 is used. The source rock is marked by an arrow, the area of this source rock that is in dark green is in the early oil maturation zone; $V_r = 0.55-0.7 \%$. 
6.6.7 Results of the simulations

The results presented in this section are from calibrated basin models (Section 6.6.6). The results extracted from the models include the burial history, timing of hydrocarbon generation, migration and accumulation and the temperature evolution for Field C.

6.6.7.1 Burial history

Two clear departures from the McKenzie (1978) stretching model are observed in the burial history plot (Figure 6.30). The first is during rifting where a large proportion of the sedimentary column for the basin was deposited very rapidly. The second is during the Late Cretaceous to Early Tertiary where subsidence is much less pronounced due to sediment starvation.
6.6.7.2 Hydrocarbon generation

The Atafona Formation source rock in the field scale model has expelled < 0.35 % of its total hydrocarbon by the present day when β factors between 1.5 and 2 are used. This would suggest the source is insufficiently rich to account for the oil in the field and suggests the hydrocarbon is derived from another source area (Figure 6.31a, b).

The basin scale model (Figure 6.31c, d) suggests that hydrocarbon starts to be generated from the lower Atafona at ~ 125 Mya for all β factors that are between 1.5 and 2. Both these models suggest that by 120 Mya ~ 60 % of the total hydrocarbon is expelled with ~ 70 % being expelled by 100 Mya. The upper Atafona source rock expels ~ 5 % of its total hydrocarbon between 10 Mya and 0 Mya when a β factor of 1.5 is used.
Figure 6.31: Transformation ratio for the field scale model when a β factor of a) 2 and b) 1.5 is applied and the transformation ratio for the basin scale model when a β factor of c) 2 and d) 1.5 is used. Note the change in scale of the y-axis between the field and basin scale models.
6.6.7.3 Migration and accumulation

No hydrocarbon is generated in the field scale model (Figure 6.31, 6.32) therefore the simulations suggest that all the hydrocarbon in the field is derived from elsewhere in the basin (Figure 6.33). The migration of hydrocarbon from the Atafona Formation into the Quissamã Formation of Field C is thought to be controlled by the time at which salt windows develop (Section 6.6.2.2); this is supported by the simulations. In order for hydrocarbon to charge the reservoir early and stop cementation by ~28°C, as suggested by the diagenetic assessment (Section 6.5), halokinesis must have commenced early to allow for the migration of hydrocarbon. In the models, because of the way the models were constructed, salt windows opened by ~ 100 Mya (Figure 6.32, 6.33); this agrees with the findings of Cobbold et al. (2001), however if halokinesis occurred later than ~ 100 Mya then the migration of hydrocarbons into the field will also be later.

The basin scale model suggests that oil and gas migrate along basement faults and vertically through the source rock from 125 Mya (Figure 6.34). However, until the deposition of the impermeable salt of the Retiro Formation (115 Mya) this hydrocarbon escapes the system. When a β factor of 2 is used in the basin scale model, gas accumulates at 114 Mya beneath the salt of the Retiro Formation and at 105 Mya oil also begins to accumulate. When the salt windows open at 100 Mya the oil and gas immediately migrate through the window, with gas also migrating through the Quissamã Formation and through the overlying shales that behave as an impermeable barrier to oil (Figure 6.34; 100 and 90 Mya, respectively).

When a β factor of 1.5 is applied gas accumulates beneath the salt of the Retiro Formation at 114 Mya and oil begins to accumulate at 103 Mya. The salt windows open at 100 Mya, which allows oil and gas to migrate through these windows and allows for oil to accumulate in the Quissamã Formation.
Figure 6.32: Salt movement from 105 to 85 Mya in the field scale model when $\beta = 2$. A continuous thickness of salt is present at 105 Mya, then by 102 Mya the salt thins and windows begin to open at 100 Mya.
Figure 6.33: Salt windows and oil (green arrows) and gas (red arrows) migration in the basin scale model, when $\beta = 2$. Initially (110 Mya) the salt is thicker to the East; this is based on the findings of Guardado et al. (1990) (Section 6.6.1.1). Salt migration has caused the development of a salt window in the West of the basin by 100 Mya. The red dashed area has been enlarged in Figure 6.34 to show the migration of hydrocarbon through this window.
Figure 6.34: Migration of hydrocarbon from the bottom of the Atafona source rock at 102 Mya, 100 Mya and 90 Mya, when $\beta = 2$. At 102 Mya a salt window was not present and therefore gas (red arrows) and oil (green arrows) accumulates beneath the salt. However, by 100 Mya a salt window opened and allowed for the effective migration of hydrocarbon into the reservoir, with the gas immediately migrating through the overlying seal and the oil accumulating in the Quissamã Formation.
6.6.7.4 Temperature evolution

The time vs. temperature profiles for the top and bottom of the Quissamã Formation when either $\beta = 2$ (Figure 6.35a) or $\beta = 1.5$ (Figure 6.35b) are very similar. Both models suggest that temperature rapidly increased due to initial rifting and rapid subsidence until $\sim 100$ Mya (Figure 6.35). At this time the temperature of the top of the Quissamã Formation is at a temperature of 40°C, whereas the bottom of the formation is at a temperature of 60°C. Then, due to the sediment accumulation rate decreasing, the temperature remains relatively constant until $\sim 20$ Mya where subsidence increases due to an increase in sediment accumulation due to the progradation of a clastic wedge derived from the uplift of the Serra do Mar mountain chain (Section 6.6.1.1) (Figueiredo and Mohriak, 1984). This increase in subsidence led to the temperature increasing in the top and bottom of the Quissamã Formation from 40°C to 60°C and from 50°C to 70°C respectively (Figure 6.35).
Figure 6.35: Temperature through time for the top and bottom of the Quissamã Formation when a \( \beta \) factor of a) 2 and b) 1.5 is applied. The dashed line marks the approximate temperature at which the last cement zone in the oil reservoir was precipitated (~28°C), this corresponds to cementation ceasing at ~ 105 Mya. The temperature profiles were obtained from the field scale model.
6.6.8 Discussion

When β factors between 1.5 and 2 are used the source rocks present in the field scale model do not reach sufficient maturity to expel the volume of hydrocarbon currently present in Field C. Therefore, the hydrocarbons that are thought to have charged the field are most likely sourced from source rocks located in deeper grabens that matured earlier (Figure 6.31, 6.33).

The basin scale model suggests that the Atafona source rock reached maturity in the Early Cretaceous (~125 Mya) irrespective of whether a β factor of 1.5 or 2 is used and by ~ 100 Mya (late Albian) > 70 % of the total organic matter in the deeper Atafona source rock has matured. This is in contrast to most published maturation models of the Campos Basin where maturation is thought to be in the Turonian-Santonian (94-84 Mya) (Beglinger et al., 2012b) or in the Late Miocene (~ 5 Mya) (Guardado et al., 1990).

The cause for the difference in the predicted time of maturation is due to a number of reasons. The published models are one-dimensional (Beglinger et al., 2012b; Guardado et al., 2000; Guardado et al., 1990; Halbouty, 2003; Mohriak et al., 1990), these models will therefore assume that all heat flow vectors are vertical and will typically neglect all time-temperature terms such as the transient and convection affects (e.g. Beglinger et al. (2012b), Mohriak et al. (1990)). Additionally, 1D models cannot accurately model the high thermal conductivities of salt domes or model changes in salt thickness due to halokinesis (Hantschel and Kauerauf, 2009) (e.g. Beglinger et al. (2012b), Mohriak et al. (1990)). Also some models do not take into account the heat generated through radioactive decay or correct for palaeowater depth; which will affect the burial history and therefore the time of source rock maturation (e.g. Guardado et al. (1990)).

Furthermore, the time at which sea floor spreading commences will also affect the maturation of the source rocks because this will affect the time of greatest heat flux (McKenzie, 1978). Sea floor spreading is included in the models of this thesis at 127 Mya (Austin Jr and Uchupi, 1982; Harry and Sawyer, 1992), this is also in good agreement with Mohriak et al. (1990) where sea floor spreading is thought to
have commenced at 130 Mya and also with the tectonostratigraphic framework provided by Petrobras (Figure 1.17). However, this disagrees with some authors that infer sea floor spreading commenced at ~ 120 Mya (Beglinger et al., 2012b; Guardado et al., 1990) and with others that infer sea floor spreading commenced at 115 Mya (Berrow et al., 2013). Typically the later the rifting phase is included in the model the later the source rocks will mature, this can be observed in the one-dimensional models of Beglinger et al. (2012b) and the two-dimensional models of Berrow et al. (2013) where the source rocks were predicted to mature during the Turonian-Santonian or during the late Albian, this is in contrast to the models presented here where the source rocks are predicted to mature by 125 Mya (Early Aptian).

The time at which salt windows form is of vital importance for all two- or three-dimensional basin models of the Campos Basin (one dimensional models do not model fluid flow) because this will control the time at which hydrocarbon can first enter the field. This is because if the source rocks mature prior to the formation of salt windows the hydrocarbon will not be able to charge the field. When a β factor of 2 is used oil begins to accumulate beneath the Retiro salt at 105 Mya, whereas when a β factor of 1.5 is used oil accumulates at 103 Mya. In both models oil enters the Quissamã Formation at 100 Mya because this is the time at which halokinesis is thought to have begun (Cobbold et al., 2001). However, if the salt windows open earlier hydrocarbon can enter Field C as early as 105 Mya.

The δ^{18}O_{VPDB} values for the calcite cement zones progressively decrease into the youngest zone and suggest a maximum precipitation temperature of ~ 28°C (Section 6.5.3). This temperature is used to determine the approximate time (through burial depth and temperature) at which cementation ceased in what became the oil reservoir (Figure 6.35). Using the model results the temperature of the upper reservoir is thought to progressively increase from 22°C to 42°C at 110 Mya to 100 Mya and was at a temperature of ~ 28°C by ~105 Mya. This would imply that salt windows must have been present at ~ 105 Mya in order for hydrocarbon to enter the field and stop cementation when the reservoir was at a temperature of ~ 28°C. This is opposed to the conclusions of other authors that suggest halokinesis began 100 Mya.
to 80 Mya (Cobbold et al., 2001) or in the Late Cretaceous (Carozzi and Falkenhein, 1985; Soldán et al., 1995) but agrees with Eichenseer et al. (1999) study for the salt on the conjugate margin of offshore Angola where it is suggested that salt windows opened in the Albian.

### 6.7 Conclusions

Field C is an isolated platform that developed on a basement high. This led to the development of high energy fabrics (grainstones and packstones) that are composed predominantly of oncoids, ooids and peloids. The distribution of fabrics in Field C is characteristic of an isolated platform with the crest of the structure being dominated by grainstones, whereas packstones dominate the flanks.

The petrophysical properties of the depositional fabrics in the oil reservoir have not been significantly altered by diagenesis, with the current reservoir quality being primarily determined by the depositional environment. This is thought to be due to the effect of oil charge stopping cementation; in some cases prior to any cementation. However, the porosity and permeability in the aquifer is much reduced for comparable grainstones and packstones and is a result of extensive cementation of the pore space.

Micrite $\delta^{18}$O$_{VPDB}$ values decrease with depth in a near linear manner. This suggests that the cements have undergone continued rock-water interaction at increasing temperature. Using a linear trend line the palaeo-geothermal gradient calculated is $\sim 33^\circ$C/km for the time at which rock-water interaction ceased. This agrees with the geothermal gradient that was calculated from the basin model results immediately after rifting and agrees with the initial geothermal gradient suggested by Mohriak et al. (1990), implying that rock-water interaction is likely to have ceased slightly after rifting.

The cements present in Field A are all relatively early cements that deposited in the vadose, the marine phreatic and the shallow burial zone. The $\delta^{13}$C and $\delta^{18}$O$_{VPDB}$ values for these calcite cements are similar to that of Albian seawater, which suggests that the cements are likely to have formed from a pore fluid similar in
composition to Albian seawater, this is in contrast to other authors that suggest the cements precipitated from meteoric water (Carozzi et al., 1983; Meyers, 1978). Therefore during the formation of the cements no fluid that is significantly different to that of Albian seawater is likely to have entered the field.

The average in-situ δ¹⁸O sub-VPDB measurement for the first non-luminescent cement zone in the oil reservoir is -0.5±0.1 ‰ (n=4), this cement zone is thought to have precipitated from subtropical Albian seawater. The highest and lowest δ¹⁸O sub-VPDB for the oldest and youngest cement zone in the oil reservoir is -0.2 ‰ and -3.9 ‰, respectively. This would suggest that all cements formed at a temperature between 10°C and 28°C. This is also supported by the lack of late burial cement phases in the field and would suggest precipitation ceased at a similar temperature to that of subtropical seawater.

The \(m\)Mg/\(m\)Ca ratio for the cements can be used as an approximate geothermometer (Table 6.2). The average \(m\)Mg/\(m\)Ca for cements in the oil reservoir is greater than that in the aquifer; this may suggest that the majority of cements formed at a shallower burial depth in the oil reservoir (~ 15°C) than in the aquifer (~ 60°C). The lowest \(m\)Mg/\(m\)Ca ratio for the oil reservoir (0.15) and aquifer (0.13) is similar; this may imply that cementation continued to a similar temperature in both the oil reservoir and aquifer however this cannot be confirmed by this study.

Oil is thought to have stopped cementation in the oil reservoir in the shallow burial zone (δ¹⁸O sub-VPDB data suggest by ~ 28°C) and preserved porosity. However in order for oil to stop cementation, oil charge must be early and prior to any significant burial, this is opposed to some authors who suggest oil charge in the Late Miocene (Guardado et al., 2000). The basin models for the Campos Basin that are presented here suggest that hydrocarbon can enter Field C as early as 105 Mya, when the reservoir is at a temperature of ~ 28°C. This hydrocarbon is not locally sourced but derived from deeper areas of the basin from source rocks in basement grabens. However, in order for this early oil charge to have stopped cementation, salt windows in the Retiro evaporate are required by ~ 105 Mya so that hydrocarbon can migrate into the Quissamã Formation. This is in contrast to some authors that suggest
halokinesis began 100 to 80 Mya (Cobbold et al., 2001) or in the Late Cretaceous (Carozzi and Falkenhein, 1985; Soldán et al., 1995) but agrees with Eichenseer et al. (1999).
CHAPTER 7
DISCUSSION AND CONCLUSIONS

7.1 Introduction

Typically a lower volume of hydrocarbon is recovered from carbonate reservoirs (10.1% for Field C (Filho, 2011)) when compared to siliciclastic reservoirs (< 62% in some reservoirs (Kabir et al., 2008)). This is due to two fundamental differences: 1) the sediment in siliciclastic reservoirs is allochthonous whereas the sediment for carbonates is autochthonous, which typically leads to more complex primary (depositional) pore systems for carbonates and 2) carbonates are more chemically reactive (Moore, 2001). This second point leads to greater diagenetic alteration to the primary pore space, which further increases pore complexity and tortuosity, making recovery more difficult. Although these differences between carbonate and siliciclastic reservoir rocks are widely accepted, the quantitative assessment of the nature and origin of the processes that control reservoir quality in carbonate reservoirs has received little quantitative documentation (Ehrenberg and Nadeau, 2005; Scholle and Halley, 1985).

This thesis presents a quantitative study on the controls on reservoir quality in three Early Cretaceous carbonate oil fields and has analysed two giant oil fields from the U.A.E. (Field A and B) in Chapters 2 to 5 and a third field from Brazil (Field C) in Chapter 6. Both these assessments commence with an analysis to determine the control that depositional environment has on final reservoir quality. The effect of diagenesis on modifying the pore volume is then determined by identifying the diagenetic constituents present, their relative order of formation and then quantifying the pore types and cements present in the fields. A detailed geochemical analysis (isotopic and elemental) is then undertaken to constrain the origin of the main pore
occluding cements and to identify the approximate temperature at which the main pore occluding cements precipitated. This understanding of the temperature at which cementation occurred is related, wherever possible, to the location of primary oil inclusions in order to understand the control of oil charge on the dynamics of cementation.

The diagenetic understanding obtained through petrographic and geochemical assessment is then integrated into basin models in order to improve their accuracy. These models are then used to predict, amongst other things the burial history, temperature evolution and the timing of source rock maturation and hydrocarbon migration. The predictions from the models can then be related to the petrographic/geochemical assessment to help further understand the origin of the processes that affect reservoir quality. For example, the modelled time-temperature evolutions of the reservoirs can be related to geochemically obtained temperatures of cement precipitation to help understand the time at which cementation ceased, the models can then be used to identify the processes that may have occurred at this time to cause cementation to cease (e.g. hydrocarbon charge). Furthermore, basin models can also be used to predict the likely origin of dissolution events identified through petrographic assessment. Therefore, by combining a detailed petrographic and geochemical assessment with basin modelling the controls on reservoir quality can be more accurately determined.

Complete descriptions, analyses and interpretations for the controls on reservoir quality within the three oil fields are provided in the preceding chapters. The purpose of this chapter is to firstly present the findings of this thesis by comparing and contrasting these three Cretaceous fields in order to identify any differences or similarities in the processes that control reservoir quality and secondly to provide a synopsis for the controls on reservoir quality and how the results of this thesis might be used to inform other studies.
7.2 Depositional Environment

The three oil fields assessed in this thesis are producing from Early Cretaceous carbonate reservoirs. Field A and B are producing from the Thamama Group and nine Hauterivian-Aptian depositional cycles (ix-i) have been analysed in this thesis from this Group. A detailed assessment of these depositional cycles has already been undertaken for Field B (Cox, 2010) and the depositional origin of these cycles is well understood (Alsharhan and Kendall, 1991; Strohmenger et al., 2006).

The typical depositional cycle described in the thesis of Cox (2010) is similar to the typical depositional cycle identified here for cycles iii-i in Field A and in the outcrop. A typical depositional cycle identified here shallows from pyritised and heavily stylolitised shales and mudstones, to orbitolinid, coated grain, rudist dominated wackestones and packstones which are overlain by pyritised mudstones and wackestones. The cycle is then capped by a bored hardground surface, which is thought to be regionally correlatable (Dickson et al., 2008). The cycles identified and described in this thesis are thought to form in a low to moderate energy environment on a carbonate ramp in a distal open marine, to proximal restricted marine depositional setting, this agrees with the observations of numerous other authors (Alsharhan and Kendall, 1991; Neilson et al., 1998; Strohmenger et al., 2006).

In contrast to the relatively low to moderate energy depositional setting for the Thamama Group, the cycles of the Albian Quissamã Formation in Field C, formed on a high energy isolated platform where bioclastic/oncolitic/oolitic/peloidal grainstones and moderate energy packstones dominate.

Once the depositional environments are understood the 3D morphology (the size, shape, orientation) of potential reservoirs can be predicted through the idealised models of LeBlanc (1972) for siliciclastics and Jung and Aigner (2012) for carbonates. These models are then populated with an understanding of the location of potential reservoirs. Typically when predicting the location of potential reservoirs it is assumed that for both siliciclastic and carbonate reservoirs the most porous and permeable lithofacies are shallow water high-energy deposits. This may be justified for siliciclastics because the reservoir quality is primarily controlled by primary
intergranular porosity and, without a significant diagenetic impact, there is normally a good relationship between porosity and permeability. This is contrary to carbonate reservoirs where multiple pore types (vuggy, intraparticle, fenestral, moldic, fracture etc.) are commonly present and the relationships between pores are highly complex, which typically leads to a poor correlation between porosity and permeability and makes the prediction of potential reservoirs more difficult.

Field C is interpreted to have formed in a higher energy depositional environment than either Field A and B. This suggests that if reservoir quality is controlled by the depositional environment alone then it may be expected that Field C will have a better reservoir quality than Field A and B; because Field C is dominated by depositionally more porous and permeable grain supported fabrics whereas Field A or B are typically composed of mud dominated fabrics (Enos and Sawatsky, 1981). However, subsequent diagenetic modifications to the depositional pore space may have greatly affected the location of potential reservoirs and therefore the 3D modelling of petrophysically important depositional fabrics would not accurately predict the location of potential reservoirs. This is especially true in carbonate reservoirs where the grain mineralogy was metastable aragonite or high-Mg calcite because the primary pore space is more prone to diagenetic alteration than for low-Mg calcite.

Some dominantly low-Mg calcite reservoirs have undergone limited diagenetic alteration and therefore the current reservoir quality is primarily controlled by the depositional environment. This is the case in the low-Mg Jurassic Smackover Formation in the North Haynesville Field in Louisiana (Ahr, 2008; Ahr and Hull, 1983; Bishop, 1968) and the Conley Field in Texas which is producing from the Mississippian Chappel Formation (Ahr, 2008; Ahr and Walters, 1985). The fields are producing from different depositional pore spaces with the North Haynesville Field producing from modified intergranular porosity in ooid-peloid-rhodolite grainstones whereas production from the Conley Field is from the intragranular porosity of crinoid-bryozoan grainstones and packstones. In both examples the depositional facies correspond well to the outline of the underlying palaeo-structural highs with the best reservoir quality being in the grain dominated
fabrics that formed typically on the crest of palaeo-structural highs. Therefore the location of potential reservoirs can be predicted by 3D modelling of the grain-dominated fabrics.

Although the North Haynesville and Conley Fields have not been significantly affected by diagenesis all carbonate reservoirs, to some extent, are affected by diagenesis. Furthermore in some cases diagenesis can completely alter the pore network, leading to there being no relationship between depositional environment and reservoir quality. Two such examples are where dolomitisation has led to the formation of intercrystalline porosity that has no relationship to depositional texture (Benson and Mancini, 1999; Kerans et al., 1994) and the second is where the reservoir is formed through diagenesis and dissolution cross cuts all pre-existing rock textures, fabrics and facies to create massive caves, caverns and karst features (Ahr, 2008); the Carlsbad Caverns in New Mexico are an example (Hill, 1990; Polyak and Provencio, 2001). These secondary processes will therefore cause the correlation between depositional fabric and reservoir quality to decrease and therefore the location of potential reservoirs cannot be predicted through the 3D modelling of petrophysically important fabrics.

All three fields assessed here were deposited from low-Mg seawater, which results in the reservoirs being relatively stable when compared with aragonitic or high-Mg calcite reservoirs. However, during the Early Cretaceous aragonitic biota did thrive, one such example is with rudists where the external portion of the valve is formed of low-Mg calcite and the internal portion of the valve is aragonitic. Therefore, typically in Field A and B the external valve is currently preserved, however the dissolution of the aragonitic valve (in the marine phreatic realm) has enhanced porosity. This event along with other diagenetic processes may have affected the location of potential reservoirs and so 3D modelling of the petrophysically important depositional fabrics may not accurately predict the location of potential reservoirs, therefore a detailed assessment was undertaken in order to determine the controls on reservoir quality.
7.3 Controls on Reservoir Quality

The depositional environment of both Field A and B greatly influences the location of the reservoirs identified in the fields. The higher energy depositional fabrics (grainstones and packstones) are the primary reservoirs for both fields and typically have a greater porosity and permeability than the underlying transgressive mudstones and wackestones. This can be observed in the plug porosity and permeability data for Field A and B where the deposits of the Transgressive Systems Tract (TST) have porosities and permeabilities of 2.5 ± 0.5 % (n=51) and 0.1 ± 0.5 mD (n=51) respectively, whereas the grain dominated fabrics of the Highstand Systems Tract (HST) have porosities of 11.9 ± 0.9 % (n=86) and permeabilities of 20.5 ± 10.9 mD (n=86). The change in depositional environment from the TST to the HST in each cycle is therefore thought to lead to the formation of multiple cyclic reservoir and non-reservoir intervals, where the reservoirs are primarily in the higher energy HST deposits and the non-reservoirs are typically within the deposits of the TST.

This is similar to Field C where the grainstones and packstones have a porosity and permeability of 19 ± 0.3 % (n=270) and 163 ± 36 mD (n=270) and the mudstones and wackestones have a porosity and permeability of 14.5 ± 1.2 % (n=20) and 2.2 ± 0.9 mD (n=20), respectively. However, in Field C grainstones and packstones dominate and mudstones and wackestones are relatively uncommon. This leads to no significant barriers to flow forming within the field, unlike in Field A and B where the low poro-perm properties of the TST, coupled with the widespread and abundant hardground surfaces have likely led to reservoir compartmentalisation.

The pores in Field C that are contributing to this overall high porosity and permeability are primarily primary (depositional) intergranular macropores with no significant volume of any other pore type being present. This is in contrast to Field A and B where the pores that are contributing to high porosities are from a number of primary and secondary (diagenetic) pore types and include intra- and inter-granular, vuggy and moldic macroporosity as well as microporosity. The dominant macro pore type is however different for the nine reservoirs assessed, with the stratigraphically
older reservoirs having a greater volume of intragranular porosity than the younger reservoirs (this is related to *Lithocodium/Bacinella* being more abundant in the older reservoirs), where intergranular porosity dominates. Secondary macro pores also contribute to the total porosity of the nine reservoirs with vuggy porosity being of a similar volume in most reservoirs; the distribution of moldic porosity is variable and is primarily controlled by the distribution of rudists.

The Lønøy (2006) porosity cut-off has been used for Field A and B to help determine which macropore type is primarily contributing to flow >1 mD. This suggests that the macropores primarily contributing to flow in the grain dominated fabrics of the reservoir are intergranular, moldic and vuggy pores, whereas flow in the muddier fabrics is typically through vuggy and intragranular pores. Therefore, due to multiple pore types contributing to permeability in Field A and B, there will be greater pore complexity which will lead to a lower permeability. This is in contrast to Field C where a single pore type dominates (intergranular porosity), which results in less pore complexity and results in higher observed permeabilities. This thesis has primarily aimed to understand the distribution of macroporosity and the effect that different macropores have on reservoir quality. Microporosity is known to be a significant contributing pore type to the total porosity, however numerous studies have focussed on this pore type (Cox et al., 2010; Deville de Periere et al., 2011; Lambert et al., 2006) and there was insufficient plug porosity and permeability data to quantify the distribution of microporosity in Field A. Furthermore the distribution and origin of microporosity in Field B has already been discussed by Cox (2010). These studies suggest that there are two phases of microporosity formation, with the first being depositional and the second thought to be due to subsequent dissolution in the burial zone (Cox, 2010; Deville de Periere et al., 2011; Lambert et al., 2006).

Although there is a sufficient volume of primary and secondary porosity in the reservoirs of all three fields for them to economic, pore occluding cements have greatly reduced their porosity. Field A and B have a diagenetic history that involves progressive cementation from the syn-depositional to deep burial environment; cementation initial commences with pyrite and microdolomite formation, followed
by fringing, equant and blocky calcite cement and saddle dolomite. A late stage
dissolution event then occurs which etches the surfaces of the calcite and dolomite
cement and also rounds micrite crystals (leading to an increase in micropore volume
(Lambert et al., 2006)). This dissolution event has significantly enhanced the
secondary macroporosity in the reservoir intervals of both Field A and B (primarily
through the generation of vuggy porosity), with the only cement known to form after
this dissolution event being authigenic microporous kaolinite.

In comparison, cementation in Field C commenced in the vadose zone and
ceased in the shallow burial zone; cementation commenced with meniscus cements
followed by fringing, syntaxial and drusy calcite cements. Like Field A and B,
dissolution has enhanced porosity after cementation ceased. However, this event is
relatively minor in comparison to that identified in Field A and B and intergranular
porosity remains the primary pore type contributing to reservoir quality.

Although multiple diagenetic phases are observed in all three fields, it has
been shown that the higher energy depositional fabrics are still associated with the
best reservoir quality in all three fields. Therefore although extensive cementation
and dissolution has occurred, the products of diagenesis in all three fields must
conform reasonably well to the depositional fabric, thus preserving higher porosities
and permeabilities in the higher energy depositional fabrics. Therefore, because
reservoir quality is in good agreement with depositional fabric, the petrophysical
properties of all three fields can be predicted by 3D modelling of the petrophysically
important/higher energy depositional fabrics.

However, although the location of potential reservoirs is related to the
location of higher energy depositional fabrics, cementation in all three fields has
greatly reduced porosity, in some cases completely occluding the porosity. Calcite
cement is the main pore occluding cement in all three fields and therefore
understanding the origin of the cement and the controls on its formation are
important to determine.
7.4 Origin of the Main Pore Occluding Cements

$\Delta^{13}C$ data were used to help identify the origin of the calcite cement present in the fields. This is because bulk micrite $\delta^{13}C$ data from Field A and B suggest the stratigraphically younger reservoirs have higher $\delta^{13}C$ values (reservoir i: $2.95 \pm 0.13 \%$ (n=20)) than the older reservoirs (reservoir viii: $2.03 \pm 0.02 \%$ (n=17)); this is typical for carbonate precipitated during the Hauterivian-Aptian (Föllmi, 2012). The average $\delta^{13}C$ and $\delta^{18}O_{VPDB}$ of bulk micrite for each reservoir can be directly related, along similar $\delta^{18}O_{VPDB}$ vs. $\delta^{13}C$ thermal alteration pathways, to the average $\delta^{13}C$ and $\delta^{18}O_{VPDB}$ of the main pore occluding cements (calcite and saddle dolomite) within the same reservoir. This implies that individual reservoirs behave as relatively confined aquifer systems, where the $\delta^{13}C$ composition of the cements is determined by the progressively increasing $\delta^{13}C$ of Early Cretaceous seawater during the Hauterivian-Aptian. Therefore the solutes required for cementation in individual reservoirs are most likely obtained through pressure solution of marine carbonate from within the reservoir in which the cements are now present.

Similar to Field A and B, the micrite and the main pore occluding cements within Field C are likely to have been precipitated from pore fluid that had a similar composition to Cretaceous seawater. Previously it was thought that the main pore occluding cements present in the Quissamã Formation were formed from meteoric pore fluid (Carozzi et al., 1983). Although some evidence does exist for subaerial exposure (e.g. the presence of meniscus cements) the $\delta^{13}C$ of the main pore occluding calcite cements and of micrite is between 2.6-3.4 % and 1.7-4.1 % respectively, which suggests that both micrite and the main pore occluding cements precipitated from pore fluid similar to Albian seawater; where $\delta^{13}C$ values of between 0.3 % to 3.7 % are typical (Price and Harwood, 2012; Veizer et al., 1999). This indicates that the micrite and the main pore occluding cements in all three fields are likely to have formed from pore fluid that had a similar $\delta^{13}C$ composition to that of Early Cretaceous seawater and was not significantly affected by meteoric pore water.
Although the δ\textsuperscript{13}C of bulk micrite for all three fields is thought to record the composition of Early Cretaceous seawater, the δ\textsuperscript{18}O\textsubscript{VPDB} of micrite is likely to have undergone rock-water interaction in all three fields. This is suggested in Field C by the δ\textsuperscript{18}O\textsubscript{VPDB} values becoming more positive at shallower depths from -2.24 ‰ at 150 m to -0.99 ‰ at 0 m in Well 2C, with the progressive increase in δ\textsuperscript{18}O\textsubscript{VPDB} at shallower depths being suitably described by a linear trend line (R\textsuperscript{2} = 0.8). Using the Kim and O'Neil (1997) equation to calculate the temperature of precipitation and assuming the same initial δ\textsuperscript{18}O\textsubscript{SMOW}, this trend line corresponds to a geothermal gradient of 33°C/km, which is similar to the initial geothermal gradient (30°C/km) for the basin (Mohriak et al., 1990); the geothermal gradient is then thought to decrease linearly to the current geothermal gradient (20°C/km). This therefore may imply that rock-water interaction ceased early in the basin’s history.

A similar decrease in δ\textsuperscript{18}O\textsubscript{VPDB} into the deepest reservoirs (highest temperature) or a similar δ\textsuperscript{18}O\textsubscript{VPDB} for the micrite in all reservoirs (similar burial history) was expected for Field A and B. However, the δ\textsuperscript{18}O\textsubscript{VPDB} of micrite for Field A and B are opposite to what was expected, with the more reduced bulk micrite δ\textsuperscript{18}O\textsubscript{VPDB} being found in the shallower reservoirs (i = -6.7 ± 0.17 ‰ (n=20)) than in the deeper, older, reservoirs (viii = -4.62 ± 0.26 ‰ (n=17)). This is thought to be a result of rock-water interaction continuing to higher temperatures in the shallower reservoirs of Field A and B. The δ\textsuperscript{18}O\textsubscript{VPDB} of micrite for Field A and B also suggest that rock-water interaction continued to a higher temperature than in Field C, where more positive δ\textsuperscript{18}O\textsubscript{VPDB} values are obtained.

Petrography reveals that all the pore occluding cements present within the Quissamã Formation of Field C are early, with no late stage burial cements identified (e.g. saddle dolomite or blocky calcite cement) and no oil inclusions observed; this is in contrast to Carozzi et al. (1983) where burial cements were identified. The δ\textsuperscript{18}O\textsubscript{VPDB} of the cements is also from -1.9 to -3.6 ‰ which would imply that the cements precipitated at a relatively shallow burial depth from pore fluid similar to, or slightly more evolved from, Albian seawater. This is in comparison to the cements in Field A and B, where deep burial cements, such as blocky and saddle dolomite cements are observed; this agrees with the findings of Neilson et al. (1998). The
\( \delta^{18}O_{VPDB} \) of these cements are also more reduced than those in Field C, with the \( \delta^{18}O_{VPDB} \) of blocky calcite cement being between -7.5 and -9.8 ‰, which may suggest that the cements formed at higher temperatures than those in Field C.

7.5 Dynamics of Cementation

In-situ \(^{m}\)Mg/\(^{m}\)Ca and \( \delta^{18}O_{VPDB} \) data also corroborate that the cements in all three fields precipitated from pore fluid with a similar composition to Cretaceous seawater and at progressively higher temperatures. Both temperature proxies become progressively lower from the oldest cements that are precipitated against the pore wall into the youngest cements in the pore centre; this is observed for all analysed samples in all three fields. This suggests that cementation took place in a relatively confined system and most likely indicates the cements precipitated at progressively higher temperatures with no noticeable addition of exotic fluid (with vastly different \( \delta^{18}O_{SMOW} \) and \(^{m}\)Mg/\(^{m}\)Ca compositions). However, as discussed previously the cements in Field A and B are likely to be derived from local marine carbonate and will therefore be buffered to some extent so that any temperatures derived from \( \delta^{18}O_{VPDB} \) data are only an approximation. Therefore, further studies need to be undertaken (such as fluid inclusion microthermometry) to help validate the \( \delta^{18}O_{VPDB} \) temperature proxy.

An extensively cemented grainstone from a non-reservoir in Field B was analysed to identify the approximate temperature at which cementation ceased. This is to help determine whether the non-reservoirs are likely early barriers to flow which will prevent significant mixing of fluid between reservoirs and allow the cements to contain a \( \delta^{13}C \) composition which suggests the cements are locally sourced (from the marine carbonate within the same reservoir in which the cements are now present; Section 7.4). The intergranular volume that is present for some grainstones in the non-reservoirs is high (~41 %). Using a typical compaction curve (Schmoker and Halley, 1982) this may suggest the early prevention of compaction, and hence the majority of cement is likely to have formed at depths < 200 m. This is also supported by high \(^{m}\)Mg/\(^{m}\)Ca ratios, with the lowest value being 0.5, and high \( \delta^{18}O_{VPDB} \) values of 0 ‰ to -1.0 ‰ for the pore occluding cements, which indicate cementation ceased at
low temperatures and from pore fluid similar to Early Cretaceous seawater. These heavily cemented grainstones combined with the shales, mudstones and wackestones of the TST have most likely resulted in reservoir compartmentalisation from a shallow burial depth and led to the formation of multiple, relatively confined reservoir systems in Field A and B.

There is a greater volume of calcite cement in the aquifer than in the coeval oil reservoir for Field A and B. This has led to the porosity and permeability of Field B being greater in the oil reservoir (10-50 %, 0.08-830 mD respectively; Cox et al. (2010)) than in its coeval aquifer (10-23 %, 0.1-4 mD respectively; Cox et al. (2010)); this has been previously observed by Neilson et al. (1998). In-situ $\delta^{18}O_{VPDB}$ data for calcite cements in coeval reservoirs of Field A suggest that cementation continued to a similar temperature in both oil reservoir i and its coeval aquifer i; -9.6 ‰ (∼111°C) and -9.9 ‰ (∼113°C) respectively, this is also supported by the independent $^{m}$Mg/$^{m}$Ca geothermometer and is in agreement with the observations of Cox (2010) for Field B. However, both the $^{m}$Mg/$^{m}$Ca ratio and $\delta^{18}O_{VPDB}$ data indicate that cementation has continued to a higher temperature in oil reservoir ii (-13.5 ‰; ∼145°C) than in the coeval aquifer ii (-9 ‰; ∼106°C) for Field A; this is in contrast to previous studies and suggests that cementation may continue for longer in the oil bearing reservoir. This would imply that a greater volume of cement formed in the aquifer over a similar (reservoir i) or smaller (reservoir ii) temperature range than in the oil reservoir. It can be observed that oil charge at -8.8 ‰ (∼104°C) reduced the volume of cement precipitated after oil charge, but did not stop cementation. This is in contrast to the aquifer where precipitation of the final pore occluding cement zone, after oil charge into the oil reservoir, greatly reduced porosity.

Although coeval reservoirs in Field A and B are likely to have a similar initial pore fluid chemistry and therefore can be directly compared, during the deposition of the nine cycles the pore fluid chemistry is likely to have changed. The first non-luminescent calcite cement zone in all the samples for Field A and B is thought to have formed in the marine phreatic environment. This cement zone was analysed via in-situ $\delta^{18}O_{VPDB}$ assessment and is in good agreement with Early Cretaceous $\delta^{18}O_{VPDB}$ secular curves (Pucéat et al., 2003). Therefore the depositional cycles are
thought to have formed during 2 glacial episodes, with the Kharaib Formation (cycles iii–ii) depositing during the interglacial episode.

The ranges in δ^{18}O_{VPDB} over which cementation occurred in the reservoirs of Field A and B have been calculated in order to remove the effect of changing Cretaceous seawater chemistry. This assessment suggests that the final pore occluding cement may have formed in the older reservoirs of the Lekhwair Formation prior to the younger reservoirs of the Kharaib and Shu’aiba’ Formations. This is in agreement with the $^{m}$Mg/$^{m}$Ca ratio, with there being fewer cement zones in the older reservoirs, and with the δ^{18}O_{VPDB} of bulk micrite being more positive in the older reservoirs; which suggests continued rock-water interaction at a higher temperature in the reservoirs of the Kharaib and Shu’aiba’ Formations (implying a sufficient remaining pore volume for rock-water interaction to continue at higher temperatures).

The cause for relatively early cementation of the older reservoirs may be related to there being a greater volume of intragranular porosity in the older cycles, whereas the pore space in the Kharaib and Shu’aiba’ Formations has a greater volume of intergranular porosity. The remaining intragranular pores in the Lekhwair Formation are typically not occluded by any form of cement which may suggest their isolation, therefore for the same volume of pore occluding cement the pores in the Lekhwair Formation will be occluded prior to those in the Kharaib and Shu’aiba’ Formation. Another possibility is that the acidic fluids that led to dissolution after the completion of calcite cementation charged the older reservoirs prior to the younger reservoirs.

Unlike for Field A and B, where cementation continued into the deep burial environment for the oil reservoirs and coeval aquifers, cementation in the oil reservoir of Field C is thought to cease in the shallow burial environment. In-situ δ^{18}O_{VPDB} for the first non-luminescent cement zone (-0.5 ± 0.1 ‰ (n=4)) in the oil reservoir of Field C suggest precipitation from Albian seawater (-2 ‰ to 0 ‰; Price and Harwood (2012), Veizer et al. (1999)). The δ^{18}O_{VPDB} then progressively decreases from the oldest cement zone into the youngest; this agrees with the $^{m}$Mg/$^{m}$Ca ratio. The range of δ^{18}O_{VPDB} obtained for calcite cement in the oil reservoir
of Field C is between -0.2 ‰ and -3.9 ‰, using the Kim and O’Neil equation and a δ¹⁸OSMOW of -1.2 ‰ for pre-glacial oceans (Carvalho et al., 1995) this implies that all the cements precipitated from water at a temperature of between 10°C and 28°C. The temperature estimate presented here is relatively low for subtropical surface water, however some studies have shown that the temperature of Albian seawater was between 10-14°C for some proto-Atlantic basins (Huber et al., 2011).

Based on petrography and geochemistry the cements present in Field C are all relatively early cements, with δ¹⁸OVPDB values suggesting that all the calcite cements in the oil reservoir had formed by ~ 28°C. This is also supported by the mMg/mCa ratios of the cements which suggest that the majority of cement precipitated at a lower temperature in the oil reservoir (~15°C; Figure 4.2) than in the aquifer (~ 60°C; Figure 4.2). This is in contrast to Field A and B, where cementation is thought to continue to the deep burial environment. Furthermore, by cementation in the oil reservoir of Field C ceasing in the shallow burial zone, the oil reservoir typically has a higher porosity and permeability than the aquifer; where cementation may have continued after oil charge. This can be observed in plug porosity and permeability data where there are higher porosities for the grainstones and packstones in the oil reservoir (grainstones: 22 ± 0.3 % (n = 173); packstones: 23 ± 0.5 % (n = 88)) than in the aquifer (grainstones: 16 ± 0.3 % (n = 143); packstones: 16 ± 0.5 % (n = 49)). This is also observed for permeability where the oil reservoir (grainstones: 1155 ± 123 mD (n = 173); packstones: 95 ± 59 mD (n = 88)) has a higher permeability than the aquifer (grainstones: 168 ± 70 mD (n = 139); packstones: 0.6 ± 0.1 mD (n = 49)).

Early oil charge in the oil reservoir of Field C is thought to have stopped cementation by ~ 28°C and prior to the formation of a significant volume of cement. However, this conclusion disagrees with some authors who have suggested the source rocks present in the basin matured in the Late Miocene (Guardado et al., 2000; Guardado et al., 1990; Nascimento et al., 1999), or in the Turonian-Santonian to the Miocene (Beglinger et al., 2012b; Mohriak et al., 1990) and therefore after the oil reservoir has been buried to temperatures > 28°C. Early oil charge is also not directly supported by the petrographic assessment presented in this thesis due to the lack of oil inclusions observed in the cements. Therefore, to determine whether early
oil charge may have stopped cementation in the oil reservoir of Field C, to constrain
the approximate time at which cementation ceased in Field A and Field C and to
identify the origin of the dissolution event in Field A, basin models were developed
for Field A and C.

7.6 Basin Modelling

The aforementioned diagenetic understanding was included into a basin
model for Field A by switching the porosity-depth curve describing the reservoir
during burial, in order to represent a greater loss of porosity at a given time due to
the formation of a large volume of pore occluding cement. The inclusion of this
diagenetic understanding has greatly improved the model by accurately predicting
the accumulation sites of hydrocarbon; this is in contrast to two other models where
atypical porosity-depth curve for carbonates was used (Schmoker and Halley, 1982)
or where the model uses standard compaction curves defined from the Petromod™
database.

Basin modelling has been used in this thesis to determine the time at which
source rocks began to expel hydrocarbon even though other, faster methods, have
been previously developed. These other methods include the Time-Temperature
Index (TTI) method presented by Waples (1980). This method is primarily based on
the Arrhenius equation (where chemical reaction rates double for every 10ºC
increase in temperature) and therefore this method predicts source rock maturity
based on the length of time a source rock is exposed to a certain temperature.
Although this method is suitable for one-dimensional models which are typically
developed for single well analyses, two dimensional basin models will more
accurately determine the time of source rock maturation and petroleum generation.
This is because these models typically predict heat flow and include an
understanding of transient and convection effects. Furthermore although the TTI
method calculates the approximate time at which the source rocks began to mature,
the time of primary migration and the migration pathways for this hydrocarbon are
not calculated. Therefore when trying to accurately model the petroleum system and
when trying to determine the time-temperature history and the time of oil charge into the reservoir two dimensional basin models are required.

The 2D models developed in this thesis predict that oil entered Field A between 55 and 45 Mya when the reservoirs were at a temperature of 85-100°C. This is similar to the temperature of oil emplacement as derived from the $\delta^{18}O_{VPDB}$ geothermometer (-8.8 ‰; ~104°C) and would suggest that the temperatures obtained from the $\delta^{18}O_{VPDB}$ temperature proxy for the cements in Field A are an accurate representation for the temperature at which cementation occurred.

The temperature evolution profiles extracted from the models for the reservoirs have then been used to constrain the time at which cementation ceased in all reservoirs. With the exclusion of oil reservoir ii, cementation is thought to continue for longest in reservoir i (~ 113°C) and therefore cementation, for the samples assessed in Field A, is thought to have ceased in all reservoirs by this temperature; this maximum temperature of cementation is in good agreement with primary fluid inclusion data (Neilson et al., 1998). The time-temperature history obtained from the model has then been used to constrain the time at which cementation ceased in Field A, with cementation thought to cease at ~ 50 Mya.

After calcite cementation ceased in the reservoirs, dissolution then etched the surfaces of the previously deposited calcite and dolomite cement and led to the formation of secondary porosity (Section 7.3). The model predicts that the dense intervals of each cycle will begin to expel oil by ~ 45 Mya. Therefore the acidic fluids causing the formation of secondary porosity may be derived from the organic acids produced by local source rock maturation. The aluminium required for the precipitation of kaolinite after dissolution can also be derived from the degradation of organic matter (Maliva et al., 1999).

To preserve the pores generated by the dissolution event to present day no further cementation or significant mechanical compaction can occur (with the possible exclusion of oil reservoir ii in Field A). The cause for this preservation may be related to the dissolution event being followed by hydrocarbon migration which may have stopped cementation. Furthermore the generation of hydrocarbon from
kerogen leads to an net increase in volume, with some authors suggesting an increase by < 25 % (Meissner, 1981); however, later studies show that this may be an underestimation (Barker, 1996; Luo and Vasseur, 1996; Spencer, 1987; Sweeney et al., 1987). This increase in pressure may prevent further chemical compaction and therefore stop cementation. Within the Cretaceous carbonate reservoirs of the Middle East significant overpressures of >2500 psi have been observed (Ehrenberg et al., 2008), which support this interpretation.

In contrast to the relatively late oil charge in Field A, which allowed cementation to continue but at a much reduced rate, oil charge is thought to enter Field C relatively early and is thought to have stopped cementation. The basin models for Field C suggest that the temperature of the reservoir increases from 22°C to 42°C between 110 Mya to 100 Mya. Therefore in order for oil to have stopped cementation by ~ 28°C, oil must enter the field at this time. Two models were developed to determine whether early oil charge may have stopped cementation in the oil reservoir of Field C. The field scale model suggests that the source rocks in the vicinity of Field C do not reach sufficient maturity for oil to charge the field at this time and therefore the maturation of local source rocks in the vicinity of Field C is unlikely to have stopped cementation. However, the basin scale model does suggest that the source rocks present in basement grabens matured by ~125 Mya, this disagrees with some authors that suggest oil charge in the Late Miocene (Guardado et al., 2000; Guardado et al., 1990; Nascimento et al., 1999) or in the Turonian-Santonian to the Miocene (Beglinger et al., 2012b; Mohriak et al., 1990).

The basin scale model predicts that oil began to accumulate beneath the evaporite of the Retiro Formation, which separates the source rock from the Quissamã Formation, by 105-103 Mya. Halokinesis of this evaporite is thought to have commenced in the Albian (Eichenseer et al., 1999) between 100 Mya and 80 Mya (Cobbold et al., 2001) or in the Late Cretaceous (Carozzi et al., 1983; Soldán et al., 1995) and allowed for the migration of hydrocarbon through salt windows. Salt windows were included in the model by 100 Mya which allowed for hydrocarbon accumulation at a temperature of ~ 42°C, however if salt windows were open by 105 Mya oil may have entered the reservoir at ~ 28°C. This conclusion suggests that
early oil charge may have stopped cementation in the oil reservoir of Field C in the shallow burial zone, however this would require salt windows, in the vicinity of Field C not to open in the Late Cretaceous (Carozzi et al., 1983; Soldán et al., 1995) but instead by the middle Cretaceous (~ 105 Mya).

In the absence of primary oil inclusions within the pore occluding cements, basin modelling reveals that initial oil charge in Field C is coincident with the temperature at which cementation ceased. Therefore, it is likely that early oil charge has preserved the high depositional porosities and permeabilities of the field. Whereas in Field A and B later oil charge, as revealed through the identification of primary oil inclusions and through the predictions made by basin models, has allowed for a more significant volume of cement to form which has greatly reduced the porosities and permeabilities of the fields. Additionally, continued cementation in the presence of hydrocarbon (as evidenced by primary oil inclusions) in Field A and B continued to reduce porosity (although at a much reduced rate) This is in contrast to Field C where oil charge stopped cementation and preserved porosity. Also, due to Field C being buried to 900 m less than Field A and B it is likely that Field C has undergone less chemical compaction which will result in the formation of less pore occluding cement and preserve porosity. Therefore, due to a combination of there initially being a lower primary porosity in Field A and B, that the fields were buried to a greater depth and due to later oil charge and continued cementation in the presence of hydrocarbon, the porosity and permeability of Field A and B is much reduced. This is in comparison to Field C which is dominated by higher energy depositional fabrics and early oil charge - which stopped cementation, and burial to shallower depths preserved a higher volume of un-occluded primary porosity.

The generation of a large volume of secondary pores has significantly contributed to the porosity and permeability of Field A and B, unlike in Field C where there is a relatively minor volume of secondary pores. Basin modelling reveals that the acidic fluids that led to this dissolution event in Field A and B are likely to be derived from source rock maturation. Without this late stage dissolution event in Field A and B it is likely that there may not have been sufficient reservoir quality present in the fields to make them economic.
7.7 Conclusions

Burial diagenesis through extensive calcite cementation and dissolution has significantly altered the primary porosity and permeability of all three fields. Geochemical analyses suggest that the main pore occluding calcite cements are derived from pore water that had a similar composition, or slightly evolved from, Early Cretaceous seawater. Additionally, in successive reservoirs for Field B the main pore occluding cements record the progressive increase in $\delta^{13}C$ typical of Hauterivian-Aptian seawater. This suggests that the cements in each reservoir are sourced locally, via pressure solution, from local marine carbonate.

The cause for the local source of cement in Field A and B is most likely because the TST for each cycle has a very low porosity and permeability, this coupled with the presence of numerous hardgrounds has led to field compartmentalisation. Additionally the current location of reservoir tops has been altered by the early cementation of the intergranular porosity, where in some grainstones the intergranular pore space was fully cemented at depths < 200 m, this has therefore aided reservoir compartmentalisation. This is in contrast to Field C where no significant barriers to flow are present.

In-situ $\delta^{18}O_{VPDB}$ data for the oldest non-luminescent cement zones in Field A and B suggest the cement zone precipitated from seawater and records two glacial episodes and an interglacial episode in the Early Cretaceous; the Kharaib formation (cycles iii-ii) was deposited during the interglacial episode. Similar to Field A and B the first non-luminescent cement zone of Field C is thought to have formed from Early Cretaceous seawater. The $\delta^{18}O_{VPDB}$ values in all three fields then become progressively more negative into the younger cement zones, suggesting continued precipitation at progressively higher temperatures.

The $\delta^{18}O_{VPDB}$ of the first cement zone in Field A and B would suggest that the $\delta^{18}O_{SMOW}$ or temperature during the formation of the first cement zone was not constant. Therefore to remove the effect of changing pore fluid chemistry the range in in-situ $\delta^{18}O_{VPDB}$ values obtained for the reservoirs were calculated. The range in $\delta^{18}O_{VPDB}$ for the Lekhwair formation (cycles ix-iv) is smaller than for the shallower
Kharaib and Shu’aiba’ Formations; this would suggest that cementation ceased in the deeper reservoirs first. This is also supported by the $\delta^{18}$O_VPDB of bulk micrite which suggests continued rock-water interaction to higher temperatures in the shallowest reservoirs, by in-situ $^{\text{m}}$Mg/$^{\text{m}}$Ca data and by there being typically fewer cement zones in the older reservoirs.

$\Delta^{18}$O_VPDB and $^{\text{m}}$Mg/$^{\text{m}}$Ca data have also been used to help understand the effect that oil charge has on the dynamics of cementation. Oil charge in Field A is thought to allow cementation to continue in oil reservoir i to a similar temperature (~ 111 °C; -9.6 ‰) as in the coeval aquifer (~ 113 °C; -9.9 ‰) but at a much reduced rate; with the final cement zone in the oil reservoir being of a much lower volume than the final pore occluding cement zone in the aquifer; this agrees with the findings of Cox et al. (2010) for Field B. However in reservoir ii it is suggested that cementation continues to a higher temperature in the oil reservoir (~ 145°C; -13.5 ‰) than in the coeval aquifer (~ 106°C; -9 ‰). The time and temperature at which hydrocarbon is thought to have entered the field has been suggested by basin modelling to be between ~ 55 Mya and 45 Mya at a temperature of between 85-100°C, this agrees with the estimated temperature of oil emplacement as suggested by the $\delta^{18}$O_VPDB temperature proxy (~104 °C; -8.8 ‰).

The time at which hydrocarbon entered the field is much later than that for Field C where early oil charge at 110 Mya to 100 Mya, when the reservoir was at a temperature of between 22 °C and 42 °C, is thought to have stopped cementation and preserved porosity in the shallow burial environment, where the most negative $\delta^{18}$O_VPDB for calcite cement is -3.9 ‰ (~28 °C). This suggests cementation ceased prior to the current reservoir temperatures (70 °C). Therefore, because of earlier oil charge in Field C, a more complex diagenetic history and a greater volume of cement is present in Field A and B.

Although cementation has greatly reduced the porosity in all three fields a late dissolution event in Field A and B has significantly enhanced porosity. Basin modelling suggests that this dissolution event is due to the effect of maturation of the kerogen within the intraformational seals at ~ 45 Mya. The maturation of organic matter will result in the generation of organic acids that can greatly enhance porosity.
and will provide the aluminium that is required for the formation of kaolinite in this newly generated pore space (Maliva et al., 1999). This suggests that the proximity of the reservoir to a potential source rock can significantly impact the current porosity expected within that reservoir and may lead to greater porosities than that predicted at depth. A late stage dissolution event is also identified in Field C, however the source rocks are not as proximal to the reservoir as for Field A and B, therefore the field has not experienced such a large scale dissolution event.

Although diagenesis has significantly affected the porosity and permeability of all three fields and especially in the aquifers of the fields, the location of potential reservoirs for all three fields can still be predicted by the 3D modelling of petrophysically important depositional fabrics (typically grain dominated fabrics). This is because the products of diagenesis in all three fields conform reasonably well to the depositional fabric, thus preserving higher porosities and permeabilities in the higher energy depositional fabrics.

### 7.8 Future Work

1) The $\delta^{18}O_{VPDB}$ of the first non-luminescent cement zone is in good agreement with marine secular curves. This would suggest that the in-situ assessment of this cement zone can be used to help determine the secular variations in marine carbonate $\delta^{18}O_{VPDB}$. It may be possible to apply this technique in different chronostratigraphic units to aid in the establishment of marine secular $\delta^{18}O_{VPDB}$ curves.

2) The more negative $\delta^{18}O_{VPDB}$ values obtained for micrite in Field A and B are related to samples where the in-situ $\delta^{18}O_{VPDB}$ values for the final pore occluding cement zone are also more negative. This may suggest that the $\delta^{18}O_{VPDB}$ of micrite can be used to determine the approximate time at which cementation ceased in relation to other samples. Therefore, it is suggested here that further in-situ $\delta^{18}O_{VPDB}$ analyses are undertaken, which are in turn related to bulk micrite $\delta^{18}O_{VPDB}$ values to help confirm if this is the case.
3) The relatively few δ\textsuperscript{13}C bulk micrite data points obtained from the Wadi Rahabah outcrop are in good agreement with the secular marine δ\textsuperscript{13}C curve. The good exposure at this outcrop provides an ideal opportunity to obtain a high resolution δ\textsuperscript{13}C curve for the Kharaib and Shu’aiba’ Formations. This curve will be of great use to help understand the origin of the negative δ\textsuperscript{13}C excursions identified in the core sections and may be of use for chronostratigraphic and correlation purposes.

4) This thesis has only obtained in-situ δ\textsuperscript{18}O\textsubscript{VPDB} analyses on the cements in the oil reservoir of Field C. Currently the m\textsuperscript{Mg}/m\textsuperscript{Ca} ratios for the cements in the aquifer suggest that the cements continued to precipitate to a similar temperature as the cements in the oil reservoir, this needs to be confirmed by further in-situ δ\textsuperscript{18}O\textsubscript{VPDB} analyses.

5) The m\textsuperscript{Mg}/m\textsuperscript{Ca} ratios for calcite cement in Field A, B and C suggests its promising use as a temperature proxy. The electron probe microanalysis of calcite cement is faster and more economic than in-situ δ\textsuperscript{18}O\textsubscript{VPDB} assessments. Therefore, to help further constrain the dynamics of cementation, further m\textsuperscript{Mg}/m\textsuperscript{Ca} analyses should be undertaken. This assessment should relate the m\textsuperscript{Mg}/m\textsuperscript{Ca} ratios to individual cathodoluminescent zones.

6) The temperature at which cementation occurred can be validated through the use of fluid inclusion microthermometry; this assessment was beyond the scope of this thesis. Future studies should compare the temperature derived from the δ\textsuperscript{18}O\textsubscript{VPDB} and m\textsuperscript{Mg}/m\textsuperscript{Ca} temperature proxies to the homogenisation temperatures for fluid inclusions to help further constrain their use as temperature proxies. Furthermore, to constrain the δ\textsuperscript{18}O\textsubscript{VPDB} and m\textsuperscript{Mg}/m\textsuperscript{Ca} temperature proxies the present day porewater δ\textsuperscript{18}O\textsubscript{SMOW} and m\textsuperscript{Mg}/m\textsuperscript{Ca} ratio should be obtained.

7) The cause for oil charge stopping cementation in Field C and allowing for cementation to continue in Field A and B is most likely a result of the different wettability states of the reservoirs; where Field C is likely oil wet
and Field A and B are mixed wet to water wet. This interpretation needs to be confirmed through petrophysical assessments in the laboratory.

8) Assuming that cementation stops in Field C due to the presence of hydrocarbon. 3D modelling of the porosity variations in Field C will allow the direction of fill to be obtained because higher porosities will be preserved in the parts of the field that were charged first.

9) Petromod™ only currently allows for the switching of a porosity-depth curve once during burial. Petromod™ needs to be improved so that the porosity-depth curve can be switched numerous times to allow for the incorporation of multiple episodes of dissolution and cementation. Furthermore, the method of incorporating diagenesis by switching the porosity-loss curves is not appropriate for cementation of the pore space, because cementation preserves the total pore volume. Therefore a new method needs to be incorporated into Petromod™ which allows for the preservation of pore volume so that the effect of cementation can be more accurately modelled.

10) Undertake source rock maturation studies on immature dense intervals of the Thamama Group. This will be used to determine the amount of organic acids that are produced during source rock maturation, to confirm whether the maturation of the source rocks can produce sufficient organic acids to cause extensive dissolution.

11) Extensive dissolution is thought to be due to the maturation of source rocks in the vicinity of Field A and B and is thought to occur at ~45 Mya. This pore space has then been preserved to present day with no pore occluding cements forming. This may be because the maturation of kerogen to hydrocarbon typically generates pressure, therefore pressure data should be sought to determine whether the reservoirs are overpressured.

12) Pore Architecture Models (PAMs) were generated and the Pore Analysis Tool (PAT) software method was then run on these models to help determine if porosity and permeability can be predicted from thin section samples (Wu et al., 2006). This assessment was undertaken on 25 samples from Field A
However, there was a significant difference between the predicted porosity and permeability and the actual, plug derived, porosity and permeability (Van der Land et al., 2012). This is thought to be a result of microporosity not being included in the model. Currently the PAMs are being improved to include microporosity and therefore further analyses should be undertaken when the work is complete; to see if there is better correlation between the predicted and actual porosity and permeability. Once calibrated, the method can then be used to predict changes to porosity and permeability during successive diagenetic steps (Van der Land et al., 2013).

13) Finally undertake further analyses on different fields in the Rub Al Khali Basin and Campos Basin to determine whether the controls on reservoir quality, identified by this thesis, are regional.
CHAPTER 8

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APPENDICES

Appendix 1: Python code for calculating porosity

The code was written in Python™ and requires the following libraries: os, sys, ImageOps, ImageFilter, matplotlib, itemgetter. This code quantifies the total area of an image that contains blue epoxy resin and is included on a CD at the back of this thesis.
Appendix 2: Field A and B stable isotope data

All the δ\textsuperscript{13}C and δ\textsuperscript{18}O\textsubscript{VPDB} data for Field A and B are presented here. The data from Field B are primarily from Cox (2010) however, new bulk micrite δ\textsuperscript{13}C and δ\textsuperscript{18}O\textsubscript{VPDB} data were collected for Field A and B using a micro-drill.

Field A

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Table 2.1: Well 1A bulk micrite data.
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Table 2.2: Well 2A bulk micrite data.

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Table 2.3: Well 3A bulk micrite data.


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Table 2.4: Well 1B bulk micrite data. New $\delta^{13}$C and $\delta^{18}$OVPDB data are marked by the grey rows, all other data are from Cox (2010).
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Table 2.7: Well 4B bulk micrite data. New $\delta^{13}C$ and $\delta^{18}O_{VPDB}$ data are marked by the grey rows, all other data are from Cox (2010).

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Table 2.8: Well 5B all data are from Cox (2010).
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Table 2.9: $\delta^{13}$C and $\delta^{18}$O$_{VPDB}$ for blocky calcite cement, all data are from Cox (2010).

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Table 2.10: $\delta^{13}$C and $\delta^{18}$O$_{VPDB}$ for calcite cement in fractures, all data are from Cox (2010).

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Table 2.11: $\delta^{13}$C and $\delta^{18}$O$_{VPDB}$ for micro-dolomite, all data are from Cox (2010).

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Table 2.12: $\delta^{13}$C and $\delta^{18}$O$_{VPDB}$ for saddle-dolomite, all data are from Cox (2010).
Appendix 3: Lithofacies

The lithofacies identified in thin section samples from Field A and B are presented in the following section.

<table>
<thead>
<tr>
<th></th>
<th>Lithofacies</th>
<th>Texture</th>
<th>Common Features</th>
<th>Reservoir Characteristics</th>
<th>Depositional Environment</th>
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</thead>
<tbody>
<tr>
<td>1</td>
<td>Orbitolinid, wackstone</td>
<td>- W</td>
<td>- abundant discoidal orbitolinids and other forams, broken rudist fragments, echinoderms, peloids, mud interlaying, moderately sorted, low biodiversity</td>
<td>- low to moderate porosity, non-reservoir</td>
<td>- HST + TST, shallow subtidal, low to moderate energy, inner ramp restricted lagoon</td>
</tr>
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<tr>
<td>2</td>
<td>Rudist, peloid packstone</td>
<td>- MDP, GDP less common W</td>
<td>- orbitolinids, well preserved rudists, echinoderms, lithocodium/bacinella, miliolids, green algae and other forams, peloids, composite grains, moderately sorted, mud interlayers</td>
<td>- moderate to good porosity, reservoir</td>
<td>- HST, shallow subtidal, high energy, above fair-weather wave base, upper ramp, near shoal crest</td>
</tr>
</tbody>
</table>
### Appendix 3

#### 3. Foraminiferal, skeletal, packstone

- **Texture**
  - MDP, GDP less common W

- **Common Features**
  - miliolids, orbitolinids, rudists, lithocodium/bacinella, gastropods, and other benthic forams such as praechrysalidina infracretacea (x)
  - peloids
  - moderately sorted
  - high biodiversity

- **Reservoir Characteristics**
  - low to excellent porosity
  - reservoir and non-reservoir

- **Depositional Environment**
  - HST + TST
  - shallow subtidal, high energy, above fair-weather wave base
  - upper ramp, near shoal crest

---

#### 4. Stromatoporoid Grainstone

- **Texture**
  - G

- **Common Features**
  - miliolids, orbitolinids, rudists, lithocodium/bacinella, echinoderms, stromatoporoids
  - peloids, composite grains
  - moderately sorted
  - high biodiversity

- **Reservoir Characteristics**
  - good to excellent porosity
  - reservoir

- **Depositional Environment**
  - HST
  - Shelf lagoon
  - Upper ramp, near shoal crest
Dolomitised wackestone

5

Texture
- W

Common Features
- rare orbitolinids, echinoderms, sponge spicules, planktonic foraminifera
- peloids
- well sorted
- low biodiversity
- grading microdolomite

Reservoir Characteristics
- good to excellent porosity
- reservoir

Depositional Environment
- HST
- lagoon

Bioclastic wackestone

6

Texture
- W less common MSP

Common Features
- green algae (dasycladacean), broken rudists, echinoderms, sponge spicules
- high biodiversity of broken bioclasts
- peloids
- moderately sorted
- mud interlayers
- rare micro-dolomite rhombs

Reservoir Characteristics
- poor to moderate porosity
- reservoir and non-reservoir

Depositional Environment
- HST + TST
- subtidal, low energy, below the fair-weather-wave base
- middle ramp
<table>
<thead>
<tr>
<th>7</th>
<th>Orbitolinid skeletal wackestone</th>
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<tbody>
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<td></td>
<td>Texture</td>
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<tr>
<td></td>
<td>- W</td>
</tr>
<tr>
<td></td>
<td>Common Features</td>
</tr>
<tr>
<td></td>
<td>- discoidal orbitolinids, echinoderms, sponge spicules, ostracods, broken rudist fragments</td>
</tr>
<tr>
<td></td>
<td>- low biodiversity and high abundance of fauna</td>
</tr>
<tr>
<td></td>
<td>- moderately sorted</td>
</tr>
<tr>
<td></td>
<td>- micro-dolomite rhombs throughout</td>
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<tr>
<td></td>
<td>Reservoir Characteristics</td>
</tr>
<tr>
<td></td>
<td>- poor porosity</td>
</tr>
<tr>
<td></td>
<td>- non-reservoir</td>
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<tr>
<td></td>
<td>Depositional Environment</td>
</tr>
<tr>
<td></td>
<td>- TST</td>
</tr>
<tr>
<td></td>
<td>shallow subtidal, low energy, upper ramp</td>
</tr>
<tr>
<td></td>
<td>- inner ramp, restricted lagoon</td>
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</tbody>
</table>

<table>
<thead>
<tr>
<th>8</th>
<th>Coated grain packstone</th>
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<td></td>
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<td></td>
<td>- P</td>
</tr>
<tr>
<td></td>
<td>Common Features</td>
</tr>
<tr>
<td></td>
<td>- echinoderms, broken rudist fragments, planktonic forams</td>
</tr>
<tr>
<td></td>
<td>- coated grains: superficial ooids, oncoids, aggregate grains, intraclasts and peloids</td>
</tr>
<tr>
<td></td>
<td>- low diversity and high abundance of fauna</td>
</tr>
<tr>
<td></td>
<td>- moderate to well sorted</td>
</tr>
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<td></td>
<td>Reservoir Characteristics</td>
</tr>
<tr>
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</tr>
<tr>
<td></td>
<td>- TST</td>
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<td>- high energy open platform above the fair-weather wave base</td>
</tr>
<tr>
<td></td>
<td>- upper ramp, near shoal crest</td>
</tr>
</tbody>
</table>
9. Thallassinoides burrowed dolomitised wackestone

**Texture**
- W

**Common Features**
- planktonic forams, gastropods, rudist fragments
- peloids
- dolomitised burrows, the burrows are surrounded by a halo of pyrite
- low biodiversity
- moderately to well sorted

**Reservoir Characteristics**
- poor to moderate porosity
- non-reservoir

**Depositional Environment**
- HST
- low energy, lagoon below the fair-weather-wave base
- middle to upper ramp

10. Miliolid, skeletal wackestone

**Texture**
- W less common M

**Common Features**
- miliolid and other benthic foraminifera, echinoderms, broken fragments of rudists
- partly dolomitised matrix
- peloids
- moderately sorted

**Reservoir Characteristics**
- poor to moderate
- reservoir and non-reservoir

**Depositional Environment**
- TST
- low energy, lagoon below the fair-weather-wave base
- middle to lower ramp
11 Foraminiferal, algal shell hash

Texture
- MDP, W

Common Features
- echinoderm fragments, sponge spicules, rudist fragments, stromatoporoids, *P. lenticularis*, miliolids, choffatella decipiens and rotalids
- partly dolomitised matrix
- grains commonly contain pyrite
- peloids, intraclasts, oncoids,
- rare well rounded quartzite clast
- moderately sorted

Reservoir Characteristics
- poor porosity
- non-reservoir

Depositional Environment
- TST
- upper to lower ramp

12 Cladocoropsis, lithocodium/bacinella packstone

Texture
- GDP less common MDP and W

Common Features
- lithocodium/bacinella, cladocoropsis, rudist, gastropod, echinoderms, *praechrysalidina* infracretacea, *bigenerina* sp and other benthic foraminifera
- peloids
- moderately sorted

Reservoir Characteristics
- poor to good porosity
- reservoir

Depositional Environment
- HST
- shoal, upper ramp
- inner shoal
13 Lithocodium/bacinella packstone

Texture
- GDP, MDP

Common Features
- lithocodium/bacinella typically observed encrusting rudist fragments, praechrysalidina infracretacea, echinoderms, salpingoporella dinarica, ostracods
- poorly to moderately sorted

Reservoir Characteristics
- moderate to excellent porosity
- reservoir

Depositional Environment
- HST
- shoal to middle ramp
- shoal to subtidal lagoon

14 Thallassinoides burrowed, skeletal packstone

Texture
- MDP less common W

Common Features
- rudist fragments, choffatella deciphens, echinoderms, sponge spicules, gastropods, nautiloculina brönnimanni
- pyritised grains
- micro-dolomitised matric
- moderately sorted

Reservoir Characteristics
- poor porosity
- non-reservoir

Depositional Environment
- HST
- subtidal, low-energy ramp below the fair-weather wave base
- shallow subtidal, low energy lagoon
<table>
<thead>
<tr>
<th>Appendix 3</th>
<th>Thorpe, 2014</th>
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</thead>
<tbody>
<tr>
<td><strong>15</strong></td>
<td>Coated grain, skeletal packstone</td>
</tr>
<tr>
<td></td>
<td><strong>Texture</strong></td>
</tr>
<tr>
<td></td>
<td>- MSP</td>
</tr>
<tr>
<td></td>
<td><strong>Common Features</strong></td>
</tr>
<tr>
<td></td>
<td>- echinoderms, ostracods, rudist fragments, miliolids, textularia sp, sponge spicule, gastropods</td>
</tr>
<tr>
<td></td>
<td>- coated grains, intraclasts, peloids</td>
</tr>
<tr>
<td></td>
<td>- moderately sorted</td>
</tr>
<tr>
<td></td>
<td><strong>Reservoir Characteristics</strong></td>
</tr>
<tr>
<td></td>
<td>- good porosity</td>
</tr>
<tr>
<td></td>
<td>- reservoir</td>
</tr>
<tr>
<td></td>
<td><strong>Depositional Environment</strong></td>
</tr>
<tr>
<td></td>
<td>- HST</td>
</tr>
<tr>
<td></td>
<td>- shallow subtidal to intertidal, high energy upper ramp above fair weather-wave base</td>
</tr>
<tr>
<td></td>
<td>- near shoal crest, and near inner shoal</td>
</tr>
<tr>
<td><strong>16</strong></td>
<td>Stromatoporoid, algal, peloidal packstone</td>
</tr>
<tr>
<td></td>
<td><strong>Texture</strong></td>
</tr>
<tr>
<td></td>
<td>- GDP</td>
</tr>
<tr>
<td></td>
<td><strong>Common Features</strong></td>
</tr>
<tr>
<td></td>
<td>- stromatoporoid, salpingoporella dinarica, sponge spicules, echinoderms, miliolids, microsolenid coral</td>
</tr>
<tr>
<td></td>
<td>- micrite grains, peloids, coated grains</td>
</tr>
<tr>
<td></td>
<td>- moderately sorted</td>
</tr>
<tr>
<td></td>
<td><strong>Reservoir Characteristics</strong></td>
</tr>
<tr>
<td></td>
<td>- poor porosity</td>
</tr>
<tr>
<td></td>
<td>- non-reservoir</td>
</tr>
<tr>
<td></td>
<td><strong>Depositional Environment</strong></td>
</tr>
<tr>
<td></td>
<td>- HST</td>
</tr>
<tr>
<td></td>
<td>- shallow intertidal, high energy upper ramp above fair weather-wave base</td>
</tr>
<tr>
<td></td>
<td>- upper ramp near shoal crest</td>
</tr>
</tbody>
</table>
17 Coated grain, peloidal, rudist packstone

**Texture**
- GDP less common MDP

**Common Features**
- sponge spicules, rudists, rotalids, miliolids, peloids, echinoderms, bigenerina sp, gastropods
- rounded micrite grains, peloids, coated grains
- moderately sorted

**Reservoir Characteristics**
- moderate to excellent porosity
- reservoir

**Depositional Environment**
- HST
- shallow intertidal, high energy upper ramp above fair weather-wave base
- upper ramp near shoal crest

18 Peloidal, packstone

**Texture**
- G, GDP, MDP

**Common Features**
- miliolids, peloids, rudists, gastropods, sponge spicules, discoidal orbitolinids
- peloids, rounded micrite grains, coated grains
- well sorted

**Reservoir Characteristics**
- moderate to excellent porosity
- reservoir and non-reservoir

**Depositional Environment**
- HST
- shallow water, low energy, below fair-weather-wave base
- lagoon or upper ramp
Appendix 3

19. Dolomitised, orbitolinid, rudist wackestone

- Texture
  - W less common MDP

- Common Features
  - sponge spicules, rudists, discoidal orbitolinids, salpingoporella dinarica (green algae)
  - dolomitised matrix
  - peloids
  - poorly to moderately sorted

- Reservoir Characteristics
  - moderate to good porosity
  - non-reservoir

- Depositional Environment
  - HST
  - subtidal, low-energy, near fair-weather-wave base
  - upper to middle ramp or lagoon

20. Mudstone

- Texture
  - M

- Common Features
  - undifferentiated foraminifera
  - well sorted

- Reservoir Characteristics
  - poor porosity
  - non-reservoir

- Depositional Environment
  - TST
  - subtidal, low-energy, below fair weather wave base
  - middle to lowe ramp
<table>
<thead>
<tr>
<th>No.</th>
<th>Sample Type</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>21</td>
<td>S. dinarica packstone</td>
<td><strong>Texture</strong>&lt;br&gt;- MDP less common W&lt;br&gt;<strong>Common Features</strong>&lt;br&gt;- salpingoporella dinarica, gastropods, rudists, lithocodium/bacinella, praechrysallidina infracretacea, echinoderms, merlingina cretacea&lt;br&gt;- angular micrite grains, peloids&lt;br&gt;- mud layers&lt;br&gt;- moderately sorted&lt;br&gt;<strong>Reservoir Characteristics</strong>&lt;br&gt;- moderate porosity&lt;br&gt;- reservoir&lt;br&gt;<strong>Depositional Environment</strong>&lt;br&gt;- HST&lt;br&gt;- subtidal, low energy, near fair-weather-wave base&lt;br&gt;- upper to middle ramp</td>
</tr>
<tr>
<td>22</td>
<td>Coated grain grainstone</td>
<td><strong>Texture</strong>&lt;br&gt;- G&lt;br&gt;<strong>Common Features</strong>&lt;br&gt;- miliolids, echinoderms&lt;br&gt;- coated grains, composite grains, oncoids, intraclasts&lt;br&gt;- well sorted&lt;br&gt;<strong>Reservoir Characteristics</strong>&lt;br&gt;- excellent porosity&lt;br&gt;- reservoir&lt;br&gt;<strong>Depositional Environment</strong>&lt;br&gt;- HST&lt;br&gt;- shallow subtidal, high energy above fair weather-wave base&lt;br&gt;- upper ramp, near shoal crest</td>
</tr>
</tbody>
</table>
Appendix 4: Point count data

The point count data discussed in Chapter 3 are included on a CD at the back of the thesis.
Appendix 5: Elemental (EPM) data

The elemental data discussed in Chapters 4 and 6 for Field A, B and C are included on a CD at the back of the thesis.
Appendix 6: Inputs for the Field A basin model

The Field A basin model uses the following inputs for each horizon/layer in the model. The first column is the event that is been modelled (deposition or erosion), with the second column showing the time over which erosion or deposition occurred. The third column shows the lithologies used for the layer; each layer uses a mixture of lithologies.

<table>
<thead>
<tr>
<th>Horizon/Layer</th>
<th>Time of deposition (Mya)</th>
<th>Lithology</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>from</td>
<td>to</td>
</tr>
<tr>
<td>Upper DibDibba</td>
<td>0.5</td>
<td>0</td>
</tr>
<tr>
<td>Lower DibDibba</td>
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<td>0.5</td>
</tr>
<tr>
<td>Erosion</td>
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<td>10</td>
<td>5</td>
</tr>
<tr>
<td>Middle Fars (2)</td>
<td>11</td>
<td>10</td>
</tr>
<tr>
<td>Middle Fars (1)</td>
<td>12</td>
<td>11</td>
</tr>
<tr>
<td>Lower Fars</td>
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<td>14</td>
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<tr>
<td>Lower Asmari</td>
<td>21</td>
<td>15</td>
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<tr>
<td>Erosion</td>
<td>25</td>
<td>21</td>
</tr>
<tr>
<td>Damman</td>
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<td>21</td>
</tr>
<tr>
<td>Rus</td>
<td>45</td>
<td>35</td>
</tr>
<tr>
<td>Upper Umm Er Radhuma</td>
<td>50</td>
<td>45</td>
</tr>
<tr>
<td>Middle (4) Umm Er Radhuma</td>
<td>51</td>
<td>50</td>
</tr>
<tr>
<td>Horizon/Layer</td>
<td>Time of deposition (Mya)</td>
<td>Lithology</td>
</tr>
<tr>
<td>----------------------------</td>
<td>--------------------------</td>
<td>----------------------------------------------------------------------------</td>
</tr>
<tr>
<td>Middle (3) Umm Er Radhuma</td>
<td>52 – 51</td>
<td>35 % Shale (typical), 24 % Dolomite (typical), 41 % Limestone (micrite)</td>
</tr>
<tr>
<td>Middle (2) Umm Er Radhuma</td>
<td>53 – 52</td>
<td>33 % Shale (typical), 22 % Dolomite (typical), 45 % Limestone (micrite)</td>
</tr>
<tr>
<td>Middle (1) Umm Er Radhuma</td>
<td>53.5 – 53</td>
<td>31 % Shale (typical), 21 % Dolomite (typical), 48 % Limestone (micrite)</td>
</tr>
<tr>
<td>Base Umm Er Radhuma</td>
<td>54 – 53.5</td>
<td>23 % Shale (typical), 18 % Dolomite (typical), 59 % Limestone (micrite)</td>
</tr>
<tr>
<td>Erosion</td>
<td>54.5 – 54</td>
<td></td>
</tr>
<tr>
<td>Upper Simsima</td>
<td>55 – 54.5</td>
<td>20 % Limestone (micrite), 10 % (ooid grainstone), 10 % Limestone (shaly), 60 % Dolomite (typical)</td>
</tr>
<tr>
<td>Lower Simsima</td>
<td>65 – 55</td>
<td>10 % Limestone (micrite), 60 % (ooid grainstone), 10 % Limestone (shaly), 20 % Dolomite (typical)</td>
</tr>
<tr>
<td>Fiqa</td>
<td>73 – 65</td>
<td>70 % Marl, 30 % Limestone (micrite)</td>
</tr>
<tr>
<td>Erosion</td>
<td>75 – 73</td>
<td></td>
</tr>
<tr>
<td>Halul</td>
<td>78 – 75</td>
<td>20 % Limestone (ooid grainstone), 60 % Limestone (Chalk, typical), 20 % Limestone (micrite)</td>
</tr>
<tr>
<td>Laffan</td>
<td>80 – 78</td>
<td>85 % Shale (typical), 15 % Limestone (shaly)</td>
</tr>
<tr>
<td>Erosion</td>
<td>82 – 80</td>
<td></td>
</tr>
<tr>
<td>Mishrif</td>
<td>84.61 – 82</td>
<td>10 % Shale (typical), 10 % Limestone (shaly), 10 % Limestone (micrite), 70 % Limestone (ooid)</td>
</tr>
<tr>
<td>Shilaif</td>
<td>87 – 84.61</td>
<td>80 % Limestone (organic rich - 10 % TOC), 20 % Marl</td>
</tr>
<tr>
<td>Mauddud</td>
<td>112 – 87</td>
<td>10 % Limestone (ooid grainstone), 50 % Limestone (micrite), 40 % Limestone (shaly)</td>
</tr>
<tr>
<td>Nahr Umr</td>
<td>114.5 – 112</td>
<td>80 % Shale (typical), 10 % Limestone (micrite), 10 % Sandstone (wacke)</td>
</tr>
<tr>
<td>Bab</td>
<td>118.5 – 114.5</td>
<td>100 % Limestone (organic rich - 1-2 % TOC)</td>
</tr>
<tr>
<td>Reservoir i</td>
<td>124 – 118.5</td>
<td>97 % Limestone (ooid grainstone), 1 % Limestone (micrite), 2 % Limestone (shaly)</td>
</tr>
<tr>
<td>Non-reservoir i</td>
<td>125 – 124</td>
<td>26 % Limestone (ooid grainstone), 74 % Limestone (organic rich - 1-2 % TOC)</td>
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<tr>
<td>Hiatus</td>
<td>125.2 – 125</td>
<td></td>
</tr>
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<td>Reservoir ii</td>
<td>127 – 125.2</td>
<td>93 % Limestone (ooid grainstone), 3 % Limestone (micrite), 4 % Limestone (shaly)</td>
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<td>Horizon/Layer</td>
<td>Time of deposition (Mya)</td>
<td>Lithology</td>
</tr>
<tr>
<td>-----------------------</td>
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<td>--------------------------------------------------------------------------</td>
</tr>
<tr>
<td></td>
<td>from</td>
<td>to</td>
</tr>
<tr>
<td>Non-reservoir ii</td>
<td>127.5</td>
<td>127</td>
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<tr>
<td>Hiatus</td>
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<td>127.5</td>
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<td>Reservoir iv</td>
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</tr>
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<td>Non-reservoir iv</td>
<td>131.11</td>
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</tr>
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<td>Hiatus</td>
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<td>131.11</td>
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<td>Non-reservoir v</td>
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<td>Horizon/Layer</td>
<td>Time of deposition (Mya)</td>
<td>Lithology</td>
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<td>--------------------</td>
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</tr>
<tr>
<td></td>
<td>from</td>
<td>to</td>
</tr>
<tr>
<td>Non-reservoir ix</td>
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<td>134.6</td>
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<td>134.72</td>
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<td>134.88</td>
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<td>Marrat</td>
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<td>Upper Minjur</td>
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<tr>
<td>Horizon/Layer</td>
<td>Time of deposition (Mya)</td>
<td>Lithology</td>
</tr>
<tr>
<td>-----------------</td>
<td>--------------------------</td>
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</tr>
<tr>
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<td>from</td>
<td>to</td>
</tr>
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<td>Middle Minjur</td>
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</tbody>
</table>

Table 6.1: Inputs used for the Field A basin model.
Appendix 7: Field C stable isotope data

The bulk isotope data used in this project were obtained by Petrobras and are presented in Table 6.1. The δ\textsuperscript{13}C and δ\textsuperscript{18}O\textsubscript{VPDB} data in Table 6.2 were collected using a steel handpick.

<table>
<thead>
<tr>
<th>Depth (m)</th>
<th>Δ\textsuperscript{13}C (‰)</th>
<th>Δ\textsuperscript{18}O (‰)</th>
</tr>
</thead>
<tbody>
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<tr>
<td>18.9</td>
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<tr>
<td>30.7</td>
<td>3.74</td>
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<tr>
<td>35.6</td>
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</tr>
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<td>41.7</td>
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<td>48.9</td>
<td>3.29</td>
<td>-1.82</td>
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<tr>
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<td>3.31</td>
<td>-2.24</td>
</tr>
</tbody>
</table>

Table 7.1: Well 2C δ\textsuperscript{13}C and δ\textsuperscript{18}O\textsubscript{VPDB} bulk micrite data, courtesy of Petrobras.
<table>
<thead>
<tr>
<th>Depth (m)</th>
<th>Cement</th>
<th>δ¹³C (%)</th>
<th>δ¹⁸O (%)</th>
</tr>
</thead>
<tbody>
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<td>Peloids</td>
<td>3.1</td>
<td>-3.1</td>
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<td>-1.9</td>
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<td>69.9</td>
<td>Peloids</td>
<td>3.8</td>
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<td>82.4</td>
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<tr>
<td>66.4</td>
<td>Drusy</td>
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<tr>
<td>97.1</td>
<td>Drusy</td>
<td>2.6</td>
<td>-2.2</td>
</tr>
</tbody>
</table>

Table 7.2: Well 2C δ¹³C and δ¹⁸OVPDB data for calcite cement.
Appendix 8: Inputs for the Field C basin model

The Field C basin model used the following inputs for each horizon/layer in the model. The first column is the event that is been modelled (deposition or erosion), with the second column showing the time over which erosion or deposition occurred. The third column shows the lithologies used for the layer.

<table>
<thead>
<tr>
<th>Horizon/Layer</th>
<th>Time of deposition (Mya)</th>
<th>Lithology</th>
</tr>
</thead>
<tbody>
<tr>
<td>Grussai</td>
<td>5 to 0</td>
<td>Limestone (micrite)</td>
</tr>
<tr>
<td>Siri</td>
<td>10 to 5</td>
<td>Sandstone (typical)</td>
</tr>
<tr>
<td>Grussai</td>
<td>20 to 10</td>
<td>Limestone (micrite)</td>
</tr>
<tr>
<td>Siri</td>
<td>30 to 20</td>
<td>Sandstone (typical)</td>
</tr>
<tr>
<td>Ubatuba</td>
<td>65 to 30</td>
<td>Sandstone (typical)</td>
</tr>
<tr>
<td>Carapebus</td>
<td>100 to 65</td>
<td>Shale (typical)</td>
</tr>
<tr>
<td>Outerio</td>
<td>106 to 100</td>
<td>Shale (typical)</td>
</tr>
<tr>
<td>Quissamã</td>
<td>108 to 106</td>
<td>Limestone (oolid grainstone)</td>
</tr>
<tr>
<td>Quissamã - dolomite</td>
<td>111 to 108</td>
<td>Dolomite (typical)</td>
</tr>
<tr>
<td>Retiro</td>
<td>115 to 111</td>
<td>Anhydrite</td>
</tr>
<tr>
<td>Erosion</td>
<td>122 to 115</td>
<td></td>
</tr>
<tr>
<td>Coqueiros</td>
<td>124.5 to 122</td>
<td>Limestone (oolid grainstone)</td>
</tr>
<tr>
<td>Atafona</td>
<td>129 to 124.5</td>
<td>Shale (organic rich, 3% TOC)</td>
</tr>
<tr>
<td>Erosion</td>
<td>130 to 129</td>
<td></td>
</tr>
<tr>
<td>Cabinuas</td>
<td>137 to 130</td>
<td>Basalt (normal)</td>
</tr>
<tr>
<td>Bottom of Section</td>
<td>137</td>
<td></td>
</tr>
</tbody>
</table>

Table 8.1: Inputs used for the Field C basin model.