

**GENESIS OF NATURAL FRACTURE SYSTEMS IN MUDROCKS
(UPPER CRETACEOUS-EOCENE), JORDAN**



By

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Dedication

This thesis is dedicated to the memory of my beloved mother, Nadia, who passed away but will always and forever be in my heart. There are no words to describe how much I miss you.

To my loving father, Salem, for his unconditional love, who always believes in my ability to be successful in my life. I am so proud to be your daughter.

To my beloved husband, Mohammad, and my wonderful kids (Salam, Sara, Jad and Jodi), who have been a great source of love and support. Thank you for being supportive and patient throughout my DPhil life. You have always supported me in a way nobody else could.

*Isra'a Salem Abu-Mahfouz
January 2019*

Declaration

I declare that the work presented in this thesis is entirely my own, except where otherwise stated. The views and interpretations expressed herein are mine, and do not represent those of any other person or group except where this is cited to be the case. The work presented in this thesis has not been submitted for any other degree or professional qualification.

Isra'a Salem Abu-Mahfouz

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Publications and presentations from this work

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Abstract

Organic-rich mudrocks are of great importance to the petroleum industry, potentially acting as source rocks, cap rocks or unconventional reservoirs. The analysis of natural fractures in mudrocks is crucial for the evaluation of their impact on fluid storage and flow. This research focuses on the investigation of natural fractures in organic-rich mudrocks, using integrated core, borehole image, field, petrographic, and geochemical techniques. The aim is to obtain a better understanding of the development of natural fracture systems in mudrocks, regarding their stratigraphic distribution, timing and formation mechanism. Natural fractures are abundant in the Upper Cretaceous-Eocene Belqa Group of Jordan. Distribution analyses of fractures resulted in an informal division of the stratigraphic section into three units; upper and lower highly fractured units and a middle weakly fractured unit. The host lithology represents an important factor influencing fracture distribution and formation; calcite-filled veins were observed in calcareous mudstone and carbonate intervals, and silica-filled veins were observed in chert intervals. Bitumen is present within natural fractures, either as macroscopic deposits, or as inclusions in the fracture cement, suggesting that hydrocarbons were present throughout and after the opening of fractures. Fractures are commonly associated with concretions in the Upper Cretaceous strata, implying that fracturing is sensitive to diagenesis. Open fractures in the succession are often confined to specific beds

minimizing the potential leakage risk due to the high fracture intensities within the upper and lower caps which can be posed where fracture networks are connected. Mechanical stratigraphic controls on fracture distribution, relationship to compaction and early cementation, as well as the presence of bitumen within fractures, suggest that fracturing was affected by different processes and has different timings. Fracture formation mechanisms can be summarized as fluid pressuring (bedding-parallel veins), early diagenesis (folded calcite veins and silica veins) and late tectonic processes (planar calcite veins and joints, and gypsum veins). Petrographic and geochemical analyses show that fractures have developed at two main stages; (1) early diagenesis (calcite veins, ~40-30 Ma, were folded by sediment compaction), followed by fracturing during burial diagenesis, and (2) late fractures (~25-20 Ma) formed during uplift and extension, or fault reactivations.

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Extended Abstract

Natural fractures in hydrocarbon reservoirs and source rocks have a significant impact on fluid storage and flow. Therefore, the analysis of natural fractures in such reservoir and source rocks allows a better understanding of the influence of fractures on hydrocarbon migration and accumulation. It is not clear how the different types of fracture formation mechanisms interact to create fractures in fine-grained rocks. The Upper Cretaceous through Eocene Belqa Group of Jordan is a potential hydrocarbon reservoir/source rock containing abundant natural fracture systems. The shallow burial depth of the strata suggests that the fractures formed under relatively low burial stress conditions, giving an opportunity to study fracture formation during shallow burial conditions.

Thus, the aim of this research is to obtain a better understanding of the development of natural fracture systems in mudrocks, regarding their stratigraphic distribution, timing and formation mechanism. The research focuses on understanding the causal links among interacting stratigraphic architecture, diagenesis, fluid migration, and fracturing.

A variety of analytical techniques were employed, including fracture characterization from core samples, borehole images and outcrops, geostatistical analysis of fracture distribution and orientation, and petrographic and geochemical analysis of fracture cement and their host rocks (beds and concretions).

Chapter 1 provides an introduction to the natural fracture systems in sedimentary rocks and the different factors that control their distribution and development. In Chapter 2, an overview of the geological and tectonic context of the Upper Cretaceous-Eocene succession of Jordan, the focus of this PhD thesis, is presented.

The key findings and related discussions of this study are provided in Chapters 3-5. Chapter 3 reports the study of subsurface natural fracture systems in the Jordan Oil Shale and over- and underlying chert-rich rocks. The study includes description and measurement of subsurface fracture properties from cores (and borehole image logs, together with petrographic and geochemical analysis of the different fractures sampled from the cores. Natural fractures with different properties are abundant in the Upper Cretaceous-Eocene succession of Jordan. A total of 11334 individual fractures were described and measured from the cores within a total core thickness of 3212m, and 4599 fractures from borehole images. The shallow maximum burial depth (c. 500-1000 m) of the succession suggests that the fractures formed under relatively low burial stress conditions, giving an opportunity to study fracture formation during shallow burial conditions, place the fracture sets in a burial/uplift context and relate fracture genesis to timing. Distribution analyses of fractures allowed the division of the stratigraphic section into three units; upper and lower highly fractured units and a middle, low intensity, fractured unit consistent with well tests indicative of very low matrix-dominated permeability.

The diagenetic development of the studied succession was primed to some degree by the original composition of the host-sediment. Veins exhibit a range of filling materials (fracture cement). The most common filling material among all vein types is, in general, calcite (~75%). However, some veins are cemented by silica (~20%) or gypsum (~5%).

The filling materials, which vary in composition and texture, usually reveal the host-rock mineralogy. Specifically, calcite-filled veins were observed in calcareous mudstone and carbonate intervals, and silica-filled veins were detected in silicified and chert intervals. Bitumen is present within natural fractures and adjacent vugs, either as macroscopic deposits or as inclusions in the fracture calcite cement. Bitumen inclusions suggest that hydrocarbons were generally present during and after the opening of most fractures. Joints (unmineralized fractures) in the Jordan Oil Shale are often confined to specific beds minimizing the potential leak risk due to the high fracture intensities within the upper and lower units which can be posed where fracture networks are connected. Fractures are commonly associated with carbonate/silica concretions in the Upper Cretaceous strata, implying that fracturing is associated with early diagenesis. Mechanical stratigraphic controls on fracture intensity, relationship to compaction and early cementation, as well as the presence of bitumen within fractures, suggest that fracturing was the result of different processes and has different timings. The mechanisms that controlled fracture formation throughout the section can be summarized as fluid pressuring (creating calcite/gypsum bedding-parallel veins), early diagenesis (e.g. folded calcite veins and planar silica veins), late diagenesis (gypsum veins), and late tectonic processes (planar calcite veins and joints associated with fault damage zones). Petrographic and geochemical analyses show that fracture timing can be divided into two main groups. These are (1) early calcite veins that were subsequently folded by sediment compaction and fractures formed during burial diagenesis, and (2) late fractures formed during uplift or associated with faulting and/or fault reactivation. Many of the regional faults seem to have been active throughout the burial and uplift history of the section, based on the presence of plastic sediment deformation and fracturing related to these faults. The different calcite structures show $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ signatures ranging from -20.70‰ to -1.0‰

and from -15.60‰ to -2.80‰, respectively. The $^{87}\text{Sr}/^{86}\text{Sr}$ values of the analysed calcite veins (planar veins, folded veins and bed-parallel veins) range between 0.7077 and 0.7080. The samples are clustered on the global sea water curve between 40 and 25 Ma, whereas the host-rock samples are grouped between 50 and 40 Ma. The radiometric (U/Pb dating) age shows that folded calcite veins are clustered around ~45 Ma, which is near the defined chronostratigraphic age of the rocks, and microfaults and planar vertical veins show an age between ~25-20 Ma.

Chapter 4 presents the study of natural fracture systems in outcrops exposing equivalent rocks, the Jordan Oil Shale and over- and underlying chert-rich rocks. The study includes field investigations of: (1) the fracture characteristics, fracture distribution, and the relationships between fractures and concretions observed in the field, (2) petrographic examination of the different fracture systems and (3) geochemical analysis of fracture cements, concretions, and their host rocks. This chapter complements the study of fractures from cores and image logs and provides a better understanding of the lateral distribution of the natural fracture systems in the study succession. The timing of outcrop fractures can be divided into two main groups based on petrographic and geochemical analyses. These are early calcite veins that were folded by sediment compaction and fractures associated with concretion formation (e.g., silica veins), and late fractures that include fractures formed during uplift or associated with faulting and/or fault reactivation, or by late diagenetic processes (e.g., gypsum veins).

Chapter 5 presents the findings of the comprehensive study of the bitumen-bearing fractures observed in the study succession (the Upper Cretaceous-Eocene rocks of Jordan), with a main focus on the source of fracture-filling bitumen and its role in the fracturing. Bitumen is present as macroscopically visible deposits and as inclusions in the cement of abundant natural fractures and adjacent vugs. The frequency of bitumen

occurrence in fractures closely correlates with Total Organic Carbon and burial depth. Petrographic and organic-geochemical analyses on bitumen samples extracted from fractures and their host rock matrix show that the fracture-filling bitumen comprises indigenous low maturity (<0.5 VRE) hydrocarbons derived from the surrounding organic-rich oil shale and has not migrated from a deeper or shallower source. Maturity indicators imply that the oil shale is in the pre-oil generation stage of early catagenesis throughout the investigated area, but with a regional increase in thermal maturity from west to east as the result of greater maximum burial depth. Bitumen fractures were developed by hydraulic fracturing as the result of fluid overpressure in the organic matter. Overpressured bitumen has acted as a fracture driver, generating bitumen veins in both the organic-rich mudstones and the adjacent chert and silicified intervals.

Chapter 6 discusses the distribution, timing and possible formation mechanisms of the different fracture systems and associated concretions observed in the succession.

Finally, Chapter 7 concludes the key findings of this research. In conclusion, this PhD research has contributed towards a better understanding of the development of natural fracture systems and the genesis of co-eval concretions in unconventional hydrocarbon reservoirs, by analysing and discussing the controls on their distribution, the timing of fracture development and the genetic formation mechanism(s) that controlled fracturing and diagenesis.

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List of Abbreviations

AHP: AL-Hisa Phosphorite Formation

ASL: Amman Silicified-Limestone Formation

BCV: Bitumen-Calcite Vein

BHI: Borehole Image

BSE: Backscattered Electron

BV: Bitumen Vein

C-I-C: Cone-In-Cone

EDS: Energy-Dispersive X-Ray Spectroscopy

FCV: Folded Calcite Vein

HCL: Hydrochloric Acid

HI: Hydrogen Index

HNO₃: Nitric Acid

JOS: Jordan Oil Shale

JOSCO: The Jordan Oil Shale Company

LF: Lithofacies

MC-ICP-MS: Multicollector Inductively Coupled Plasma Mass Spectrometry

MCM: Muwaqqaer Chalk-Marl Formation

NRA: Natural Resources Authority

OPTV: Optical Televiwer

QSC: Qirma Calcareous Sandstone Formation

RE: RockEval

SEM: Scanning Electron Microscope

S_{hmin}: Minimum Horizontal Stress

S_{HMAX}: Maximum Horizontal Stress

Sr: Strontium

TOC: Total Organic Carbon

TS: Total Sulphur

U/Pb: Uranium/Lead

URC: Umm Rijam Chert-Limestone Formation

VRE: Vitrinite Reflectance Equivalent

WSC: Wadi Shallalah Chalk Formation

WUG: Wadi Umm Ghudran Formation

Chapter 1: Introduction

1.1 Research Background

A natural fracture is a structural discontinuity can be present in different shapes, sizes and geometries, and originates as a result of a loss of the cohesion of rocks due to a natural fracturing mechanism during burial or uplift. In general, the compressive strengths (tensile, cohesive and unconfined) of a fractured rock unit are significantly lower than of those of unfractured rock unit of the same type (e.g., Hoek, 1983; Barton et al., 1985; Schultz, 1996).

Brittle fractures are characterised into three main types, based on their mode of displacement (Pollard & Segall, 1987; Scholz, 2002; Bons et al., 2012; Spence et al., 2014) (Figures 1.1 and 1.2). These are (1) mode I fractures (opening-mode), in which the displacement vector of the fracture is perpendicular to the fracture plane (Figure 1.1a), (2) Mode II fractures (in-plane shear), known also as a sliding shear mode, reflecting a case where the displacement vector of the fracture is parallel to the fracture plane and parallel to the fracture propagation direction (Figure 1.1b) and (3) Mode III fractures (out-of-plane shear), known also as tearing shear mode; the displacement vector of the fracture is parallel to the fracture plane and perpendicular to the fracture propagation direction (Figure 1.1c).

Opening-mode fractures, which are the main focus of this study, are known as extensional fractures, and usually propagate along a plane perpendicular to the minimum-compressive principal stress (Secor, 1965). They can be present either in the form of uncemented fractures (i.e., joints) or as cemented fractures (i.e., veins). Veins may still be redeveloped or reopened along the interfaces between the vein walls and

the vein cement, due to the low tensile strength of these borders (Gale & Holder, 2008, 2010).

Opening-mode fractures are thought to form where pore pressure exceeds the least compressive total normal stress and the appropriately-oriented failure strength of the rock in tension, but they may also form at pore pressures significantly less than the overburden stress magnitude, and by minor extensional or volumetric stresses, or thermal stresses, particularly under shallow burial conditions (Eichhubl et al., 2001; Eichhubl & Aydin, 2003; Olson et al., 2009). They can also be present in association with shear fractures (faults). The abundance of opening-mode fractures has been reported to increase in shear zones, particularly towards the fault core (e.g., Agosta & Aydin, 2006).

Natural fractures, open or cemented, vary in their morphology, mineralogy, orientation, and distribution in different geological settings (Figure 1.3). Fracture characteristics, including type, orientation and internal structures, can provide significant information on fracture development controls such as paleo-stress fields, kinematics of deformation, and fluid pressure (Urai et al., 1991; Angelier, 1994; Twiss & Unruh, 1998; Hilgers & Urai, 2002a; Bons et al., 2012). Unlike open fractures, mineralised veins, as certain natural fractures, preserve a good record of the original fracture properties, providing a unique opportunity for examining natural fractures in mudrocks (Gillespie et al., 2001).

Rock properties (such as elastic properties: bulk modulus and Poisson's ratio) are considered as a main control on fracture development and distribution (Gross et al., 1995; Underwood et al., 2003; Bourne, 2003; Cooke et al., 2006; Laubach et al., 2009; Couples, 2013). The difference in the lithology and mechanical properties of host strata have been observed to control the fracture height and spacing of different types of fractures in layered rocks, and this can be, in turn, significant for the fracture formation

mechanisms (Ladeira & Price, 1981). Fracture spacing has been widely believed to be related to the thickness of host beds in different types of rocks (e.g., Ladeira & Price, 1981; Narr & Suppe, 1991; Gross et al., 1995; Bai & Pollard, 2000; Underwood et al., 2003; Schöpfer et al., 2011). Figure 1.4 presents the different settings of fracture systems in layered rocks.

Various mechanisms may control fracture formation in mudrocks, either individually or in combination, including fluid overpressure, tectonic events, diagenesis, and force of crystallization (Secor, 1965; Cosgrove, 1995; Parnell & Carey, 1995; Volland & Kruhl, 2004; Gale & Holder, 2010; Meng et al., 2017).

Tectonically-imposed fracturing can be caused by regional or local tectonic stresses at different times of the basin history (Gudmundsson, 1987; Volland & Kruhl, 2004). In addition, stress release or cooling during uplift can generate tectonic fractures (Lorenz & Finley, 1991; Gale & Holder, 2010).

Fracturing caused by fluid overpressuring, known as hydraulic fracturing, is usually initiated when fluid pressure exceeds the least principal stress and the tensile strength of the sediments (Hubbert & Willis, 1957). In such type of fracturing, the fracture usually propagates perpendicularly to minimum horizontal stress (S_{hmin}) (Figure 1.5). This type of fracturing in mudrocks has been related to different processes, such as fluid expulsion during diagenetic processes or primary oil migration (Mandl & Harkness, 1987; Freed & Peacor, 1989; Parnell & Carey, 1995; Katahara, 2006).

The development of natural fractures in organic-rich shales has become recently a focus of a large amount of research (e.g., Gillespie et al., 2001; Curtis, 2002; Gale et al., 2007; Engelder et al., 2009; Gale & Holder, 2010; Walton & McLennan, 2013; Gasparri et al., 2014; Gale et al., 2014; Hooker et al., 2017a, 2018; Alfahmi, 2018; Abu-Mahfouz

et al., 2019a,b), due to the growing interest of exploration in unconventional shale plays. The analysis of natural fractures in such hydrocarbon reservoirs allows a better understanding of the influence of fractures on fluid storage and flow. However, constraining the genesis and timing of natural fractures and understanding their implications on fluid flow are still challenging subjects. It is not clear how the different types of fracture formation mechanisms interact to generate subsurface fractures. Therefore, this thesis research focuses on understanding the causal links among interacting stratigraphic architecture, diagenesis, fluid migration, and fracturing, both in the subsurface (from core and borehole image logs) and in the field from outcrop analogues, with a major aim of placing fractures in a structural and basin evolutionary context and evaluating their effects on hydrocarbon migration.

However, fracture characterization based solely on the outcrop analogues can be irrelevant or even worse, misleading, primarily with regards to length-aperture scaling, cement fill and trace morphology. Outcrop studies can without doubt help to constrain fracture distribution, and complement the analysis of subsurface fractures that can be studied through core, seismic and borehole image logs (Alfahmi, 2018). Fractures retrieved from the core and borehole images are important to get an insight into subsurface reservoir evolution.

Natural fractures are abundant in the Upper Cretaceous through Eocene strata of Jordan, but vary in character and have different properties. These differences present a unique opportunity to examine the different fracture properties, place the fracture sets in a burial/uplift context and relate fracture genesis to timing and distribution.

1.2 Oil Shale Definition

In general, oil shales are defined as fine-grained sedimentary rocks that contain preserved organic matter in the form of kerogen (Duni, 2006), which is insoluble in organic solvents. Kerogen is the source of the liquid hydrocarbons (shale oil). The organic richness of an oil shale rock is expressed by Total Organic Carbon (TOC, wt%), or by the oil yield. The organic matter preserved in the oil shale is typically composed of carbon, hydrogen, oxygen, and a small amount of sulphur. It is usually present mixed with some minerals such as fine-grained silicates and carbonates. Bitumen, an organic matter that is soluble in organic solvents, can be also present in oil shales.

Oil shale comprises a lesser amount of organic matter than coal. In addition, the organic matter in oil shale has a hydrogen/carbon atomic ratio up to 1.8 times lower than for crude oil and up to 3 times higher than for coals (Duni, 2010). A variety of organisms, such as the remains of algae, pollen, spores, and plants cuticles and corky fragments of herbaceous and woody plants usually form the organic components of oil shale (Alali et al., 2003; Duni, 2010).

1.3 The Jordan Oil Shale

The Jordan Oil Shale (JOS) is typically described as an organic matter-rich, bituminous marl and limestone that was deposited in a highly reducing shallow marine environment during the Late Cretaceous (Maastrichtian) to Late Eocene Periods (Abed & Arouri, 2006; Alnawafleh, 2007; Alnawafleh & Fraige, 2015; Alqudah et al., 2015, März et al., 2016). The thickest oil shale interval belongs to the Muwaqqar Chalk Marl Formation.

Oil shale in Jordan is considered as one of the most significant natural resources (Sunna', 2004). The JOS is of high quality and is characterised as one of the richest oil

shales in the world (Jaber et al., 1997; Hamarneh, 1998; Dyni, 2006). It is considered to be one of the world's top 10 largest oil shale reserves (Figure 1.6), despite the high sulphur content. The JOS is widespread, underlying >60 % of the Jordan region, with an estimated total in-place resource up to 5242 million tons of oil shale, which is equivalent to 34172 million barrels of shale oil (Dyny, 2006; Alali et al., 2006; Dyni, 2010). Oil Shale deposits are reported in about 24 main localities (Figure 1.7). Eight of these localities have the most important deposits, and they are located in west-central Jordan, ~20-75 km east of the Dead Sea. These are El-Lajjun, Attarat Umm Ghudran, Juref ed Darawish, Sultani, Wadi Maghar, Khan ez Zabib, Suwaqa, and Wadi Thamad (Dyny, 2006).

The TOC of the JOS ranges from 3 to 30 wt% (Hakimi et al., 2016). Sulphur is usually associated with the organic matter, with minor occurrences as pyrite (Abed & Arouri, 2006). The major mineral components of the JOS are calcite, quartz, kaolinite, and apatite, with small amounts of dolomite, pyrite, illite, feldspar, and gypsum. The organic matter has high oil generative capacities, with Hydrogen Indices ranging between 600 and 900, and is characterised as a Type I/II-S kerogen due to its elevated sulphur contents between 2 and 12 wt% (Abed et al., 2005; Hakimi et al., 2016; Sokol et al., 2017; Al Aljariri Alhesan, 2018; Abu-Mahfouz et al., 2019a).

1.4 Significance of the Study

This study was conducted to examine the natural fracture systems in unconventional shale reservoirs, and to set the fractures in a structural and basin evolutionary context in order to evaluate their effects on hydrocarbon migration and potential for economic production. The wealth of the data provided by JOSCO (Jordan Oil Shale Company), including the open access to the core samples, wireline petrophysical logs and borehole

images of the wells drilled across Jordan (an area of about 22000 Km²) in the period between 2010-2014, combined with field works carried out in central and south Jordan, has allowed a unique opportunity to characterize the natural fracture systems in the Upper Cretaceous-Eocene succession of Jordan that has not been previously studied in detail for this purpose. Therefore, the present study provides a general framework of the structural evolution of the JOS and the over- and underlying chert-rich units. In addition, this study is considered as one of the first studies to utilise U-Pb dating techniques to constrain the absolute timing of fracture development in carbonate rocks.

The fracture systems in the JOS are ideal for a study of natural fractures because: (1) the Late Cretaceous through Eocene time interval in Jordan and surrounding countries created a potential petroleum resource containing abundant natural fractures, and hence this study has both a scientific and resource significance at the same time; (2) there is a range of fracture types present in a lithologically diverse succession, hence offering opportunities to study the interaction between tectonic and diagenetic fracture formation mechanisms in unconventional hydrocarbon plays.

1.5 Research Methods

This study is based on investigations of fractures from cores, borehole image logs, and outcrops. The characteristics of the different natural fracture systems were examined from cores and image logs. Field work afforded (1) a comprehensive study of the regional and local geological settings, including tectonic and sedimentary structures, and stratigraphy, and (2) characterization of the natural fracture systems in the equivalent interval and correlation with the subsurface fractures. The detailed work is presented in Figure 1.8. A variety of analytical techniques were used, each of which was geared towards constraining the structural evolution of the JOS rocks, and

understanding their burial and diagenetic evolution. These techniques are listed as follows:

- (1) Geological description of core samples, including: (a) identification of fracture types, morphologies and mineralization, (b) measurement of fracture geometries, apertures and spacing, (c) analysis of crosscutting relationships between fractures, and (d) measurement of strain around fractures and concretions;
- (2) Geostatistical analysis of fracture distributions and correlation of fracture logs with geochemical and wireline data (TOC, gamma ray, and density logs). These tools are important in predicting geology away from well control, for integration into geological models, and to feed into structural and depositional settings of the rock formations. Techlog software was used to conduct this work.
- (3) Study of natural fracture systems at outcrop analogues: (a) fracture description and measurement, (b) examining fracture stratigraphy, (c) sampling of the different fracture cements and host rocks, (d) studying the relationships between fractures and the diagenetic features such as concretions; and (e) measuring the strain field around chert and carbonate concretions;
- (4) Petrographic analysis of fractures and their host rocks to constrain the timing of the fracture systems. The thin sections were examined using optical microscope and scanning electron microscope (SEM); and
- (5) Geochemical analysis, including (a) stable (oxygen and carbon) and radiometric (strontium and U/Pb) isotopic analysis of fracture cements and host rocks, X-ray diffraction analysis of some fracture cements and (c) organic geochemical analysis (organic biomarkers) of migrating organic matter and host rocks.

The detailed methodology of each analysis is provided in each result chapter (Chapter 3, Chapter 4 and Chapter 5).

1.6 Research Aims

In general, this research aims to investigate the distribution, timing and formation of natural fracture systems in mudrocks, and place the fractures in a structural and basin evolutionary context.

Generic questions to be addressed by this research are:

- (1) Does early diagenesis lead directly to the development of natural fracture systems?
- (2) Is there any relationship between cementation/orientation and stress?
- (3) Is there a link between stratigraphic, diagenetic, and structural development?

Specifically, the research aims to address the following questions:

- (1) What are the geological controls of fracture abundance and spatial distribution?
- (2) What is the timing of development of different fracture types?
- (3) What is/are the genetic mechanism(s) of fracture formation?
- (4) What are the relationships between fractures and hydrocarbon maturation and migration?

1.7 Thesis Outline

Chapter 1 introduces the research background, explains the main significance of conducting this study, and summarises the research methods and aims.

Chapter 2 provides an overview of the geological and tectonic settings, structures, stratigraphy, and depositional environments of the Upper Cretaceous-Eocene strata of Jordan.

Chapter 3 reports the study of subsurface natural fracture systems in the JOS and over- and underlying chert-rich rocks. The study includes description and measurement of fracture properties from cores and borehole image logs, and petrographic and geochemical analysis of the different fracture systems observed in the cores.

Chapter 4 presents the study of natural fracture systems in outcrops exposing the JOS and over- and underlying chert-rich rocks. The study includes field investigations of: (1) the fracture characteristics, fracture distribution, and the relationships between fractures and concretions observed in the field, (2) petrographic examination of the different fracture systems and (3) geochemical analysis of fracture cements, concretions, and their host rocks.

Chapter 5 presents the findings of the comprehensive study of the bitumen-bearing fractures observed in the study succession (the Upper Cretaceous-Eocene rocks of Jordan), with a main focus on the source of fracture-filling bitumen and its role in the fracturing.

Chapter 6 discusses the distribution, timing and possible formation mechanisms of the different natural fracture systems observed in the study succession.

Chapter 7 summarizes the key findings of this research.

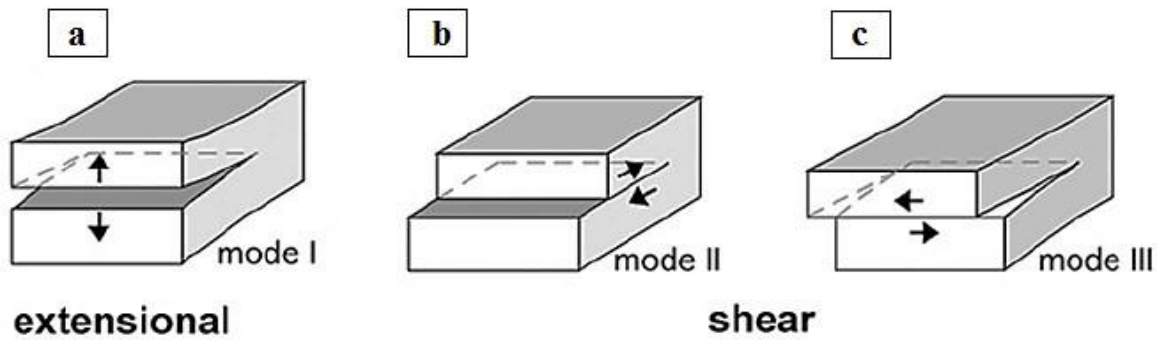


Figure 1.1 A schematic drawing showing the possible modes of failure in the rocks. Mode I, perpendicular to the fracture plane; Mode II, the displacement vector of the fracture is parallel to the fracture plane and parallel to the fracture propagation direction; Mode III, the displacement vector of the fracture is parallel to the fracture plane and perpendicular to the fracture propagation direction.

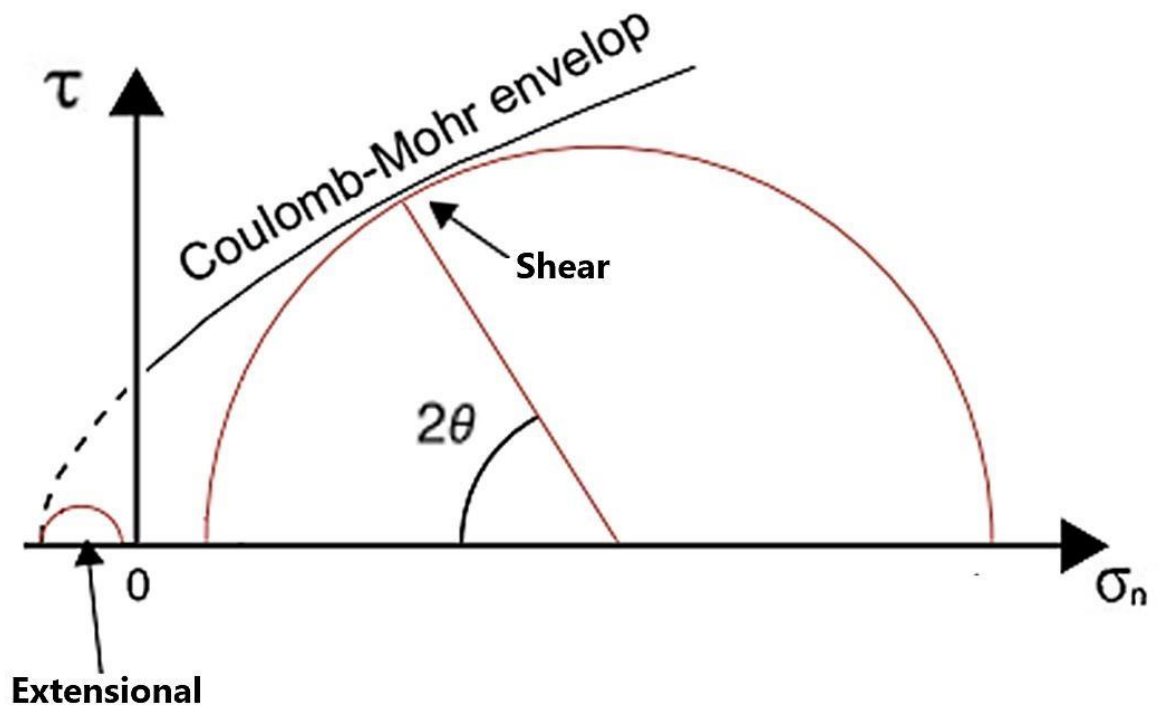


Figure 1.2 The Coulomb-Mohr failure criteria demonstrating development of fractures under normal and shear stress fields in rocks. τ , Shear Stress; σ_n , Normal Stress; ϑ , the Angle of the Failure Plane.

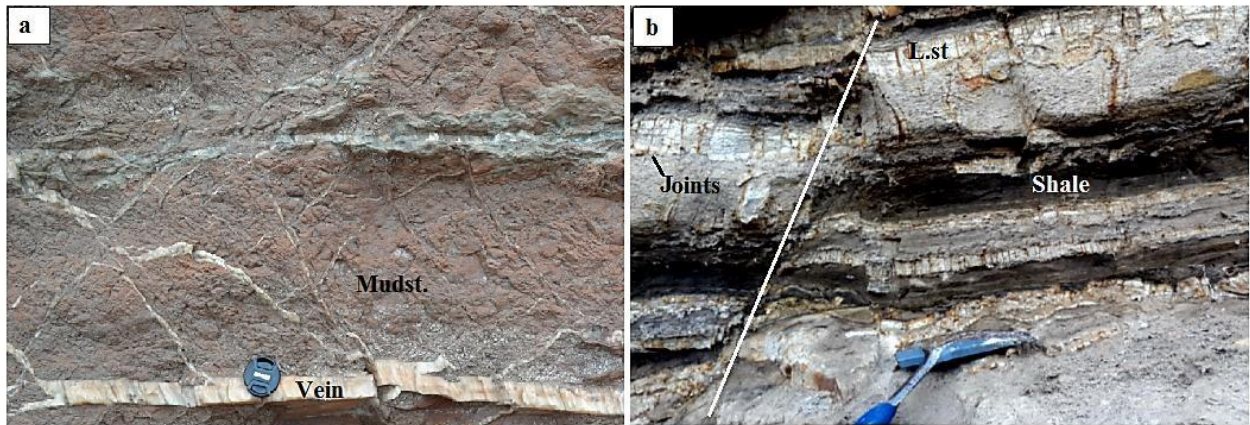


Figure 1.3 Field photographs showing examples of different fracture systems, including veins in mudstone, SW England (a), and joints and a fault (white line) in carbonates in Sicily, Italy (b). L.st, Limestone.

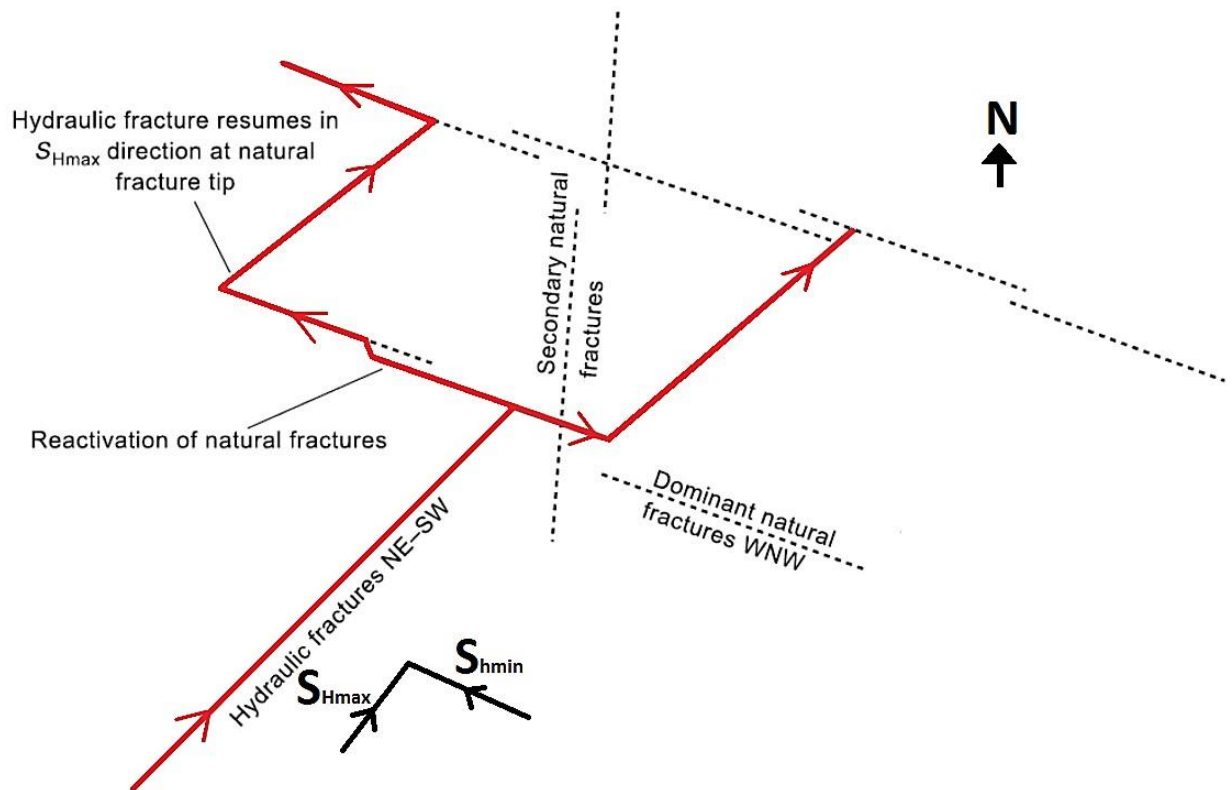


Figure 1.4 Illustration of hydraulic-fracture propagation in naturally-fractured rocks (modified from Gale et al., 2007). Hydraulic fractures (red line) propagate perpendicular to minimum horizontal stress, S_{hmin} , passing through natural fractures (dashed-black lines) with a large approach angle, and reactivates pre-existing natural fractures with a relatively small approach angle. The arrows indicate the propagation direction of hydraulic fractures.

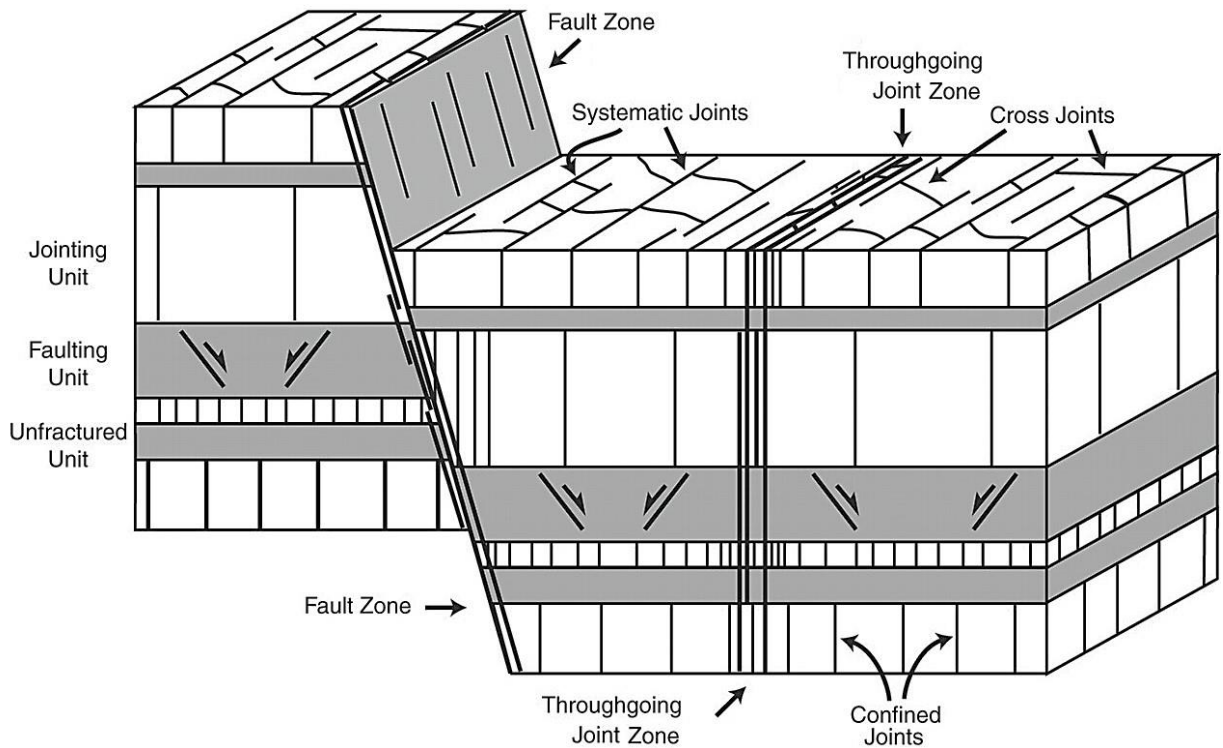


Figure 1.5 A model showing the fracture architectures in layered rocks (modified from Gross & Eyal, 2007).

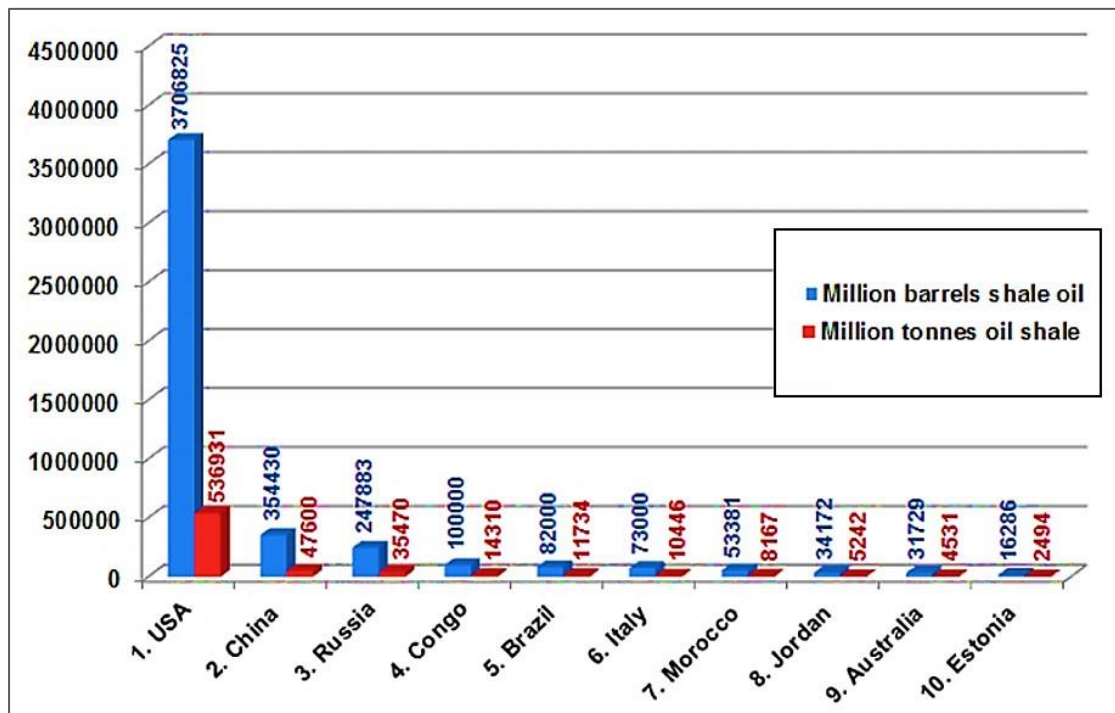


Figure 1.6 The top 10 countries with largest oil shale reserves in the world at the end of 2008 (modified from Dyni, 2010).

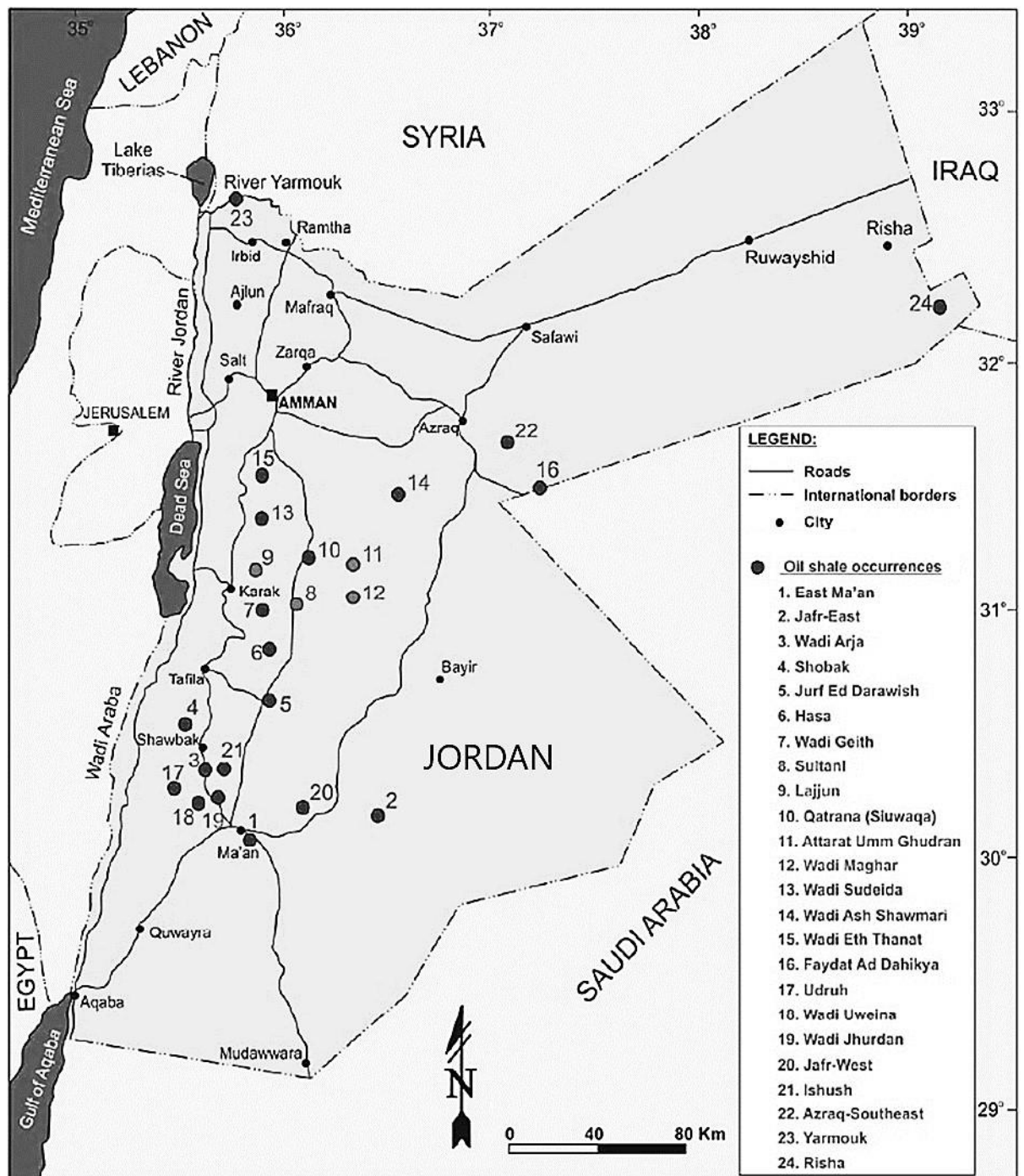


Figure 1.7 Surface and subsurface of oil shale occurrences in Jordan (modified from Hamarneh, 2006).

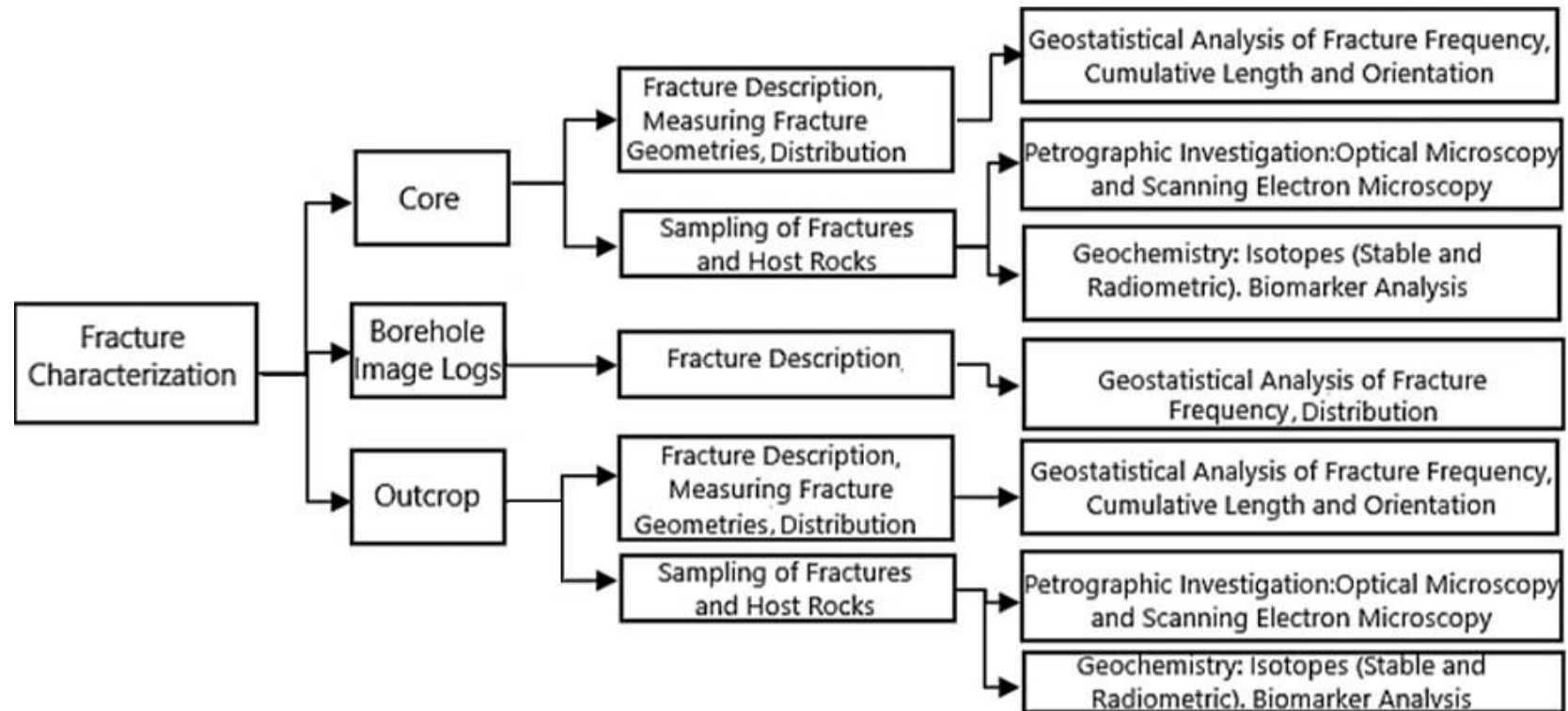


Figure 1.8 Workflow of the present study.

Chapter 2: Geological Context

2.1 Introduction

This chapter provides an overview of the geological and tectonic settings, structure, stratigraphy, and depositional environments of the Upper Cretaceous-Eocene Belqa Group of Jordan. The stratigraphic units studied include the JOS and the over- and underlying chert-rich units.

2.2 Geological and Tectonic Setting

Jordan is located on the north-western margin of the Arabian Plate (Figure 2.1). During the Late Cretaceous-Paleocene, it was part of a shallow shelf on the southern margin of the Neo-Tethys Ocean (Powell & Moh'd, 2011). The tectonic evolution of the Arabian Plate was affected by major regional structures, including the African-Arabian rift system, the collision of the Arabian and Eurasian Plates and the Arabian dome (Barberi et al., 1970; Garfunkel, 1988; Burke, 1996) (Figure 2.1).

The African-Arabian plate separation was reconstructed and interpreted to have occurred approximately 25 Ma ago (Joffe & Garfunkel, 1987; Steckler et al., 1988), causing the closure of the Tethys Ocean in the northeast (Stern & Johnson, 2010). This rifting of the African and Arabian Plates led to the opening of the Red Sea (southwest Arabia) and the Gulf of Aden (South Arabia) (Figure 2.1).

On the northeastern side of the Arabian Plate, the present-day relative motion of the plate towards the north, results in the collision of Arabia with the Eurasian and Anatolian Block, forming the East Anatolian Fault, the Bitlis suture, the Zagros Fold Belt, and the Makran

Trench (Figure 2.1). The Zagros Fold Belt forms a major part of the Alpine Orogeny associated with the Neo-Tethys closure, essentially started during the Late Cretaceous (Alavi, 1994).

Two major transform fault zones demarcate the Arabian Plate along its eastern and western margins. These are the Dead Sea Transform Fault in the west and the Owen Transform Fault in the east (Figure 2.1).

The upper crust of the Arabian Plate comprises Precambrian crystalline basement rocks known as the Arabian Shield, the uplifted basement of the Arabian Plate (Figure 2.1). These are overlain by Phanerozoic sediments that extend over the whole region except for the southwest, which has been uplifted by the rifting of the Red Sea, and eroded, and covered by Cenozoic flood basalts (Stern & Johnson, 2010). Both sedimentary and volcanic rocks have formed the Arabian Platform, which covers most of the Jordan region. The sedimentary succession is mainly carbonate and siliciclastic deposits, with occasional evaporites (Figure 2.2). The Precambrian and Palaeozoic rocks are exposed in south Jordan and in small parts along the Wadi Araba-Dead Sea-Rift (Bender, 1974) (Figure 2.3). They unconformably underlie the Mesozoic and Cenozoic sediments that outcrop all over Jordan. Neogene basalt covers large areas of the northeastern Jordanian desert (Bender, 1968; 1974) (Figure 2.3).

Different geological events took place in the Late Cretaceous on the southern continental margin of Tethys (Flexer et al., 1986; Cohen et al., 1990; Powell, 1989; Powell & Moh'd, 2011). These include the local extensional tectonics in Sinai (Bachmann et al., 2010), compressional/transpressional (Levant Plate/Syrian Arc) and extensional deformation (Arabian Plate), karstic erosion, isostatic uplift, and gravitational slump-folding.

The Syrian Arc, which is considered as a main structural element in the region, is a NE-SW intraplate fold-thrust belt that extends from the central Sinai through Palestine northward to central Syria and Lebanon (Chaimov et al., 1992; Shahar, 1994) (Figure 2.1). This system of folds has significantly influenced the regional paleotopography (Alqudah, 2014). This led to the formation of anticlines, creating a belt of local highs during the Late Cretaceous. On the contrary, the synclines represented local basins that form an appropriate environment for the deposition of oil shale (Abu-Jaber et al., 1989; Ziegler, 2001). The formation of the northern part of the Syrian Arc (Palmyride Fold Belt), which developed as a basin during the Jurassic, has been controlled by the opening of the Neo-Tethys (Chaimov et al., 1992). The main deformation stage of the Syrian Arc took place during Eocene-Miocene, and seismic activity continues to the present-day (Chaimov et al., 1992).

The structure and sedimentation of the Jordan region were primarily affected by the tectonic evolution of the area (Abu-Jaber et al., 1989). The structural style was influenced by the opening of the Red Sea and the Gulf of Aden (Diabat, 2009). The Arabia-Eurasia plate movements and collision have resulted in considerable deformation along the continental margins, causing uplift of some areas and the development of syn-sedimentary tectonics (Butterlin et al., 1993), including restricted sub-basins formed due to the resulting subsidence (Lopes & Cunha, 2007) (Figures 2.3, 2.4 and 2.5). Reconstruction of the uplift and subsidence history of the resulting structures can help evaluate the development of the sub-basins with respect to the faults (Alqudah et al., 2015).

The several different stress regimes produced by the interactions of the different tectonic elements in the region have created different strain patterns. The major uplift and erosion events that affected the region include the uplift and erosion in the Late Jurassic-Early

Cretaceous, uplift and erosion in the Late Cretaceous and the uplift in the Cenozoic (Figures 2.6 and 2.7). Small uplifts distributed in central and southeastern Jordan are described to have been caused by tangential compression (Bender, 1974). The Late Cretaceous regional exhumation was caused by the Alpine shortening (Abu-Jaber et al., 1989) (Figure 2.6). Horizontal shortening and arch formation have controlled regional tectonism in Jordan (Eyal & Reches, 1983; Abu-Jaber et al., 1989). Rifting occurred in the Late Jurassic-Early Cretaceous (Bandel, 1981; Powell & Moh'd, 1993) was accompanied by intraplate continental volcanism (Lang & Mimran, 1985). The Early Cretaceous marks a phase of post-rift flexural subsidence of the passive continental margin (Powell & Moh'd, 2011). A full summary of the main tectonic events including the uplifts and the subsidence that affected the region of Jordan between Late Cretaceous and Paleogene is presented in Figure 2.6.

The Jordan region is characterized by four main geomorphological features, namely the Precambrian Basement Complex in the south, the Late Cretaceous to Early Tertiary Azraq-Sirhan Depression in the east adjacent to the border between Jordan and Saudi Arabia, the Neogene Dead Sea Transform associated with pull-apart basins to the west, and the Rutbah Uplift with widespread Neogene volcanic field to the north (Lüning & Kuss, 2014) (Figures 2.3 and 2.4).

Jordan contains four major basins, including the Al-Jafr Rift Basin to the south, the Dead Sea Basin in the west, the Azraq-Sirhan Basin in the Central and south east, and the Paleozoic Risha Basin in the north east (Figures 2.3, 2.4 and 2.5). Paleogeographic investigations suggest that the Azraq-Sirhan Basin, which forms a big proportion of the area of focus, started to develop in the Early Cretaceous. Consequently, the Mesozoic strata in this basin

experienced several events, including significant deformation, regional uplift, erosion and volcanic activity (Abu-Jaber et al., 1989).

The Dead Sea Transform, which forms the boundary between the Arabian and Sinai Plates (Garfunkel & Ben-Avraham, 1996) and borders Jordan from the west (Figures 2.1 and 2.3), is the most significant tectonic feature in the region. It has played a main role in controlling the stratigraphy and structures of the area (Diabat, 2013). The Dead Sea Transform Fault extends from the Taurus Mountains of the Alpine Orogenic Belt in Turkey in the north to the northern rim of the Red Sea in the south (ca. 1000 km long) (Figures 2.1 and 2.8). The fault has undergone ~107 km of strike-slip displacement since the time of its formation in the Early Miocene (Freund et al., 1970; Garfunkel, 1981; Quennell, 1983, 1984; Vita-Finzi, 2001). Its separation is estimated to have occurred in two main stages based on matching the structural and sedimentary features offset across the fault zone (Quennell, 1951). The first separation of ~63 km occurred during the Miocene, and the second separation took place during the late Pleistocene. Another parallel conclusion was developed by Freund et al. (1970) and Garfunkel (1981) based on measuring concurrent motion in the adjacent Red Sea.

The Dead Sea Basin contains a total basin fill approximately 10 km (Ginzburg & Ben-Avraham, 1997). The eastern and western flanks of the Dead Sea Rift are asymmetrical with considerably greater uplift and riftward flexing, and higher elevations of the eastern side compared to the western flank (Zak & Freund, 1981; Wdowinsky & Zilberman, 1996; Feinstein et al., 2013) (Figures 2.9, 2.10 and 2.11). Uplift is interpreted to be part of a regional rearrangement along the western Arabian platform margin occurring at the time of Red Sea rift initiation (Feinstein et al., 2013).

2.3 Structural Elements

The main structural elements in Jordan comprise the N-S striking Dead Sea rift valley to the west, NW-SE striking normal faults (e.g., the Fuluq, Ramtha-Wadi Sirhan, Amra, Wadi Dabi, and Karak-Fayha Faults) and grabens (e.g., the Azraq and Sirhan Grabens), a set of E-W striking, right-lateral strike-slip faults (e.g., the Zarqa Ma'in and Suwaqa Faults) and gentle folds related to the Syrian Arc (Beydoun et al., 1994; Al-Zoubi & Ben-Avraham, 2002; Diabat, 2009; Segev et al., 2014) (Figures 2.1, 2.3, 2.4, 2.5 and 2.8). The northwest-striking faults are the longest faults in the area. For instance, the Karak-Fayha Fault is approximately 300 km long (Bender, 1974) (Figure 2.8).

Active faulting during the Late Cretaceous-Eocene represented by the Zarqa Ma'in Fault, the Suwaqa Fault, the Salawan Fault, the Karak-Fayha Fault and the Sirhan Fault System has contributed to the development of five main sub-basins in Jordan (Figures 2.3, 2.4 and 2.5). These are the Azraq-Hamza Sub-basin (Central East), the Sirhan Sub-basin (South East), the El-Lajjun Sub-basin (Central West), the Al-Jafr Sub-basin (South), and the Yarmouk Sub-basin (North) (Figure 2.3). Sediments were accommodated by local extension along the NW-SE striking normal faults (Abu-Jaber et al., 1989; Lüning & Kuss, 2014; Hooker & Cartwright, 2016). Sedimentation has occurred within the fault-bounded sub-basins (Alqudah et al., 2015), with an average sedimentation rate approximately 1.2 cm/Ky on the stable ramp areas. In the rifted Azraq-Hamza Sub-basin in the central east, it was however approximately 6.5 cm/Ky (Powell & Moh'd, 2011). The maximum burial depth of the sediments within the basins is suggested to reach approximately 1-1.5 km (Figure 2.7).

The Azraq-Hamza Sub-basin is part of the Azraq-Sirhan Graben System covering the central and southeast parts Jordan (Figures 2.3 and 2.4). The Azraq-Sirhan Graben System formed by a phase of normal dip-slip movement of the Sirhan and Fuluq Faults (Naylor et al. 2013; Alqudah et al., 2015). The graben was further divided into the Azraq-Hamza and Sirhan Sub-basins by the E-W trending Suwaqa Fault (Beydoun et al. 1994; Alsharhan & Nairn 2003) (Figure 2.3). This sub-basin is restricted by the Sirhan, Siwaqa and Zarqa Ma'in Faults (Alqudah et al., 2015) (Figures 2.4 and 2.8). During the Maastrichtian-Middle Eocene, subsidence along the Sirhan Fault system enabled the accumulation of thick sediments in the Azraq-Hamza Sub-basin. This is reflected in the eastern part of the sub-basin where sediments are twice the thickness of its western equivalent block (Alqudah et al., 2015).

The Sirhan Sub-basin, which extends along the northwest of Saudi Arabia (equivalent to the Al-Jawf Basin; Aoudeh & Al-Hajri, 1994), is a NW-SE trending basin associated to the Sirhan Faults System (Figures 2.3, 2.4 and 2.8). Oil shales of variable thickness accumulated in the sub-basin during Maastrichtian-Eocene time, indicating that sedimentation was influenced by the NW-SE fault system, as a result of graben subsidence (Alqudah et al., 2015). Missing Paleocene biozones on the footwall of the Sirhan Faults indicates that this fault system was active during the Paleocene.

The El-Lajjun Sub-basin is a N-S trending graben (Figure 2.3). The graben is considered as the most important graben in central Jordan where considerable thicknesses of oil shale deposits are preserved in the lower part of the Muwaqqar Chalk Marl formation (Abu Ajamieh, 1980, Shawabkeh, 1991). The graben controls a morphological depression bounded to the east and west by faults striking mostly N-S (NRA, 1990). The graben covers an area

of approximately 20 km², and up to 50 m thick. Syn-tectonic sediments were deposited in the graben during the Late Cretaceous-Early Paleocene. The graben is marked by a wide topographic low consisting of downfaulted Muwaqqar Chalk Marl Formation overlain by Pleistocene gravels. Differences in the throw of the graben bounding faults have resulted in irregular graben floor that is sloped slightly westwards.

The Al-Jafr Basin is demarcated by two significant regional faults (Figure 2.3), the E-W striking dextral strike-slip Salawan Fault and the NW-SE striking normal Karak-Fayha Fault. The Karak-Fayha Fault is believed to be a surface form of a Precambrian rooted fault that experienced multi-stages of reactivation (Amireh et al., 2001; Naylor et al., 2013; Beik et al., 2017). According to the Alqudah et al. (2015), the Karak-Fayha Fault underwent a reactivation phase during the Maastrichtian-Paleocene and this led to the locally high rates of accumulation of oil shales in the downthrown block settings. In addition, Maastrichtian-Paleocene successions of about 130 m bituminous marls were deposited, indicating an active phase of the two faults during the Maastrichtian and Early Paleocene (Alqudah et al., 2015).

The Yarmouk Sub-basin, which is located in north Jordan (Figure 2.3), is associated with the fold system resulting by Syrian Arc system. Late Cretaceous compressional stress is indicated by the accumulation of Cretaceous oil shales within the basin (Alqudah et al., 2015).

The Dead Sea Transform Fault is composed of two main fault segments along the western border of Jordan: Wadi Araba Fault (south of the Dead Sea) and Jordan Valley Fault (north of the Dead Sea) (Figure 2.8). It also comprises other related transpressional and transtensional structural features, including pull-apart fault escarpments and pressure ridges (Garfunkel, 1981; Atallah, 1992; Ben-Avraham et al., 2008, 2010; Al-Awabdeh et al., 2016).

Folding occurred during the Late Cretaceous, related to the development of the Syrian Arc (Chaimov et al., 1992), but is not well developed east of the rift (Arabian Plate) as it is to the west within the Levant Plate (Cohen et al., 1990) (Figure 2.1).

The Dead Sea Transform Fault is characterized as the main locus of seismicity in the area, with strain fractionating by strike-slip and normal-fault earthquake focal mechanisms (Al-Tarazi et al., 2006; Hofstetter et al., 2007). Small-scale structures such as veins, joints and faults are widespread on both sides of the Dead Sea Transform (Eyal & Reches, 1983; Diabat et al., 2004); reflecting an active system of intraplate brittle deformation within the upper crust (Claps & Gross, 2016). These mesostructures have allowed the analysis of paleostress regimes (e.g., Angelier, 1979). Based on this analysis, two main paleostress regimes have been deduced since the Middle Cretaceous in different geographic locations in Jordan, including the Dead Sea stress field (Miocene-present) with NNW-SSE aligned Maximum Horizontal Stress (S_{HMAX}) (Figure 2.12), and the Syrian Arc stress field (Cretaceous to Pliocene) with WNW-ESE trending S_{HMAX} (Eyal, 1996; Eyal et al., 2001; Diabat et al., 2004, Diabat, 2009; Al Khatib et al., 2010) (Figure 2.13). The area was also subjected to SE-NW compression stresses through the Oligocene-Miocene that have contributed to the reactivation of pre-existing faults and weakness zones, and creating new fractures and discontinuities.

The Zarqa Ma'in Fault is an important structural element in the study area striking E-W (Figures 2.3 and 2.8). It extends from the Dead Sea for about 50 km to the East. This fault originally developed by extension, but was later reactivated in dextral shear (Diabat & Masri, 2002). The fault was also affected by the Neogene volcanic activity.

The E-W striking Suwaqa Fault (Figures 2.3, 2.4 and 2.8) crosses Jordan from the Sirhan Graben in the East to the Dead Sea in the west and represents the southern margin of the Azraq Basin (Diabat, 2009). The fault formed due to tectonic activities related to the opening of the Red Sea and the sinistral movement of the Dead Sea Transform (Diabat, 2006). Dextral movement occurred along the Suwaqa Fault during the anticlockwise rotation of the Arabian Plate. The fault was controlled by a stress field recognized by N-S to NNW-SSE compression and E-W to ENE-WSW extension, which is consistent with that known in the Middle Miocene-Recent and associated to the tectonic activity of the Dead Sea Transform (Ibrahim et al., 2014).

The area of central Jordan, which represents a large proportion of the study area, was mainly controlled by the extension of the Zarqa Ma'in and Suwaqa Faults (Obeidat & Rimawi, 2017) (Figure 2.3). It contains three main depocentres: the Harrana (the western part of the Azraq-Hamza Basin) and the El-Lajjun Basins (Central West) and the Azraq Basin (Central East) (Figure 2.3). The other part of the study area is situated in the Al-Jafr Basin to the south and the Sirhan Sub-basin (Southeast) (Figure 2.3).

The central and the southern regions of Jordan, the areas of main focus of the present study, were also restricted and their structural evolution was affected by a set of NW-SE regional faults, including Fuluq, Ramtha-Wadi Sirhan, Amra, Wadi Dabi, and Karak-Fayha Faults (Figures 2.3 and 2.8).

2.4 Stratigraphic Framework of the Belqa Group

The outcrop of Upper Cretaceous rocks covers almost the entire Jordan Plateau (Figure 2.3). The lower boundary of the Late Cretaceous sedimentation in the northern, western and

central parts of Jordan is marked by a marine calcareous facies association, while in the southeast, deposition of sandy sediments occurred in a continental environment (Bender 1974) (Figure 2.14). Bituminous strata are present at different stratigraphic levels in numerous areas. However, the Upper Cretaceous-Paleogene chalk-marl succession contains the highest bitumen content (Figure 2.14).

The low levels of deformation in the Cretaceous-Eocene succession in Jordan and adjacent areas, together with the continuous exposure of the succession and the massive subsurface data collected from boreholes distributed across the region, offer an outstanding information source of the development of depositional sequences in a three-dimensional geometry (Powell & Moh'd, 2011).

2.4.1 Lithostratigraphy

The Belqa Group of Jordan, which ranges in age from the Coniacian to the Eocene (Powell 1989; Powell & Moh'd, 2011), consists mainly of limestone, chalk, marl, phosphorite and bituminous marl (shale). The group comprises six geological formations (Figure 2.14; Table 2.1). These are (from older to younger); the Wadi Umm Ghudran Formation (WUG), the Amman Silicified Limestone Formation (ASL), the Al-Hisa Phosphorite Formation (AHP), the Muwaqquer Chalk-Marl Formation (MCM), the Eocene Umm Rijam Chert-Limestone Formation (URC), and the Wadi Shallalah Chalk Formation (WSC) (Parker 1970, Powell, 1989).

The Belqa Group is underlain by the Ajlun Group and overlain by the Qirna Calcareous Sandstone Formation (QSC) in the East (Figure 2.14). This study focuses on three formations from the Belqa Group. These are, in ascending order, the AHP, MCM and URC Formations, with occasional occurrence of the ASL (Figures 2.15, 2.16 and 2.17).

2.4.1.1 The WUG Formation (Coniacian-Lower Campanian)

The WUG Formation is mainly a chalky unit (Table 2.1). It is subdivided into three members, namely, from older to younger, the Mujib Chalk, the Tafilah Member and the Dhiban Chalk (Powell & Moh'd 2012). The three members can be observed throughout central Jordan (Parker 1970; Powell, 1988; Moh'd, 1992; Powell & Moh'd 2012). The formation top is recognised at the first massive chert bed of the basal ASL formation (Alsharhan & Nairn 2003). The lower contact is gradational and is defined where basal, detrital chalk of the WUG rests on the Ajlun Group carbonates.

The three-member sequence is an example of relatively deeper (chalk) and shallower (varied facies) water-depths in a pelagic, mid- to inner-ramp environment (Powell & Moh'd 2012).

2.4.1.2 The ASL Formation (Campanian)

The ASL Formation predominantly consists of chert, dolomite, limestone and phosphorite (Table 2.1). It is mainly subdivided into two units. A lower unit consists mostly of silicified limestones, and an upper unit comprises phosphorites. The lower unit contains bedded cherts, including brecciated chert (Figure 2.16b, c and e). The chert changes into a phosphatic chert in the upper part of the ASL sequence. Nodular chert is also present in this formation, in the upper part of the sequence (Figure 2.16d). The base of the ASL is defined underneath the first massive thick-bedded chert and above the marly chalks of the WUG Formation (Powell, 1989) (Figure 2.16e). The top boundary is gradational and is distinguished by the increase in phosphate and marl (Alsharhan & Nairn 2003). The maximum thickness of the formation is up to 70 m. The ASL hosts one of the local aquifers in Jordan (Abed, 2000; Powell & Moh'd, 2011). The formation crops out in Jordan from the north to the southeast. The age of this formation, based on the existence of *Baculites* cf. *ovatus*, has been determined to be

Campanian (Bender, 1974; Abed & Kraishan, 1991; Abed, 1994; Haggart, 2000; Pufahl et al., 2003) (Figure 2.14).

The ASL was mainly deposited in a subtidal to shallow shelf environment (Powell, 1988, 1989; Powell & Moh'd, 2011) and is locally characterized by small-medium scale syn-sedimentary folds (Bandel & Salameh, 2013).

2.4.1.3 The AHP Formation (Campanian-Lower Maastrichtian)

The AHP formation (Figure 2.17) is typically assigned to the Campanian-Lower Maastrichtian (Bender, 1974; Powell, 1989; Sadaqah, 2000). It comprises phosphorite, phosphoritic chert and limestone, fossiliferous limestone (coquina), bituminous marl, and chert (Table 2.1). The thickness of the formation is up to 70 m in Wadi Mujib area (Masri, 1996), and 60 m in the El-Lajjun area, central Jordan.

The formation is composed mainly of three members; a lower member (the Sultani Phosphorite Member), a middle member (the Bahiya Coquina Member) and an upper member (the Qatrana Phosphorite Member) (Figure 2.14). The Sultani Phosphorite Member consists of alternating phosphatic limestones, marl, phosphatic marl, and bedded phosphatic chert. The Bahiya Coquina Member, a distinguished marker unit of the Upper Cretaceous AHP in central Jordan, is composed of two main coquina beds (~7 m thick each in average) occurring between the phosphorite members. These beds are mainly composed of *in situ* oyster banks and level-bedded (off-bank) oyster accumulations (Powell & Moh'd, 2011). The Qatrana Phosphorite Member is about 10 m thick. It is the host unit for the high grade phosphorites in central Jordan and it is to some extent, bituminous. It consists mainly of phosphatic limestone, marly limestone, marl, phosphate, and phosphatic chert with concretions of chert and micritic limestone.

The lower contact of this formation is gradational and is perceptible by the difference in slope between the soft marl and chalky marl and the hard chert and limestone of the underlying ASL formation (Alsharhan & Nairn 2003). The formation top is defined at the base of the massive yellowish green chalky marl of the MCM.

The formation generally contains abundant bivalves, oysters, fish remains, and foraminifera, indicating a shallow-water pelagic depositional environment (Powell, 1989; Powell & Moh'd, 2011). The enrichment of phosphorite in this stratigraphic unit is an attribute to the major phosphorite deposition event that has taken place at the southern margin of the Tethys Ocean during the Late Cretaceous time (Fleurance et al., 2013).

2.4.1.4 The MCM Formation (Maastrichtian-Eocene)

The MCM Formation, which contains the richest oil shale (Total Organic Carbon up to 30 w%) interval in Jordan in its lower part, overlies the AHP and underlies the URC (Figure 2.16a). Generally, this lithostratigraphic unit is diachronous (Maastrichtian/Paleogene) (Alqudah et al., 2014a; Ali Hussein et al., 2015; Huggett et al., 2017) younging from SE to NW Jordan.

The formation consists mainly of chalky marl, marl, bituminous marl (oil shale), marly limestone, and chert (Figure 2.18). The thickness of this formation varies significantly over short lateral distances (Powell, 1989; Pufahl et al., 2003). The maximum thickness of the MCM at outcrop is up to 332 m in the Yarmouk area in the North (Parker, 1970), 100 m in Wadi Mujib area (Masri, 2010), 150 m in central Jordan (Barjous, 1986). Whereas, in the subsurface, the thickest interval of this formation is present in the Azraq-Hamza Sub-basin in central-east Jordan and it is mainly restricted to the centre of the sub-basin (Alqudah et al.,

2015). This typically comprises 200-300 m of bituminous marly limestones (Bender, 1975; Powell, 1989; Obeidat & Rimawi, 2017).

The typical MCM formation is composed of three main units. The upper unit consists of light greenish-grey marl and chalky marl intercalated with black chert beds. The middle unit is composed mostly of homogenous dark greenish-grey chalky marl and bituminous marl (oil shale) beds. The lower unit consists of interbedding of dark greenish-grey bituminous marl, marly limestone, phosphatic limestone and chert beds and nodules. The oil shale is normally associated with phosphorites (Abed & Sadaqah, 1998; Pufahl et al., 2003; Abed et al., 2007). Bituminous marl, phosphatic limestone and chert beds mark the boundary between the MCM and the AHP. The top is defined by the first appearance of the hard, chalky limestone of the URC formation (Andrew, 1992). Carbonate and chert concretions are also present within the three lithological units in the marl and marly limestone beds with high abundance in the lower and the upper units (Bender, 1975; Powell, 1989; Andrews, 1992).

The deposition of the MCM took place in a moderate to deep marine pelagic environment (Powell, 1989) (Figure 2.19). During the deposition of this formation, the region of Jordan witnessed a major marine transgression resulting in the outer continental shelf (Powell & Moh'd, 2011) and deposition of the MCM in a pelagic environment reflected by the abundant planktonic foraminifera especially to the North (e.g. Yassini, 1979).

The deposition of both the oil shale and phosphorites has taken place in a high bioproductivity system attributed to the paleo-upwelling on the southern, shallow epicontinental shelf of the Tethys Ocean (Abed et al., 2005; Lüning & Kuss, 2014). The deposition of the oil shale has occurred in basins where limited circulation of groundwater has led to hyper-salinity,

reducing conditions and good preservation of organic matter (Abed et al., 2005) (Figure 2.19).

2.4.1.5 The URC Formation (Eocene)

The URC Formation, the uppermost formation of the studied section (Figure 2.16a), is composed of alternating beds of chalky limestones, chalky marl, marly limestones, limestones, and abundant chert beds and nodules. The chalky marl and marly limestone beds are present in the lower part of the formation adjacent to the boundary with the underlying MCM formation. Carbonate concretions are abundant on the boundary with the MCM formation (Figure 2.20). Shell fragments, fish teeth, bone remains, and microfossils, such as plankton and foraminifera, are present in this formation. These litho- and bio-facies indicate a deep-water environment of deposition of the URC formation (Powell, 1989; Powell & Moh'd, 2011). Both the URC and the underlying the MCM formation are thicker in the north of Jordan, and they gradually decrease in thicknesses towards the south and southeast (Powell, 1989).

2.4.1.6 The WSC Formation (Middle-Upper Eocene)

The WSC age ranges from middle to upper Eocene (Figure 2.14). The formation is composed mainly of chalk with marl, glauconitic and micritic limestone (Figure 2.21). The formation is around 150 m thick. The main exposure of the WSC is in Wadi Al-Shumer to the north (Figure 2.21).

Correlations between the above described lithostratigraphic units show a general deepening from the Harrana Basin (Central West) to the Azraq Basin (Central East) and an increase in the thickness, particularly of the MCM formation, towards the east. This is due to the major

syn-sedimentary normal faulting (NW-SE) that led to the formation of a deeper depositional setting in the Azraq Area (Figures 2.4, 2.5 and 2.22). The significant variation in thickness over short distances suggests syn-depositional subsidence through the Maastrichtian and Paleogene times in fault-bounded basins (Alqudah, 2014).

2.4.1.7 Basalt rocks

Neogene-Quaternary basalts are exposed in four main places in Jordan, including the Jordanian Harrat basalt in North-eastern Jordan, which is a part of Harrat Al-Shaam and occupying an area of 11,400 Km² (Bender, 1974; Al-Malabeh, 1993; Al-Malabeh, 2003), Harrat Irbid and Umm Qais basalts, North Jordan (Abu-Mahfouz, 2009; Al-Malabeh, 2004; Abu-Mahfouz et al., 2016), the Dead Sea rift-associated basalts, and isolated basalt effusions in central Jordan (Al-Malabeh, 1993) (Figure 2.3). Basalts in the western and north-eastern part of Jordan cover some parts of the study areas. In the West, the basalts belong to the eastern margins basalts of Jabal Shihan volcano (Ibrahim et al., 2014), which is located approximately 80 km south of Amman. The north-eastern part is covered by a large part of Harrat Al-Shaam Volcanic field. The basalt flow is primarily composed of blocky, massive, and light grey basalt (Masri, 2010).

2.4.2 Lithostratigraphic Classification of the studied Section

To provide a stratigraphic framework for the present structural study and because of the complex diachronous and repetitive nature of the oil shale interval (Alqudah et al., 2015) within the studied succession, the four previously defined geological formations (ASL, AHP, MCM, and URC) are described, herein, from the cored interval as three units (Lower, Middle and Upper). The classification is mainly based on lithotypes and the presence (lower and

upper units) or absence (Middle Unit) of chert beds/nodules as well as the enrichment of organic matter in the marl beds (Figure 2.23):

Lower Unit: a heterogeneous unit composed of bituminous marls, chalky marls and phosphate, intercalated with chert and limestone beds and concretions. Chert concretions and bands are typically 90% quartz and the remaining 10% is calcite. Chert represents 15-20% of the rock volume of this unit. Bioturbation is moderate. This lithological unit corresponds with the AHP and the bottom part of the MCM (Figure 2.23).

Middle Unit: a homogenous unit consists of bituminous marls and chalky marls, with mainly calcite cement and microfossils cemented by calcite. A few horizons of carbonate concretions are also present in this unit. Quartz is very low within this unit. Bioturbation is high. This is the middle unit of the MCM (Figure 2.23).

In central Jordan, in Attarat Umm-Ghudran area, bituminous chalky marls in this unit show very low TOC values (<1wt%), and are described as thermally altered rocks that have undergone high temperature, low-pressure metamorphism (Fleurance et al., 2013), termed pyro-metamorphism (Grapes, 2006). The organic-matter content of this interval was endogenously combusted by hydrothermal solutions, forming marbles containing a suite of high-temperature minerals. This metamorphism is clearly observed in the core samples (see Chapter 3.4.1.1.7 for more detail).

Upper Unit: This unit mainly comprises intercalations of limestones, chalky limestones, chalky marls, silicified limestone, and chert as bands and concretions. Bioturbation is high. This unit matches the top part of the MCM formation and the entire URC formation (Figure 2.23).

2.4.3 Biostratigraphy of the Belqa Group

The age of the Belqa Group ranges from Late Coniacian to Late Eocene based on outcrop and subsurface data (Powell & Moh'd, 2011) (Figure 2.14). The group therefore spans the Cretaceous/Cenozoic boundary, which is marked by a depositional hiatus in the top part of the MCM (Yassini, 1979). The age of the lowermost chalk of the WUG was assigned to the Santonian (Parker, 1970; Bender, 1974). However, recent studies in the Negev and central Jordan suggest a Late Coniacian age for the basal chalk unit (Garfunkel, 1978; Reiss et al., 1985; Farouk et al., 2016). The uppermost unit of the Belqa Group, which is composed of marl, chalky limestones with chert concretions overlain by nummulitic wackestones and sandy carbonates at Dahikiyah (East Jordan) is considered to be Late Eocene in age (Bender, 1974; Powell & Moh'd, 2011).

Alqudah et al. (2015) have used calcareous nannofossils to define the biostratigraphic framework of the oil shale through the wells that are also the focus of the present study (Figure 2.24). They have assigned an Early Maastrichtian to late Middle Eocene age of these oil shales. However, they found, based on their locations across Jordan that the age range differs between wells. The southwest wells have purely Maastrichtian oil shales, whereas Paleocene-Early Eocene age was assigned to the southeast wells (Figure 2.24). The age of the Middle Unit Oil Shale in the wells located in east Jordan ranges between Maastrichtian-Early Eocene, while in the central Jordan wells this interval is Eocene in age, specifically ranging from Early to Middle Eocene (Figure 2.24). The absence of the Upper Cretaceous strata in central Jordan may be due to the wells not penetrating deeper because the drilling stopped at the base of Eocene or Paleocene (Alqudah, 2014). No evidence for any oil shales older than Early Maastrichtian in the studied succession was presented.

2.5 Sedimentation and Depositional History

During the Late Cretaceous-Paleocene, Jordan was positioned on a shallow marine carbonate platform, south of the Neo-Tethys Ocean (Barjous & Mikbel, 1990; Ziegler, 2001; Powell & Moh'd, 2011; Ali Hussein et al., 2015). Shelf carbonate deposition was dominant during this period (Figure 2.25). Sedimentation during this period was characterised by depositional sequences deposited in a hemi-pelagic/pelagic carbonate ramp setting (Powell & Moh'd, 2011, 2012). The entire succession represents a passive continental margin sequence comprising the Neo-Tethys margins (Powell & Moh'd, 2011). The depositional sequences vary from carbonate platform in the South and open shelf with hemi-pelagic mid- to inner-ramp facies associations in the central region to deep ramp and pelagic settings in the North. This gradation in depositional environments (Powell, 1989; Powell & Moh'd, 2012) resulted from slight variations in slope geometry on the carbonate ramp (Burchette & Wright, 1992).

In the Late Cretaceous (Coniacian-Early Campanian), sedimentation was characterised by a combined lithofacies association of chalk, chert and phosphorite, deposited in water depths range from shallow to moderate (ca. 30-100 m) in a ramp setting (Powell & Moh'd, 2012).

The remarkable change in sedimentation, from rimmed carbonate shelf to hemi-pelagic/pelagic ramp, is linked to the mid-oceanic rifting of Neo-Tethys, tilting, intra-craton deformation and subsidence of the platform (Sass & Bein, 1982; Robertson & Dixon, 1984; Flexer et al., 1986; Powell & Moh'd, 2011; Powell & Moh'd, 2012). All these events were connected with fluctuations in oceanic biogenic productivity and upwelling currents on the southern Neo-Tethys margin which, sequentially, resulted in hemi-pelagic and pelagic sedimentation (Almogi-Labin et al., 1993; Kolodny & Garrison, 1994; Powell & Moh'd,

2012). Oceanic upwelling and high productivity of organic matter led to the deposition of chalk, chert, marl and phosphate (Bender, 1974; Flexer et al., 1986; Pufahl et al., 2003; Powell Moh'd, 2012) (Figure 2.19).

The pelagic outer-ramp lithofacies association composed of chalk (calcareous nanoplankton, diatoms and planktonic foraminifera), locally with up to 50% detrital phosphatic skeletal fragments, with fish teeth and bivalve fragments at the base (Magaritz, 1974; Powell, 1989; Abed & Kraishan, 1991). In contrast, the hemi-pelagic mid- to inner-ramp association consists of intercalations of chalk, porcellanite, chalky marl, chert and phosphorite (Magaritz, 1974; Powell & Moh'd, 2012).

Marine conditions that dominated during the Late Cretaceous-Paleocene ended when the area experienced an uplift in the Early Oligocene (Alqudah et al., 2014b). This was followed by regional faulting concentrated mainly along reactivated faults and former weakness zones (Diabat & Masri, 2005; Obeidat & Rimawi, 2017). The wealth of subsurface data and the exceptional three-dimensional exposure of the Cretaceous-Palaeogene strata in Jordan allowed for the reconstruction of different sequences of deposition and linking of coeval events across the 'passive' continental margin (Powell & Moh'd, 2011).

Several factors have controlled depositional settings and stratigraphic architectures on the southern passive continental margin of Tethys. These include changes in climatically driven eustatic sea level (Haq et al., 1988; Sharland et al., 2001), the formation of the Neo-Tethys Ocean and its spreading centre (Figure 2.26) and the intraplate tectonic distortion of the Arabian Neoproterozoic Shield and the associated sedimentary rocks, to the South and East, which affected the flux of siliciclastic sediment (Powell & Moh'd, 2011).

Paleogeographic conditions, changes in accommodation potential, sediment and nutrient supply, organic productivity, and sea-water physical and chemical conditions have also influenced the development of the environments of deposition (Powell & Moh'd, 2011; Claps & Gross, 2016).

Climate has significantly influenced the depositional environments and controlled their development with time (Powell & Moh'd, 2011). Evidence from fossils suggests global high temperatures and high atmospheric CO₂ levels during the Cretaceous, with no substantial polar ice caps, indicating Greenhouse conditions. During the Late Cretaceous, a global cooling trend has been attributed to declining atmospheric CO₂ (Spicer & Corfeld, 1992; Jarvis et al., 2006). It is also deduced, based on reconstruction of plate tectonics that the Arabian Craton moved northwards from an equatorial position in the Cretaceous to ca. 20° north palaeolatitude by Maastrichtian (Scotese, 1991; Stampfli & Borel, 2002) (Figure 2.27). This suggests a possible low-moderate latitude, warm climate during the deposition of the Cretaceous succession in Jordan (Powell & Moh'd, 2011).

Jordan is positioned within the arid sub-tropical zone (Alsharhan & Nairn, 1997) (Figure 2.27) and is characterized by a varied climatic regime (Baaske, 2005). The climate in the northwestern parts is Mediterranean with a precipitation rate of approximately 300 mm per year (Bender, 1974). In the South and East the climate is arid, with annual precipitation less than 50 mm (Basaske, 2005).

Depositional sequences can be recognised with respect to global sea-level curves (Figure 2.28) for the Cretaceous (Haq et al., 1988; Sharland et al., 2001; Haq & Al-Qahtani, 2005) and sea-level curves for Levant (Flexer et al., 1986), northeast Egypt and parts of south

Jordan (Kuss, 1992; Bachmann & Kuss, 1998). The relative sea-level curves in Jordan displays features with changes to the global trend (Figure 2.28).

Generally high sea levels dominated during the Late Cretaceous, keeping most of the Jordan region flooded where the shallow marine sediments of the Belqa Group were deposited. During Cretaceous to Eocene, the Neo-Tethys Ocean has infrequently transgressed, in the period of the global sea-level rise over the margins of the Arabian Craton, in the South and East (Figures 2.26 and 2.28). Carbonate platform sequences (Albian-Eocene) can be traced in the southwest, almost matching the palaeoslope, to the Sinai Peninsula and the Eastern Desert of Egypt (Lewy, 1975; Kuss & Bachmann, 1996; Kuss et al., 2003; Bachmann et al., 2010; Farouk et al., 2016).

A major transgression occurred in the period between the Early Cenomanian and Early Paleogene, resulting in deposition of the marine carbonate sediments of the Ajlun and Belqa groups in Jordan (Powell, 1989) (Figure 2.28). During the Coniacian, a regional rise in sea level and marine transgression occurred. This led to drowning of the Cenomanian-Turonian rimmed carbonate-shelf by Late Coniacian times (Powell, 1989; Schulze et al., 2003; Powell & Moh'd, 2011, 2012).

The global trend of the Late Coniacian-Maastrichtian is one of rising global sea level and marine onlap. During the Campanian, the regression on the inner ramp seems to match, to some extent, with a lowstand in Early Campanian, but not with the sea-level rise suggested for the Late Campanian (Powell & Moh'd, 2011). The sea-level during the Maastrichtian shows an overall regressive trend, with two major regressions during the Middle and the Late Maastrichtian (Baaske, 2005). Moreover, marine onlap characterised by the Maastrichtian

pelagic transgressive sequence (MCM) was deposited during global sea-level falling (Haq et al., 1988).

The Late Paleogene witnessed a final regression and closure of the Neo-Tethys Ocean related to Alpine earth movements in the Syrian Arc, including initial stages of Dead Sea rifting, as the African-Arabian Plate moved northwards against the Eurasian Plate (Scotese, 1991; Stampfli & Borel, 2002; Powell & Moh'd, 2011). Regression of the Neo-Tethys Ocean took place in Late Eocene and was characterised by shallow-water nummulitic shoals and heterolithic, siliciclastic/carbonate facies. Variations in sea level between the Oligocene and present day are estimated to have reached up to 150 m (Haq et al., 1988).

In general, the studied geological succession represents the development from a shallow-water ramp with mostly carbonate deposition (ASL) into a gradually deeper open shelf (AHP and MCM), culminating in a shallower shelf progradation (uppermost MCM). This depositional setting was ultimately drowned and overlain by the deeper deposits of the hemipelagic URC formation, due to a relatively regional-scale rise of the sea level (Powell & Moh'd, 2011).

The MCM formation which is the host of the richest oil shale unit (Figure 2.18), was deposited in an enormous and uniform moderate-deep marine pelagic environment (Powell, 1989) with the recurrent alternation of chalky marlstones and chalky limestones (Figure 2.19). The bituminous marl and marly limestone units deposited in anoxic basins in the deepest parts of graben structures (Ali Hussein et al., 2015; Alqudah et al., 2015; Hakimi et al., 2016; Hakimi et al., 2018). This mainly took place at the base of a density-stratified

watermass in which reducing, hypersaline bottom conditions were characterised by high algal activity (Speers, 1969; Spiro et al., 1983).

It has been thought that the large open shelf system was characterized by the development of locally subsiding basins, such as the El-Lajjun Sub-basin in central Jordan (Abu Ajamieh, 1980; Abed et al., 2005). The basins were most likely fault-bounded, where restricted circulation, richness of nutrients, and a stratified watermass favoured the accumulation of organic-rich deposits and their subsequent preservation (Figure 2.19). However, these basins may also be created by gentle, syn-formal folding ('basin and swell') related to intraplate compressional deformation of the Arabian Craton (Powell & Moh'd, 2011) (Figure 2.25). Surface water has low nutrient contents, owing to removal by biota, but deep water is usually rich in nutrients; therefore, restricted basins act as a trap for nutrients. This explains the high productivity levels (high TOC) and very high concentrations of trace metals (Aqleh, 2015; Huggett et al., 2017).

2.6 Diagenesis

Diagenetic processes played an important role in the Cretaceous-Paleogene strata of Jordan (Powell & Moh'd, 2012; Huggett et al., 2017; Hooker et al., 2017a). The presence of chert beds and nodules and displacive carbonate concretions (see Chapter 4 for more details) with local dolomitization within the succession indicate that diagenetic processes already occurred in an early stage.

Different studies (e.g., Reiss et al., 1985; Soudry et al., 1985; Kolodny & Garrison 1994; Powell & Moh'd, 2012) have interpreted the formation of carbonate concretions in the lowermost part of Belqa Group and its relevant unit on the other side of the Dead Sea to have

resulted from early diagenesis of an original siliceous/carbonate ooze. Chert in the section indicates that the parent sediment was possibly fine-grained calcareous/siliceous ooze, wherein the carbonate was then replaced by silica during early phase of diagenesis. The high productivity of silica in restricted basins may be referred to deep-water inflow, and surface-water outflow (Ali Hussein et al., 2014b; Huggett et al., 2017), and to upwelling on the Arabian Platform (Powell & Moh'd 2011).

It is thought that biogenic opal dissolves in sediments and, when conditions become suitable, opal-CT forms from pore-waters before it dissolves and is re-precipitated as microcrystalline quartz (Behl & Garrison, 1994; Kolodny & Garrison 1994; Powell & Moh'd, 2012). However, the occurrence of brecciated laminated chert fabrics in the ASL, suggests a development of an intermediate phase in which a gelatinous precipitate forms before the crystallization of opal-CT (Bener, 1971). Many studies support the idea that the process of brecciation of Campanian cherts in the region, equivalent to the ASL chert, occurred during early diagenesis (e.g., Steinitz, 1981; Fink & Reches, 1983; Powell & Moh'd, 2011, 2012).

Biogenic-silica transition from opal-A to opal-CT and then to quartz can produce volumetric changes in semi-lithified silica or carbonate sediments (Kastner et al., 1977; Behl & Garrison, 1994), including deformation structures (Eichhubl & Behl, 1998; Davies et al., 1999; Davies, 2005; Hooker et al., 2017a). It has been argued by Hooker et al. (2017a) that silica diagenesis has led to fracturing through the Upper Cretaceous-Paleogene succession. They suggested that volume redistribution or loss attributed to the opal-quartz transformation (OpalA-OpalCT-Quartz) during early burial have controlled the formation of fractures so that chemical reactions can result in development of fracture pore space that can be filled by diagenetic minerals (Hooker et al., 2017a). See chapter 3 for more detail.

2.7 Regional Controls on Oil Shale Deposition

During the Maastrichtian-Paleogene, the geology of most of the Middle East region (including Jordan, Palestine, Syria, Lebanon, Turkey, Iraq, Egypt, and Saudi Arabia) was influenced by three main tectonic events: the Syrian Arc, the Al-Jawf rifting and the Alpine Orogeny (Alqudah et al., 2015) (Figure 2.6). Maastrichtian oil shale deposits are widespread in many of these countries (Figures 2.29 and 2.30) and are documented in many studies: in Syria and Iraq (Jassim & Gailani, 2006), in Egypt (El Kassas & Khaled, 1997) and in Jordan (Yassini, 1979). This wide distribution of oil shales during the Maastrichtian time is believed to be associated to the Syrian Arc controlling the regional sea level and the oil shale deposition in the Epicontinental Sea along southern Neo-Tethys margin (Alqudah et al., 2015). A major hiatus was caused by the Rutbah Uplift that occurred in Iraq and east Jordan in the Late Maastrichtian (Jassim & Gailani, 2006) (Figure 2.29).

During the Paleocene, the deposition of oil shale continued in Lebanon, Syria and Jordan, controlled by the Al-Jawf rifting that affected the regional topography (Alqudah et al., 2015). Several active-rifting events have influenced the Sirhan Sub-basin and the Al-Jawf Graben (Aoudeh & Al-Hajri, 1994). In the Paleocene, the rift underwent invigoration, forming the sub-basins, and uplift of other blocks (Figure 2.6). The sequences in Syria, Iraq, central east Jordan and Qatar show sedimentation breaks during the Early Danian (Alsharhan & Nairn, 2003).

During the Early Eocene, the deposition of oil shale was enhanced toward the east of Jordan with rising sea level (Jassim & Gailani, 2006; Alqudah et al., 2015). During the Middle

Eocene, deposition at regional scale can be documented in central Jordan, Syria and Iraq (Alqudah et al., 2015).

2.8 Summary

This chapter has summarised the geological setting of Jordan and shed the light on the Late Cretaceous-Eocene time. During this time interval, the area of Jordan was part of a shallow epicontinental sea where a shallow-water carbonate platform generally hosted the deposition. Arabian Plate tectonism influenced sedimentation and structural development in the area. The Upper Cretaceous-Eocene Belqa Group reflects a general progressive, marine onlap, disrupted by short-term, regressive lowstands. The JOS which belongs mainly to the MCM Formation was deposited in a moderate to deep marine pelagic environment. Diagenetic processes were also thought to have taken place during the early burial phase based on the existence of chert beds/nodules, displacive carbonate concretions and local dolomitization within the studied lithostratigraphic succession.

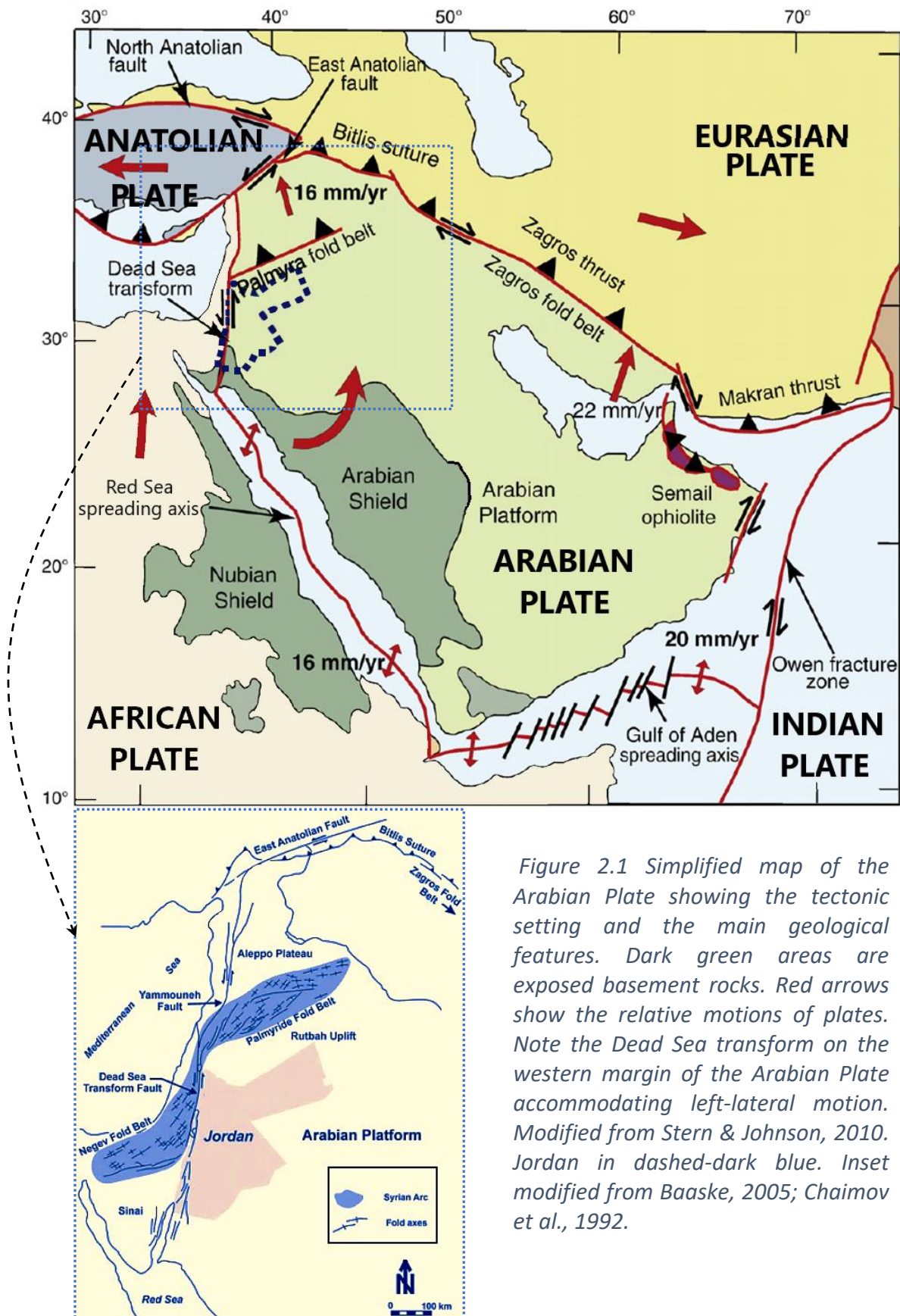


Figure 2.1 Simplified map of the Arabian Plate showing the tectonic setting and the main geological features. Dark green areas are exposed basement rocks. Red arrows show the relative motions of plates. Note the Dead Sea transform on the western margin of the Arabian Plate accommodating left-lateral motion. Modified from Stern & Johnson, 2010. Jordan in dashed-dark blue. Inset modified from Baaske, 2005; Chaimov et al., 1992.

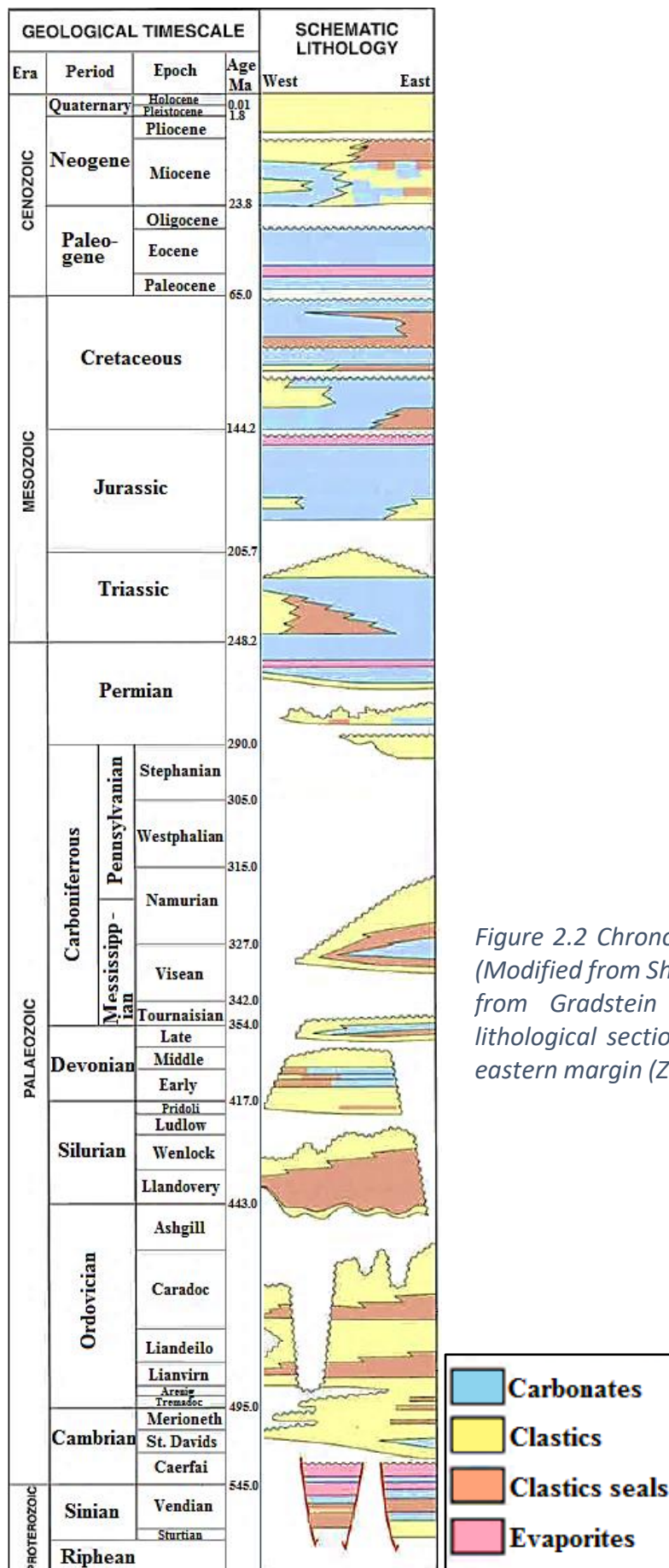


Figure 2.2 Chrono-stratigraphy of the Arabian Plate (Modified from Sharland et al., 2001). Absolute age is from Gradstein & Ogg (1996). The schematic lithological section runs from central Arabia to the eastern margin (Zagros) and provides gross lithology.

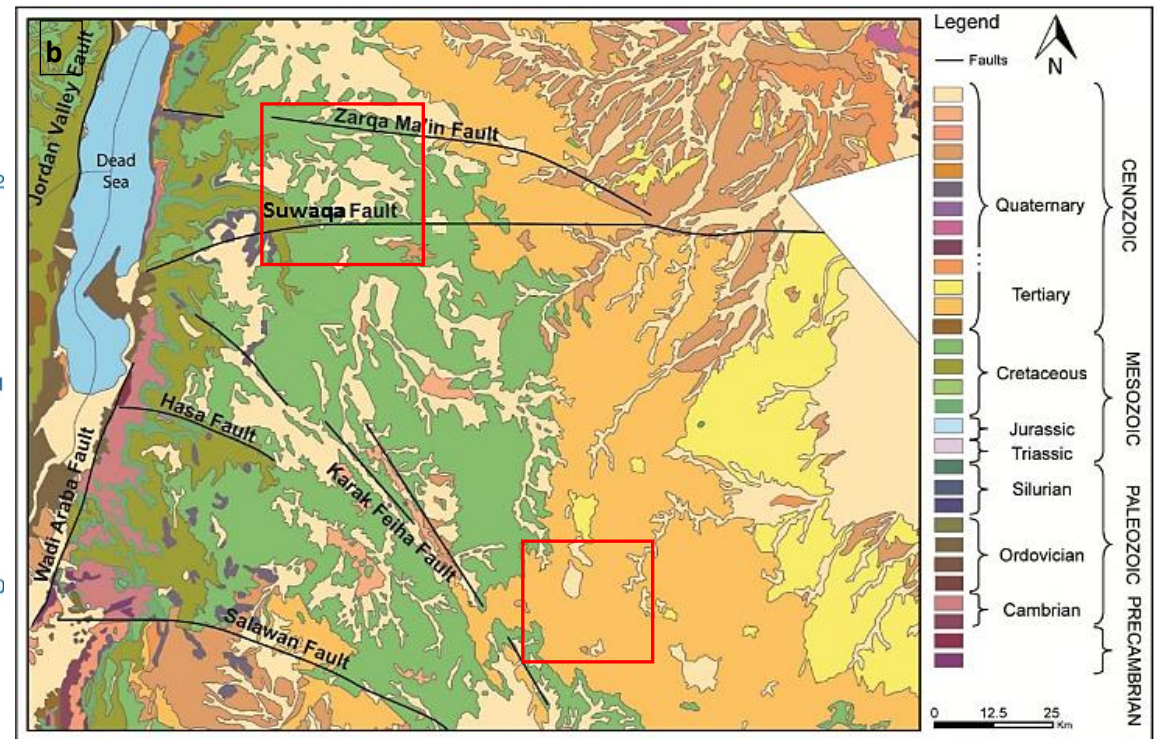
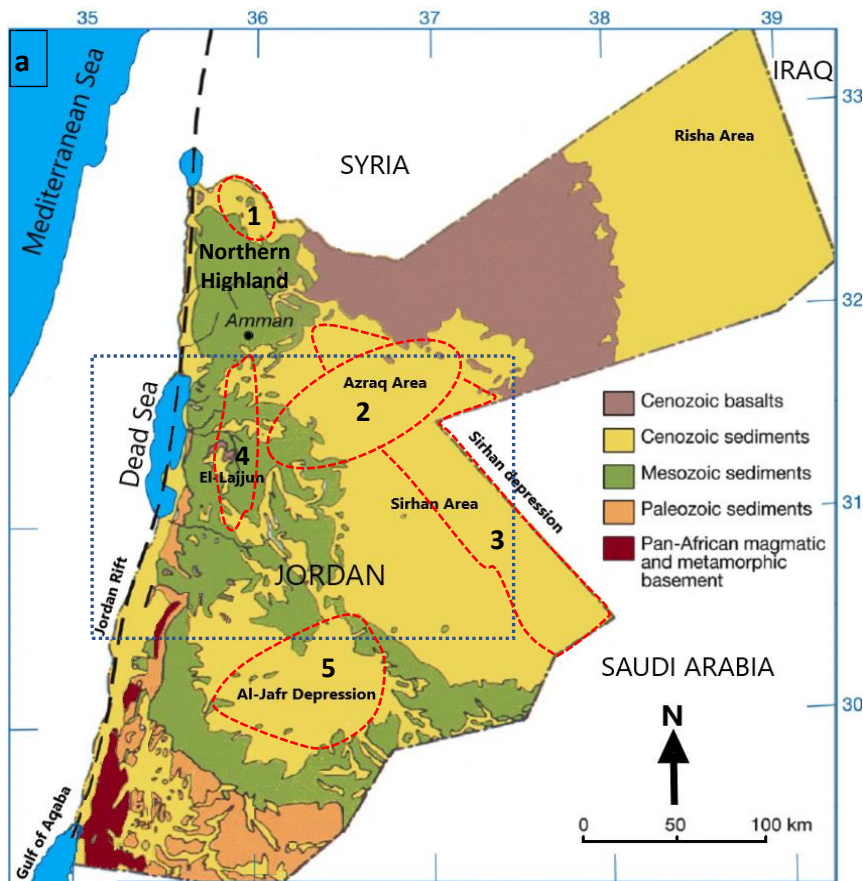


Figure 2.3 a. Simplified geological map of Jordan. Modified from Förster et al., 2010. Red-dashed shapes outlines the main sedimentary basins (1: Yarmouk Basin, 2: Azraq-Hamza Basin, 3: Sirhan Basin, 4: El-Lajjun Basin, 5: Al-Jafr Basin). Sub-basins outline after Alqudah et al., 2015. b. A zoom-in of the blue-dotted rectangle outlined in (a) showing the geology and main structural elements. Modified from Förster et al., 2010; Fleurance et al., 2013. Red squares show the outlines of Figure 2.13a (central Jordan) and b (south Jordan).

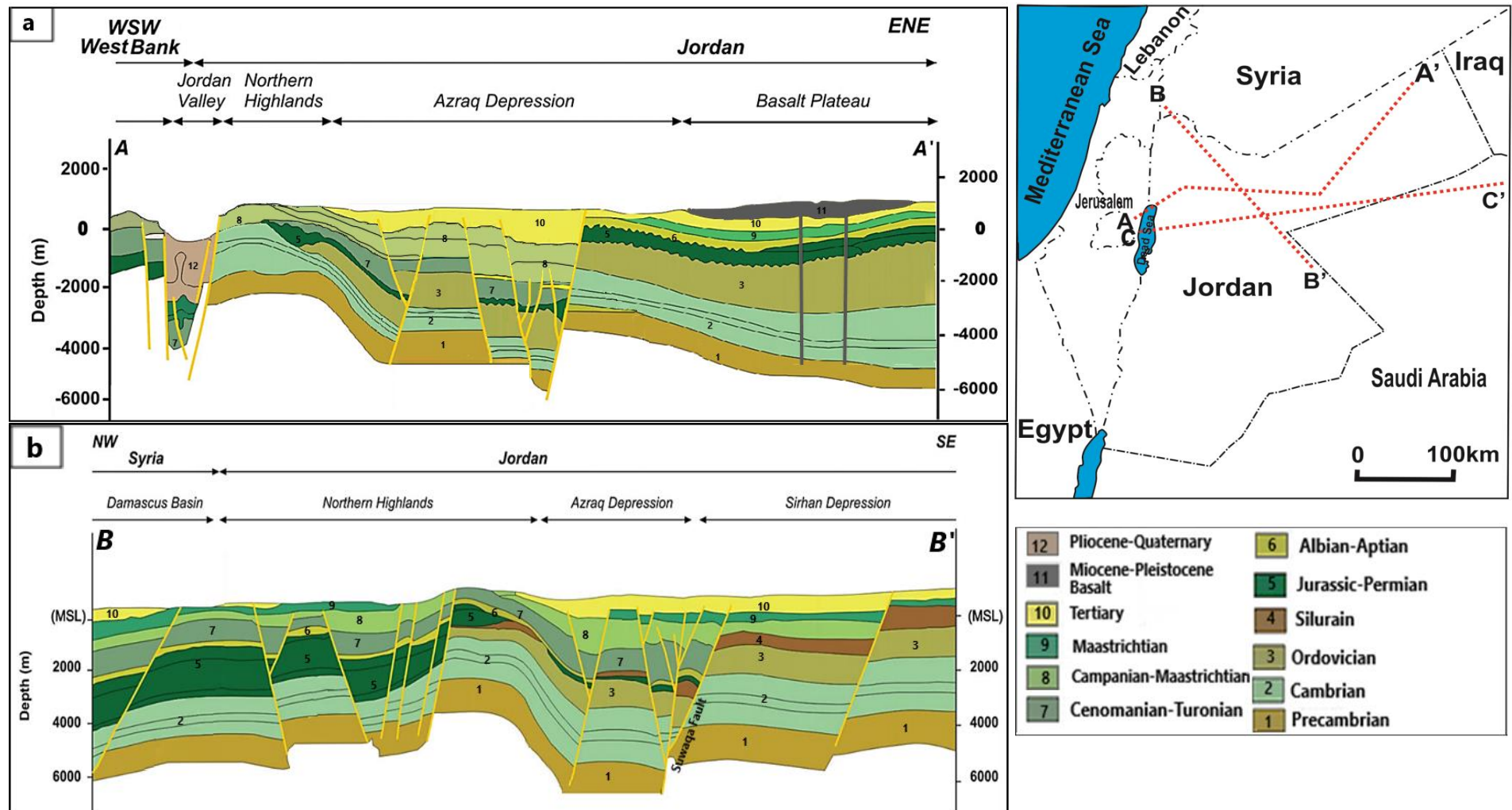


Figure 2.4 Schematic E-W structural cross section section. b. Schematic N-S structural cross section section. Modified from Lüning & Kuss, 2014 and Natural Resources Authority of Jordan (NRA), 2001.

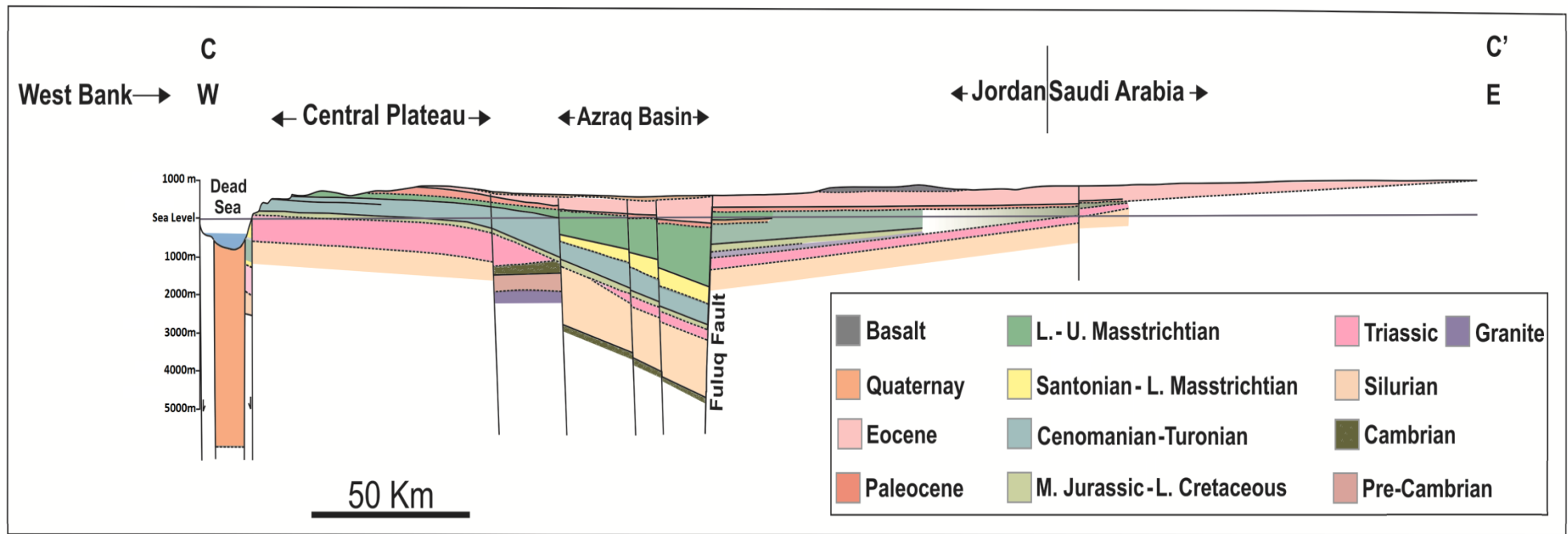


Figure 2.5 A schematic E-W cross section. Modified from Alder, 1985. See the map in Figure 2.4 for the cross-section outline

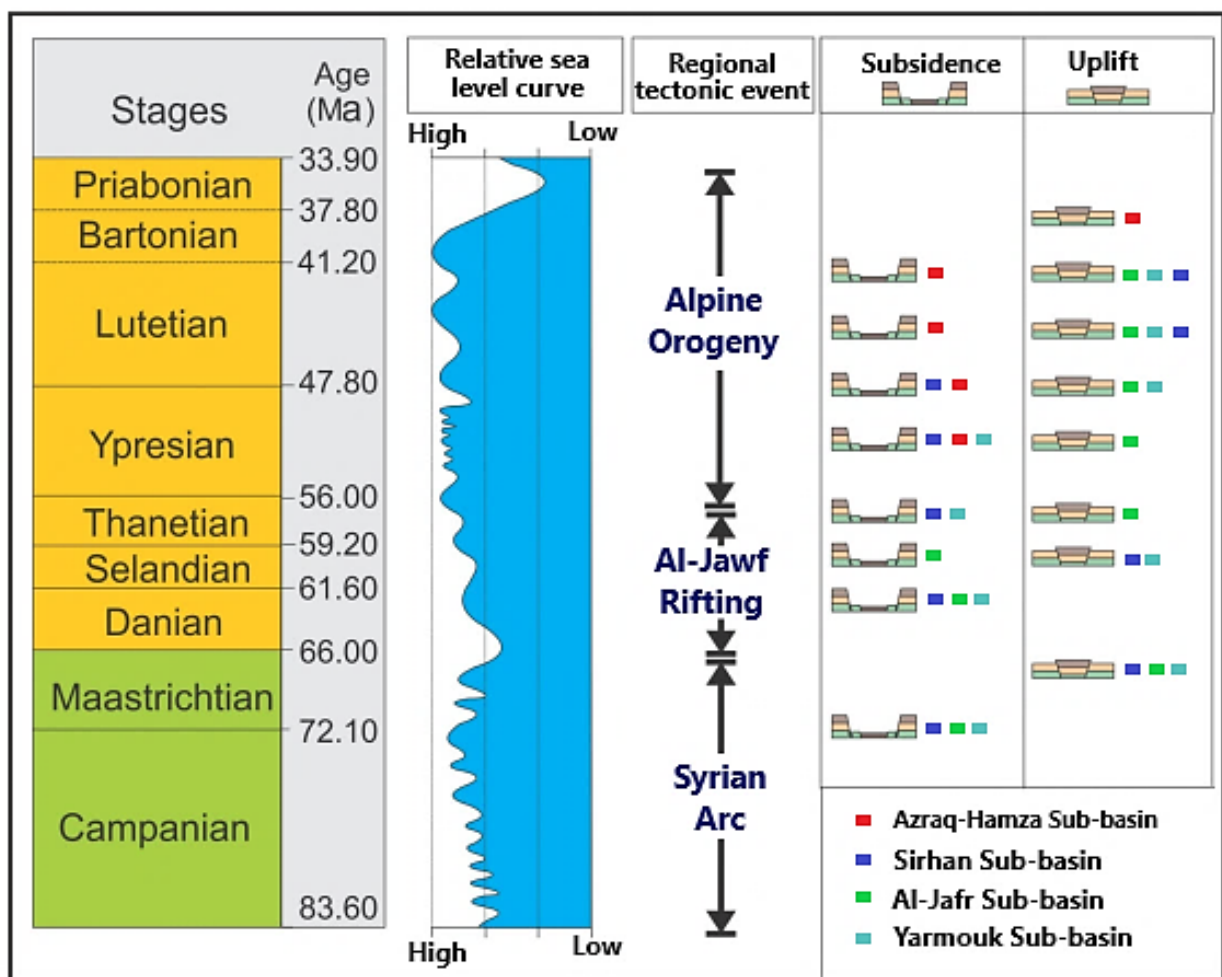


Figure 2.6 A summary of the tectonic events controlled the region during the Late Cretaceous-Paleogene time. Modified from Alqudah, 2014.

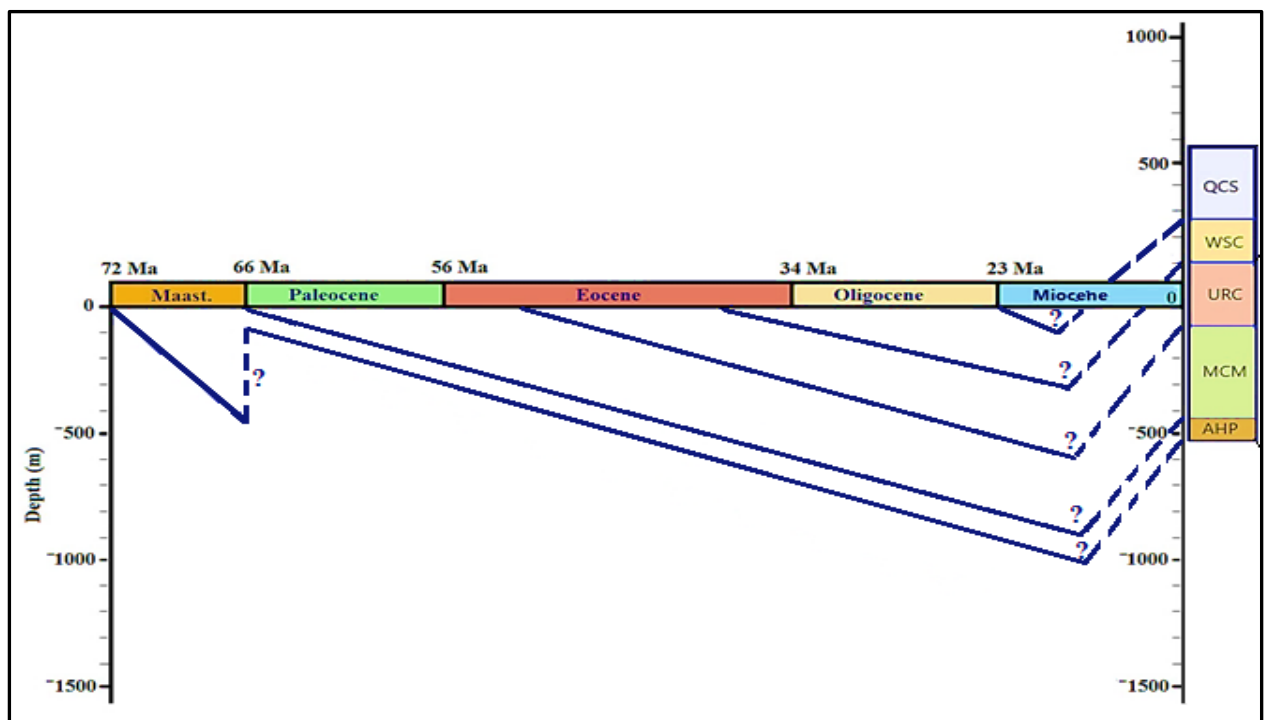


Figure 2.7 A schematic burial depth curve of the Upper Cretaceous-Eocene succession in central Jordan.

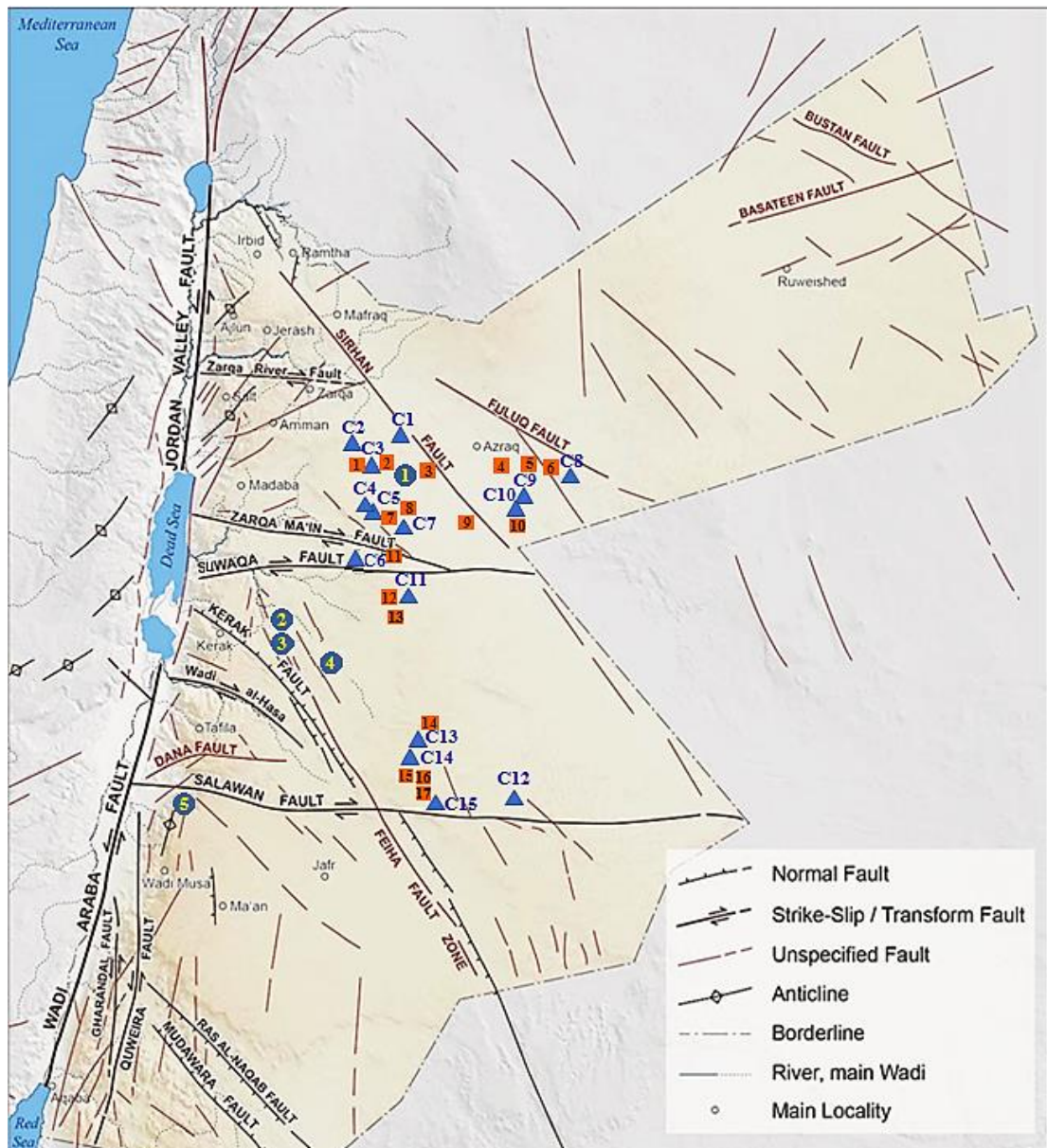


Figure 2.8 Structural map of Jordan, showing the main structural elements in the region and the locations of the study wells (orange squares; 1-17), cored wells (blue triangles) and outcrops (blue circles: 1, Harrana; 2, EL-Lajjun Quarry; 3, Qatrana Roadcut; 4, Sultani Quarry; 5, Wadi Al-Bustan). Modified from Diabat, 2012.

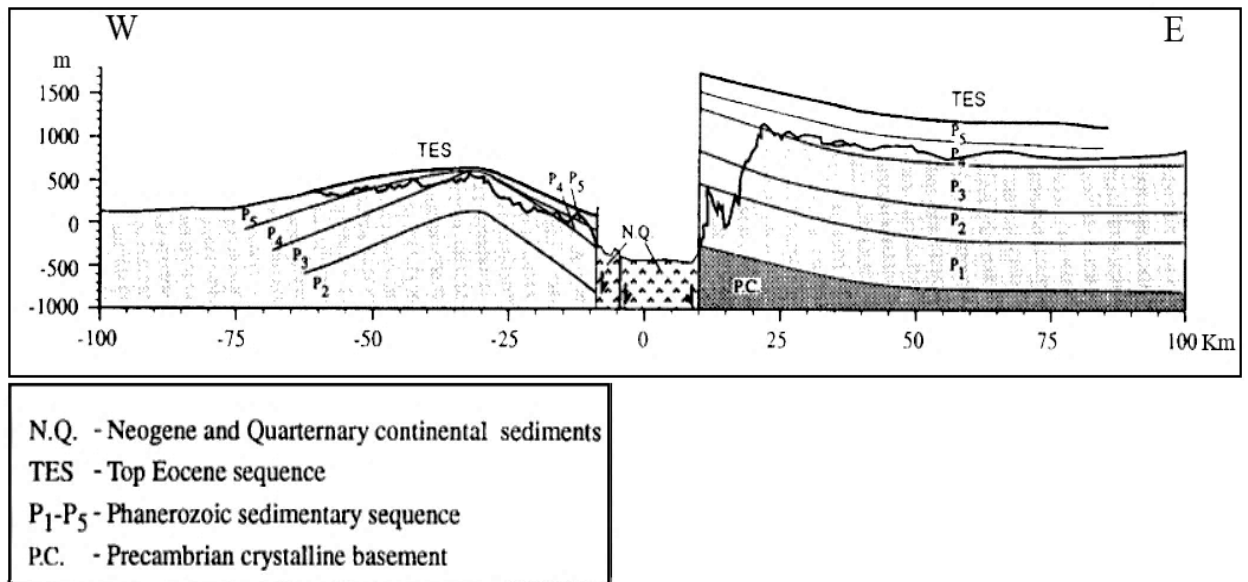


Figure 2.9 A schematic geologic section across the Dead Sea Rift showing the large-scale asymmetric structure across the rift. Modified from Wdowinski & Zilberman (1996).

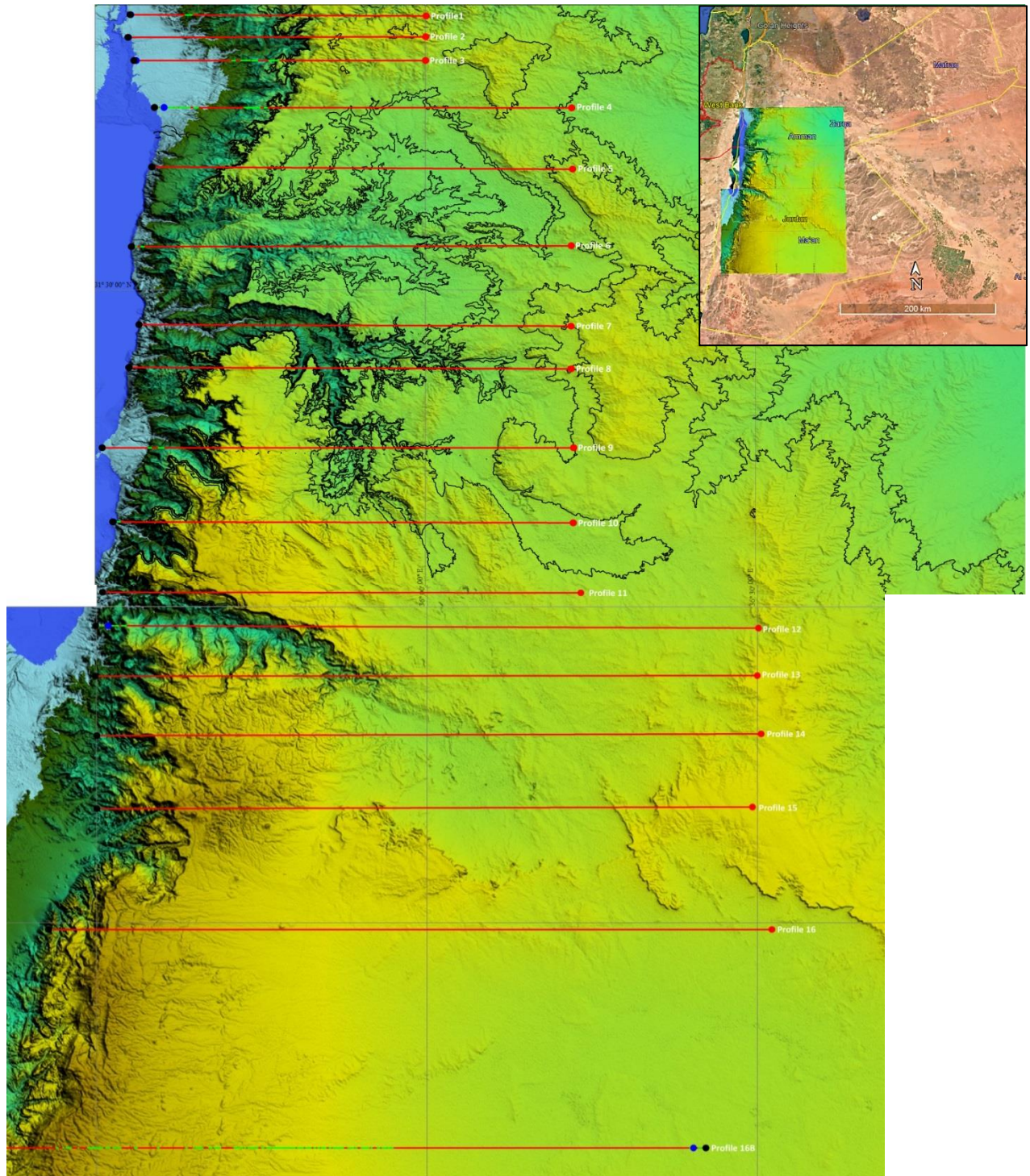
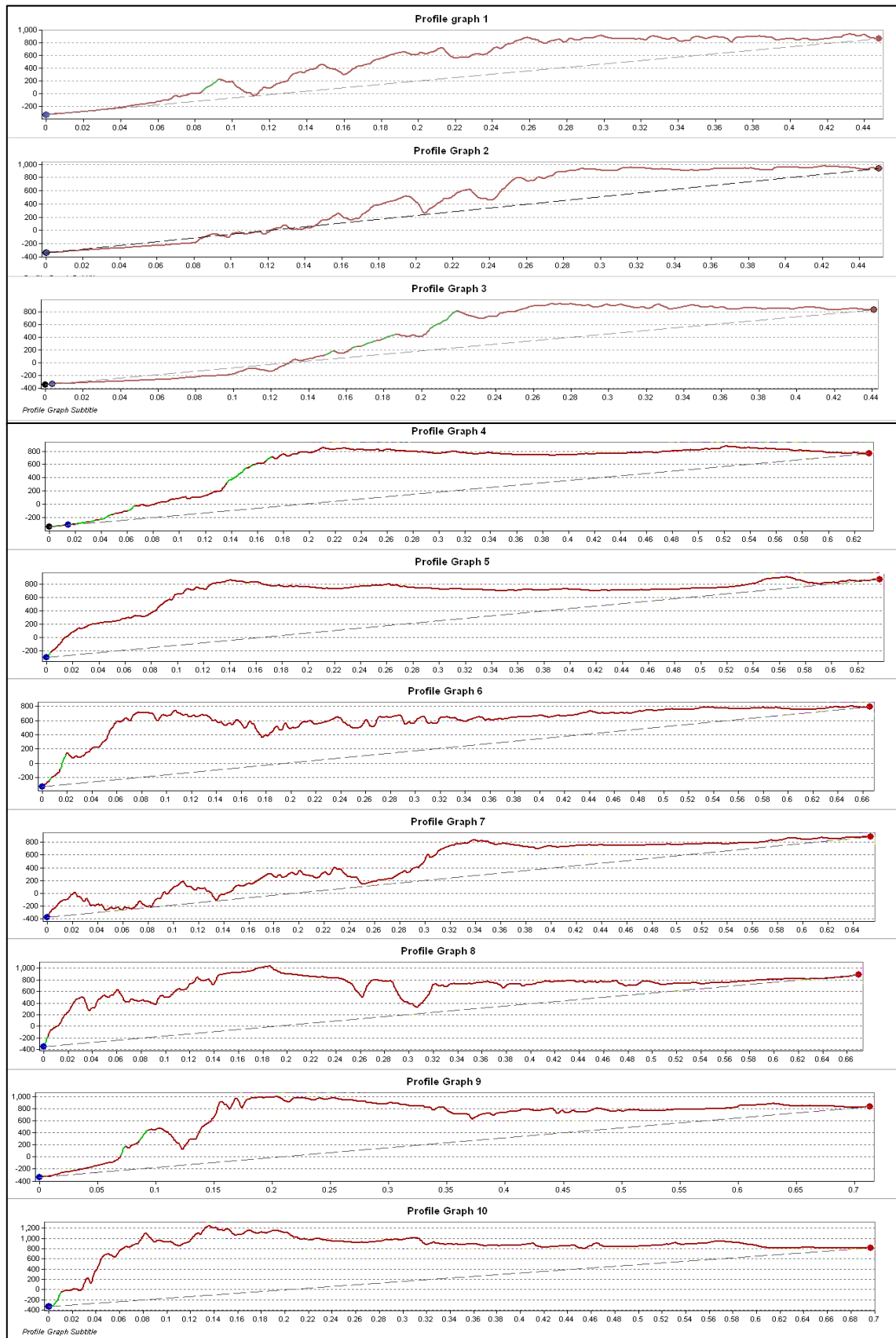


Figure 2.10 Digital Elevation Map (DEM) of the areas of focus. E-W red lines outline the elevation profiles presented in Figure 2.11.



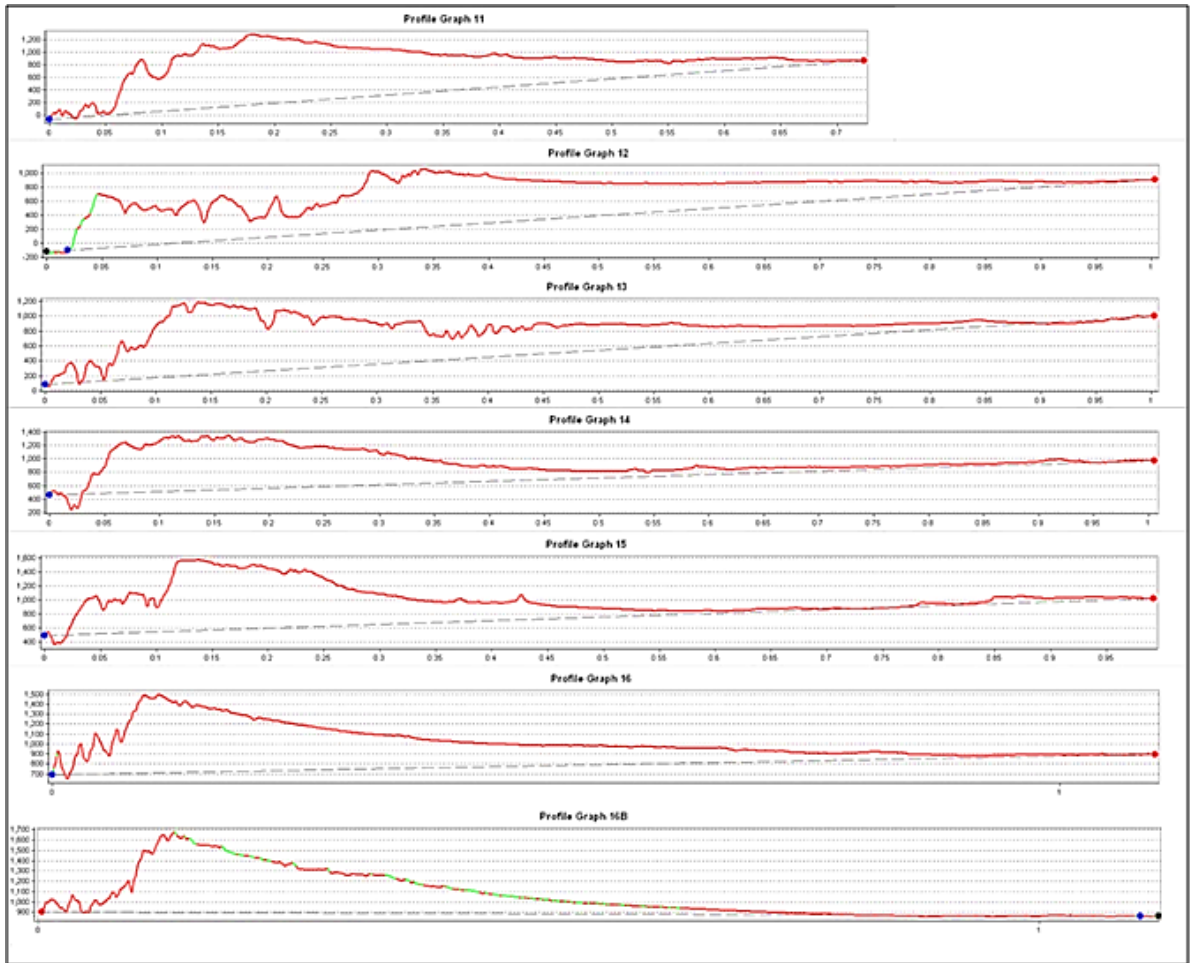


Figure 2.11 Elevation profiles of the areas shown in Figure 2.10. See previous page for all profiles.

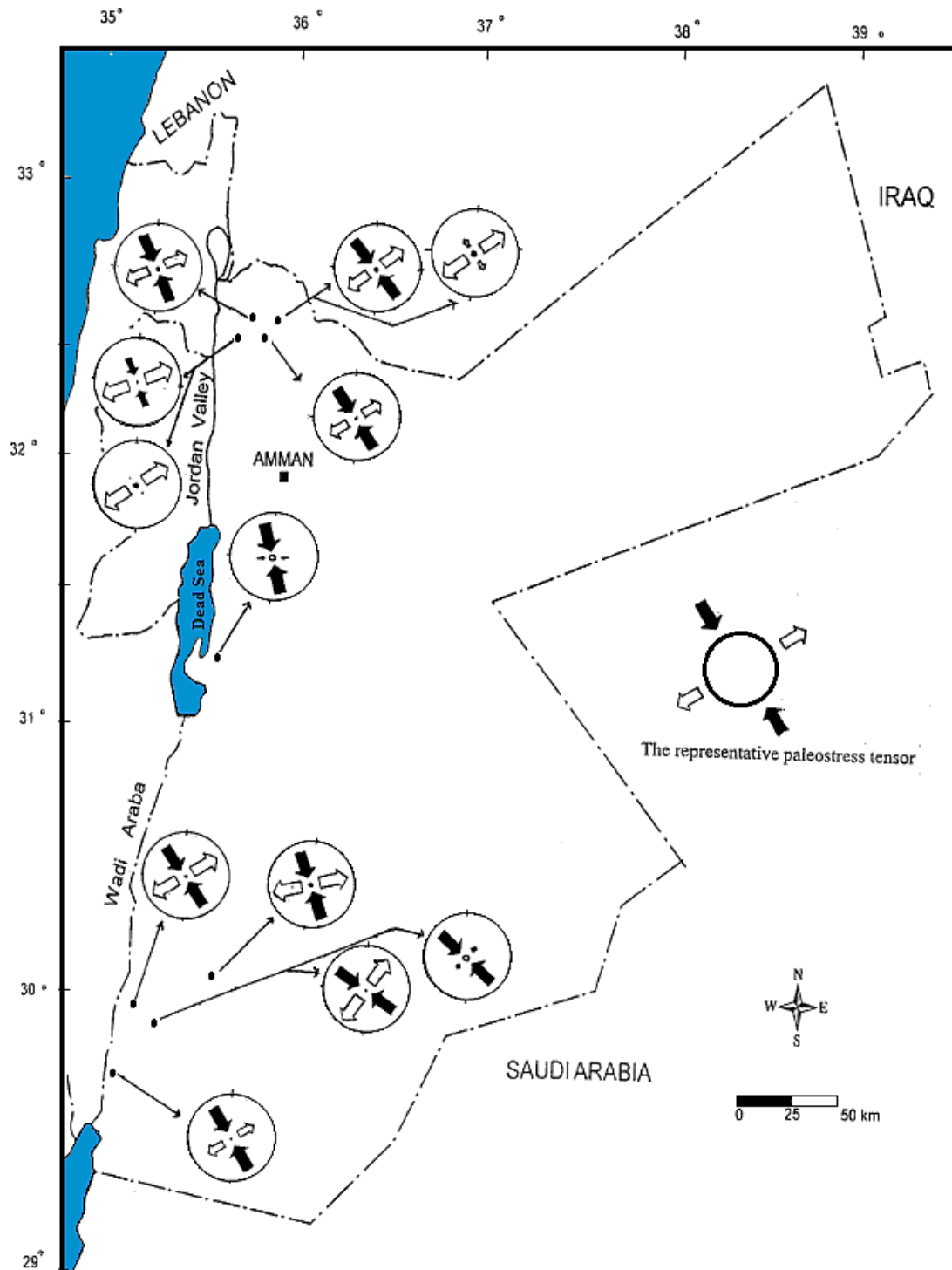


Figure 2.12 Distribution of the stress field for the Cretaceous rocks at the eastern edge of the Dead Sea transform in Jordan. The stress field is characterized by NNW–SSE compression (black arrow: σ_1) and ENE–WSW extension (white arrow: σ_3). It is related to Middle Miocene–Recent sinistral movement along the Dead Sea Transform and the opening of the Red Sea. Modified from Diabat et al., 2004.

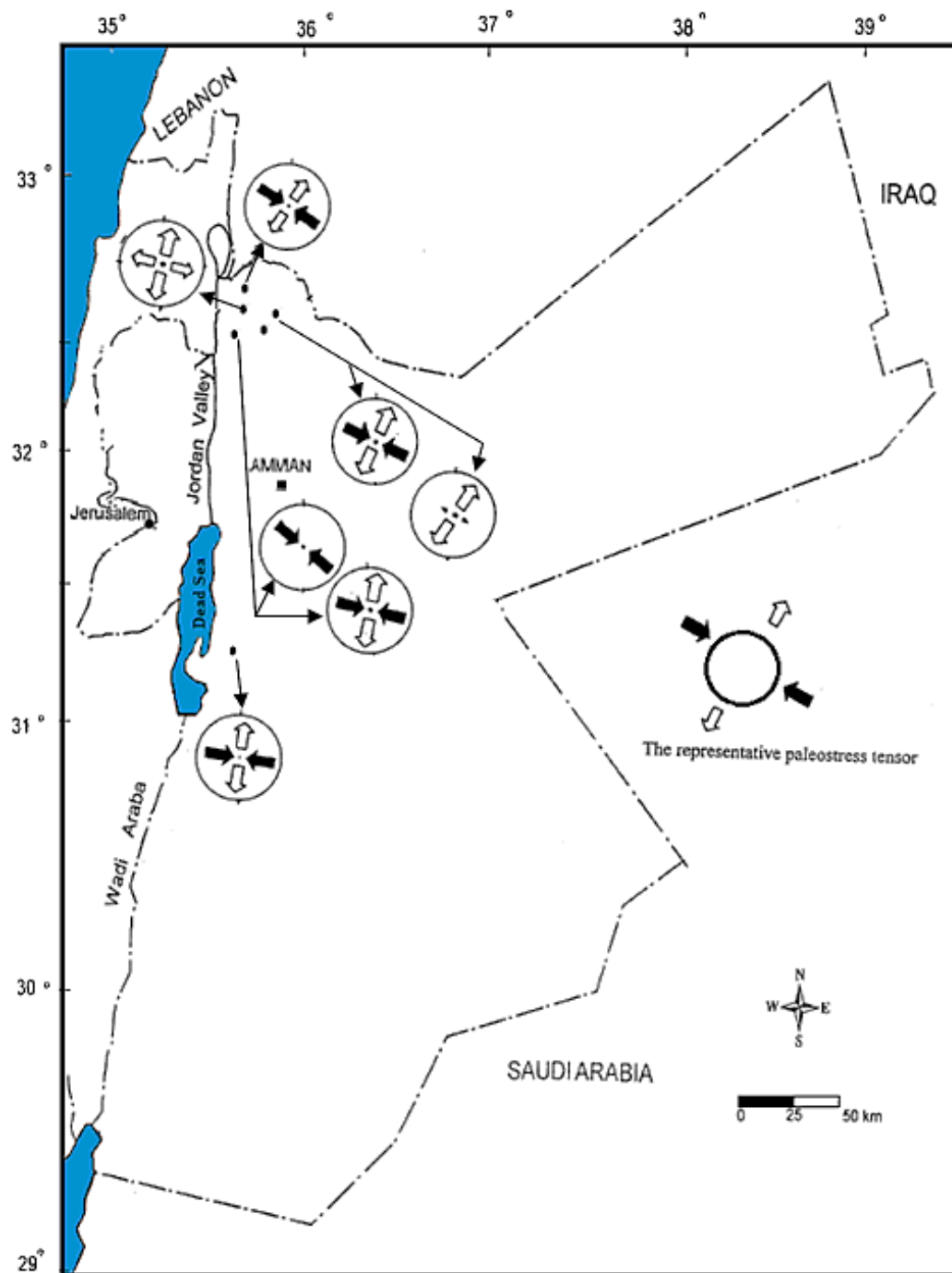


Figure 2.13 Distribution of the stress field for the Cretaceous rocks that is characterized by WNW–ESE compression (black arrow: σ_1) and NNE–SSW extension (white arrow: σ_3), and is mainly restricted to the northern part of Jordan. Modified from Diabat et al., 2004. This stress field is related to the development of the Syrian Arc fold belt.

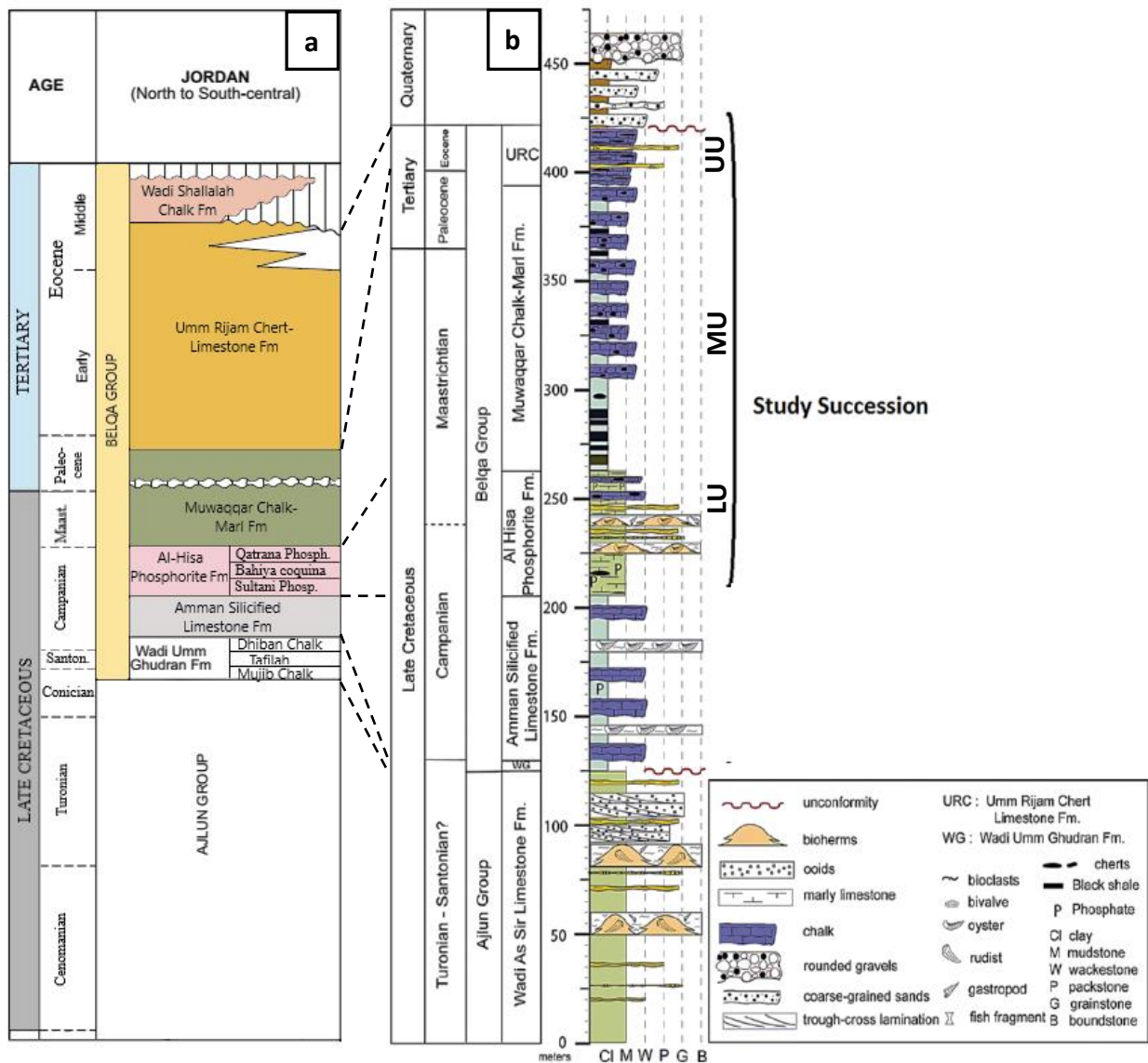


Figure 2.14 a. Lithostratigraphy of the Upper Cretaceous and Paleogene in Jordan from north to south (modified from Powell, 1989, and Powell & Moh'd, 2011). b. Chrono-lithostratigraphy of Central Jordan. Modified from Fleurance et al., 2013. LU, Lower Unit; MU, Middle Unit; UU, Upper Unit.

Chronostratigraphy		Lithostratigraphy		Lithology (Fossils)	Depositional environments
Quaternary	Holocene			Rounded to well-rounded gravels, sands and silts	Recent alluvial, wadi sediments (present day drainage network)
	Pleistocene			Travertine, poorly sorted angular to rounded clasts gravels (mainly cherts), sands and silts	«Hot-spring», alluvial, mudfats
Tertiary	Eocene		Umm Rijam Chert- Limestone Fm.	Chalky limestone, thin-bedded chert, packstone, grainstone, sparse phosphate (Fish fragments, gast ropods, nummulites, calcareous nanoplankton, silicoflagellates)	Deep-water pelagic, locally hypersaline, locally shallow lagoons
	Palaeocene		Muwaqqar Chalk-Marl Fm.	Chalk, bituminous-organic-rich (more than 1% TOC)- chalky marl, limestone and chert concretions (Bivalves, ammonites, gastropods, fish fragments, calcareous nanoplankton)	Deep-water pelagic ramp (locally small anoxic basins with density stratified water column)
Cretaceous	Maastrichtian				
	Campanian	Belqa Group	Qatrana Phosphorite Mb.	Phosphate, chert, chalk, chalky marl, oyster-rich coquina and bioherms (Bivalves including large oysters, gast ropods, fish and reptile fragments, ammonites, silicoflagellates, radiolaria, calcareous nanoplankton, foraminifera)	Shallow-water pelagic ramp, oyster bioherms on inner ramp, stagnant bottom: winnovated phosphorite
			Bahiyya Coquina Mb.		
			Sultani Phosphorite Mb.		
		Al Hisa Phosphorite Fm.	Amman Si licified Limestone Fm.	Chert, microcrystalline limestone, chalky marl, sparse phosphorite, oyster-rich coquina (Bivalves including oysters, sparse ammonites, fish and reptiles fragments, silicoflagellates, radiolaria)	Shallow-water, pelagic inner ramp
			Wadi Umm Ghudran Fm.	Chalk, thin-bedded chert, sparse phosphate, dolomitic grainstone (Bivalves, gastropods, fish fragments, calcareous nanoplankton)	Pelagic ramp (mid- to inner), shallow shoreline
	Turonian	Ajlun Group	Wadi As Sir Limestone Fm.	Micrite, wackestone, packstone, sparse ooidal grain stone, chert nodules (Bivalves, rudists, gastropods, foraminifera)	Rimmed shallow carbonate shelf, rudist patch reefs shoaling inner shelf

Table 2.1 A summary of stratigraphic nomenclature: lithostratigraphy, lithology and depositional environments of the studied succession in Central Jordan. Fm.: Formation, Mb: Member. Modified from Fleurance et al., 2013 (data compiled from Abed, 1982; Abed & Kraishan, 1991; Powell & Moh'd, 2011; Pufahl et al., 2003).

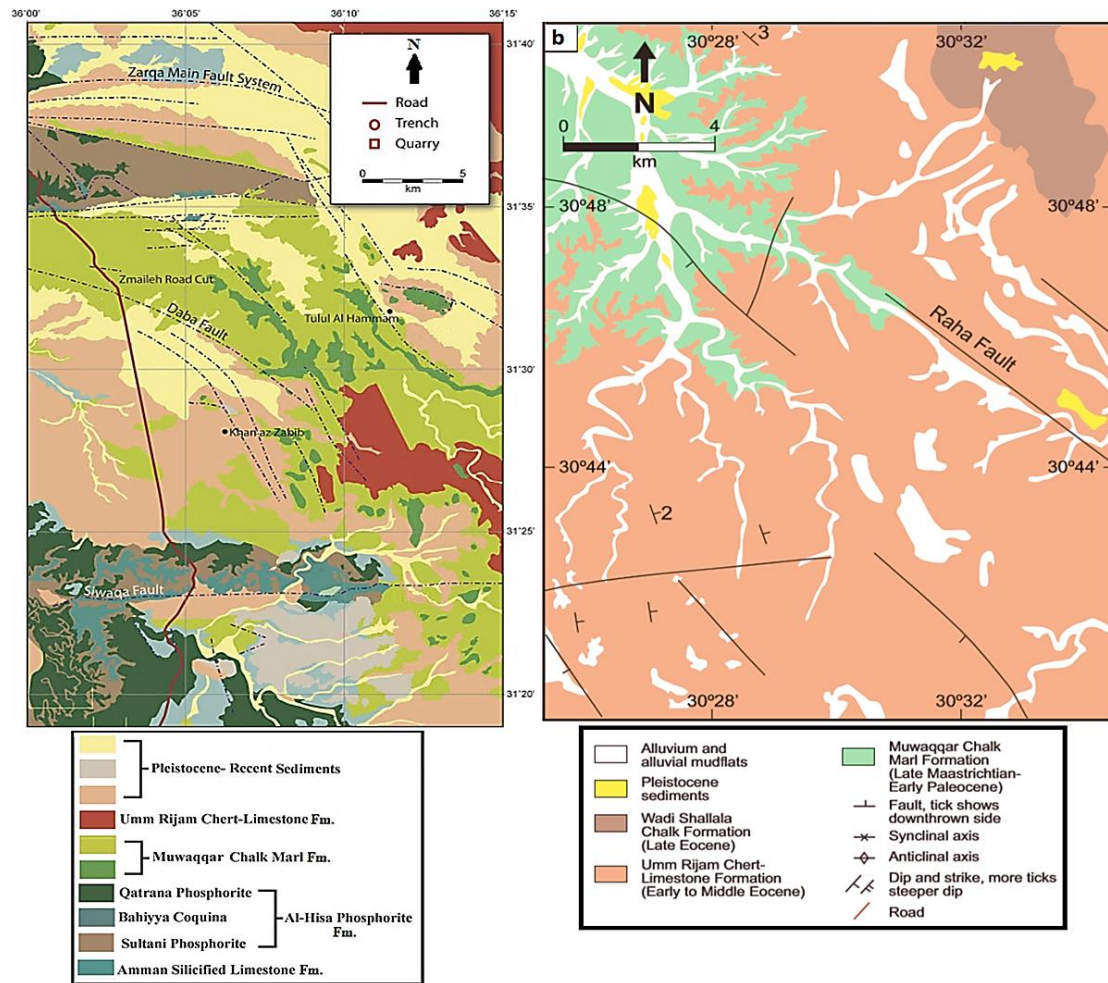


Figure 2.15 a. Geological maps of the area outlined in the red squares in Figure 2.3b showing the geological formations and the main faults in central Jordan (a) and south Jordan (b). (a) is modified from Khoury et al., 2015 and (b) compiled and modified from Abu-Lihie, 1985; Kherfan, 1986; Gharaibeh, 2010; Ali Hussein et al., 2015.

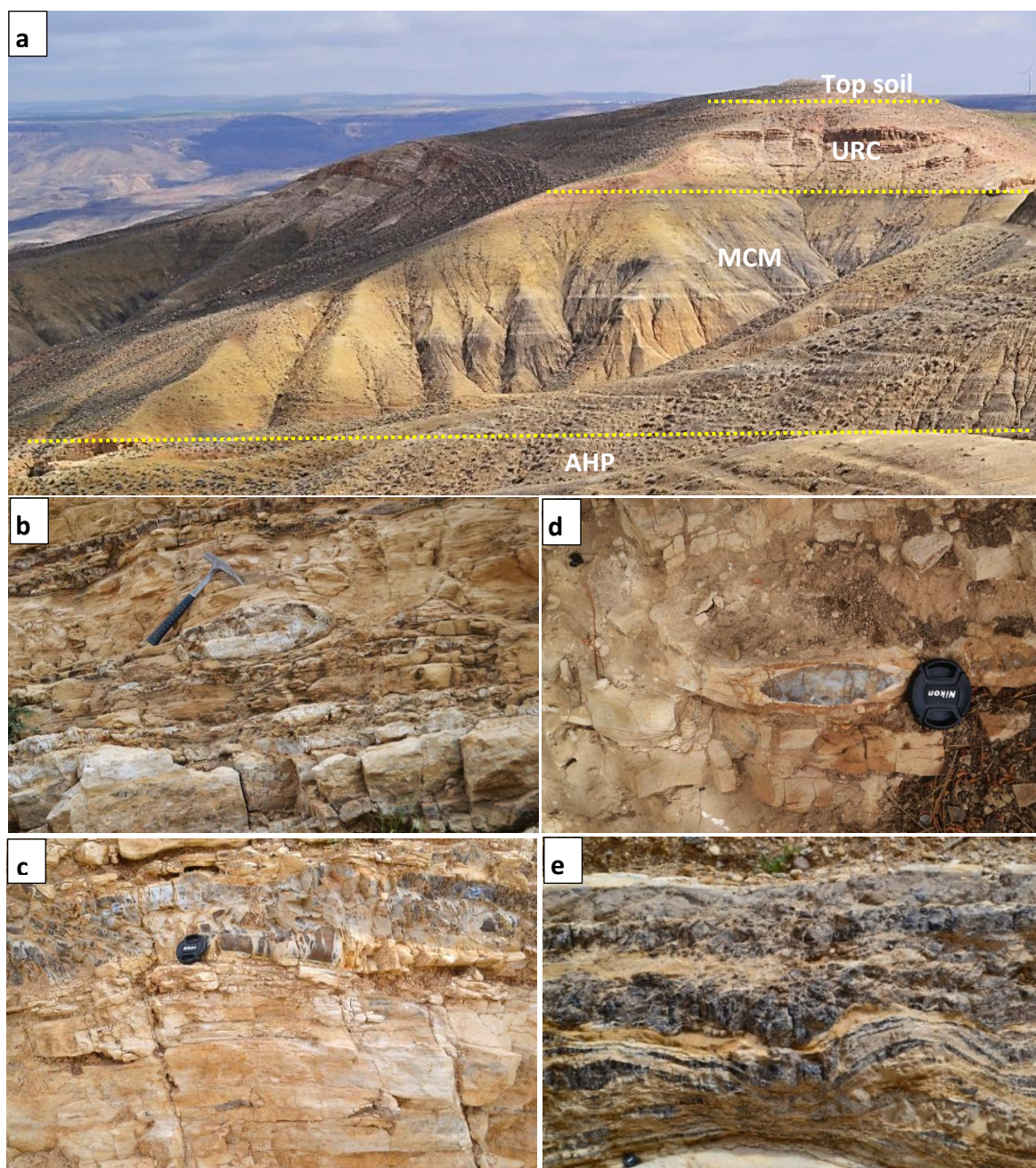


Figure 2.16 Field Photographs of the studied outcrops a. the AHP, MCM and URC formations exposed at Wadi Al-Bustan, south Jordan. b-d. Examples of nodular chert and chert brecciation in the ASL Formation in central Jordan. d. Folding in chert beds in the ASL Formation in Wadi Al-Bustan, south Jordan.



Figure 2.17 A road-cut view of the AHP formation along the Qatrana-Karak Highway.

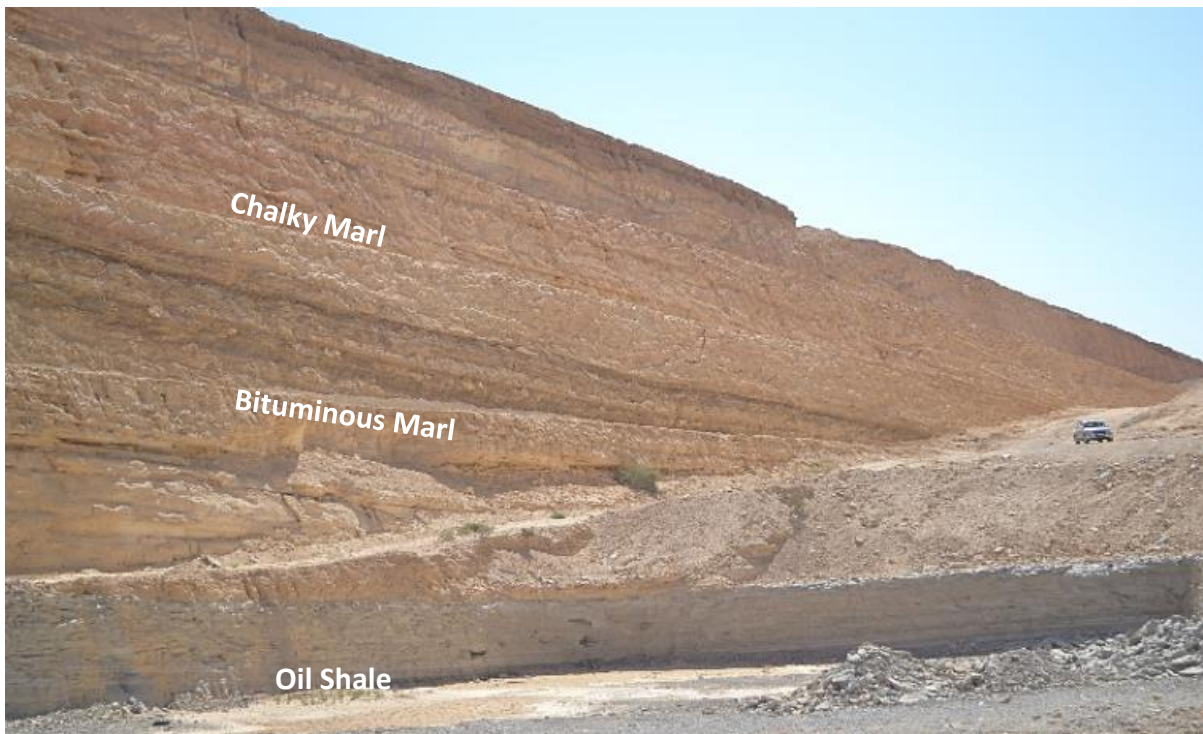


Figure 2.18 A quarry section exposing the MCM formation in the Sultani Area, central Jordan.

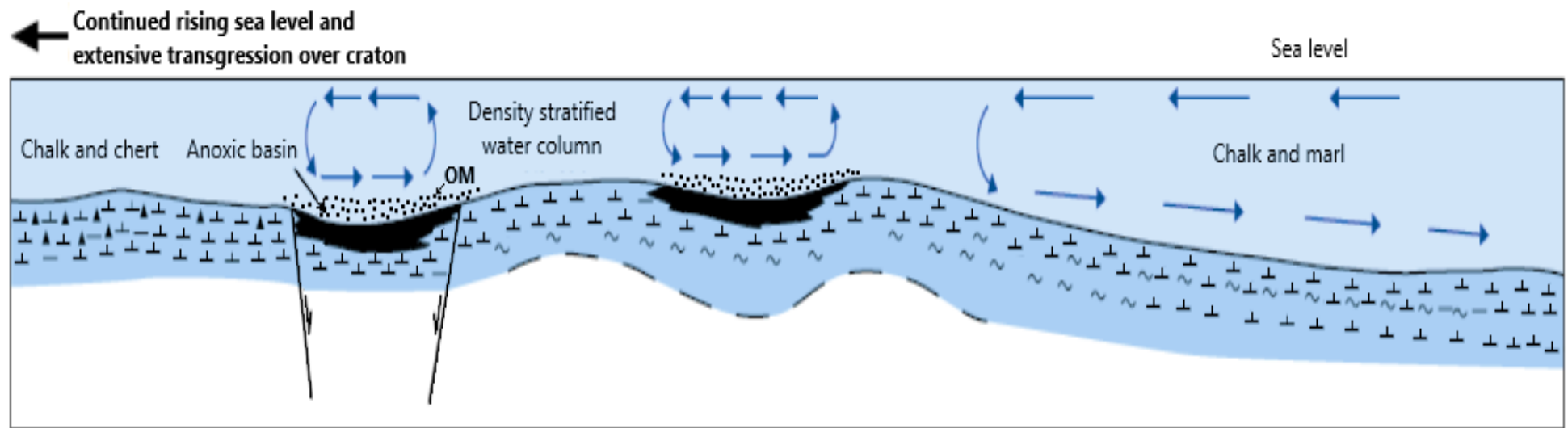


Figure 2.19 ESE-WNW cross-section with paleogeography and schematic depositional settings for the deposition of the MCM during the Maastrichtian and Tertiary times. Restricted sub-basins with differential subsidence formed along the enormous open shelf environment between Jordan and Levant, where the organic-rich marlstones and limestones were deposited during this time (Modified from Powell & Moh'd, 2011).

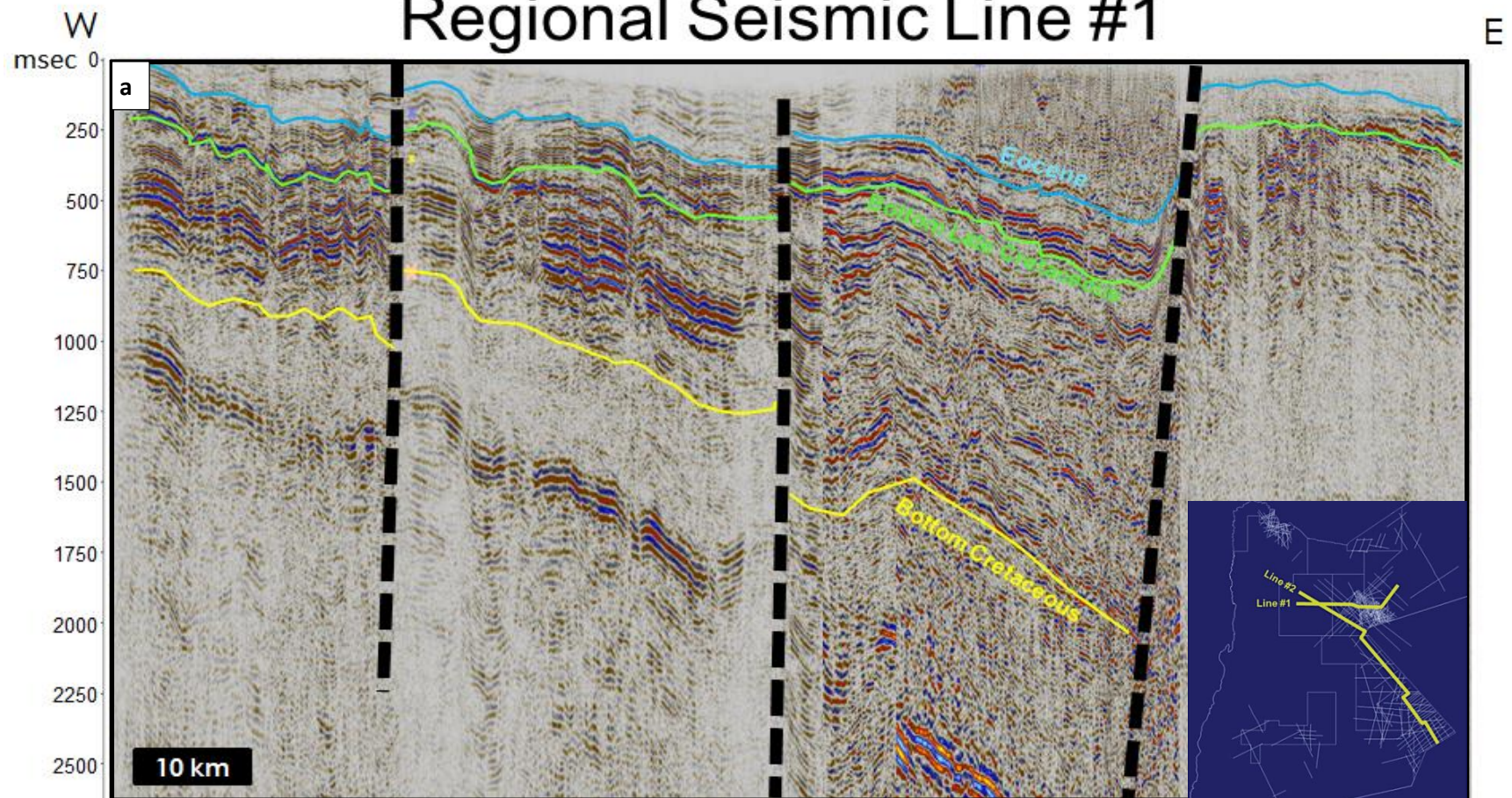


Figure 2.20 Field photograph of the contact between the URC and MCM formations showing the carbonate concretions zone, Harrana-central Jordan.



Figure 2.21 Field photograph facing east, showing the WSC Formation (WSC) in Wadi Al-Shummer, north Jordan.

Regional Seismic Line #1



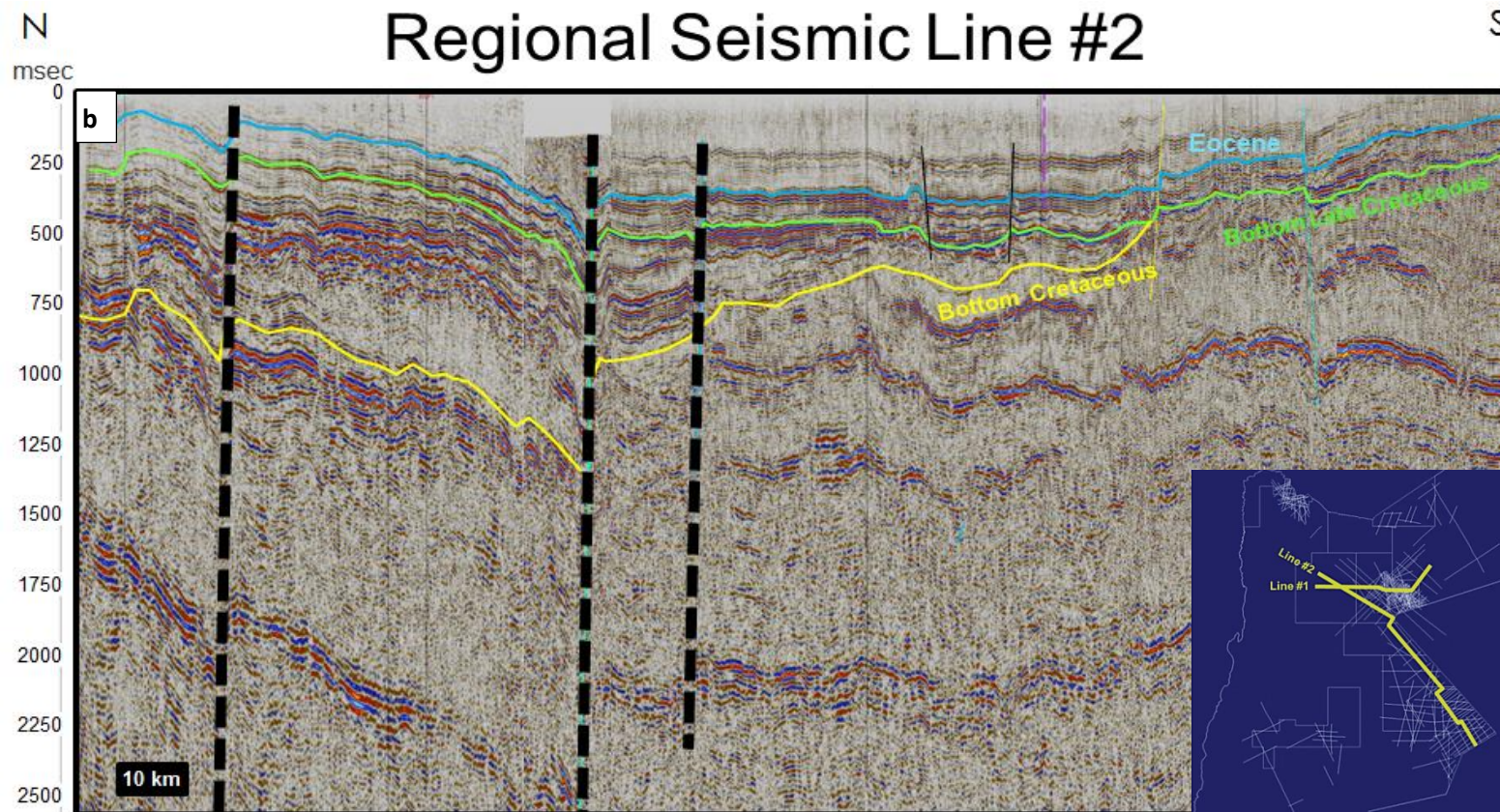


Figure 2.22 Seismic cross sections across Jordan: E-W (a, see previous page) and N-S (b) showing the increase of thickness the studied section from West to East and from south to north (modified from Shell internal report). The seismic interpretation was based on well data.

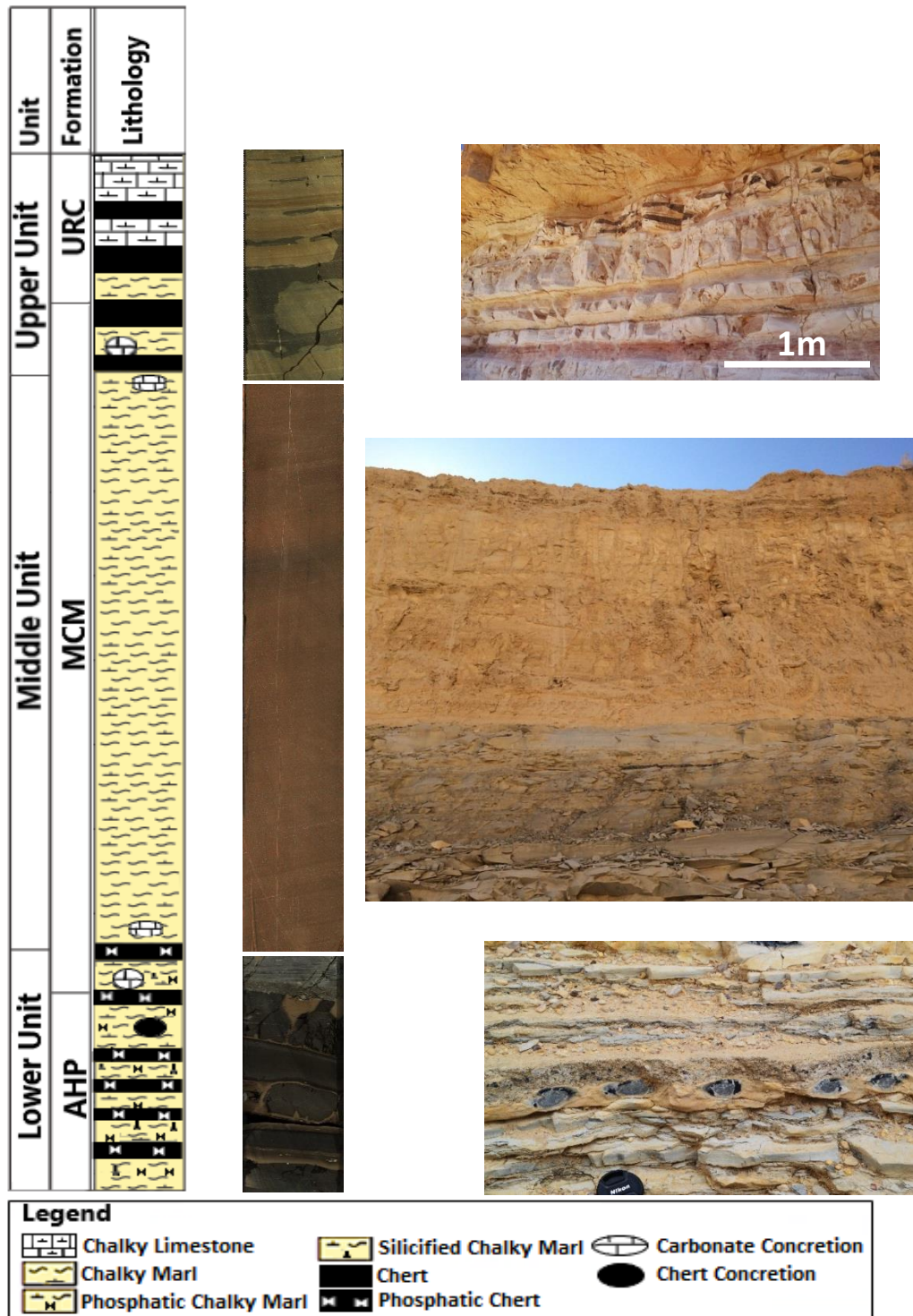


Figure 2.23 Schematic drawing of representative columnar section, with core and outcrop photographs representing each interval, showing the lithology of the studied cored intervals. The width of the core photograph is 10cm.

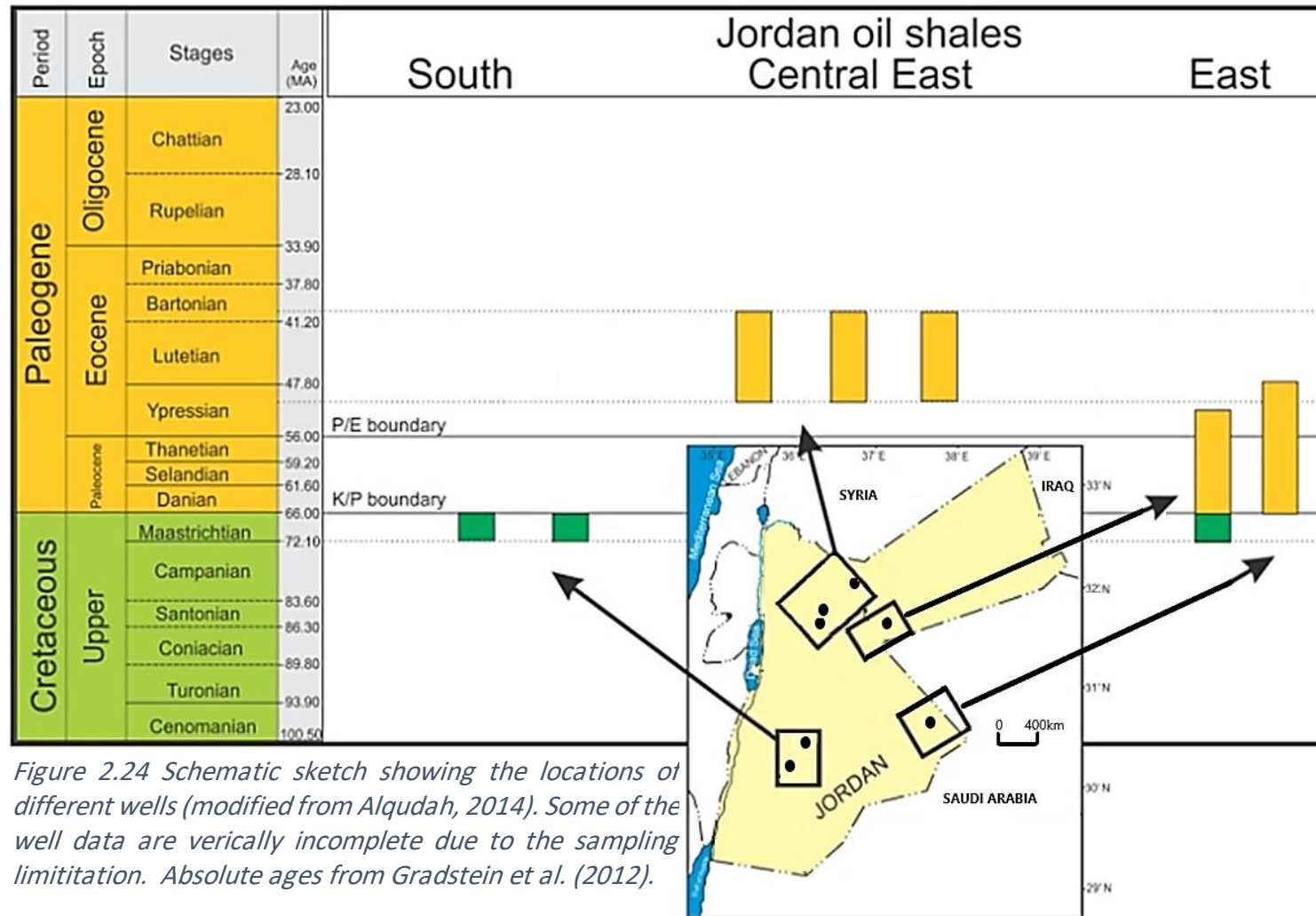
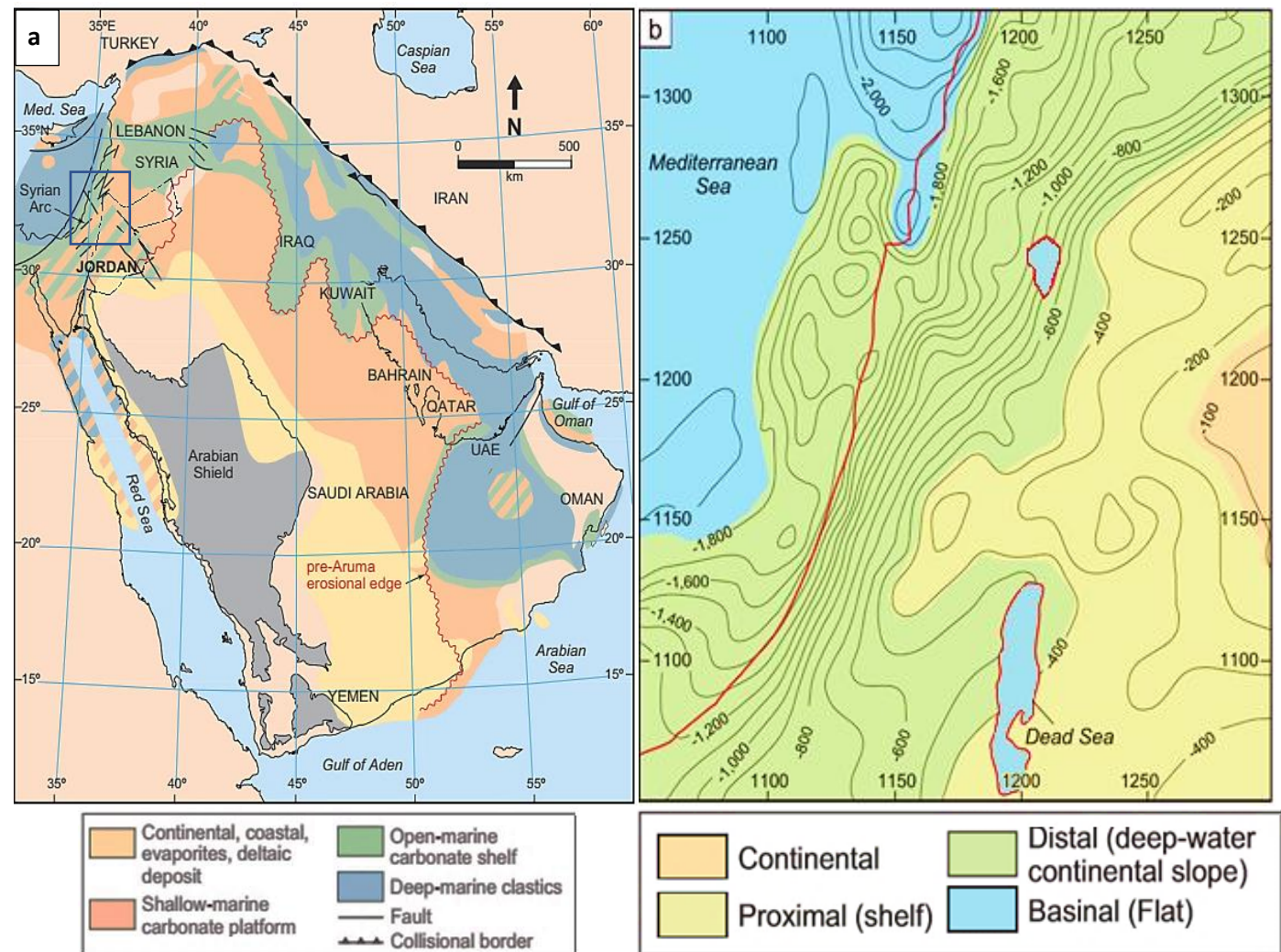


Figure 2.25 Paleofacies of the Arabian Plate during the Upper Cretaceous-Lower Paleocene. Modified from Ziegler, 2001; Ali Hussein et al., 2015. b. A Middle-Late Eocene paleo-bathymetric map showing the different parts of the carbonate ramp outlined in blue square in (a) (modified from Ali Hussein et al., 2015). Isobathymetric lines in meters, the datum is the Eocene mean sea level. The coordinates are in Palestine Grid (modified from Segev et al., 2011).



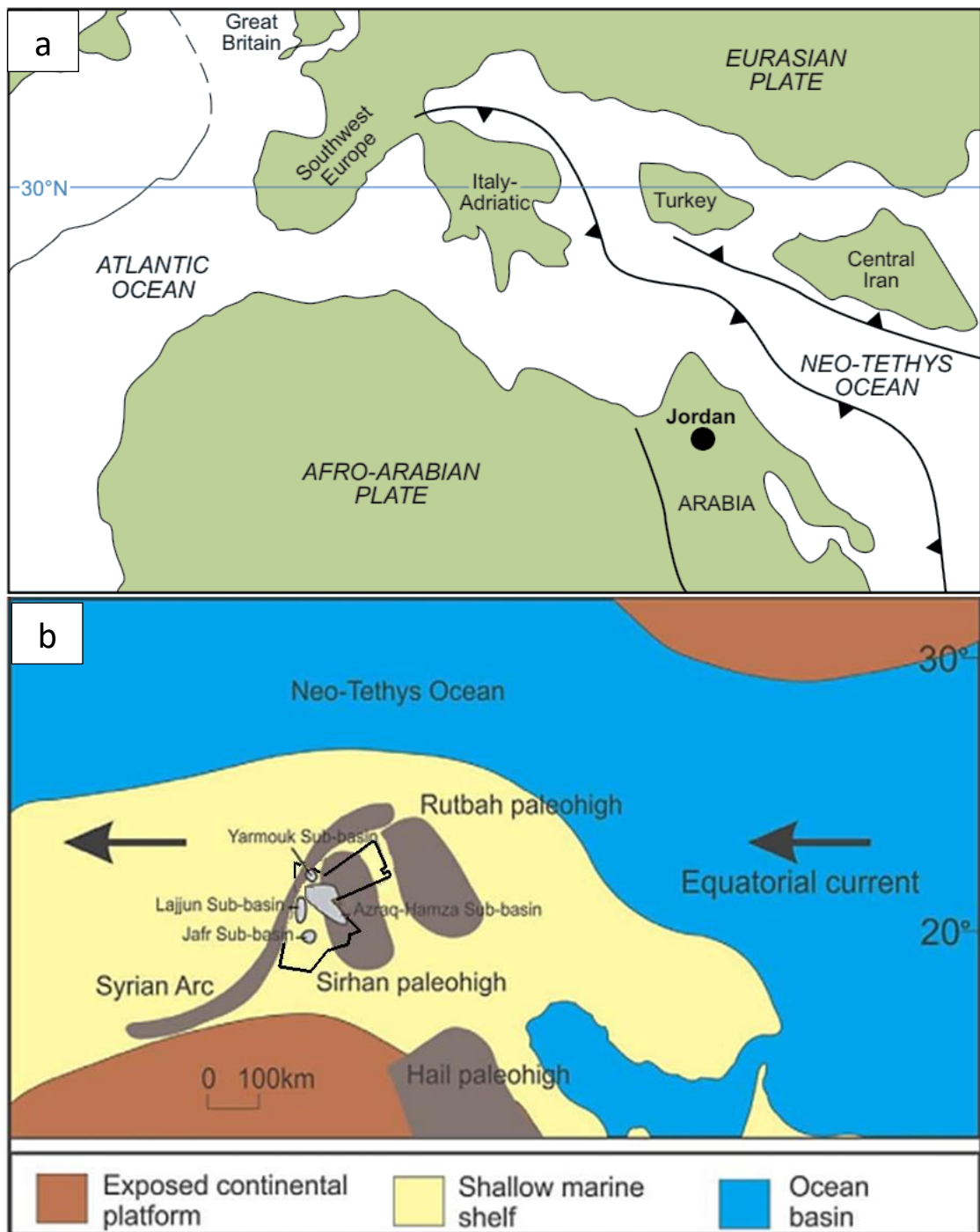


Figure 2.26 a. Late Cretaceous palaeogeographic reconstruction (modified from Powell & Moh'd, 2011; Scotese, 1991) placing Jordan at the Southern margin of the Neo-Tethys Ocean. Lines with triangles indicate subduction/obduction zones. b. Paleo-map shows major currents, paleohighs and sub-basins in Jordan during the Late Cretaceous-Paleogene (Modified from Abed, 2013; Alqudah, 2014).

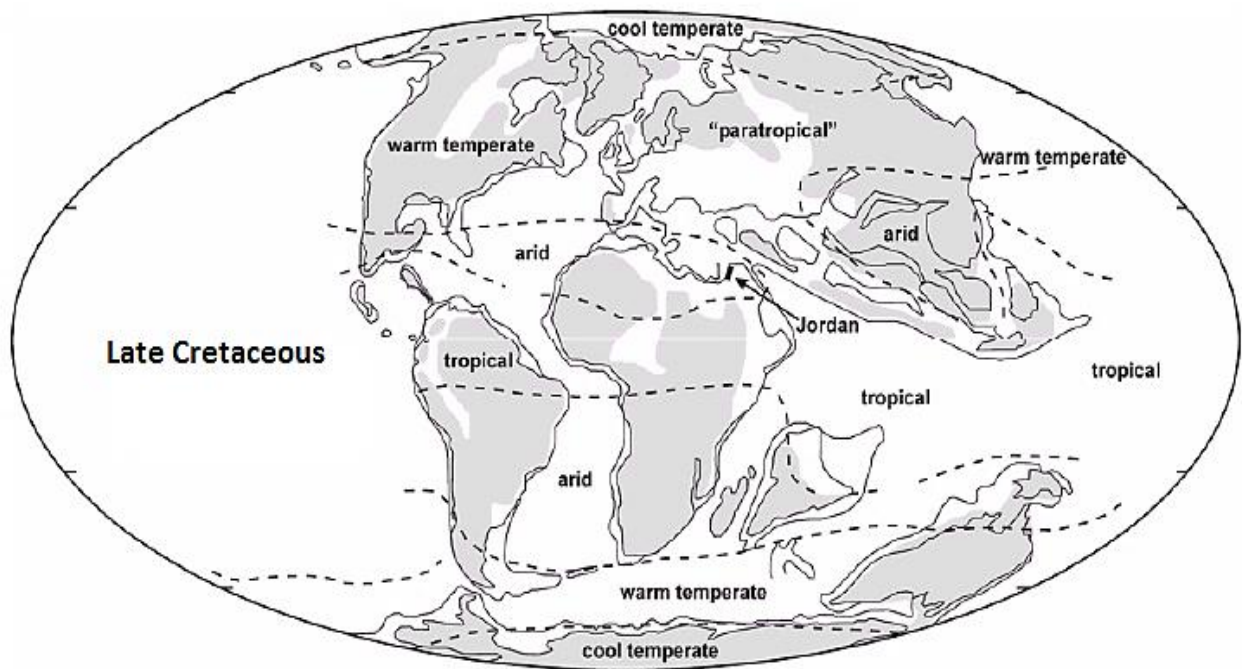


Figure 2.27 Global climate zones during Late Cretaceous (modified from Scotese, 2000). The grey-shaded areas represent terrestrial continental areas.

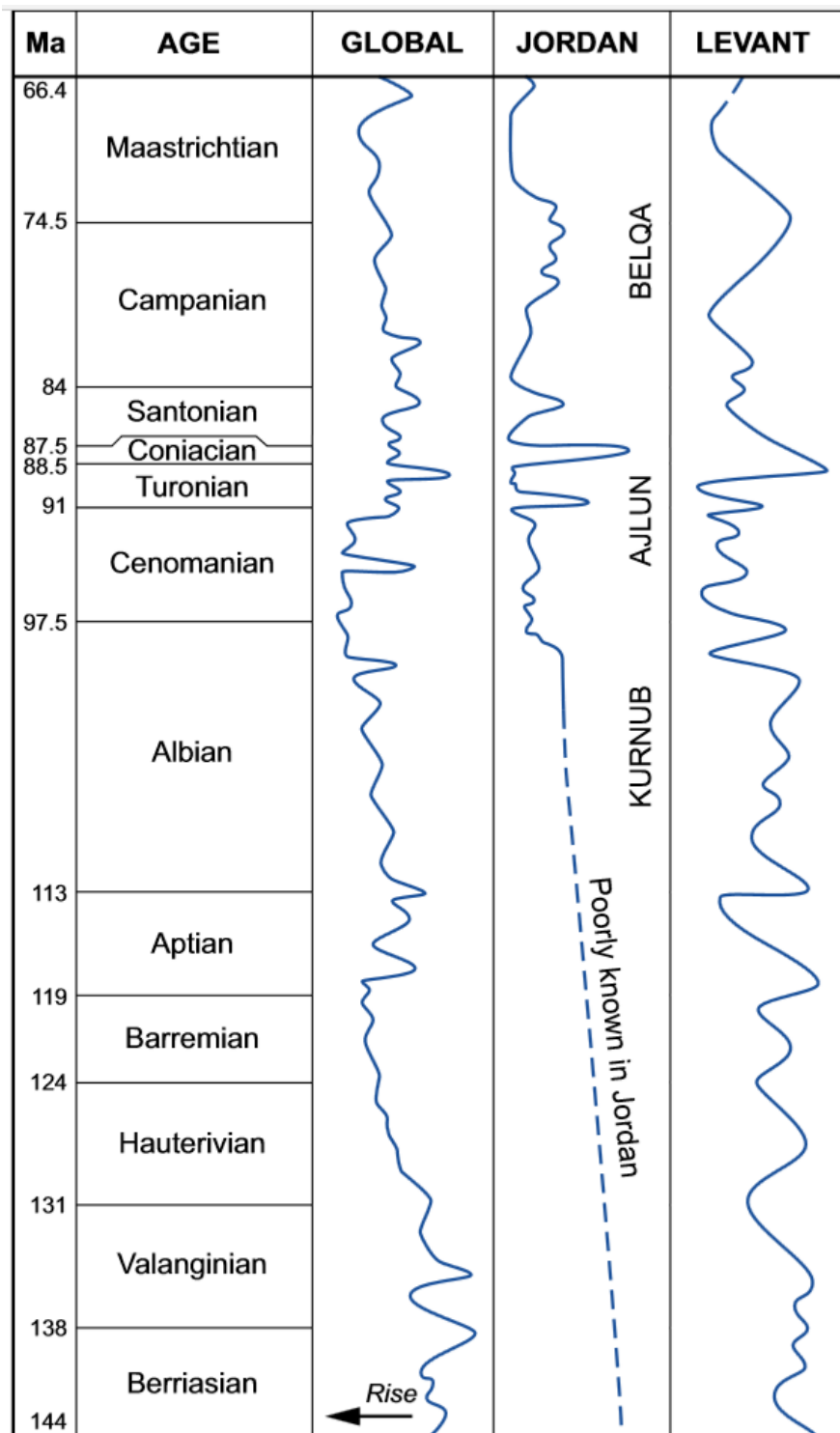


Figure 2.28 Comparison of relative sea-level curves for Jordan and Levant (the southern Neo-Tethys) compared to the global sea-level curve (Haq et al., 1988). Modified from Powell & Moh'd, 2011. Chronostratigraphic framework is from Scott et al. (1988).

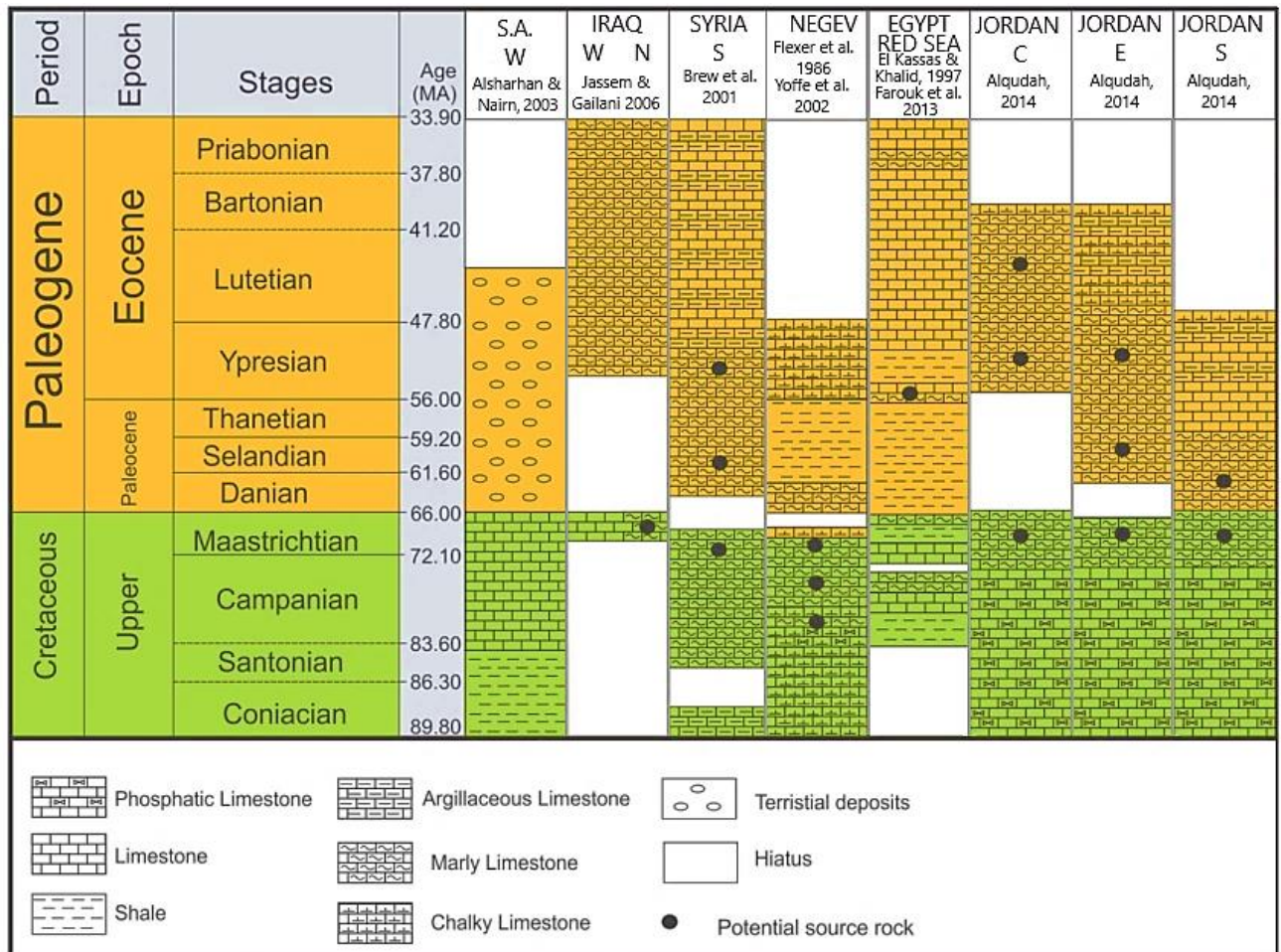


Figure 2.29 Regional correlations of lithological units in the region including potential source rocks. S.A.: Saudi Arabia. Modified from Alqudah, 2014. Lithological units are taken from Alsharhan & Nairn (2003), Jassem & Gailani (2006), Brew et al. (2001), Flexer et al. (1986), Yoffe et al. (2002), El Kassas and Khalid (1997), Farouk et al. (2013) and Alqudah (2014). A low sedimentation rate in the Sirhan Sub-basin (central and east Jordan), being high at the same time in the Jafr Sub-basin (south), corresponds with an early, late Maastrichtian regressive event and probably, with the uplift of the Sirhan Sub-basin.

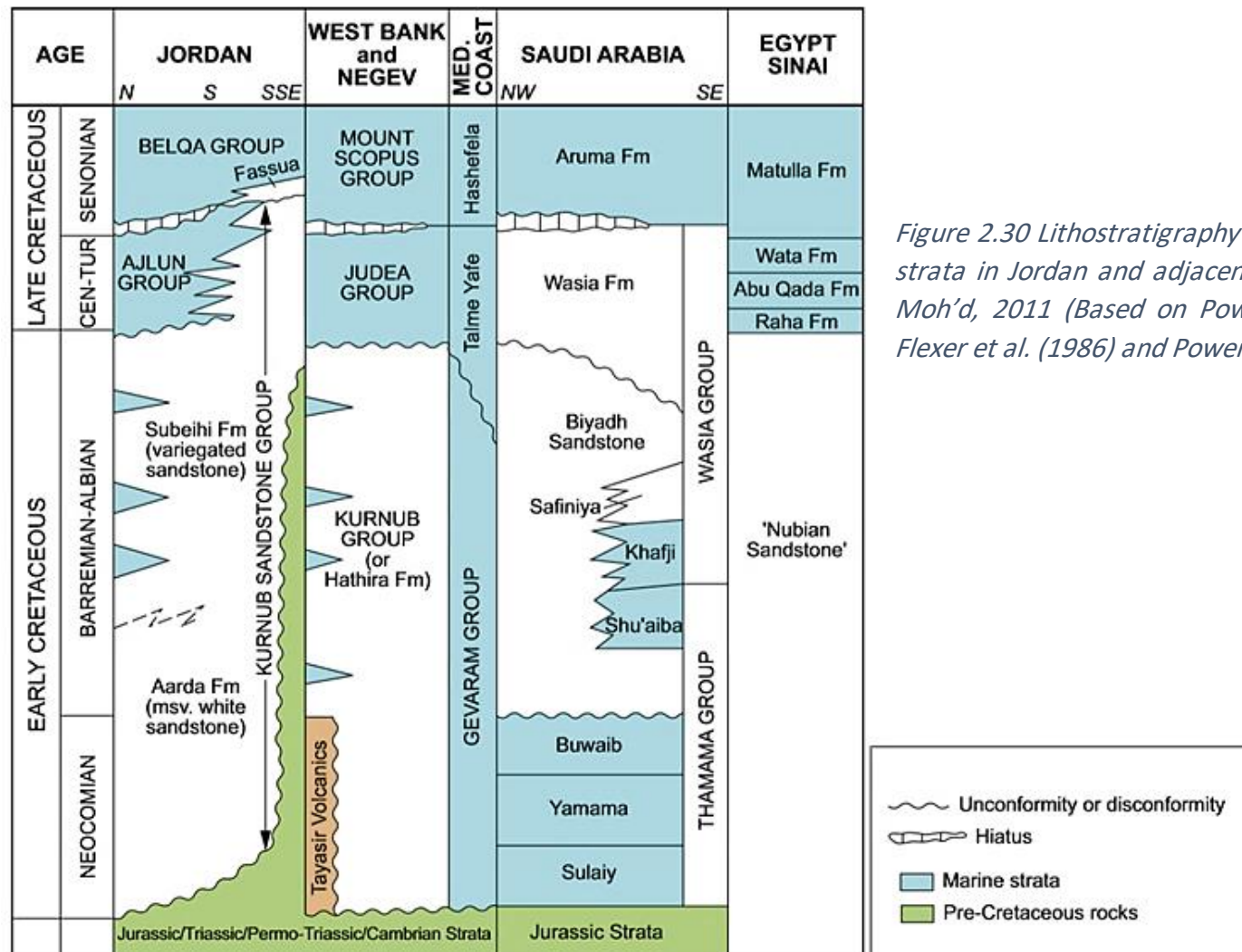


Figure 2.30 Lithostratigraphy and sequences of the Cretaceous strata in Jordan and adjacent areas. Modified from Powell & Moh'd, 2011 (Based on Powell (1989), Bartov et al. (1972), Flexer et al. (1986) and Powers (1968)).

Chapter 3: Characterization of Subsurface Natural Fracture Systems and Constraining the Timing and Mechanisms of their Development in the JOS Reservoir

3.1 Introduction

‘Reservoir fractures’ is a general name used to define naturally-occurring discontinuities in rock units (Nelson, 2001). Characterization of natural fractures in subsurface reservoirs can be achieved by several different techniques (Nelson, 2011; Spence et al., 2014). These vary from measuring fractures from reflection seismic surveys to core and borehole images. Among all these techniques, data collected from core samples can be the most reliable data. Data collected by other techniques are indirect, and could be subject to limited resolution or calibration (e.g., Nelson, 2001). 3D seismic data allow the study of stratigraphy and large-scale structures in subsurface rock units (e.g., Cartwright & Huuse, 2005), but they are ineffective in the characterization of natural fractures for which this study focuses on. Borehole images can be used to characterise fracture intensity (e.g., Wu & Pollard, 2002; Tokhmchi et al., 2010; Alfahmi, 2018), and to measure fracture aperture (Luthi & Souhaité, 1990) and orientation (Genter et al., 1997; Lagraba et al., 2010).

Generally, most studies have dealt with only a single technique to describe and measure natural fractures in subsurface rocks. Investigation of the fracture properties from different sources is therefore important to get a better understanding of their distribution, timing and formation mechanisms. This chapter presents the findings of the analysis of subsurface natural fracture systems that were described and measured from both cores and equivalent borehole images of wells drilled into the Upper Cretaceous-Eocene succession of Jordan. The two data sources provide a robust database to study fractures in the subsurface with regard to their types, morphologies, geometries, intensities, orientations, vertical distributions, spacing, and mineral fill.

The chapter aims mainly to characterise the natural fracture systems from core and borehole image logs by (1) investigating their spatial distributions, (2) constraining the timing of fracture development, (3) understanding their formation mechanisms, and (4) placing the fractures in a structural and basin evolutionary context.

The contents of this chapter are organised under six main sections. The first section provides an overview of the methods used to describe and measure fracture properties from the core and borehole images, and the geochemical and petrographic methods used to characterise fracture cement and their host rocks. The second section provides a lithofacies definition of the studied interval, the JOS and the over- and underlying chert-rich units. The third section is divided into two main parts: (1) the findings of the analysis of core fractures including their macroscopic and microscopic descriptions, fracture spacing, fracture distributions, and the geochemical characteristics of fractures and their host rocks and (2) the results of fracture characterizations from borehole image logs including their types, distributions and orientations. The fourth section presents the results of the analysis of strain around fractures and concretions. Discussion of the timing, distribution and formation mechanisms of the different subsurface fracture systems present in the JOS is provided in the fifth section. The sixth section summarises the findings of this chapter.

3.2 Methods and Datasets

This section introduces the methods and datasets of the chapter. Different analytical techniques were used, each of which is geared toward characterising fracture systems and constraining the structural evolution of fine-grained rocks.

Natural fractures are usually described by parameters that include geometry, aperture, mineralogy (mineral fill), intensity, distribution, and orientation (e.g. Nelson, 2001; Laubach

& Ward, 2006; Olson et al., 2009; Gasparrini et al., 2014; Gale et al., 2014). The methods used in this study include (1) characterising subsurface fractures from the core by measuring and describing fractures from core samples, and petrographic and geochemical analysis on both fracture cement and their host rocks and (2) mapping out fractures in borehole images for both cored and non-cored wells (cuttings wells). Before applying these techniques, the core lithofacies were defined in each measured interval to provide a framework of the collected fracture data.

Borehole drilling, coring and imaging processes were done by the same operation team and the same company during a short period of time between 2010 and 2014. Mud was used as the drilling fluid during sinking of the wells.

3.2.1 Methods of Fracture Analysis from Cores

3.2.1.1 Fracture Description Methods

The core analysis has included describing and measuring natural fractures in fifteen vertical cores in Jordan (see Figure 2.8, Chapter 2). Fractures were identified as naturally occurring based on the presence of any filling materials (e.g., mineral cement or bitumen). Uncemented fractures (joints) observed in the core were distinguished from those drilling-induced fractures by comparing them with their extent in the borehole images. The true strikes of fractures are not presented because cores are not oriented. The fifteen examined cores include five located in the Harrana area in the central west (C1-C5), two cores (C6C7) located in the central area in Attarat Umm-Ghudran and Qaser Tuba, three cores located in the Azraq area in the central east (C8-C10), two cores located in the Sirhan area in the southeast (C11-C12), and three cores located in the Bayer area in the southwest (C13-C15) (see Figure 2.8, Chapter 2).

The cored interval ranges from 150 m to 300 m depth below ground surface per well, with a core diameter of about 10 cm. Fracture type, morphology, geometry, spacing, connectivity, and mineralogy were described from the core. The frequency, aperture and length of different fractures were measured using the “Quick Look” method developed by Gross (2014). The rationale of using this method was the large numbers and lengths of the studied cored sections. This method includes counting and measuring fractures within a one-meter length core box comprised of 4 trays. Different fractures were described and measured within cored intervals defined by the top and bottom depths of each tray (Figure 3.1a-b). Specifically, this data collection method has involved identifying the fracture type and its cement (filling material), measuring fracture lengths, apertures (opening displacement) and spacing (Figure 3.1c), studying cross-cutting relationships, and quantifying the compaction around fractures and concretions (wherever found). A hand lens (10X) was used to study fracture tips. Apertures of the detected opening-mode fractures in all cores were measured using a digital caliper.

Figure 3.2 shows an example of the raw data collected from cores. The data were compiled in tables for each depth interval classified according to fracture type (veins, joints and faults/breccia). The fracture frequencies and cumulative lengths of each fracture type in each given interval were normalized to the length of recovered core (Figure 3.3). Ultimately, fracture data were plotted as a function of depth together with other core data (lithostratigraphic, geochemical and wireline data). The cored interval was divided into different mechanical units based on the fracture distribution (see section 3.4.1.5).

3.2.1.2 Petrographic Analysis Methods

Thirty-three thin sections with a thickness of 30 μm were prepared from fracture and host rock samples collected from the fifteen studied cores to study the mineralogy and

microscopic properties of the different fracture types and their host rocks. Optical analysis of thin sections of the selected samples was made in cross-polarized light with a Nikon Optophot Microscope linked to a digital camera of high resolution.

Selected thin sections were polished and carbon coated for study under a scanning electron microscope (SEM). The thin sections were examined using a FEI Quanta 650 FEG scanning electron microscope. Backscattered electron (BSE) images were created to study mineral compositions and textures of the samples. Elemental mapping was performed using an energy-dispersive X-ray spectroscopy (EDS) attached to the SEM. All SEM analyses were performed at 20 kV and a working distance about 10 mm.

3.2.1.3 Geochemical Analysis Methods

Geochemical data of the JOS, including TOC, Total Sulphur (TS) and maturity data were acquired for cored sections from JOSCO, Jordan. Every data point represents an interval of homogenized crushed sample taken from a one-meter homogeneous lithological unit. TOC and TS were analysed using a calibrated LECO-CS elemental analyser. The maturity analysis was performed using a Rock-Eval 6 Standard Analyzer Unit from Vinci Technologies.

Geochemical analyses including stable ($\delta^{13}\text{C}$ and $\delta^{18}\text{O}$) and radiogenic ($^{87}\text{Sr}/^{86}\text{Sr}$, U-Pb) isotopes were carried out on the cements of calcite-cemented veins and their host rocks. The aims of performing these analyses are to evaluate the source of fracture carbonates by comparing the different isotopic ratios of fracture carbonates with those of the host rock matrix, and to constrain the timing of fracture development of the different fracture types.

3.2.1.3.1 Carbon and Oxygen Isotopes ($\delta^{13}\text{C}$ and $\delta^{18}\text{O}$)

Stable oxygen and carbon isotopic compositions in carbonates can provide valuable information on the temperatures and source of fluids at the time of precipitation (Mook, 1971; Glig et al., 2008; Boni & Mondillo, 2015). Stable isotopic analysis of calcite veins and host rocks was used in this study with the aim of identifying the source of carbonates in these veins and constraining the timing and genesis of veins. Stable isotopes of $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ were analysed for studying the different types of calcite structures (planar veins, folded veins, cone-in-cone, and faults) and their host rocks.

Samples were collected from the three lithological units (Lower, Middle and Upper) of the fifteen cores in order to cover a wide range of host rock lithologies and structures (see Appendices). Forty-two pairs of powdered samples were extracted from the different types of calcite structures and their host rocks, using a hand-held foot-operated microdrill. To avoid any cross contamination, the microdrill was cleaned thoroughly between uses with deionised water and high-pressure air. Samples were cleaned using hydrogen peroxide (H_2O_2) and dried in the oven at 40 °C for at least 2 hours. Concentrated phosphoric acid (H_3PO_4) was added to the samples to produce the CO_2 to be analysed.

Samples were analysed isotopically for $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ using a MAT 253 mass spectrometer at the isotope laboratory of the Open University. The isotopic ratios were determined relative to the Pee Dee Belemnite (PDB) standard (Hoefs & Hoefs, 1997).

3.2.1.3.2 Strontium ($^{87}\text{Sr}/^{86}\text{Sr}$) Isotope

The isotopic composition of strontium in carbonates can help in the reconstruction of the geological, diagenetic and tectonic history (Jenkyns et al., 1995; Veizer et al., 1999; Wierzbowski, 2013; Zakharov et al, 2018).

Twenty-nine pairs of samples from the pre-extracted, powdered calcite veins and host rocks samples that were used for the stable isotopes analysis were also used for the $^{87}\text{Sr}/^{86}\text{Sr}$ isotopic analysis. Samples were weighed in teflon vials. These vials were thoroughly cleaned by acetone, deionised water, nitric acid (HNO_3), MQ water, and hydrochloric acid (HCL), respectively (see Appendices). Samples were dissolved in 0.5M acetic acid. All solutions were then transferred to a clean Teflon vial and dried out. 2M HNO_3 was then added to each sample to prepare for chemical separation. Strontium was separated using the automatic PrepFAST system (procedure adapted from Romanillo et al., 2015). This included isolating the strontium from the sample matrix used for the isotopic analysis. After separating the samples chemically, the analysis of strontium isotope ratios ($^{87}\text{Sr}/^{86}\text{Sr}$) were carried out at the Department of Earth Sciences at the University of Oxford using a ‘‘Nu Plasma’’ Multicollector Inductively Coupled Plasma Mass Spectrometry (MC-ICPMS). The data were normalised to the value obtained from the analysis of the SRM987 isotope standard ($^{87}\text{Sr}/^{86}\text{Sr} = 0.710248$).

3.2.1.3.3 U-Pb Radioactive Dating

This analysis was used to attempt to place absolute age constraints for the development of calcite-cemented structures. U-Pb systematics can age-date calcite deposited during fracture opening, thus constrain fracture timing. The U-Pb dating technique allows precise dating of the tectonic and diagenetic events, and constrains the absolute timing of different structures (Hansman et al., 2018, Beaudoin et al., 2018; Parrish et al., 2018).

The analysis of U-Pb dating was possible because the high uranium concentration in the studied samples has allowed the calculation of the uranium ratios from the calcite present in the different calcite structures. Uranium substitutes for calcium in carbonate cements, allowing the ability to age-date the carbonate veins from the uranium-rich calcite (Hooker et al., 2018).

Ages of the different calcite-filled structures were determined by isotope dilution using a mixed ^{236}U - ^{204}Pb - ^{130}Th spike (A. Mason 2018, personal communication) and a first-generation Nu Plasma MC-ICP-MS with a DSN100 desolvator and a c. 50uL/min or 75uL/min self-aspirating PFA nebuliser.

Samples of calcite crystals were extracted from the vein calcite using a small diamond saw, and transferred to acid-cleaned 15 ml polypropylene bottles. The samples were then sonicated repeatedly in 18 Mohm water with rinsing between each wash. The samples were then cleaned twice by distilled 2% HNO_3 with sonication to remove any residual dirt. After each wash, the samples were thoroughly washed with 18 Mohm water and sonicated to remove any residual acid and dislodged surface material.

Approximately 30 uL/drop (two drops) of the spike was then added directly to the acid cleaned samples and gently shaken to mix the sample. Once the visible reaction was complete, the solution was diluted to around 15 ml with 18 Mohm water, and fully shaken to make it homogeneous. The solutions were immediately analysed, with no preconcentration of U and Pb.

3.2.2 Fracture Analysis from Borehole Image Logs

The second technique of subsurface fracture analysis was by mapping and measuring fractures from Borehole Images (BHIs) of thirty-two wells (including the fifteen cored wells) distributed across Jordan (see Figure 2.8, Chapter 2). Techlog software was used to perform this analysis. The importance of this technique is that it is not only effective in extracting fracture information from the cored wells prior to correlation with the data collected from core samples, but also allows mapping of fractures from the non-cored wells, significantly expanding the fracture dataset. Data obtained from this analysis includes dip, strike, frequency and fracture aperture.

Borehole imaging was done using an Optical Televiewer (OPTV) instrument by the JOSCO team as a fundamental part of wireline examinations conducted on the JOS. Using visible light optics, the OPTV probe delivers a continuous, detailed true colour image of the internal borehole wall. The light source used provides high power straight light. The OPTV yields good resolution of images above the water table; however, it is ineffective when it penetrates the water table due to the turbulent wellbore conditions. The OPTV image log is a very effective database to study fractures in the non-cored wells, particularly for the wells with relatively shallow depths of the studied section. The image log dataset consists of 14 wells from the Harrana area, 8 from the Azraq area, 3 from the Sirhan area and 7 from the Bayer area (See Figure 2.8, Chapter 2). This analysis allowed measurement of the fracture frequency, distribution and orientation.

3.3 Lithofacies Definition

A complete lithostratigraphic framework of this study is presented in Chapter 2 (section 2.4.2). This section introduces the main lithofacies described from core. Twelve main lithofacies (LF) were defined through the cored section (Figure 3.4). The most commonly identified lithofacies in the section are summarised as follows:

LF1: Silica-bedded marlstone; this facies unit characterizes the lower and the upper units of the studied core succession. It is composed of an interbedded alternation marlstone, and chert bands and concretions.

LF2: Bioturbated horizon (Thalassinoides-burrowed bed) can be obviously distinguishable in the Lower Unit. This indicates high energy lithofacies.

LF3: Chalky marlstone; this lithofacies association mostly represents the Middle Unit of the section, and can also be present in the lower and upper units.

LF4: Phosphatic marlstone; this is the principal characteristic facies of the Lower Unit of the section.

LF5: Phosphate-rich fossiliferous limestone; mainly present in the Upper Unit, and composed of phosphatic and skeletal packstone, with dominant foraminifers and fish fragments, with a moderate degree of sorting.

LF6: Laminated organic-rich dark greenish grey marlstone; this characterises the Middle Unit and mainly formed of laminated marl. This developed horizontal lamination may indicate a quiet depositional environment (low energy).

LF7: Massive organic-rich dark greenish grey marlstone; this is also observed in the Middle Unit, and is recognised by the massive appearance of marl.

LF8: High-energy limestone; this lithofacies is characterised by poorly sorted bioclastic limestone and is mostly rich in foraminifera. Hummocky structure and cross-lamination are clearly observed.

LF9: Low-energy limestone: this facies association is characterised by well-sorted bioclastic limestone.

LF10: Hemipelagic chalky limestone; this facies association represents the lower part of the URC formation, and it is composed of thick beds of marly limestones with chalk, rich in nanofossils and foraminifera.

LF11: Brecciated marble; this facies association is only present in the Upper Unit of the core from Well C6, and is composed of (thermally altered carbonates and high temperature minerals) that has been brecciated in many zones.

LF12: Organic-poor very light grey marlstone; this is a specific facies association of the chalky marl beds on the boundary between the middle and upper units in the core from Well

C6 in central Jordan. The marl beds seem to have lost their organic matter content due to alteration by hydrothermal solutions.

3.4 Results of Subsurface Fracture Characterization

3.4.1 Natural Fractures Observed in Core

Previous studies have identified a number of fracture types within the JOS and the over- and underlying chert-rich units (Claps & Gross, 2016 internal report; Hooker et al., 2017a). This study goes further, expanding on these preliminary studies which are based on only four cored wells to a regional study carefully characterising natural fractures in fifteen, widely-distributed, cored wells, and relating the fracture types to their distribution, timing and genesis.

3.4.1.1 Fracture Description and Mineralogy

Various types of structures were observed and measured from the cores including veins (cemented fractures), cone-in-cone structure, joints (uncemented or open fractures), bitumen-bearing fractures, faults, and dilatational breccias (Figures 3.5, 3.6 and 3.7). They are present at various length scales ranging from seismically observable faults to small faults and joints/veins, through to microfractures. Veins are easily distinguishable structures due to the distinct colours of their filling minerals. They are classified into two main types based on their morphologies (a) planar veins and (b) folded veins (Figure 3.6).

Veins exhibit a range of filling materials (fracture cement). The most common filling material among all vein types is, in general, calcite (~75%). However, some veins are cemented by silica (~20%) or gypsum (~5%) (Figures 3.5 and 3.6). Bitumen is present in some fractures, either entirely filling the fracture, or coexisting with calcite in planar calcite

veins (Figures 3.7e-f). Planar veins are recognised by the diversity of their filling materials (calcite, silica, or a combination of calcite and silica) whereas folded veins are usually observed filled by calcite. Fibrous bedding-parallel veins are filled either by calcite or by gypsum (Figure 3.6b and d). Cone-in-cone structures are only filled by calcite (Figure 3.6a). Faults can be filled by calcite. This may indicate that there is a relationship between the fracture cement and its shape (see Discussion; section 3.7).

Fracture mineralogy usually reveals the lithology of the host rock. Folded calcite veins are only observed in calcareous mudstone layers; calcite veins are common in calcareous marls, marly limestone and limestone; and silica-filled veins were detected in chert and silicified intervals (Figures 3.5 and 3.6). This suggests that these veins have been formed in a chemically-closed system (Metcalf & Rochelle, 1999), with a supersaturated fracturing fluid. Veins formed in closed systems have been previously widely recognised for calcite veins in chalks and shales (e.g., Czerniakowski et al., 1984; Blyth et al., 2004; Hooker et al., 2017b) and for quartz veins in metamorphic rocks (e.g., Yardley & Bottrell, 1992; Cartwright et al., 1994; Widmer & Thompson, 2001).

3.4.1.1.1 Folded Veins

Folded calcite veins are only observed in mudstone beds close to, or positioned between, chert beds or concretions (Figure 3.8a-c). They are present as either isolated veins (Figure 3.5d) or predominantly, as a cluster of veins in mudstone (Figure 3.5c). No folded veins were observed in coarse-grained beds such as resedimented beds.

Folded veins are cemented primarily by calcite (Figures 3.5c-d and 3.8). The veins are observed to be arranged in a vertical to sub-vertical geometry (Figures 3.5c-d and 3.8). In most cases, the upper and lower tips of the folded vein usually align on the same vertical

plane (Figure 3.5c-d). Generally, folded veins appear to have undergone a large reduction in vein height. This reduction is estimated to be about 50% (Hooker et al., 2017a).

The aperture size of the vein is usually linked with the degree of folding; thick veins are less folded than thinner veins (Figures 3.8 and 3.9). In other words, the thinner vein has recorded higher shortening strain than the thicker ones (see section 3.5 for more detail).

The vein tips are more angular in thin veins than in thick veins (Figures 3.5c-d and 3.9). Folded calcite veins are much shorter than planar veins, with a current height generally less than 10 cm (Figure 3.5). They are typically closely spaced; clusters of several folded veins can be observed within a single bed across the core width (Figure 3.5c). The greatest folding of the folded vein body is seen in the parts that are with immediate contact with chert beds/concretions (3.8a). The vein folding becomes less towards the vein tips (Figure 3.8).

Systematic observations of microfractures associated with folded veins can be seen in the core. Microfractures are highly clustered in the parts with a greater bending of the folded vein and within the inner arcs of the folded veins (Figure 3.8d). The microfractures are also filled by calcite.

Folded veins are mostly found in non-paralleled arrangements (Figure 3.8c). The veins are extremely folded in the mudstone beds whilst they are planar (unfolded) where they transect chert beds or concretions (Figure 3.8a-b). This reveals a significant relationship in the timing of fractures and their relations to diagenesis (see Discussion; section 3.7).

3.4.1.1.2 Planar Veins

Planar veins are the most common vein type observed along the cores. They are present in all rock lithologies within the three stratigraphic units (Lower, Middle and Upper), but with different intensities (see Section 3.4.1.2.4).

Planar veins vary in thickness from submillimetre-scale (hairline veins) to centimetre-scale thick veins. They are present in core as either simple, individual veins (Figures 3.5a-b, 3.9b and 3.10) or complex interconnected networks of mm-scale apertures veins (Figures 3.6c and 3.9a). These veins are observed in approximately parallel arrangements wherever present as vertical/sub-vertical veins and where they are not associated with chert concretions (Figures 3.5b and 3.10d).

Planar veins are recognised by their straight shape (Figures 3.5a and 3.10d), with occasional irregularity (Figure 3.5b). Stepping in the vein can be seen where the vein transecting two different lithologies (Figure 3.10c). This irregularity is an attribute of opening-mode fractures but is usually unobservable due to the small-scale aperture of these veins.

The veins are observed either as a simple, isolated vein (Figures 3.5a, 3.10b and 3.11) or as a group of interacting planar veins (Figure 3.6c and 3.10a). Morphologically, the veins show a range of tip morphologies varying from single tapering ends (Figures 3.5a and 3.10) to branching ends (Figure 3.10a). Vein aperture decreases towards the vein tips from maximum values that are close to uniform along fracture lengths (Figures 3.5a-b and 3.10a-b). Planar veins are mineralised by either calcite or silica. The majority of the observed planar veins are cemented by calcite (~70%).

3.4.1.1.2.1 Planar calcite veins

Planar calcite veins observed in the core are mostly orthogonal to bedding (Figures 3.10 and 3.11), particularly if the veins are isolated. They are best described as being widely dispersed along the core; typically, only a single vein is observed in any short, defined interval of the core (Figure 3.5a). Planar calcite veins are commonly found in marls rich in organic carbon. These veins are filled by calcite either entirely or partially (Figure 3.10). Entirely calcite-filled veins form the largest proportion among both types.

Interestingly, planar calcite veins occasionally exhibit crack-seal texture (Ramsay, 1980) in the core sample, particularly in the thicker veins (Figure 3.11), albeit that this texture is usually distinguishable under the microscope (see section 3.4.1.6.5). The crack-seal texture implies multiple-stage vein growth.

Solid bitumen is present within planar calcite veins filling the vacant segment of the vein (Figure 3.10c).

3.4.1.1.2.2 Planar silica veins

Planar silica veins are usually present in chert and silicified beds or concretions. They are found as horizontal or vertical veins (Figure 3.6c). These veins are usually cemented by silica. However, some vertical silica veins may also contain calcite cement (Figure 3.5b). Horizontal silica veins differ from vertical silica veins only by the absence of calcite cement (Figures 3.5b and 3.6c).

3.4.1.1.3 Bedding-Parallel Veins

Bedding-parallel veins are observed in all cores and they exhibit a range of internal structures that can be observed in the core samples. These vary from fibrous bedding-parallel veins (“beef”) to cone-in-cone veins (Bons et al., 2012; Cobbold et al., 2013) (Figure 3.6). Based on their filling materials, bedding-parallel veins are observed in three types. These are calcite bedding-parallel veins, gypsum bedding-parallel veins, and calcite bedding-parallel veins (Figure 3.6).

3.4.1.1.3.1 Fibrous bedding-parallel veins (Beef)

Fibrous bed-parallel veins observed in the cores are well developed in organic-rich mudstones and are mainly filled by calcite or gypsum (Figures 3.6a-b and d). These veins have clear sharp contacts with the host rocks (Figure 3.6). The veins are very light (white or light grey) in colour and can be easily recognized within the dark host shales. The veins are

usually observed along the lamina boundaries (Figure 3.6). This may reflect that these veins were created along the bedding due to strength anisotropy.

In all cores, bedding-parallel calcite veins usually exhibit clear blocky calcite crystals orthogonal to the vein walls (Figure 3.6b), whereas bedding-parallel gypsum veins comprise fibrous cement with a median line (Figure 3.6d). Generally, bedding-parallel veins often display a regular aperture along the vein segment seen within the core width. Vein tips were not often described from the core due to the limitation in the width of the vertical cores. Only a few examples of vein tips of bedding-parallel gypsum veins were observed in the core, where the vein spacing decreases towards the tips (Figure 3.12a-b). In such a case, the bedding-parallel veins appear as sets of short, thin lenses of en echelon vein arrays can be seen within the core.

3.4.1.1.3.2 Cone-in-Cone

Cone-in-cone intervals (Figures 3.6a and 3.13), calcite aggregates of conical shapes pointing upward, are observed in the core samples in cm-scale. In all investigated cores, cone-in-cone intervals were commonly observed within organic-rich mudstones (Figures 3.6a and 3.13). They are mainly composed of conical calcite crystals and abundant hostrock inclusions (Figure 3.6a and 3.13a). The boundaries of cones are usually traced by the fine-grain host rock material. The lower contact of the cone-in-cone structure is usually sharp (Figures 3.6a and 3.13a).

Cone-in-cone structures occur either as horizontal strands (Figure 3.13b-c) within mudstone beds or as a zone of cone-in-cone structure bending around carbonate concretions (Kiriakoulakis et al., 2000; Selles-Martinez, 1996; Gaines & Vorhies, 2016). The latter was not observed in the JOS cores. Interestingly, vertical strands of the cone-in-cone structure were occasionally seen in the core interlinking the horizontal cone-in-cone occurrences (Figure 3.13c). These vertical strands can be distinguished by their braid-like shapes. Some

of these vertical cone-in-cone occurrences were also observed in the core as isolated structures (Figure 3.13d). It is not clear whether such vertical examples are geometrically associated with any lateral segments not seen in core.

3.4.1.1.5 Bitumen-bearing Fractures

Bitumen is present within or associated with all of the fracture types either as macroscopic solid bitumen in the un-mineralised fractures entirely filling the fracture (Figure 3.7e-f), or as inclusions within the vacant segments of planar calcite veins (Figure 3.10c). Bitumen is only found within planar calcite veins; no bitumen inclusions have been observed either in folded calcite veins or in silica veins. Some bitumen veins are observed to be cross-cutting or running adjacent and parallel to silica veins (Figure 3.14a-b). There are a few observations of bitumen inclusions associated with shear fractures (faults), such as those observed in the Lower Unit of the core from Well C7 at a depth of 667 m. These few cases mainly represent bitumen that occurs within the fault cement or opening-mode fractures that are associated with fault damage zones. Bitumen is also commonly present in vugs within the host rock containing bitumen fractures (Figure 3.14d). It is not clear whether such vugs are connected by fractures not seen in core.

Bitumen-bearing fractures are not typically as straight and regular as mineralized veins, and their aperture varies irregularly (Figure 3.14a-c). Some bitumen fractures branch through the host rock causing partial brecciation (Figure 3.14d). The different types of bitumen-bearing fractures are described in more detail in Chapter 5, with more focus on the provenance of the fracture bitumen and its role in fracturing.

3.4.1.1.6 Joints (*Planar Uncemented Fractures*)

Uncemented fractures or open fractures (which are termed joints in this study) are predominantly present in the beds with the most brittle characteristics (Figures 3.7a-b). They are always planar and are often confined to beds (Figure 3.7a-b), particularly where the lithology exhibits large competence contrasts. This may be attributed to the differences in rock properties (stiffness) between beds (Evans et al., 1989; Bai & Pollard, 2000; Bai et al., 2000; Bourne, 2003). Joints are occasionally observed partially filled by bitumen (Figure 3.7b). Open fractures are observed in marl of the Middle Unit only in the core from Well C10 (Figure 3.7g).

Joints are generally present in parallel arrangements (Figure 3.7a-b), but they show radial distribution when present inside chert or carbonate concretions.

3.4.1.1.7 Shear Fractures (*Faults and Breccia Zones*)

Shear displacement of faults were recognised in the core by many features such as slickensides, fault stepping and offsets of lamination (Figure 3.7c and 3.15a-b). The faults observed are usually normal, with fault planes ranging from cm-scale to m-scale in height. They are present either as single faults or as segmented microfaults (Figures 3.7c and 3.15a). Faults observed in the core display displacement ranges from mm- to cm-scale; large displacements are impossible to measure due to the limited core width.

Faults are observed in the core in two main morphologies. These are single-plane faults (Figures 3.7c and 3.15a) and faults that display fault zones (Figure 3.15e-f). Each measured fault zone, in the latter, includes an identifiable fault core and two deformed zones within the fault walls (hangingwalls and footwalls), which are termed fault damage-zones (e.g. Segall & Pollard, 1983; McGrath & Davison, 1995; Kim et al., 2003, 2004) (Figure 3.15e-

f). Although the complete fault zone may not be observed in the core due to the limited width, the core from Well C7 has fortunately provided three fault zones within the depth interval from 660 m to 690 m (Figure 3.15e).

The fault-damage zones of the observed faults include a cluster of opening-mode fractures varying between sealed and unsealed fractures (Figure 3.15e). The core of each fault is composed of different fault-related features such as cataclasites, gouge and breccia and mostly cemented by calcite with bitumen inclusions (Figure 3.15).

Breccia Zones are observed in the core in multiple examples, and are usually associated with faults (Figure 3.15e). Breccia zones consist commonly of rock fragments cemented by calcite (Figure 3.15c-d). The geometries of breccia zones exhibit a large component of dilation (Figures 3.7d and 3.15c-e). This may suggest development under elevated fluid pressure conditions (Sibson, 1986; Nygard et al., 2006; Shukla & Sharma, 2018).

The highest percentage of breccia zone was observed in the core from Well C6, mostly, in the Upper Unit, where the host rocks are completely metamorphosed (Figure 3.4) (see Chapter 2.4.2 for more information).

3.4.1.1.8 Drilling-Induced Fractures

Drilling-induced fractures are fractures created in the core by the stress generated during the drilling process (Li & Schmitt, 1998; Zhang et al., 2010, Gale et al., 2018). Such fractures are characterized by their high angle to the well bore and are parallel to S_{HMAX} (Gale et al., 2018) (Figure 3.16). They can be easily distinguished in the cored intervals from other natural fractures by a petal-median-line morphology (Figure 3.16). Since this type of fracture is not naturally-generated and is described as a drilling-induced fracture, it is ignored in this study.

3.4.1.2 Fracture Geometry, Frequency and Distribution

The total number of fractures (i.e., fracture frequency) and the average values of fracture height and width (aperture size) measurements of all fracture types measured from all cores are presented in Table 3.1. Figure 3.17a also shows the counts of all opening- and shear-mode fractures measured from the cores. The total number of all calcite veins, including planar calcite veins and folded calcite veins described from all cored intervals are presented in Table 3.2.

3.4.1.2.1 Fracture Length

The fracture frequencies are plotted against fracture cumulative lengths of opening-mode fractures in all cores (Figure 3.18). Opening-mode fractures observed from the cores range from cm-scale to m-scale in height. Core observations show that opening-mode fractures are commonly confined to their host beds with fracture height similar to the thickness of the fractured bed (Figures 3.6a-c and 3.10a, b and d). This is consistent with the observations of many core and outcrop fracture studies (e.g. Helgeson & Aydin, 1991; Gross & Engelder, 1995; Bai & Pollard, 2000). Planar veins show lengths higher than folded calcite veins (Figure 3.5). They vary in length from cm- to m-scale, whereas folded veins exhibit cm-scale lengths, with a maximum measured length of 10 cm. The height of a folded vein is represented by the distance between the two tips (Figure 3.5d).

The lengths of bedding-parallel veins are not surely identified in the core because in most cases, only small segments of bedding-parallel veins, with lengths equal to the core width, were measured (Figure 3.12).

3.4.1.2.2 Fracture spacing

Both fractured and non-fractured beds are observed in all core intervals. Non-fractured beds in the core might have a wide fracture spacing that is hard to measure due to the limited core width. Spacing of opening-mode fractures in the core is usually linked to the thickness of the fractured bed (Figure 3.19); fractures are closely spaced in thin beds, whereas greater spacings are measured in the thick beds. Some thin beds are observed to have a high density of closely-spaced thick fractures (Figure 3.5c). Cemented fractures are commonly more closely spaced than uncemented fractures. Folded veins are more closely spaced than planar veins; a cluster of folded veins can be seen within the core width (Figure 3.5c).

Shear fractures were normally observed in the core as discrete structures (Figure 3.7c), thus, measurement of fracture spacing was not applicable.

3.4.1.2.3 Fracture Aperture

Aperture data of opening-mode fractures measured from all cores are presented as average values in Table 3.1. The fracture aperture values are plotted against frequencies of opening-mode fractures in all cores to examine the aperture distribution (Figure 3.20). Scaling of fracture cumulative lengths and aperture sizes are plotted in Figure 3.21. The aperture size of all fracture types varies from mm- to cm-scale. Apertures of opening-mode fractures measured from the core typically narrow towards fracture tips, particularly for both folded and planar veins (Figures 3.5 and 3.8). They usually exhibit tapering tips.

Subhorizontal veins show fracture tips bending towards adjacent veins and in some cases, merging with them (Figure 3.12). The highest cumulative fracture aperture size is measured in the silicified and chert units in the Lower and Upper units. This emphasises the possible control of rock properties on the variation of the aperture size of fractures.

Folded veins have aperture values relatively higher than planar veins measured from all cores. This variation in the values of fracture aperture sizes may suggest different mechanisms and timing of fracture development.

Fracture aperture to Height (D/H) scaling relationship of opening-mode fractures (veins and joints) reveals a positive linear relationship (Figure 3.21). In general, fracture heights and aperture sizes are relatively higher in the places that show opening-mode fractures linked to major faulting than for those that are not related to faults. Fractures in the cores from C9, C10 and C11 have the highest averages of aperture sizes (Table 3.1).

3.4.1.2.4 Fracture Frequency and Distribution

The results of fracture frequency and cumulative length are plotted versus depth (Figures 3.22-3.28) to investigate the vertical distribution of fractures of each core in the different studied localities. Construction of fracture logs involved plotting the frequencies and cumulative lengths for each type of fracture (veins, joints, faults) in a specific colour. Fracture data were also displayed together with the wireline logs and geochemical data (Figures 3.24-3.28).

The core fracture dataset includes 11334 individual fractures within a total core thickness of 3212m (Table 3.1). Opening-mode fractures (veins and joints) are the most common fractures observed in the core (Figure 3.17a). They represent a total of 11079 fractures of all measured fractures. Among all fracture types, veins are the most abundant fracture type in all cores (Figure 3.17a; Table 3.1). Planar calcite veins form the biggest proportion between all calcite vein types observed; they form 83% of all measured veins in the cores.

Folded calcite veins form 9% of all measured veins in the cores (Figure 3.17b; Table 3.2). Calcite and silica bedding-parallel veins are observed in all cores, while gypsum bedding-parallel veins are only observed in frequent organic-rich mudstone intervals in the core from

C6 within the depth interval between 143m and 156m (Figure 3.12). Bitumen-bearing fractures are clustered in the lower silicified/chert intervals of the cored interval in all the studied cores (see Chapter 5).

Fractures observed in cores from the Harrana area (see Figures 3.23-3.26) are mostly cemented veins, whereas in the Azraq area, open fractures (joints) are more abundant, particularly in the Lower Unit (see Figures 3.27 and 3.28). In general, fracture intensity of all opening-mode fractures tends to increase with the proximity to the regional faults, and in the cores that include microfaults in their samples (Figure 3.29). In the latter, the opening-mode fractures mainly cluster in the fault damage zones.

Generally, the lower and upper units of the cored interval are significantly fractured in all of the Azraq and Harrana cores in central Jordan (C1-C12 and C14) (Figures 3.23-3.28). Cores C13 and C15, from Bayer (South Jordan), display fractured lower and middle units but a slightly-fractured Upper Unit (Figure 3.23). The Middle Unit, which is composed of chalky marl and bituminous mudstone, has commonly low fracture intensity for all cores except for C6, C10 and C11 in which the Middle Unit is relatively high fractured (Figures 3.23-3.26). This unit lacks the presence of chert and silicified intervals, whereas in the lower and upper units, chert and silicified intervals are highly abundant.

Fractures in the Lower Unit of the core are mainly opening-mode fractures (Figures 3.24-3.28). Veins in this unit are commonly filled by calcite. Bitumen can be found separately in fractures or together with calcite in calcite veins (see Figure 3.14). Both bitumen veins and calcite folded veins occur only in this unit. Fractures in the Lower Unit are extremely developed and clustered when they are associated with chert concretions (Figures 3.7a and 3.8a-c).

Fractures are occasionally observed in Middle Unit of the core and where present, they are mostly sealed (i.e., veins) (Figures 3.24-3.28). Core C10 however, shows relatively higher

fracture intensity compared to the Middle Unit of the other cores (Figure 3.28). Fibrous bedding-parallel gypsum veins are observed in this unit and are clearly observed in the core form C6 in organic-rich mudstone intervals (Figure 3.12). Several small fault zones are observed a few meters below these gypsum veins intervals. Carbonate concretions are infrequently observed in this unit.

In the Upper Unit, both opening-mode fractures and shear fractures (faults) are present (Figures 3.24-3.28) and are mostly cemented by calcite. Faults and breccia zones are restricted to this unit.

3.4.1.2.5 Fracture Relationships

Different relationships between fractures were observed in the core; the crosscutting relationships, fracture interaction and interconnection (connectivity). Crosscutting relationships are clearly observed in the core between the different types of opening-mode fractures (Figure 3.30). Bedding-parallel calcite veins are observed in the core to crosscut folded calcite veins (Figure 3.30a). Bitumen veins are also observed to crosscut planar silica veins (Figure 3.14a-b). No systematic crosscutting relationships between calcite veins and bitumen veins were documented. Planar calcite veins crosscut silica veins in many examples in the cored intervals (Figure 3.30b).

Interestingly, no crosscutting relationships are observed between fractures of the same type. This suggests that different fractures have developed in different stages.

Interaction between fractures is observed in the core, although the core width is relatively limited. This is manifest in the closely spaced and clustered fractures. In general, fractures are more interconnected where they are associated with chert nodules (Figure 3.7a). The interaction between fractures is obviously recognised in the core between the vertical and horizontal silica veins, bitumen-bearing fractures as well as, amid open fractures (Figures 3.6c, 3.14a-c and 3.31). Such interactions result in a complex fracture network that may

enhance connectivity (Meng, 2017). Folded veins exhibit no type of interaction or intersection. They occur usually as isolated or parallel arrangements with no interaction or connection to other fractures (Figure 3.5c).

3.4.1.3 Concretions and Concretion-Related Fractures

Concretions are abundant in the core intervals and are observed as four types (Figure 3.32). These are carbonate concretions (Figure 3.32a, d-e, g, and m), chert concretions (Figure 3.32k-l and n-o), composite concretions composed of a double-zone of carbonate and silica (Figure 3.32b, f, j and p), and carbonate concretions with chert shaped like triangles at the two edges of the concretion (Figure 3.32c). The complex concretions are present in two forms: (1) a chert concretion core enclosed by carbonate zone (Figure

3.32b, f and p) and (2) a carbonate concretion core surrounded by chert zone (Figure 3.32j). A complete concretion body is not always visible in the core sample; some concretions only show parts of their bodies in the core (Figure 3.32d). Wherever the complete body of carbonate and chert concretions is observed in the core, the concretions are of cm-scale in size (Figure 3.32a and c).

Both symmetrical and asymmetrical concretions are observed in the cores. Based on their shapes, the different types of symmetrical concretions observed can be grouped into four main types. These are spherical, elliptical, lensoidal, and tabular concretions (Figure 3.32). Chert concretions commonly show elliptical and lensoidal shapes, although irregular chert concretions also occur (see Figure 3.32n). Carbonate concretions occur mostly in symmetrical shapes, and are usually observed in spherical or elliptical shapes (Figure 3.32g and m). Lensoidal carbonate concretion are rarely found (Figure 3.32a).

Sedimentary laminae can be traced in most of the observed carbonate concretions (Figure

3.32g-h). This suggests a replacive growth of these concretions (Raiswell, 1971; Gaines & Vorhies, 2016).

Both carbonate and chert concretions are, in many cases, observed to be associated with natural fractures (Figures 3.7a, 3.8a-b, 3.14a, and 3.32j-k). There are three main relationships between fractures and concretions are observed. These include fractures confined to the concretion body (Figures 3.7a, 3.14a and 3.32i and m), fractures bend around the concretion (Figure 3.32e) and fractures crosscut the concretion (Figures 3.8a-b and 3.32e, n and o). In the latter, fractures usually tend terminate toward the concretion centre, with a highest fracture aperture at the concretion margin and a tapering tip towards the concretion centre (Figure 3.32i, m and o). In such a case, fractures are always cemented. The host-sediment beds, in many examples, reveal differential compactions around the concretions (Figure 3.32p). This may indicate replacive concretion growth.

3.4.1.4 Fracture Orientation and Inclination

Based on their orientations, natural fractures observed in the core were subdivided into three main categories including vertical fractures (e.g., vertical planar veins, vertical folded veins, and vertical joints), sub-vertical or oblique fractures (e.g., sub-vertical silica/calcite veins and microfaults), and horizontal fractures (e.g., bedding-parallel calcite veins and bedding parallel-gypsum veins) (Figure 3.33). A combination of two or three fracture types can also be present, forming a complex fracture network (Figure 3.6c and 3.14a).

Such fracture networks are usually observed in chert and silicified beds. Cone-in-cone structures are usually spreading within the host rocks laterally, but vertical strands are also present (Figure 3.13c-d).

Most vertical and sub-vertical opening-mode fractures display dip values range between 70° and 90° (Figure 3.33), whereas for horizontal and sub-horizontal fractures, they range

between 0° and 10°. Horizontal fractures are usually aligned parallel to the bedding planes (Figure 3.6).

Planar veins are observed as vertical, sub-vertical or horizontal veins (Figures 3.5a-b, 3.10 and 3.11). Folded calcite veins are usually vertical, with fold axes parallel to bedding (Figures 3.5c-d and 3.8). This suggests that these veins have been vertically shortened. Some calcite veins show radial orientations, particularly if they are associated with concretionary bodies (Figure 3.34). Shear fractures (faults) display dip ranges between 50° and 80° (Figure 3.35), with normal displacement.

The above-described fracture orientations were measured based on the pre-knowledge of that the cores have been drilled vertically by JOSCO, despite no information on core deviation being available. In addition, bedding is normally at right angles to the core. Due to the uncertain orientation of core, the true fracture strike was not constrained.

3.4.1.5 Definition of Mechanical Units in Core

The term “mechanical stratigraphy” was defined by Gross et al. (1995) to characterise different geological units based on the change in intensity of fractures, matching the lithologic or stratigraphic boundaries. The studied succession is generally subdivided into three main mechanical unit. These mechanical units strongly match the previously defined lithostratigraphic unit (Lower, Middle and Upper). The lower and upper units display high fracture intensity, whereas the Middle Unit show lower fracture intensity (Figures 3.243.28). More precisely, the three mechanical units were further subdivided into five mechanical subunits based on the rock properties and vertical distribution and intensity of different fracture types along the cored intervals. These subunits were also compared to the fracture data extracted from cored and non-cored wells in order to get a better understanding of the

lateral extent of these mechanical units. The depth intervals, thicknesses, and fracture intensities for the five mechanical units determined for each core are presented in Table 3.3.

Mechanical units are defined within the three lithostratigraphic units (Lower, Middle and Upper), identified in Chapter 2, for all the studied cores (Figure 3.36). The first three defined mechanical subunits (MU1-MU3) in all cores overlaps the Upper Unit. MU1 comprises mainly hemipelagic limestone, silicified limestone and chert. This unit has relatively high fracture abundance (Figure 3.36). MU2, which is ~10 m in average thickness, is composed mainly of chalky limestone and marly limestone and it has no fractures (Figure 3.36). MU3 comprises marly limestone, limestone and marl. This unit shows high fracture intensity in all cores except for core C9, which shows low fracture intensity (Figure 3.36). The fourth mechanical subunit (MU4) is represented by the homogenous organic-rich marl and chalky marl, which forms the Middle Unit. This unit displays low fracture in eleven cores (C1-C5, C7-C9 and C12-C15), but it shows a higher fracture intensity in four cores (C6, C10 and C11). This increase in fracture intensity is linked to the proximity of these cores to major faults (Figures 3.29). The fifth mechanical unit (MU5) is composed of phosphatic marl, silicified limestone and chert. This unit is defined as highly fractured mechanical unit in all cores, and is equivalent to the Lower Unit (Figure 3.36).

The high intensity of fractures in MU1, MU3 and MU5 is attributed to the lithology and rock properties of these units, since they are mainly composed of limestone and silicified limestone with abundant chert beds and concretions (Figure 3.4).

3.4.1.6 Microscopic Description of fractures

Microscopic investigation revealed varied types of host rock lithologies, fracture cements, internal structures and fracture interactions, indicating a complicated history of deposition, and fracture development.

3.4.1.6.1 Host Rock

Microscopic examination of the host rocks from the three stratigraphic units studied shows that the mudstone in the lower and middle units mainly consists of calcite, quartz, organic matter and apatite, with locally considerable amounts of chalcedony, and dolomite (Figures 3.37 and 3.38). Quartz is commonly of microcrystalline nature. Pyrite and clay are occasionally observed. Calcite has cemented most microfossil tests in the mudstone (Figure 3.37a). Radiolaria are locally abundant in the Lower unit (Figure 3.37c). Dolomite and Mg-rich carbonate are found in the calcareous matrix as distinct crystals (Figure 3.37a-c). The mudstone cement in the Lower Unit comprises mainly calcite and microcrystalline quartz. The quartz cement is pervasively present in three main types. These are pore-filling microcrystalline cement, micro-nodules, and as replacement of earlier calcite bioclasts (Huggett et al., 2017).

Microcrystalline quartz cement typically is of a biogenic source, with biogenic non-crystalline opal-A as radiolaria or diatoms (Figures 3.37c and 3.38). Relict of radiolaria and silica spheres (opal-CT) (Figures 3.37c and 3.38) are also found in the silica cement, indicating a biogenic origin of the cement. This observation is consistent with geochemical and petrographic studies of the same rocks (e.g., Fleurance et al., 2013; Huggett et al., 2017). The mudstone in the Lower Unit shows generally low silica content wherever chert concretions are present. Bitumen is observed as small particles, with no visible structure in variable shapes in the matrix of the organic-rich mudstones, in both the lower and middle units (Figure 3.37b-c). It is commonly observed in this unit infilling microfossil tests or in fractures.

3.4.1.6.2 Chert Concretions

Chert concretions are composed mainly (~95%) of microcrystalline quartz (Figure 3.39). Microfossil “ghosts” can be seen in chert concretions (Figure 3.39). Locally small amounts of calcite and pyrite are also present in the chert (Figure 3.39a). The chert concretions mainly show passive cementation of sediment. The concretions typically display gradational boundaries (Figure 3.39b).

Chalcedony-cemented radiolaria and foraminifera were observed inside the concretions (Figure 3.39). At the margin and outside the chert concretion, microfossils are cemented by calcite (Figure 3.39b). The overall texture suggests a progressive process of silicification that might have occurred during or after calcite cementation of microfossil tests.

3.4.1.6.3 Carbonate Concretions

Carbonate concretions are composed mainly (>90%) of calcite (Figure 3.40). The calcite is of both detrital and authigenic origin, although it is hard to infer the amount of the authigenic material (Huggett et al., 2017). Apatite occurs in cavities of fossil fragments (Figure 3.40). Carbonate concretions investigated in the Lower Unit of core are of replacive type. This is evidenced by the preserved sedimentary structure and laminae inside the concretion body, the differential compaction of the host-sediment beds and by the folding of grains around fossils inside the concretion (Figure 3.41b).

3.4.1.6.4 Folded Calcite Veins

Microscopic examination of folded calcite veins shows that these veins are entirely cemented by calcite (Figure 3.42). In most cases, the cement of folded veins shows crackseal texture (Figure 3.42c). The veins exhibit elongate-blocky texture. The crack-seal bands are folded with the fold of the vein (Figure 3.42c). This suggests that the sealing of the vein occurred

at an early stage. Calcite cement of the folded veins shows twin planes that occurred during the folding of the vein (Figure 3.42g). This is evidenced by the radial direction of the twin planes along the vein body (Figure 3.42g).

The veins do not contain any traces of bitumen. This indicates that bitumen veins postdate the development of folded veins. Stylolite is commonly observed in the fold hinges (Figure 2.42h).

Thick folded veins exhibit branching in the vein tips (Figure 3.42d). Thick folded veins are observed to be coexistent with thinner folded veins in the same interval (Figure 3.42a). Thin folded veins, similar to the macroscale veins, show vein folding greater than thicker veins (Figure 3.42a). Microfractures are observed within the inner arcs of both thick and thin folded veins in the form of micro-calcite veins (Figure 3.42). They are more intense in thick veins compared to thinner veins (Figure 3.42).

Microveins within the inner arcs are also folded, with fold axes aligned parallel to the fold axis of the main folded veins (Figure 3.42b-c). This implies a single direction and mechanism of formation and folding. The microveins appear as propagations of the crack-seal texture, but with a decrease in the thickness of bands towards the vein centre (Figure 3.42c).

Segments of silica veins are observed cross-cutting folded veins in many places (Figure 3.42e-f). This indicates that the silica veins postdate the folded calcite veins.

3.4.1.6.5 Planar Calcite Veins

At the microscopic scale, planar calcite veins are mainly characterised by a crack-seal texture (e.g., Ramsay, 1980), representing multiple generations of calcite crystals with elongate-blocky/syntaxial texture (Figure 3.43a-b and e). They appear under the microscope as several parallel bands of calcite cements (Figure 3.43a-b). Late sparry calcite crystals are observed in some of the studied samples (Figure 3.43b-c). Micro inclusions of bitumen and

host rock are observed within the cement of planar calcite veins aligned along crack-seal interfaces (Figure 3.43a-c and e). Microscopic analysis of bitumen-bearing calcite veins normally reveals at least two stages of carbonate cements.

Planar calcite veins tend to change their direction with the change of rock lithologies (Figures 3.41a and 3.43g-h). In other words, the veins change their direction if they cross two rock intervals with two different rock properties (i.e., marl and limestone, or marl and chert).

3.4.1.6.6 Planar Silica Veins

Silica veins are cemented by fibrous chalcedony or microcrystalline quartz (Figure 3.44). The chalcedony veins mainly exhibit fibrous textures (Figure 3.44). Vertical and subvertical silica veins are usually filled by microcrystalline quartz, and they exhibit blocky texture, whereas horizontal silica veins are cemented by chalcedony that is displaying fibrous antitaxial texture (Figure 3.44a-c). Micro-silica veins are always planar. They are usually present as a complex network in silicified and chert beds or concretions (Figure 3.44).

Microcrystalline-quartz veins usually crosscut fibrous chalcedony veins (Figure 3.44a-c). In general, silica cement within the silica veins is entirely free of bitumen inclusions. Multiple examples of horizontal silica veins are documented crosscut by bitumen veins (Figure 3.44d).

3.4.1.6.7 Bedding-Parallel Veins

3.4.1.6.7.1 Fibrous Bedding-Parallel Veins (Beef)

The internal structure of bedding-parallel veins and the growth direction of calcite crystals usually provide information about the opening mechanism(s) of vein (Bons et al., 2012). Fibrous bedding-parallel veins can generally be classified based on the direction of fibres

growth as antitaxial, ataxial and syntaxial veins (Bons, 2000; Hilgers & Urai, 2002a). Within a fibrous vein, usually, a median line is present along the centre of vein, and usually forms of small calcite crystals, rapidly outgrown during initial stages of vein development (Hilgers & Urai, 2002a).

Fibrous bedding-parallel calcite veins observed in the core are mainly described as syntaxial/elongate-blocky fibrous veins (Figure 3.45a-b). Calcite crystals in the syntaxial vein show crystal growth from the vein wall towards the vein centre. The veins display a median line with host rock inclusions (Figure 3.45b).

Bedding-parallel gypsum veins exhibit antitaxial growth in which the gypsum fibres grow from the median line towards the vein wall (Figure 3.45c-d). These veins are present in organic-rich mudstone intervals. These gypsum veins are usually observed in the core sample as sub-horizontal to horizontal.

3.4.1.6.7.2 Cone-in-Cone Structure

Cone-in-cone structures are composed of conical stacks of calcite fibres (Figure 3.45e-j). Not all of the cones show complete conical shapes; some cones are incomplete or appear as trimmed by other cones or the host sediment (Figure 3.45e-i). The conical shapes display height values ranging from mm- to cm-scale. The cones exhibit regular apical angles, with a narrow range (24°-30°). Usually, the cone-in-cone aggregates terminate in the host sediment with an upper, un-identified boundary, resulting in a gradual contact with the host sediment (Figure 3.45i). The lower boundary is, in contrast, clearly distinct (Figure 3.45i). The cone-in-cone contains abundant of fragments/inclusions of the host sediments trapped between the calcite cones as solid inclusions (Figure 3.45e and g-i). The boundaries of cones can be traced by the dark solid inclusions that separate the cones from each other (Figure 3.45g-i).

3.4.1.7 Geochemistry

3.4.1.7.1 Stable ($\delta^{13}\text{C}$ and $\delta^{18}\text{O}$) Isotopes

The $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ compositions were analysed for cements of each type of calcite structures (folded veins, planar veins, fibrous bedding-parallel veins, cone-in-cone, and faults) and their host rocks collected from the JOS cores (Table 3.4).

The calcite structures show a wide range of $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ compositions; from -20.70‰ to -1.0‰ and from -15.60‰ to -2.80‰, respectively (Figures 3.46 and 3.47). The $\delta^{13}\text{C}$ signatures show negative values, ranging from -14.90‰ to -7.50‰ for folded veins, from -20.70‰ to -4.0‰ for planar veins, from -9.10‰ to -5.10‰ for fibrous bedding-parallel veins, from -10.46‰ to -5.98‰ for cone-in-cone, and from -5.80‰ to -1.0‰ for faults (Table 3.4). The $\delta^{18}\text{O}$ isotopic composition values of planar calcite veins are between 14.50‰ to -4.60‰. They vary from -11.65‰ to -2.80‰ for folded veins, from -13.58‰ to -7.20‰ for fibrous bedding-parallel veins, -9.60‰ to -2.90‰ for cone-in-cone, and from -15.60‰ to -13.80‰ for faults (Table 3.4).

In general, vein carbonate cements show $\delta^{13}\text{C}$ isotopic composition values are comparable with their host rocks, although the vein carbonates are relatively more depleted in carbon than the host rock of each vein type (Figure 3.48).

Plots of $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ compositions of samples demonstrate the covariance of oxygen and carbon isotope signatures in all the calcite veins (Figures 3.46 and 3.47). Lower $\delta^{18}\text{O}$ values commonly correspond to higher $\delta^{13}\text{C}$ values.

The results show that folded veins and bedding-parallel veins are isotopically distinct from planar calcite veins and faults. This suggests two isotopically distinct groups of structures; early calcite structures and late calcite structures.

3.4.1.7.2 Strontium ($^{87}\text{Sr}/^{86}\text{Sr}$) Isotope

The $^{87}\text{Sr}/^{86}\text{Sr}$ values of the analysed calcite structures (planar veins, folded veins and bedding-parallel veins) range between 0.7077 and 0.7080 (Table 3.5; Figure 3.49). The host rocks of these structures have also a relatively similar range of $^{87}\text{Sr}/^{86}\text{Sr}$ values (0.7078-0.7080) (Table 3.5; Figure 3.49).

The calcite vein samples are clustered on the global sea water curve between 40 and 25 Ma, whereas the host rocks samples are grouped between 50 and 40 Ma (Figure 3.50). This suggests that the age of the vein cements is relatively younger than their host rocks, which may indicate diagenetic dissolutions in this period.

Planar veins reflect ages younger than that of both the folded veins and cone-in-cone structures (Figure 3.50). Folded veins show ages that are relatively closer to the age of the host rock samples.

3.4.1.7.3 U-Pb Radioactive Dating

U-Pb calcite geochronology is another technique used in this study to absolutely constrain the timing of development of different fracture systems. The relatively high concentrations of uranium in the samples allowed a unique opportunity to successfully apply this technique on the samples and to age-date the different fracture systems in the studied succession. The results of all calcite-cemented structures dated are shown in Table 3.6. The geochronological results reveal multiple generations of crystallization age of calcite structures (Figure 3.51). The data show that folded calcite veins are clustered around ~45 Ma, which is near the defined chronostratigraphic age of the rocks. One bedding-parallel vein sample displays a timing similar to the folded vein samples, and one shows a younger timing (Figure 3.51). Cements from microfaults and planar vertical veins reflect later age dates (Table 3.6; Figure

3.51). Data from planar calcite veins show a cluster along the ~25 Ma line, whereas faults show a radiometric age near 20 Ma (Figure 3.51).

3.4.2 Fracture Characterization from Borehole Image Logs

The main aim of using the BHIs in this study is to characterize fractures in the subsurface and attempt to correlate them with the fracture data measured from the cores. In addition, the BHIs provide a means to describe and measure fractures from the non-cored wells since the OPTV images will allow description of the different types of fractures and allow measurement of their orientation in both cored and non-cored wells.

3.4.2.1 Fracture Description

Many different structural features on the borehole wall are observed through the borehole image log, including opening-mode fractures (veins and joints) and shear fractures (Figure 3.52). Veins are observed in the BHIs by their contrast in mineralization, whereas unmineralized fractures (joints) are more difficult to identify (Figure 3.52 a-c, i and k-l). Joints are detected in the BHIs as dark lines (Figure 3.52f). Bitumen-bearing fractures are observed as black leaner traces (Figure 3.52d). Bedding planes are also clearly observed in the BHIs (Figure 3.52f and j) (Naeimipour, 2016).

Veins observed in the BHI include planar veins and folded veins (Figure 3.52a-b and k-l). Planar veins are detected in different rock lithologies such as mudstone, chert and carbonate, with a considerable cluster in chert beds and concretions (Figure 3.52g-h). Folded veins, as in the core, are only observed in mudstone beds that are close to or located between chert beds or concretions (Figure 3.52g-h). Folded veins are observed to be folded in the mudstone and they are straight wherever transecting chert (Figure 3.52k). A similar observation was also made from the core (see Figure 3.8a-b).

Vertical, continuous striations are sometimes observed in the BHI (Figure 3.52f). These are observed in the BHIs as white traces parallel to the borehole axis. Such traces are not observed in the equivalent core intervals.

3.4.2.2 Fracture Frequency and Distribution

The fracture population described from the high-resolution BHIs are mainly of opening mode types. They form a total of 4599 fractures. These include 2125 veins and 2474 joints (Figure 3.53). A strong match is recognised in the fracture distribution between the core and BHIs; fractures mainly cluster in the lower and upper silicified/cherty unit of the study succession. This gives confidence for the reality of the fracture analysis based on BHI data. Veins are more clustered than the joints (Figure 3.52).

3.4.2.3 Fracture Orientation

Veins (especially planar veins) and joints, both exhibit a pronounced NNW-SSE fracture trend (Figure 3.54). This is parallel to the trend of the regional normal faults in the area (see Figure 2.8; Chapter 2) and consistent with the present-day S_{HMAX} orientation. This strongly suggests that most of the planar fractures including planar calcite veins and joints observed in the BHIs are of tectonic origin. While planar calcite veins are usually present in parallel arrangement (Figure 3.55a), folded calcite veins, however, are observed to be in an irregular arrangement (Figure 3.55b).

In general, veins exhibit a smaller orientation range than joints (Figure 3.54). Fracture orientations in the Harrana area are distributed more heterogeneously than in the Azraq area (Figure 3.54). This may be an attribute of the proximity to regional faults in the Harrana area, including the Suwaqa and the Zarqa Ma'in faults (see Figures 3.29, and Figure 2.8 in Chapter2).

3.4.2.4 Concretions

Both chert and carbonate concretions are abundant in the BHI logs (Figures 3.52c and k, and 3.55b-c). They are observed in different shapes, including spherical, elliptical and lensoidal. Examples of carbonate concretions are observed, as in the core, to have two triangular regions at their margins that are filled by chert (Figure 3.55c). Irregular carbonate and chert concretions are also observed (Figure 3.55c). Fractures are observed normally to be associated with concretions (Figure 3.55c).

3.5 Strain Analysis

Strain analysis from core offers an opportunity to examine the amount of deformation that rock units have undergone during burial and uplift (Fossen et al., 2016). It can also provide significant information about the mechanisms of deformation. Loose sediments compact through progressive burial (Garrison, 1981), or when experiencing diagenesis and lithification in different degrees. Mud is highly compactable; it can compact approximately 50% or more of its original volume (Chapman, 1983). Many deformational features and structures such as layer thinning and bending of laminae around concretionary bodies (i.e., differential compaction), stylolite, or deformation of pre-existing structures can result by compaction. Pressure solution and stylolite are good indicators of features resulting from compaction of carbonate rocks such as limestone (Norman, 2015).

Strain was measured in the present study around veins and concretions in order to evaluate the potential role of compaction and diagenesis in fracturing. The distinct lamina boundaries in the beds that host concretions (Figure 3.56a-b) allow a unique opportunity to calculate the compaction degree around concretion (Hooker et al., 2017a).

The compaction degree of the folded veins was calculated, with the assumption of an original planar form and with respect to the surrounding hostrock grains. The compaction ratio is calculated for folded veins (C_v) from core as follows (Hooker et al., 2017a):

$$C_v = \frac{L_0}{L_1}$$

Where,

L_0 : is the length of the original vein before folding (original un-compacted length). This was measured by tracing the meandering trace of the folded vein.

L_1 : is the straight distance between the tips of the current folded vein (final compacted length) (Figure 3.56c-d).

The actual length of the vein was carefully measured by avoiding any spacious lengths resulting from slabbing or inclination. The compaction ratios for the folded veins show a narrow range (from 1.24 and 2.57), with an average of about 1.75 (Table 3.7; Figure 3.57). Folded veins from the Azraq area show the highest compaction ratios (average of 1.87) compared to compaction ratios of other folded veins measured from Harrana and Bayer (Table 3.7). The thickness of the folded vein significantly influences the compaction ratio; thin veins were folded much more than thick veins within a single bed (Figure 3.56c).

Compaction strain analysis was also performed around replacive chert and carbonate concretions observed in the core. The compaction here is represented by the deformation occurring in the lamina thinning around concretions (Figure 3.56a-b). The compaction ratio of the thicknesses of the deformed beds in the strain field was measured, both within and around concretions and in the non-deformed part of the same beds. According to Hooker et al., (2017a), the concretion compaction ratio (C_N) can be calculated as follows:

$$C_N = \frac{T_N}{T_L}$$

Where,

T_N : is the original thickness of a stratigraphic package across the concretion vertically.

T_L : is the compacted thickness of the same package lateral to the concretion.

The results of this analysis are displayed in Table 3.8. Compaction ratios quantified for the chert concretions display a range between 1.33 and 3.67 (with an average of 2.18), whereas for the carbonate concretions, they range between 1.71 and 6.58 (with an average of 3.33). Chert concretions and folded calcite veins show similar ranges of amount of compaction (see Discussion, Section 3.6.3.1).

Strain was also analysed within fault damage zones in order to characterize the strain fields within the deformed rock units. This analysis was performed by using a linear scanline measuring technique (Marrett, 1996; Ortega et al., 2006). The scanline here is a line highlighted on the core sample with a length equal to the core width. Different regular intervals were used between scanlines. These include 5, 10 and 20 cm intervals. For each scanline, the opening displacements of different fracture types crossing the scanline were measured, and the strain percentage is calculated by the following equation:

$$\% \epsilon = 100 * \Sigma L / \Delta L$$

Where

$\% \epsilon$: Strain Ratio

ΣL : is the sum of all opening displacements of all fractures crossing a scanline

ΔL : is the difference between the scanline length and ΣL

An example of the measurements made using the scanline method is displayed in Figure 3.58. The results of strain analysis were plotted versus depth (Figure 3.59).

3.6 Discussion

3.6.1 Fracture Distribution

3.6.1.1 Vertical Distribution

Vertical core and borehole image logs afforded a robust database to study the different controls on the vertical distributions of the different fracture systems in the studied succession. Data from both cores and BHIs show that the fracture distribution is in general controlled by the rock lithology (i.e., rock properties). It is widely documented that the variation in the mechanical properties of rock units can create heterogeneity in fracture intensity and spacing (e.g., Gross & Engelder, 1995; Laubach et al., 2009; Evans et al., 2012). The lithologic control on fracture distribution is clearly reflected by two key observations. Firstly, most of the opening-mode fractures observed are confined to single beds or to individual concretions. At the bed scale, fractures usually terminate at the bed boundary, particularly with the change of the lithology between beds (Figures 3.5 and 3.7a-c). The second observation is reflected by the high clustering of fractures in the chert and silicified units, compared to the low fracture intensity in the marl units (Figure 3.36). This predominance of fractures in beds with high mechanical stiffness (brittleness) indicates that the development of fractures has been influenced by the physical and mechanical properties of the rock, represented by the small-scale mechanical stratigraphy. It also indicates that both direct (e.g., volume change, silicification) and indirect (e.g., local overpressure caused by chert formation) diagenetic processes have controlled fracturing at least locally, in the succession. Diagenesis is also considered as an important control on the properties of fracture systems (Laubach et al., 2009; Spence et al., 2014), including their distribution, particularly for the folded calcite veins. Preferential fracturing in chert and silicified intervals may result from variations in mechanical stratigraphy with time produced by diagenetic

changes occurring in the rock (Shackleton et al., 2005; Laubach et al., 2009). The interbedding between chert and mudstone beds in the Lower Unit is believed to be the result of diagenetic alteration of locally abundant concentrations of radiolaria and diatoms (Hooker et al., 2017a). The chert is believed to have developed by replacing carbonate components in the original host material with silica from biogenic components (Powell & Moh'd, 2012; Hooker et al., 2017a; Huggett et al., 2017), which is consistent with the low silica content of the adjacent mudstones. Supportively, the exclusive presence of folded calcite veins in the Lower Unit in all cores in the mudstone intervals positioned between chert or silicified intervals (see Figure 3.8), suggests a diagenetic control on the development of these veins (see Section 3.6.3 for more detail).

The vertical anisotropy in fracture distribution allowed the division of the section into three different mechanical units (Lower, Middle and Upper) that are subdivided further into five mechanical subunits (MU1-MU5), based on the presence or absence of fractures (Figures 3.24-3.28 and 3.36; Table 3.3). Mechanical stratigraphy divides layered rocks into mechanical units based on their physical properties such as tensile strength and elastic moduli (e.g. Laubach et al., 2009). The stiffness character of the chert and silicified intervals in the highly fractured mechanical subunits (MU1, MU3 and MU5) is argued here to play a significant role on fracture development in these units. Veins in the Middle Unit (MU4) seem to be widely spaced and were not observed due to the limited core width. This is evident from the field observations from the equivalent interval (see Chapter 4 for more details).

Another important control on fracture intensity and spacing is the bed thickness (Price, 1966; Ladeira & Price, 1981; Pollard & Aydin, 1988; Wu & Pollard, 1995; Bai et al., 2000; Schöpfer et al., 2011); fractures show smaller spacing values in thin beds, whereas higher spacing values were measured in thick beds. The strong, positive linear relationship between

fracture spacing and bed thickness (Figure 3.19) reflects the control of mechanical properties of rocks on fracture distribution. The high density of closely-spaced thick veins in some thin beds indicates a further opening of pre-existing fractures has occurred due to increasing stress (Bai et al., 2000).

3.6.1.2 Lateral Distribution

Regional correlation between wells, including both core and cuttings wells, provided a good means to test the lateral distributions of the different fracture systems (see Discussion Chapter, Chapter 6, for regional fracture distribution). The cored intervals in all cores show that the fractures are abundant in the Lower and Upper units and low fracture intensity in the Middle Unit (Figures 3.23-3.28).

An important regional control on fracture intensity is the influence of major faults (Spence et al., 2014). Some fracture systems in the studied succession, such as planar calcite veins and shear fractures show high fracture intensity in the wells that are located close (<5 Km) to major faults (Figure 3.29). This indicates that the development of such fractures in these areas was influenced by tectonic stress. This is also supported by the preferred NNW-SSE fracture orientation of planar fractures (i.e., planar calcite veins and joints) measured from both BHIs, which shows fracture trend (Figure 3.54) that is parallel to the trend of the regional normal faults in the area (Figure 3.29) and consistent with the present-day S_{HMAX} orientation. This result is consistent with the results obtained from core, particularly from the geochemical and petrographic analysis that constrained these fractures as tectonic fractures.

3.6.2 Fracture Timing

In most studies, the timing of the natural fracture systems can be accomplished only in a relative sense by the study of the crosscutting relationships between fractures, from

examination of the internal structure of veins and their crystal growth, or by the study of the stable isotopic signatures of veins and their host rocks (Beaudoin et al., 2015; Hooker et al., 2018). A few studies have constrained the absolute timing of veins or fault-related veins using U-Pb dating (e.g., Uysal et al., 2011; Kolodny et al., 2014; Hansman et al., 2018, Beaudoin et al., 2018; Parrish et al., 2018) and faults (e.g., Roberts & Walker, 2016; Nuriel et al., 2017). Most of these studies, however, used the U-Pb to constrain the veins and faults from carbonate or sandstone rocks, but not in mudrock-hosted fracture systems. The present study combines all the above mentioned approaches in order to constrain the timing of the diverse fractures systems observed in the Upper Cretaceous-Eocene strata.

3.6.2.1 Crosscutting Relationships

Crosscutting relationships are important for constraining the relative timing of fracture development (Beaudoin et al., 2015; Hooker et al., 2018). The different crosscutting relationships observed between the examined fractures reveal several phases of fracture development. The lack of crosscutting relationships between fractures of the same type indicates that each type of fracture has developed in an individual stage. At both the microscopic and the core scales, the crosscutting relationship between vertical bitumen-bearing fractures and silica veins (Figures 3.14a-b and 3.44d) indicates that bitumen-bearing fractures postdate silica veins. The crosscutting relation between planar calcite veins silica veins reported in many examples (Figure 3.30b) implies that calcite veins postdate silica vein. The absence of crosscutting relations between calcite veins and bitumen-bearing fractures together with the presence of bitumen inclusions in the cement of calcite veins indicate that both calcite veins and bitumen-bearing fractures are coeval (Figure 3.60). In addition, the crosscutting relationship between bedding-parallel calcite and folded calcite veins implies that folded veins are older than bedding-parallel veins (Figure 3.30a).

3.6.2.2 Stable and Radiometric Isotopes

Fracture cement provides a record of fluid composition and temperature (Hooker et al., 2015), which in many cases help in constraining the timing of fracture development. It is also believed that cemented fractures can form in a closed system, within a hydrostatic regime (Blyth et al., 2004; Verlaquet et al., 2011), or can open in an open system, among significant fluid advection (Elburg et al., 2002; Nollet et al., 2009; Fall et al., 2014). In the case of the JOS, the compatibility in the isotopic results between vein cement and host rock lithology implies a fracture formation mechanism in a chemically-closed system. This is reflected by the results obtained from both the stable and strontium isotopes that show relatively similar ranges between the vein-filling carbonate and hostrock carbonate of each vein type examined (Figures 3.48 and 3.49), suggesting that calcite precipitated from local paleofluids.

In addition, most of the calcite veins were detected in the mudstone intervals, which have a low permeability nature, combined with the observation that the calcite veins are most commonly observed to be entirely filled (i.e., not partially) suggests that the cementation of the veins have mainly occurred during their opening.

Stable isotope signatures of the calcite veins in the Upper Cretaceous-Eocene strata allowed the constraint of the timing of vein formation and the source of carbonates and fluids contributed to the calcite precipitation. The carbon and oxygen isotopic compositions of the early structures (i.e., folded calcite veins) are consistent with the low-temperature precipitation of carbonate derived from organic material, while the isotopic results of the planar calcite veins are consistent with higher-temperature precipitation and dissolution and reprecipitation of host-rock carbonates.

The results of the isotopic analysis, including the stable ($\delta^{13}\text{C}$ and $\delta^{18}\text{O}$) and radioactive ($^{87}\text{Sr}/^{86}\text{Sr}$ and U-Pb) isotope results divide the calcite structures examined into two main groups based on their timing: (1) early calcite structures (e.g., folded veins, bedding-parallel veins) and (2) late calcite structures (e.g., planar calcite veins, micro-faults) (Figures 3.48, 3.50, 3.51, and 3.60). Early structures such as folded calcite veins measured from the studied strata, based on the U-Pb age-dating, are strongly comparable to the chronostratigraphic age of rock deposition (Figure 3.51).

To summarise, crosscutting relationships between fractures have revealed that fractures have formed in chronological order; folded veins formed in the following by bedding-parallel veins, silica veins, then planar calcite veins and bitumen veins (Figure 3.60). Interestingly, the different geochemical techniques used in this study to constrain the timing of fracture development of the different fracture systems observed in cores and BHIs, show great consistency in their results and provide a further order of the sequence of events. Both stable and radiometric isotopes show that the fracture systems have two main different timing constraints; (1) early fracture development resulted in folded calcite veins and bedding-parallel veins and (2) late fracture development resulted in planar calcite veins, joints and micro faults.

3.6.3 Fracture formation mechanisms

The different characteristics and timing of natural fracture systems observed in the Upper Cretaceous-Eocene succession of Jordan have been argued above. In this section, the discussion will be focused on the possible genetic mechanisms of the different fracture systems. The fracture formation mechanisms of the different structures observed in the studied succession are discussed in detail in Chapter 6.

3.6.3.1 Folded Calcite Veins

It has been widely argued that diagenetic processes can provide an explanation of the fracturing in mudrocks (e.g., Laubach et al., 2009; Hooker et al., 2017a). In the case of the JOS, folded calcite veins are interpreted here to be formed amid significant sediment compaction resulting from volume change caused by silica diagenesis. The presence of folded veins in the Lower Unit of all cores distributed across Jordan (East, West and South) suggests that the sediments in all basins have undergone compaction at the same time and at a relatively similar burial depth. The wide distribution of folded veins implies a mechanism that is not dependent on local structural context, but rather it is closely related to primary lithology and diagenesis.

Volume loss during silica diagenesis has been thought to usually produce silica-filled fractures (Eichhubl & Boles, 1998; Maher & Shuster, 2012). However, it is not absolutely evident what type of cement should fill fractures created by volume loss that results from silica diagenesis (Hooker et al., 2017a). The present study argues that the volume loss or redistribution associated with silica diagenesis has led to opening the folded calcite veins. The exclusive calcite filling of the folded veins indicates that the pore water was supersaturated by calcite. Silica was present together with calcite as cements of the host mudstone (Hooker et al., 2017a), as revealed by the calcite-filled radiolaria tests (Figures 3.37 and 3.38). It seems that the silica cement was depleted locally during the silicification process from the host mudstone to form the chert intervals. The chert concretions therefore are thought to have been produced from local silica remobilised from the mudstone beds. This is also evidenced by the presence of replaced carbonate tests "ghost" (Figure 3.39) and the sedimentary laminae (Figure 3.8a-b) preserved inside these concretions.

The fracture distribution, early timing (U-Pb age date compared to chronostratigraphic age), and strain analysis of the folded veins are all supportive of silica diagenesis as the main trigger of fracturing. The mudstone is of high organic matter content. The original components of silica that were mobilized during fracturing are thought to be of biogenic origin based on the local occurrence of radiolaria and the stratified nature of chert beds (Powell & Moh'd, 2012; Huggett et al., 2017; Hooker et al., 2017a). In addition, the bulk rock strontium ratios are close to the global Paleogene seawater values (Figure 3.50). The chert seems to have been created from the transformation of biogenic opal-A to opal-CT and quartz (Knauth, 1994). This is evident from the regular intervals of chert in the Lower Unit with the presence of radiolaria. Folded calcite veins are usually present within mudstone beds adjacent to chert concretions (Figure 3.8a-c). They are observed to be extremely folded in the mudstone beds whilst they are planar where they cut chert beds or concretions (Figure 3.8a-b). Biogenic silica transition from opal-A to opal-CT and then to quartz during early burial usually results in volumetric change in semi-lithified silica (Kastner et al., 1977; Behl & Garrison, 1994), either as by volume redistribution or volume loss. This volume change can cause deformational structures in the rocks (Eichhubl & Behl, 1998; Davies et al., 1999; Davies, 2005; Hooker et al., 2017a). Silica diagenesis could have occurred at this shallow burial depth at warm bottom water temperatures (Huggett et al., 2017). Precipitation of silica phases is highly controlled by temperature; hydrous opal forms at low temperature and transforms to the more stable quartz at higher temperature (Williams et al., 1985). The temperature for opal-A to opal-CT transformation ranges between 18 and 56 °C, with respect to the clay content of the sediment (Keller & Isaacs, 1985; Knauth, 1994; Huggett et al., 2017; Jurkowska et al., 2019). However, opal-A to opal-CT transformation also occurs at lower temperatures (~5 °C) (Bennett et al., 1999). The low clay content of the marl beds

raises the likelihood of opal-CT and quartz precipitation at low temperatures (Huggett et al., 2017).

Strain analysis from core offers an opportunity to examine the amount of deformation that rock units have undergone during burial and uplift (Fossen, 2016). It can also provide significant information about the mechanisms of deformation. The consistency between the compaction ratios of chert and folded veins (Tables 3.7 and 3.8) suggests that the folded veins and chert concretions were relatively formed within the same stage, at a relatively early stage in the compaction history.

Geochemically, the homogeneity between the vein cement and the host rock lithology suggests a fracture system that developed under limited advective fluid migration, in a generally chemically closed system, consistent with the low permeability of these fine-grained rocks. The development of cement is generally controlled by the dissolution and precipitation of their forming minerals (Huggett et al., 2017). The early folded calcite veins have slightly depleted $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ (Figure 3.47). $\delta^{18}\text{O}$ signature reflects both the fluid composition and the precipitation temperature, with progressively lighter oxygen with elevating temperature. The $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ isotopic signatures of folded veins are consistent with low temperatures of precipitation. The carbon was probably produced by microbial degradation of organic matter, which was abundant in this setting. This shows that bacterial activities have contributed to the precipitation of carbonate in the veins (Marshall, 1982; Hooker et al., 2017a; Meng et al., 2017), building up local early overpressure associated with methanogenic bacteria (Meng et al., 2017), that consume organic material at shallow burial (Hooker et al., 2017a). The resulting overpressure has contributed to the fracturing in this case. The high organic content in the sediment could have also caused a reduction in the effective permeability for fluids that resulted in also an increase overpressure.

At the microscale, indistinct carbonate cement also filled microfractures as they opened and were folded. The study of microfractures helps in understanding the mechanism and timing of their development (Anders et al., 2014). The development of microfractures associated with folded veins is argued here to be synchronous with sediment compaction and vein folding. The unidirectional fold axis of both folded veins and their associated microfractures (Figure 3.42b) suggests that they both were subjected to the same direction and possibly the magnitude of compaction and vein folding.

The progressive folding of the folded calcite veins reflected by the folding of the crackseal bands with the folding of veins, indicates that the sealing of the veins occurred at an early stage, during the vein formation. The crack-seal texture also implies concurrent opening and sealing of the veins with calcite (Hooker et al., 2017a). The folding of the calcite twin-planes and their radial arrangement along the vein body (Figure 3.42g) provide additional evidence that the sealing of the folded veins occurred during the development of the vein.

A final line of evidence for the role of diagenesis in controlling the development of folded veins is the non-parallel arrangement and the tendency for wide dispersion of strike values of folded veins. This indicates that vein formation was not influenced by tectonic stresses, but instead diagenetic processes were the main trigger, which create in most cases, a disorganized or non-parallel arrangement of fractures (Maher et al., 2017).

3.6.3.2 Planar Calcite Veins

The timing of the development of planar calcite veins identified from the isotopic analysis of their cement displays these veins formed late and coincides with the age of faults (Figures 3.47, 3.50 and 3.51). They have isotopic ratios compositions comparable to late calcite in fault zones. Planar calcite veins have relatively more depleted oxygen than other calcite-cemented structures, indicating higher temperature of precipitation (Figure 3.46). These

veins have $\delta^{13}\text{C}$ signatures closer to those of seawater, which may reflect dissolution/reprecipitation of host-rock carbonate materials. The $^{87}\text{Sr}/^{86}\text{Sr}$ data of planar calcite veins is also consistent with the carbon isotopic values, implying a dissolution/reprecipitation of carbonate. Strontium is preserved in the calcite during the dissolution and reprecipitation process (Elburg et al., 2002).

The veins are interpreted here to have formed by tectonic extension. The increase of the intensity of these veins with the proximity to major faults (Figure 3.29) indicates that the development of these veins was influenced by tectonic stress conditions. In addition, the vein orientation also shows a general trend consistent with the trends of major faults in the region (Figure 3.54), indicating a tectonic genetic formation of these veins. Another line of evidence is that these veins are usually present in the core and the BHIs in parallel arrangements (Figures 3.10 and 3.55) compared to the folded veins and concretion-related veins, which tend to be arranged in non-parallel patterns. This suggests that the development of planar calcite veins was influenced by greater differential horizontal stress (Hooker et al., 2017a), that might also be expected at greater burial depth, in a tectonically imposed stress field, but not by fluid overpressure or diagenesis. Fluid overpressure usually results in irregular fracturing or brecciation on the rock (Parnell & Carey, 1995). In contrast to the folded calcite veins, planar calcite veins are interpreted to postdate sediment compaction based on the presence of the late sparry calcite crystals (Figure 3.43b-c). In addition, the twin planes of planar calcite veins are thinner than those the thick twins in the folded calcite veins. The elongate-blocky texture of the planar calcite veins reveals a moderate fracture opening rate (Prajapati et al. 2018) (Figure 3.43). Elongate-blocky crystals demonstrate favoured orientation at the final phase of vein growth. This indicates a growth competition during the development of vein (Hilgers & Urai, 2002a; Prajapati et al., 2018). Planar calcite

veins display crack-seal texture which suggests repeated incremental opening and sealing during vein development.

3.6.3.3 Bedding-Parallel Veins

Bedding-parallel veins are possibly forced to open by an increase in the internal pressure. This can result from the mechanical difficulty of reaching vertical extension in rocks under significant overburden. Therefore, horizontal fracturing likely indicates significant fluid overpressure in the studied succession. Fibrous bedding-parallel veins and cone-in-cone structures in shales are thought to develop by natural hydraulic fracturing caused by overpressure during primary oil migration (Stoneley, 1983; Rodrigues et al., 2009). Therefore, such veins start to develop along bedding planes once the shale enters the oil generating window and produces enough amounts oil and gas (Parnell et al., 2000; Lash & Engelder, 2005; Cobbold et al., 2013).

Bedding-parallel fibrous veins occurring as flattened or lenticular segments were found in the organic-rich calcareous mudstone (shale) of the Upper Cretaceous-Eocene succession of Jordan. Fibrous calcite veins are likely to form by fluid advection through open space in the vein (Hilgers & Urai, 2002a; Bons et al., 2012), with the presence of fluids supersaturated by calcite. The internal structure of fibrous veins allows outstanding opportunities to understand fracture development by analysis of the kinematics of vein formation (opening process), the source of fluids and the mode of fibre growth (Bons et al., 2012). However, there is a lack of comprehensive studies that combine different methods such as sedimentology, petrography and isotopic geochemistry to explain the genesis of fibrous calcite veins in shales (e.g., Marshall 1982; Al-Aasm et al., 1993), although significant information on the diagenetic process and evolution of the formation fluids in host sediments were documented (Yu et al., 2015).

At the microscale, fibrous bedding-parallel calcite veins observed in the core display syntaxial fibrous texture (Figure 3.45a-b), in which calcite crystals tend to grow from the vein wall towards the vein centre. This texture leads to the interpretation of that the veins may have formed at relatively shallow burial depth, where fluid overpressure generated by the calcite-supersaturated fluids may have created the bedding-parallel vein space (Yu et al., 2015, Meng et al., 2017).

Hydrocarbon (bitumen) observed as inclusions in the bedding parallel calcite veins suggests that the bitumen was present during the formation of these veins (Figure 3.45a-b).

In the case of the fibrous bedding-parallel gypsum veins, the veins show antitaxial texture (Figure 3. 45c-d), which suggests that the veins developed under dilation coexistent with mineral precipitation (Bons et al., 2012; Gasparrini et al., 2014). The development of gypsum veins requires sulphate-rich fluids to precipitate gypsum, which in the case of the JOS can be derived from the sulphur-rich host mudstone.

Cone-in-cone structures are widely documented and are believed to develop early under shallow overburden load, where sediments deform in a plastic deformational way (Woodland, 1964; Selles-Martinez, 1996; Cobbold et al., 2013; Hooker & Cartwright, 2016). They are important in understanding the diagenetic evolution of sedimentary basins (Cobbold et al., 2013). Hooker and Cartwright (2016) argued that the obvious vertical displacement of the host material in the JOS core samples within the observed cone-in-cone suggests that the overburden compressive load was balanced during calcite precipitation. This could have been achieved by either static fluid overpressure (SellésMartínez, 1996), seepage forces (Cobbold & Rodrigues, 2007; Maher et al., 2017), or the force of crystallization (Richardson, 1923; Franks, 1969) during early diagenesis at shallow burial depth, while the overburden stress is low. However, the counterbalancing could have also

occurred during deep burial, with significant fluid overpressure or at late stage, during exhumation of the rocks (Hooker & Cartwright, 2016). The present study concurs with the first assumption in which cone-in-cone formed during shallow burial.

This is evident from the geochemical isotopic analysis of this structure and the host rock. The timing of the development of cone-in-cone, based on U-Pb dating indicates that it formed at an early stage (Figure 3.51). This early structure has slightly depleted $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ similar to the case of the folded veins (Figures 3.46-3.47). The $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ signatures are consistent with low temperature of precipitation from original formation water (seawater). Another piece of evidence for early formation of this structure is that the cone-in-cone pattern is likely to form if the calcite fibres and microfossils were rigid and host material poorly consolidated (Hooker & Cartwright, 2016). This could only occur at early burial. The development of the cone-in-cone seems to have started by a deformation of the host material forming a cone structure, which could have triggered other cones to form in the shape we see now (Figures 3.6a and 3.45e-j), in patterns similar to those fractures forming as en échelon arrays (Olson & Pollard, 1991; Hooker & Cartwright, 2016).

3.6.3.4 Bitumen-Bearing Fractures

Low permeability and a high organic content are critical to fracturing of rocks (Mandl & Harkness, 1987). The development of bitumen-bearing fractures in the studied succession seems to have been controlled, or at least assisted, by hydraulic fracturing, driven by lithostatic fluid overpressure in the organic material. The splaying of bitumen veins and brecciation in the low-permeability, organic-rich mudstone (Figure 3.14c), are good examples of hydraulically-fractured rock due to fluid overpressure (Parnell & Carey, 1995; Passchier, 2000). Fracturing driven by fluid overpressure with regard to maturation of organic matter has been reported in many studies (e.g., Jochum et al., 1995; Zhang et al.,

2016). It is assumed that volume expansion of kerogen during thermal maturation is the main trigger for fracture formation (Michels et al., 1994; Figueroa Pilz et al., 2017). In the case of the JOS, the generated bitumen is of early maturity. However, the yield of this early expelled bitumen seems to have been high enough to cause a volume change that resulted in fracturing.

The aperture irregularity of bitumen veins in silicified and chert intervals (Figure 3.14a), and the change of vein direction when they propagate in different beds indicate that fracturing was controlled by the rock properties (Bons et al., 2012).

Petrographically, the development of bitumen filaments into well-developed micro veins within the organic-rich mudstone beds also indicates fracturing due to high pore pressure (Figure 3.38).

Chapter 5 presents a comprehensive description and discussion of the development of bitumen-bearing fractures.

3.6.3.5 Shear Fractures

The late shear fractures (faults) observed in the core are consistent with significant tectonic deformation of the rocks, based on the vertical distribution of these faults in the cored intervals (Figures 3.24-3.28), their relationships to the patterns of major faults and the amount of displacement along cored faults. The presence of faults in the Upper Unit of the cored intervals indicates a late formation of these structures. This is also supported by the geochemical analysis performed on calcite cement filling these faults. They are present as either single faults, segmented microfaults or a fault zone, with a fault core and damage zones (Figures 3.7c and 3.15a and e-f). The fault-damage zones of the observed faults include a cluster of opening-mode fractures mainly, planar veins and joints, but not early folded calcite veins (Figure 3.15e). Some segmented microfaults observed are overlapped,

indicating fault propagation by segment linkage (Cartwright et al., 1995). The radiometric dating of calcite-cemented faults (Figure 3.51) reveals that the faults have a younger age (20Ma) than planar calcite veins. This may indicate that the basin has experienced the formation of opening-mode fractures and then proceeded into faulting or fault reactivation. The presence of bitumen in the fault core is an indicator of that a fault was permeable during hydrocarbon migration.

3.7 Conclusions

- 1) A wide variety of natural fractures has been observed and measured from core and image logs. These include veins, cone-in-cone structures, joints, bitumen-filled fractures, and faults. Opening-mode fractures are the most common observed fractures.
- 2) Rock properties and burial depth play a main role in controlling fracture distribution and development.
- 3) Fracture minerals are related to the host-rock lithology and therefore, the system is considered to be chemically closed at the time of fracture development and sealing.
- 4) Petrographic examinations and crosscutting relationships reveal that natural fractures in the JOS can be divided into two main groups: (a) early fractures including folded calcite veins and bedding-parallel veins and (b) late tectonic fractures including planar calcite veins and faults.
- 5) $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ signatures of the early structures (-14.90‰ to -7.50‰) are consistent with the low-temperature precipitation of carbonate derived from organic material while isotope results of late planar veins (-20.70‰ to -4.0‰) are consistent with

higher-temperature precipitation and dissolution and reprecipitation of host-rock carbonates.

- 6) Different mechanisms have controlled fracture formation and development in the Cretaceous-Eocene section that involves the Jordan Oil Shale and under- and overlying chert-rich units. These include fluid pressure (e.g., bedding-parallel calcite veins), early diagenesis (e.g., folded calcite veins and planar silica vein) and late tectonic processes (e.g., planar veins, joints and faults).
- 7) Diagenetic processes have occurred throughout the burial history, and started at an early stage (during sediment compaction).
- 8) Folded veins formed very early, prior to any significant lithification of the host sediment and were later folded by vertical compaction during early diagenesis.

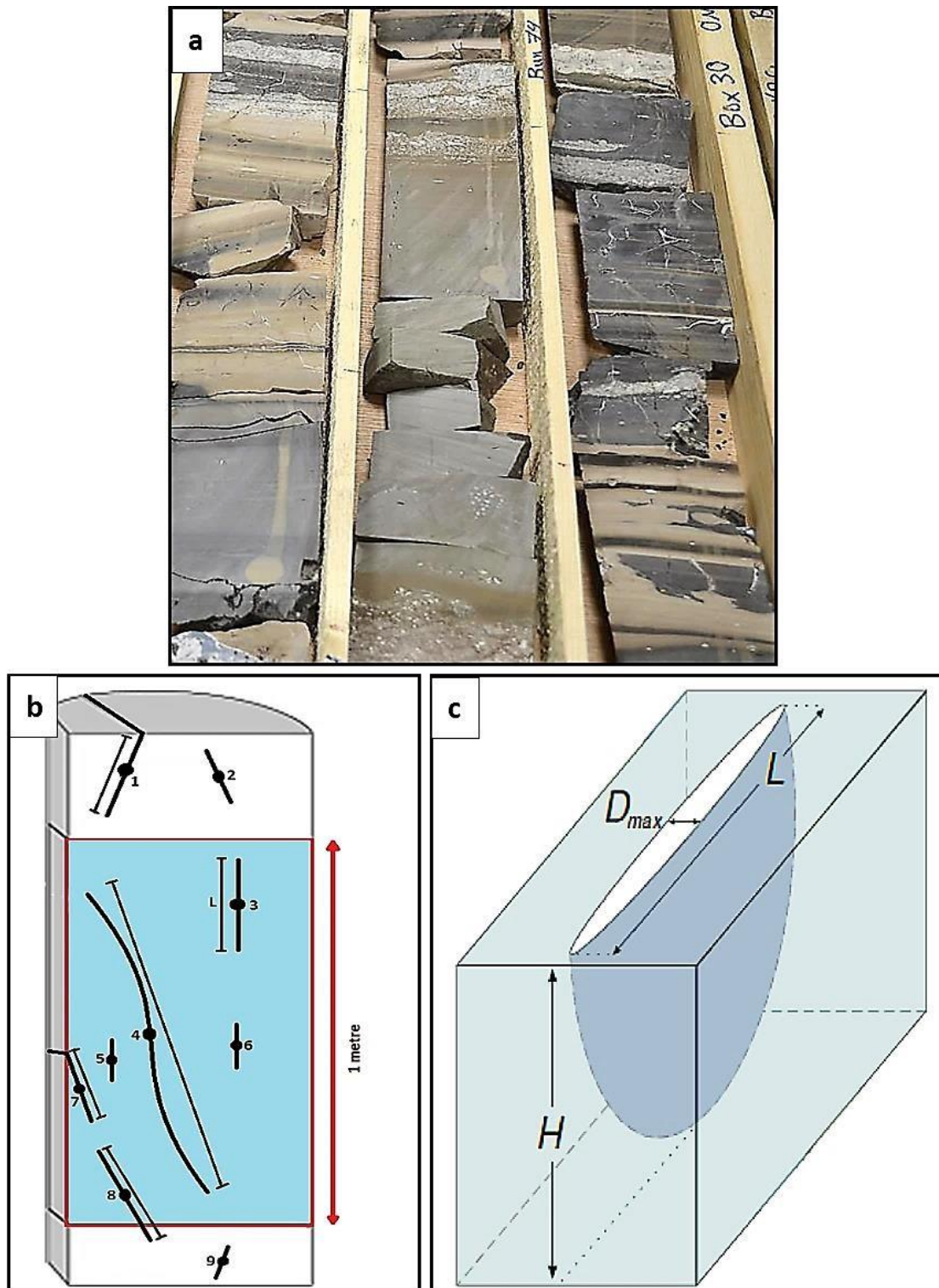


Figure 3.1 Fracture description and measuring methods from core. a. Photograph of slabbed core in a one-meter core box. b. Illustration of fracture geometry in the core. Note the fracture frequency in the blue cored interval is six. Lines showing the vertical fracture length measured on the slabbed surface. c. Opening-mode fracture model showing fracture length, L ; aperture (opening displacement), D and fracture height, H (modified from Klimczak et al., 2010).

				VEINS				JOINTS				OTHER
Box-Tray	Top Depth (m)	Bott Depth (m)	Recovered Core (cm)	#	Length (cm)	Apert (mm)	Fill & Other	#	Length (cm)	Apert (mm)	Other	Faults, Comments etc
3-2	117.36	118.30	58.0	1	17	1.26	calc.	0				
4-1	118.30	119.23	94	1	10.7	1.53	calc.	1	2.5	1.14	open.	—
4-2	119.23	120.13	90	1	2.5	—	Calc. hairline					Diameter 92 cm. (indicated)
				2	6.5	—	Calc. hairline					
				3	2.5	0.61	calc.					
5-1	120.13	121.0	60									
5-2	121.0	121.90	90	1	4.0	—	Calc. hairline	1	4.5	0.48	open	
				2	4.0	—	Calc. hairline	2	4.9	0.30	open	
				3	8.0	—	hair line					
6-1	121.90	122.89	96.5					1	10.0	0.45	open	
								2	7.0	0.49	open	
								3	3.5	0.52	open	
6-2	122.89	124.0	94.0	1	4.5	1.14		1	6.0	—	open	Broken
								2	4.5	0.54	open	
								3	11.5	0.30	open	
								4	3.5	0.57	open	
7-1	124.0	124.89	89.0	1	3	—	hair line	1	20.0	0.62	open	
								2	21.0	0.68	open	
								3	15.5	0.90	open	
								4	4.0	0.65		
7-2	124.89	125.83	73.0	1	8.0		hair line	1	11.0	0.53	open	
				2	2.8		hair line					

Figure 3.2 Example of a datasheet of raw fracture data measured from the core.

Interval	Top Depth (m)	Bott Depth (m)	# Veins (cemented)	Cumulative Vein Length (cm)	# Joints	Cumulative Joint Length (cm)	# Faults	Cumulative Fault Length (cm)	# Features (Veins, Joints, Faults)	Cumulative Length All Features (cm)	# Fractures & Faults	Cumulative Length (cm)
245	860.57	861.75	0	0	4	15.4	1	8.7	5	24.1	3	18.1
246	861.75	862.35	0	0	6	24.8	0	0	6	24.8	5	21.2
247	862.35	863.30	0	0	1	12.4	0	0	1	12.4	1	12.4
248	863.30	864.15	0	0	0	0	0	0	0	0	0	0
249	864.15	865.12	1	5.3	0	0	0	0	1	5.3	1	5.3
250	865.12	866.12	0	0	0	0	0	0	0	0	0	0
251	866.12	867.12	0	0	0	0	0	0	0	0	0	0
252	867.12	868.00	0	0	0	0	0	0	0	0	0	0
253	868.00	868.75	0	0	0	0	0	0	0	0	0	0
254	868.75	869.72	0	0	0	0	0	0	0	0	0	0
255	869.72	870.65	5	26.3	13	47.5	0	0	18	73.8	13	52.1
256	870.65	871.65	0	0	2	27	0	0	2	27	2	27
257	871.69	872.58	0	0	12	72.9	0	0	12	72.9	10	60.7
258	872.58	873.29	6	22.8	29	59.1	0	0	35	81.9	33	76.2
259	873.29	874.23	0	0	4	12.3	0	0	4	12.3	2	7.6
260	874.23	875.10	0	0	3	6.1	0	0	3	6.1	2	4.8
261	875.10	876.00	2	5.6	3	6.6	0	0	5	12.2	5	12.2
262	876.00	877.01	0	0	0	0	0	0	0	0	0	0
263	877.01	878.00	0	0	5	14.7	0	0	5	14.7	3	9.6
264	878.00	879.00	0	0	0	0	0	0	0	0	0	0
265	879.00	880.00	0	0	0	0	0	0	0	0	0	0
266	880.00	881.00	2	6.2	25	95.3	0	0	27	101.5	20	82.7
267	881.00	882.02	0	0	1	6.3	0	0	1	6.3	1	6.3
268	882.02	883.02	0	0	1	2.5	0	0	1	2.5	1	2.5
269	883.02	884.04	0	0	0	0	0	0	0	0	0	0
270	884.04	885.00	1	2.6	0	0	0	0	1	2.6	0	0

Figure 3.3 Sample of data table from core C 10.

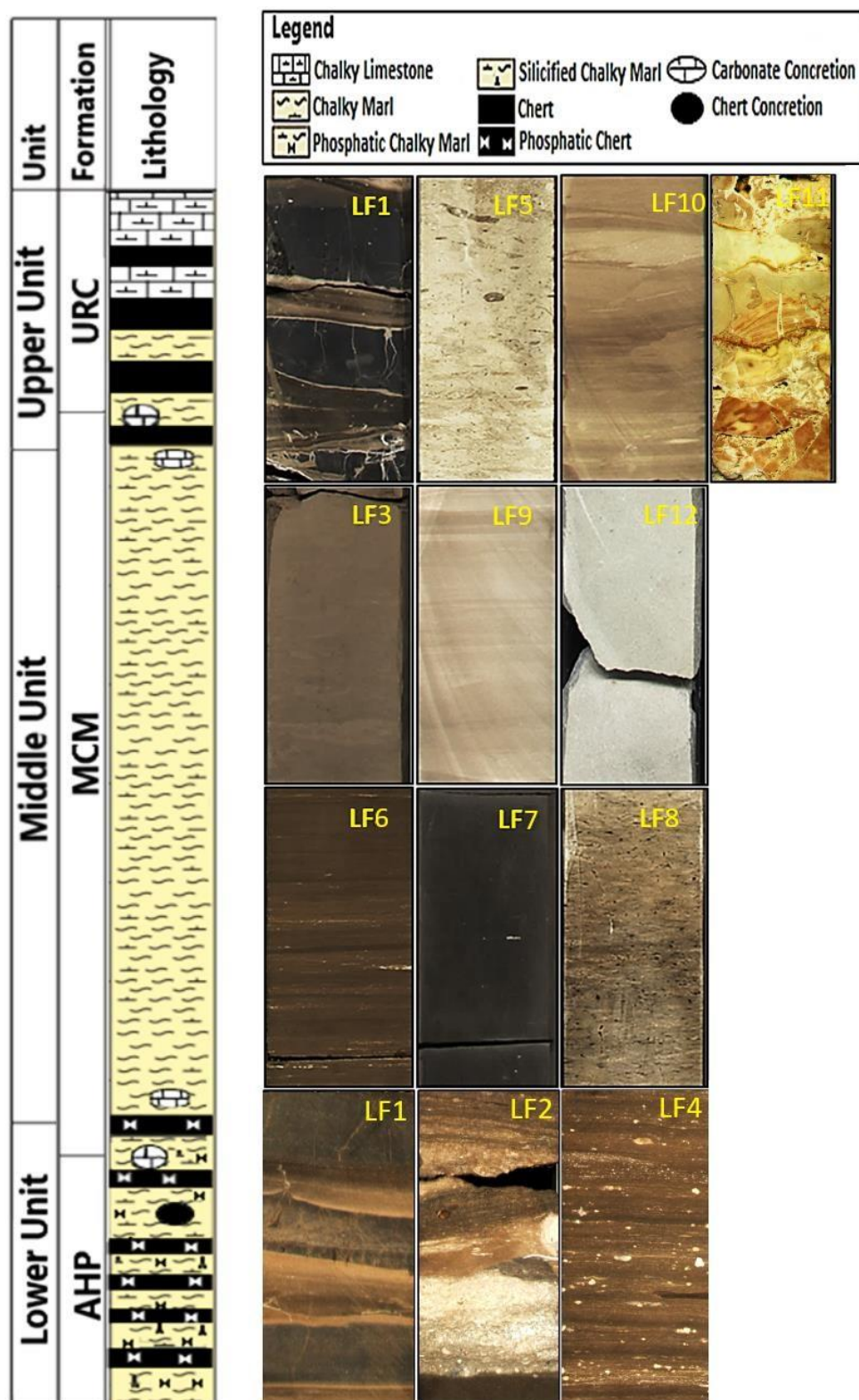


Figure 3.4 A Schematic drawing of representative columnar section showing the lithostratigraphy of the studied cored intervals, and core photographs of the twelve lithofacies associations (LF1-LF12) defined along the section. See text for the names of lithofacies associations. Core photograph width is 10 cm.

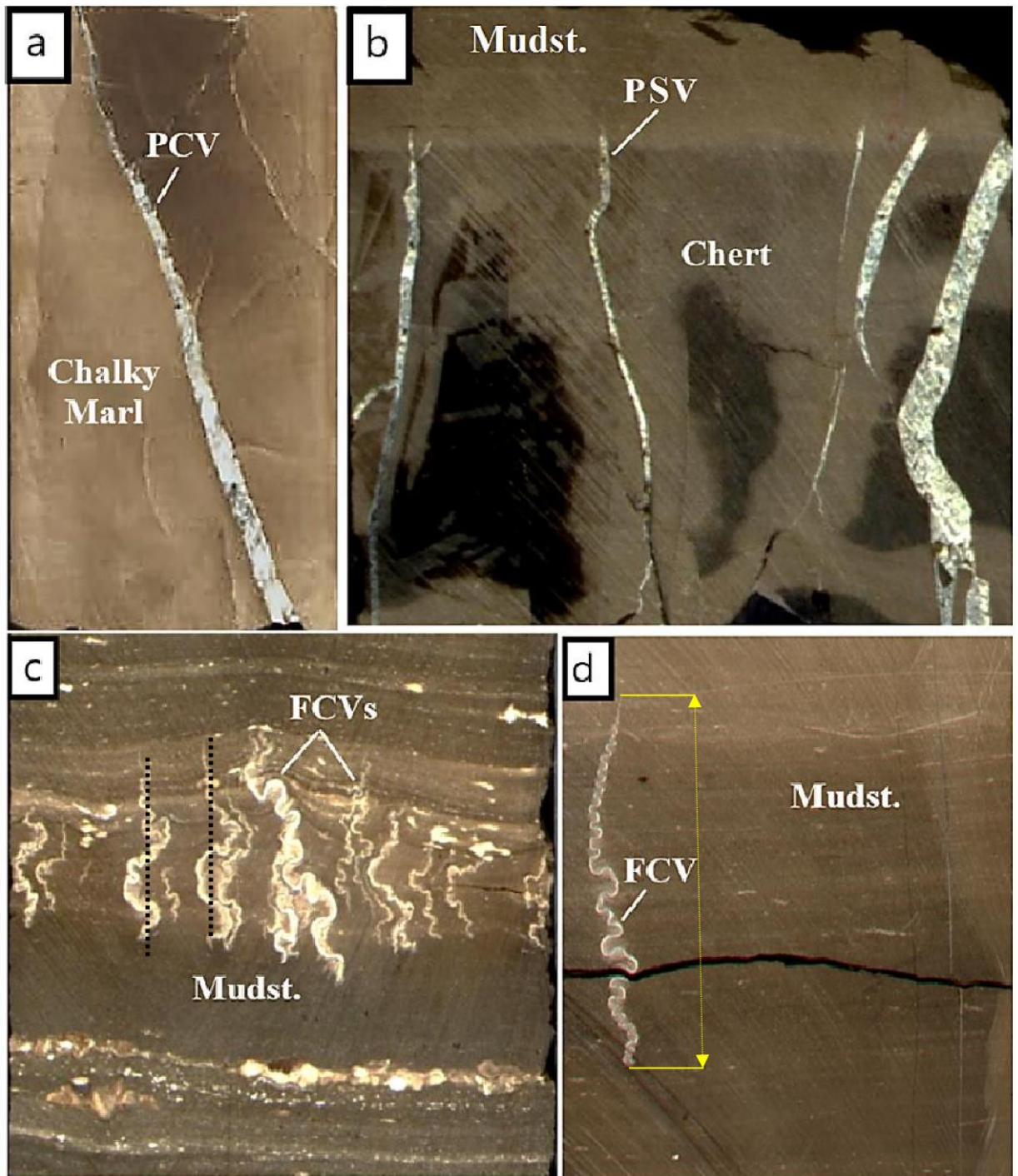


Figure 3.5 Core photographs of slabbed cores showing examples of calcite veins in the cores. PCV, Planar Calcite Vein; PSV, Planar Silica Vein; FCV, Folded Calcite Vein. a. Planar calcite vein in chalky marl in the Upper Unit. b. Planar silica vein in chert bed in the Lower Unit. Note that calcite crystals are also present in some parts of the vein. c. A cluster of folded calcite veins in organic-rich mudstone bed in the Lower Unit. Note that the two tips of the folded veins drop on the same vertical plane (black-dotted lines). The laminae are folded around the vein upper tips. d. A single folded calcite vein in an organic-rich mudstone bed in the Lower Unit. Yellow arrow displays the length of the vein. Core photograph width is 10 cm.

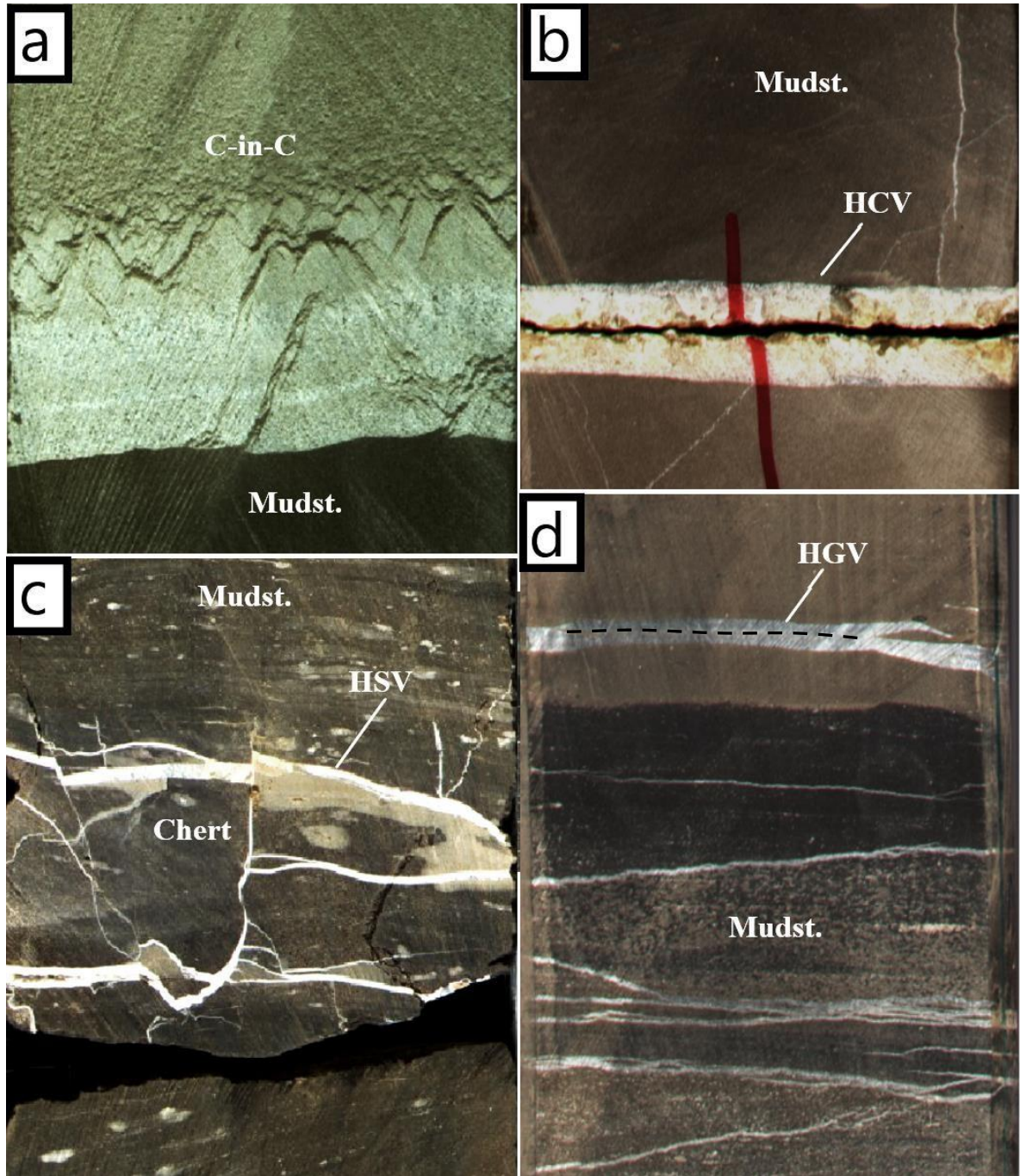


Figure 3.6 Core photographs of slabbed cores showing examples of the different types of bedding-parallel veins observed in the Lower Unit of the different cores. C-in-C, cone-in-cone; HCV, Horizontal Calcite Vein; HSV, Horizontal Silica Vein; HGV, Horizontal Gypsum Vein. a. Calcite cone-in-cone in an organic rich mudstone interval. b. Bedding-parallel calcite vein in calcareous mudstone showing a blocky texture. c. A network of bedding-parallel silica veins interconnected by vertical silica vein. d. Bedding-parallel gypsum veins in an organic-rich mudstone. Note the median line in the upper thick gypsum vein (black-dashed line). Core photograph width is 10 cm.

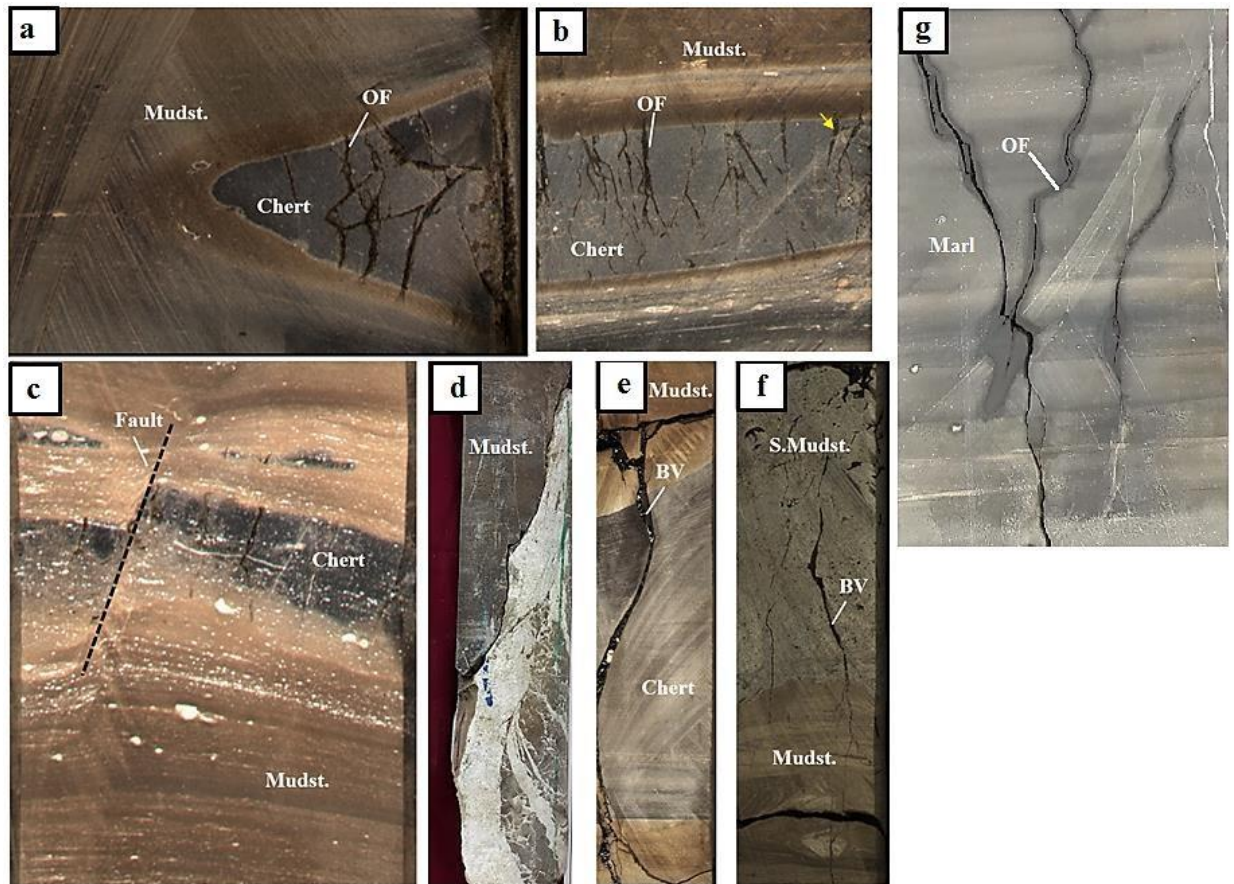


Figure 3.7 Core photographs of: a-b. Joints (open fractures) observed in the chert intervals in the Lower Unit. Note the partial presence of bitumen in open fractures (yellow arrow) c. Normal fault in mudstone and chert in the Upper Unit. d. Dilatational breccia zone in mudstone in the Upper Unit. e-f. Bitumen veins in chert and silicified intervals in the Lower Unit. g. Open fracture network in marl interval in the Middle Unit of the core from Well C10. The sample was wetted with water to display fracture permeability (photo was taken following drying of the non-fractured portion of the core). Core width is 10 cm.

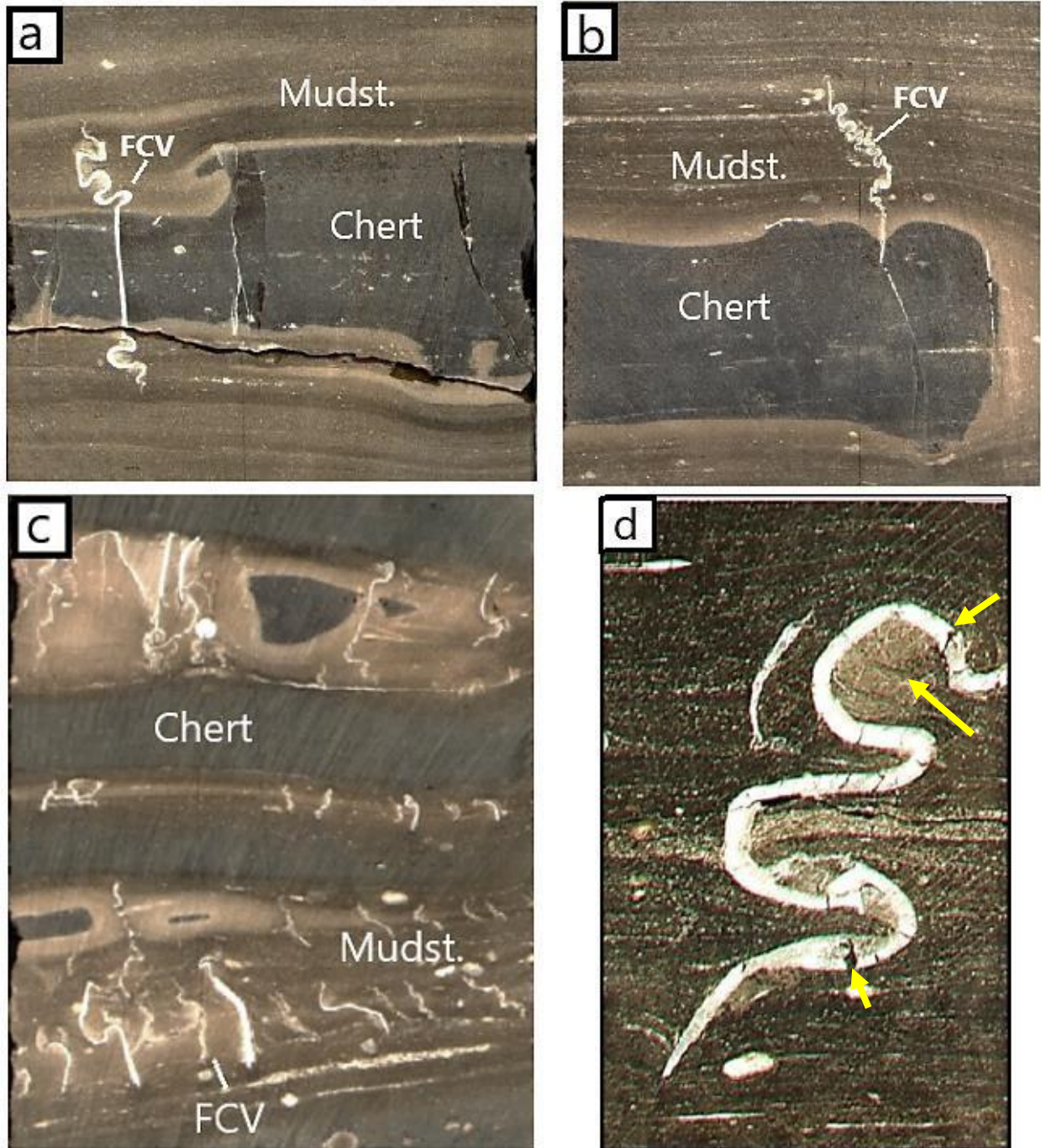


Figure 3.8 Core photographs showing the presence of folded calcite veins close to or within chert beds and concretions (a-c) in the Lower Unit. Note the preserved sedimentary laminae (white arrows) inside the chert concretions in (a and b). d. Folded calcite veins showing microfractures (yellow arrows) in the inner arc and on the vein body. FCV, Folded Calcite Vein. Core photograph width is 10 cm (a-c) and 5 cm in (d).

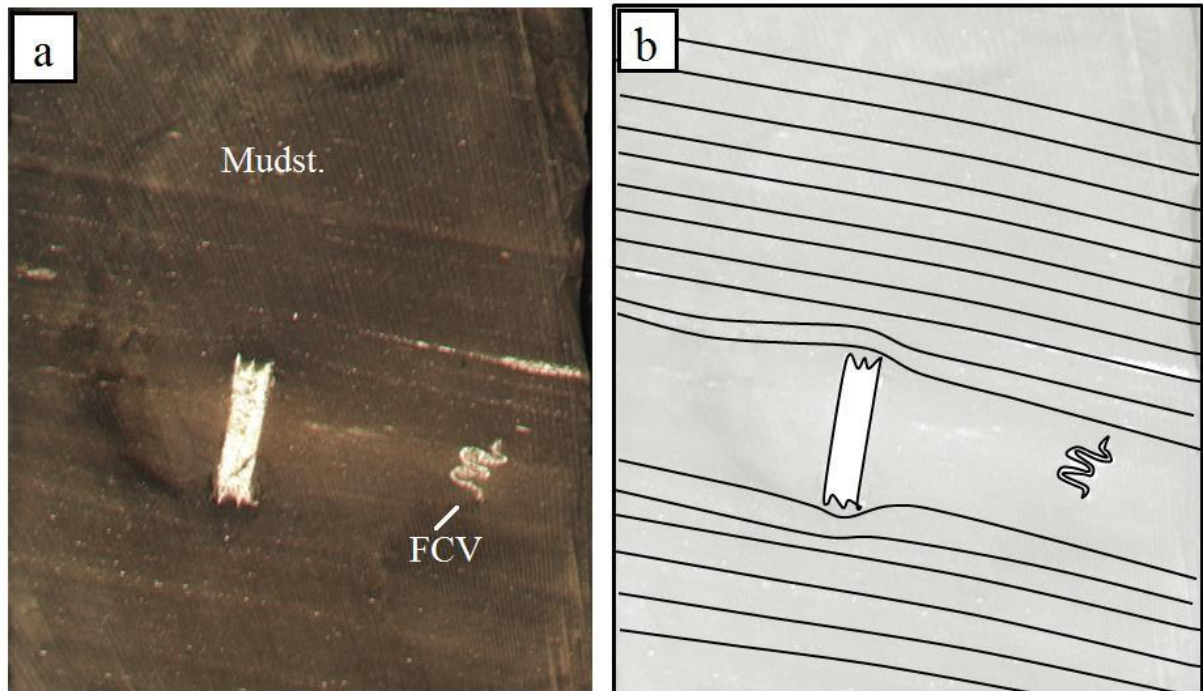


Figure 3.9 a. A core photograph showing an occurrence of a thin folded calcite vein and thick straight calcite vein in the same bed. FCV, Folded Calcite Vein. b. Schematic drawing of (a). The thick vein seems to have been more competent, and therefore, it was not susceptible to folding. Note the bending of laminae at the thick vein tips. Photograph width is 10 cm.

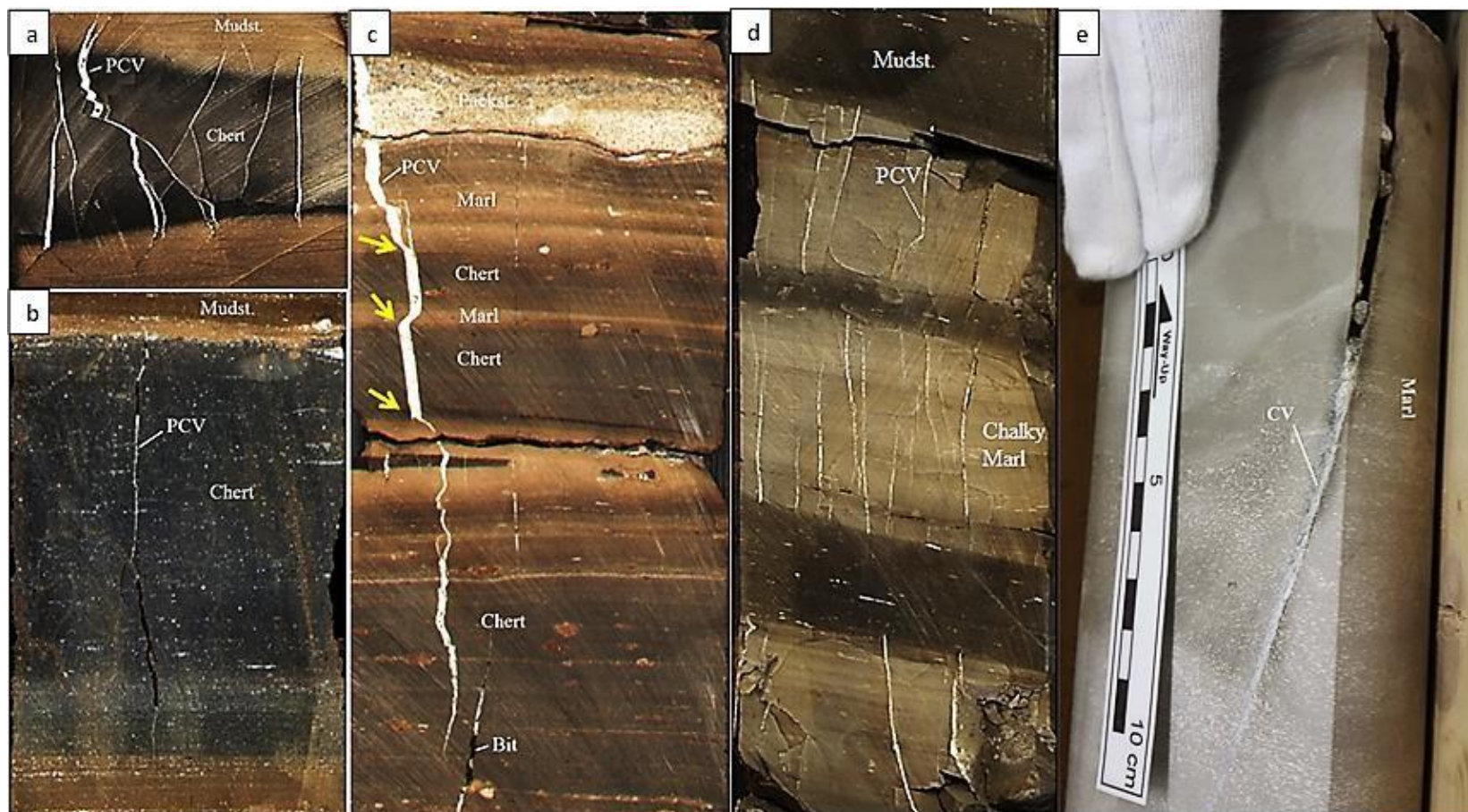


Figure 3.10 Photographs of slabbed cores showing the different properties of planar calcite veins. PCV, Planar Calcite Vein; Bit, Bitumen; Packst, Packstone; CV, Calcite Vein. a. complex planar veins in chert bed in the Lower Unit of core. Note the interaction between veins and the branching/splaying tips of some veins. b. Partially-filled planar calcite vein confined to chert bed. c. Planar calcite vein propagates in different rock lithologies. Note the vein stepping with change in lithology from chert to marl (yellow arrows). d. Parallel arrangements of bed-confined planar calcite veins. e. Partially-filled calcite vein. Note the large calcite crystals in the upper part of the vein.

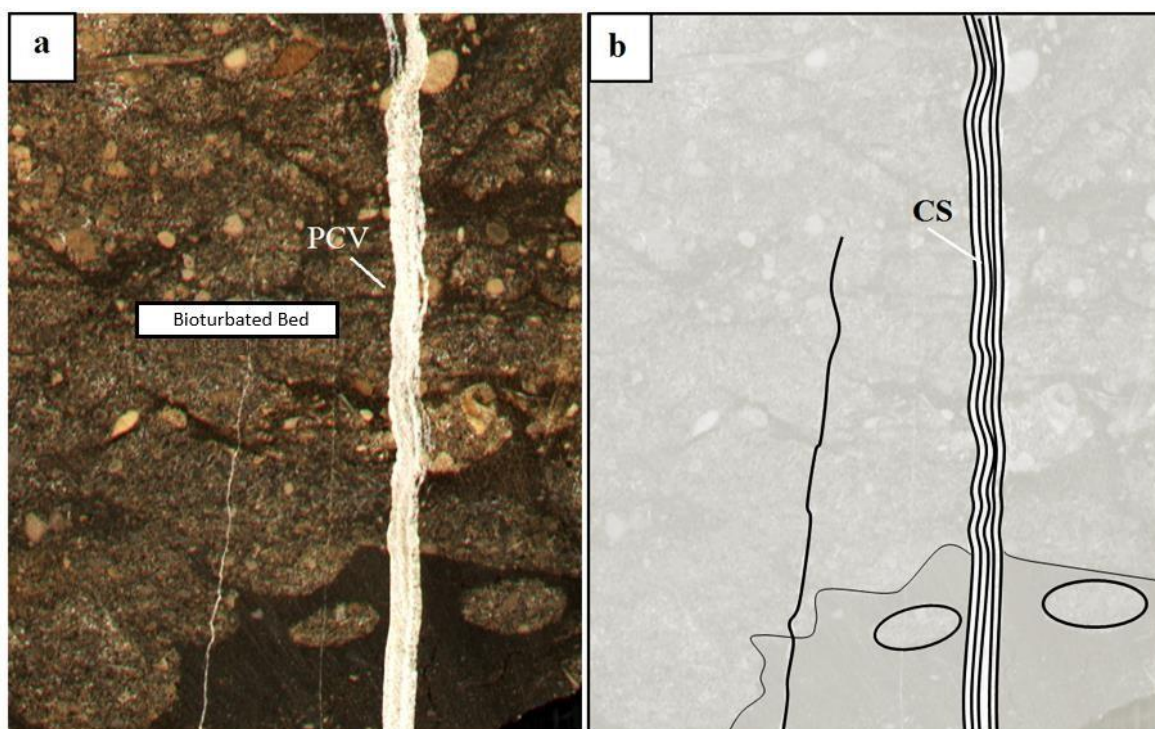


Figure 3.11 a. Core photograph of planar calcite vein showing crack-seal texture in a bioturbated *Thalassinoides*-burrowed horizon in the core sample in the Lower Unit. b. Schematic drawing of (a). Black ellipses are burrows in cross section. CS, Crack Seal. Photograph width is 10 cm.

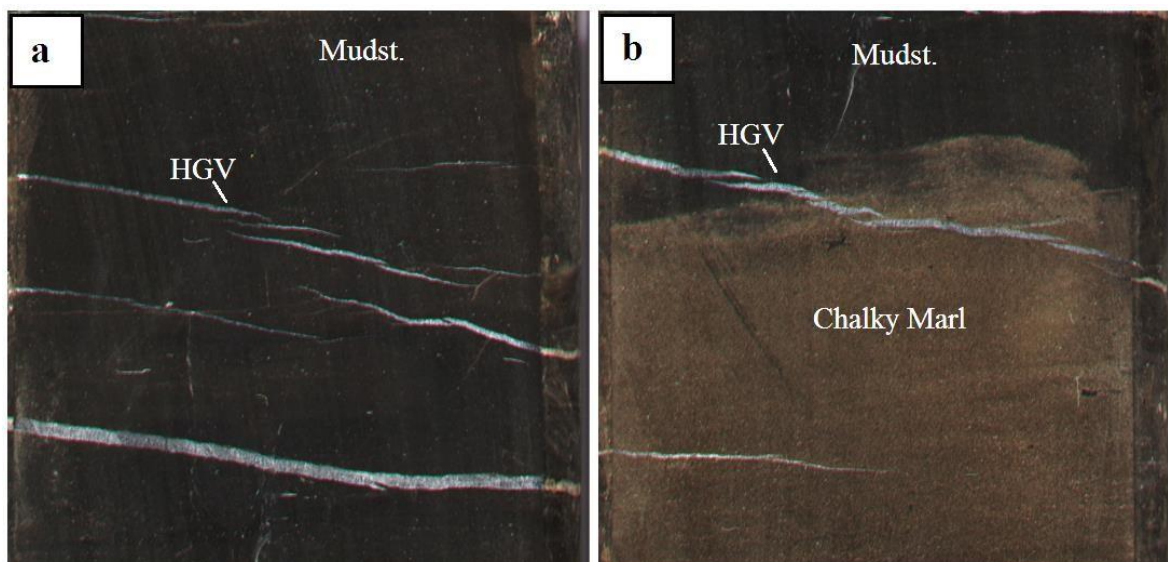


Figure 3.12 Core photographs of bedding-parallel gypsum veins. Width of each photograph is 10 cm.

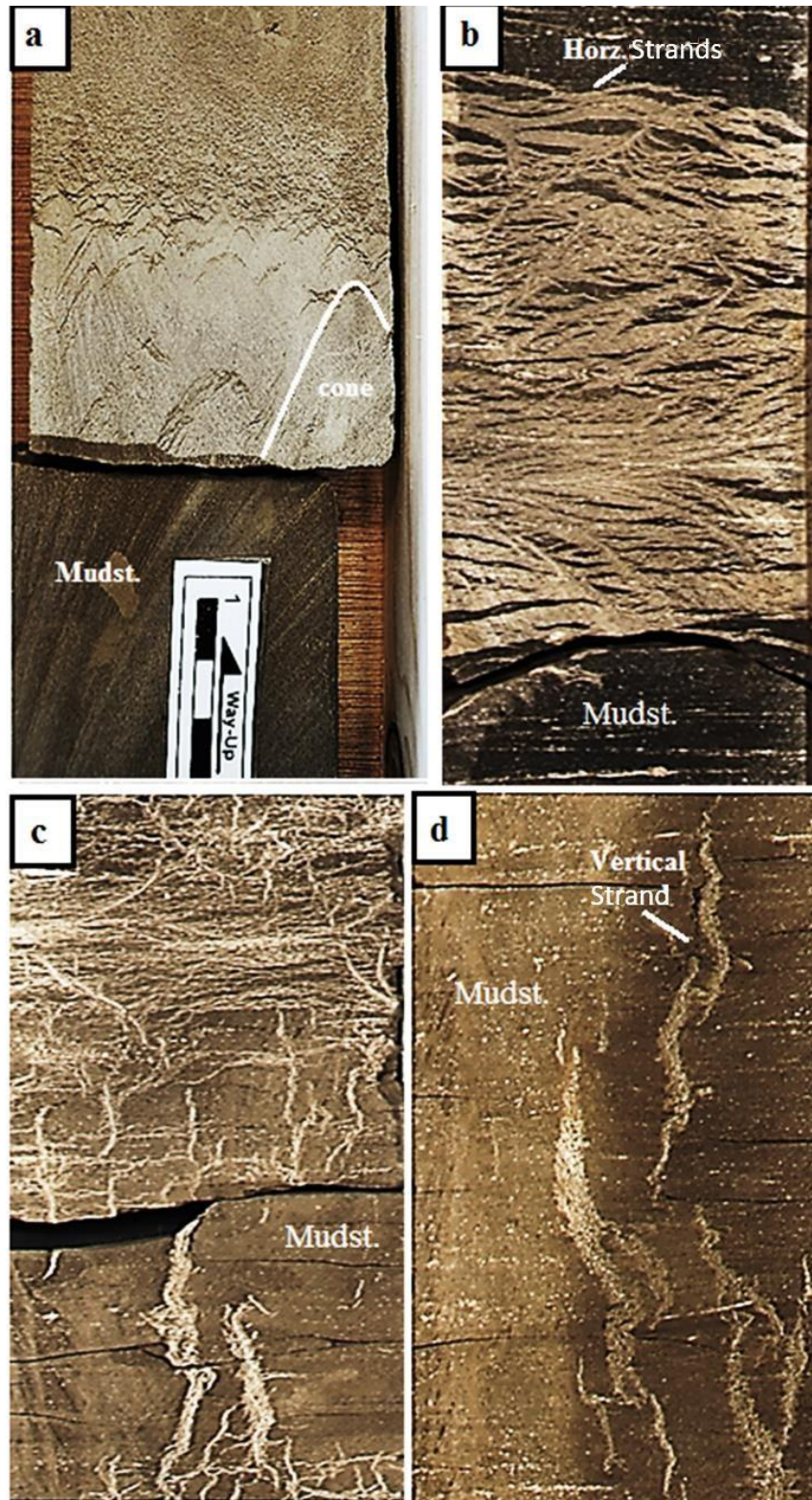


Figure 3.13 Core photographs of examples of cone-in-cone structures. a. Cone-in-cone showing obvious conical shapes. b. Horizontal strands of cone-in cone. c. Horizontal and vertical cone-in-cone strands. d. Isolated vertical strands of cone-in-cone in the core from C2 at depth 428 m.

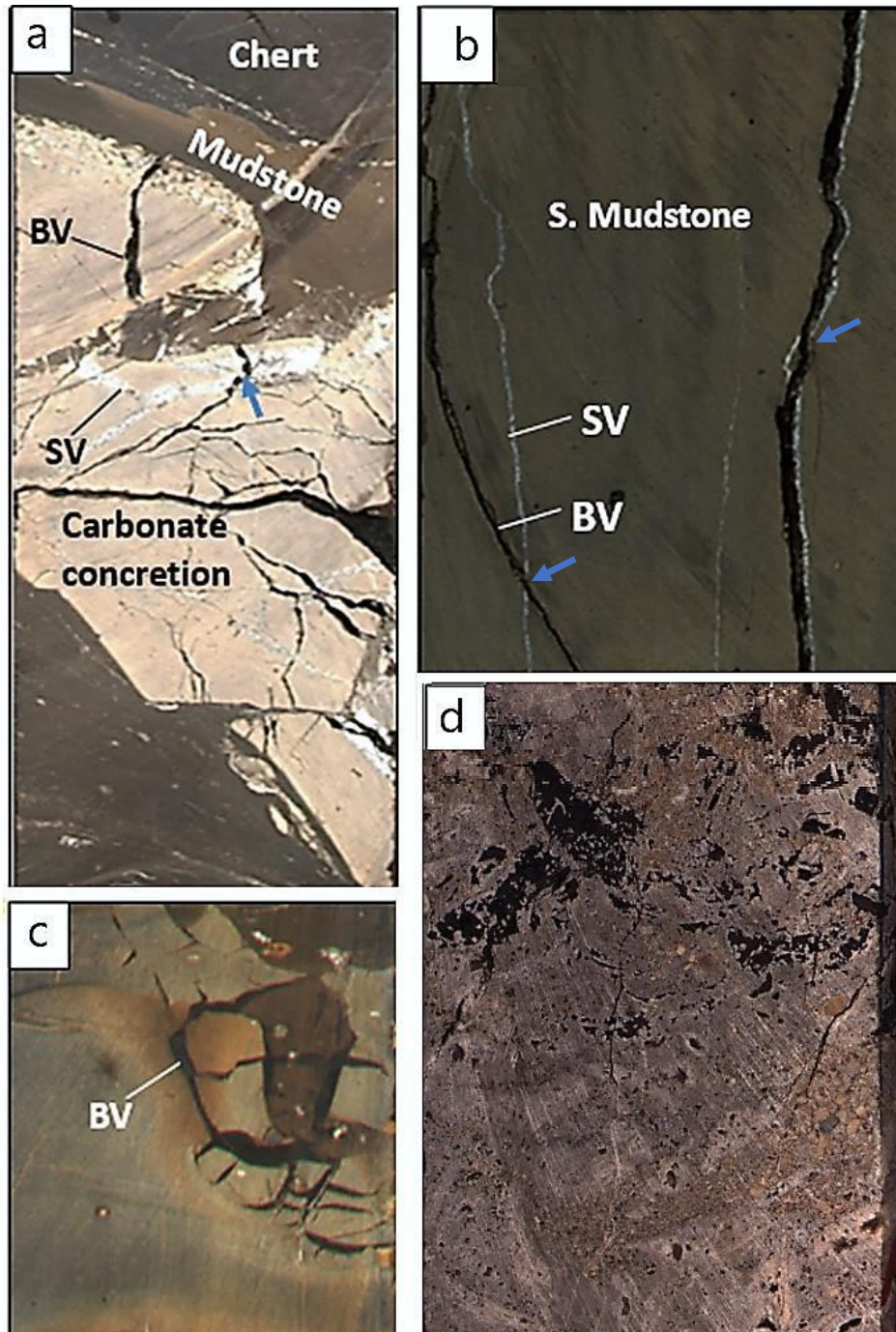


Figure 3.14 Core photographs of bitumen bearing fractures and vugs observed in the Lower Unit. SV, Silica Vein; BV, Bitumen Vein. a. Bitumen veins crosscutting silica veins (blue arrow) in carbonated concretion. b. Bitumen veins crosscutting (blue arrows) and running parallel to silica vein in silicified mudstone. c. Brecciation of silicified mudstone by bitumen veining. d. Bitumen vugs in silicified limestone interval. Core width is 10 cm.

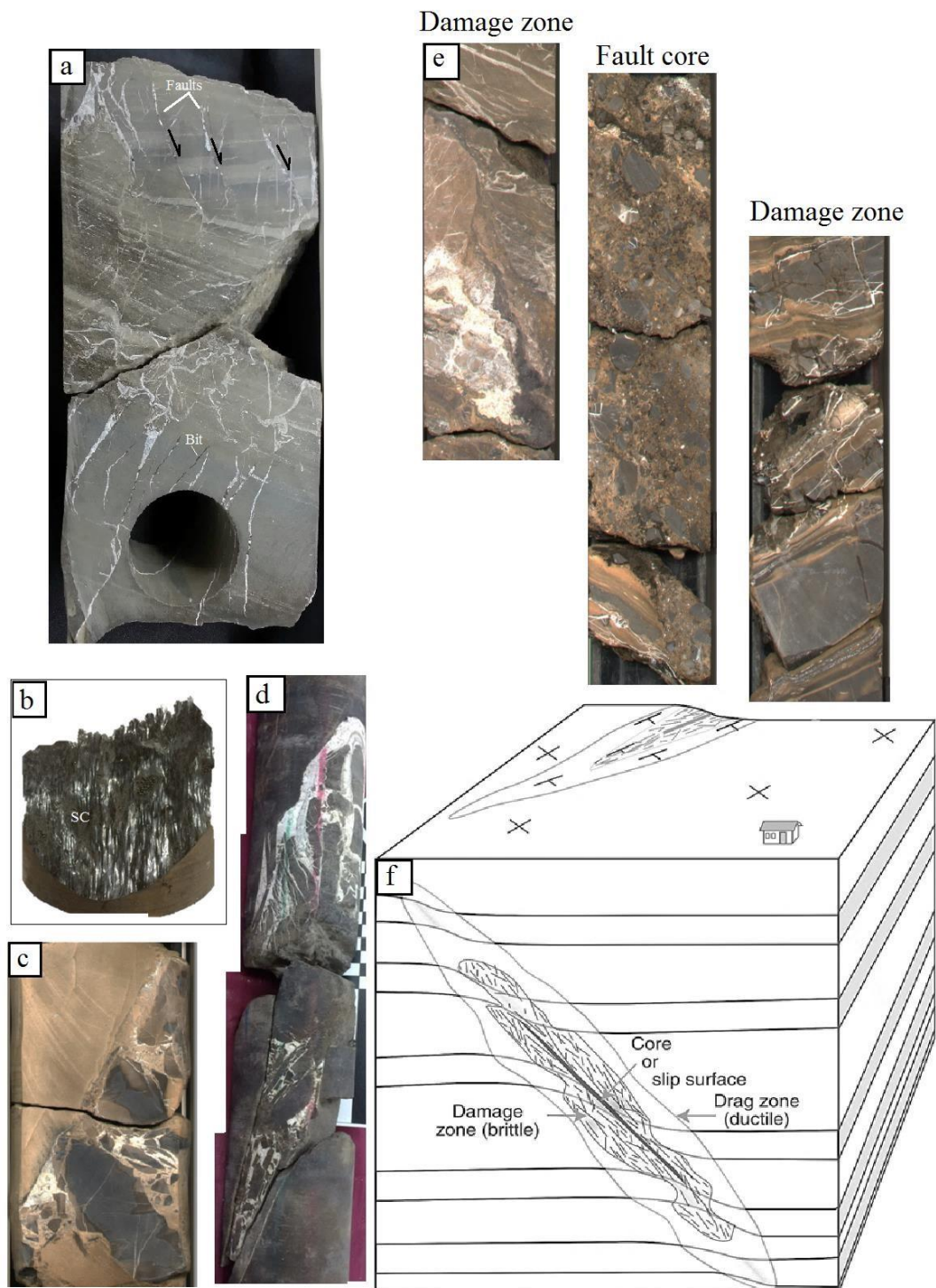


Figure 3.15 Examples from the cores of faults and fault-related structures. Bit, Bitumen; SC, Slickensides. a. Microfaults showing normal offset in laminated silicified marl. Note the stepping in laminae. b. Smoothly-polished and striated surface (Slickenside; SC) along fault plane in mudstone in the core from C10. The striations show the direction of movement. c and d. Dilatational calcite-supported breccia zones in chalky marl and bituminous mudstone, respectively. e. Fault zone with fault core (contains cataclasites, gouge and breccia cemented by calcite with bitumen inclusions) and two damage zones observed in the core from C7 at depth 667 m. f. Conceptual sketch of the fault zone in (e). Core photograph width is 10cm.



Figure 3.16 Example from the core samples of petal-like fractures induced by drilling.

Core Name	Cored Interval Thickness (m)	Core Location	Total Number of Veins	Average Cumulative Length of All Veins (cm)	Average Vein Aperture (mm)	Total Number of Joints	Average Cumulative Length of All joints (cm)	Average Joint Aperture (mm)	Total Number of Faults & Breccias	Total Number of All Fractures	Average Cumulative Length of All Fractures (cm)	Average Aperture of All Fractures (cm)
C1	201	CW	90	17	0.9	5	16	0.6	0	95	33	1.5
C2	169	CW	55	30	0.6	15	9	0.5	0	70	39	1.1
C3	255	CW	464	29	1.3	201	19	1.1	5	670	48	2.4
C4	217	CW	800	25	1.4	336	10	1.3	3	1139	35	2.7
C5	232	CW	991	22	0.5	269	11	0.4	19	1279	33	0.9
C6	246	CW	513	34	1.6	164	20	0.7	29	706	54	2.3
C7	293	CW	593	40	1.7	460	38	1.0	84	1137	78	2.7
C8	185	CE	875	20	0.5	728	16	1.0	30	1633	36	1.5
C9	276	CE	450	19	1.9	195	14	2.1	5	650	33	4
C10	263	CE	506	13	1.7	755	12	1.8	43	1304	25	3.5
C11	137	SE	1343	45	1.5	550	35	2.3	21	1914	33	3.8
C12	161	SE	75	21	1.0	8	15	1.6	1	84	36	2.6
C13	202	SW	82	49	0.5	31	30	0.9	4	117	79	1.4
C14	183	SW	128	20	0.5	84	21	0.8	3	215	41	1.3
C15	192	SW	184	14	0.4	129	12	0.3	8	321	26	0.7
Total	3212		7149			3930			255	11334		

Table 3.1 Fracture dataset from all of the studied cores. Note that cumulative length is measured form each one-metre core interval

Core Name	No. of all Veins	No. of folded veins	No. of planar veins
C1	55.00	7	48
C2	90.00	35	55
C3	464.00	26	438
C4	800.00	135	665
C5	991.00	71	920
C6	513.00	119	394
C7	593.00	50	543
C8	875.00	11	264
C9	450.00	38	412
C10	506.00	6	500
C11	1343.00	37	1306
C12	75.00	7	68
C13	82.00	16	66
C14	128.00	28	100
C15	184.00	27	157
Total	7149.00	613	5927
Percentage (%)		8.60%	82.9%

Table 3.2 Number of all planar and folded calcite veins observed in all cores

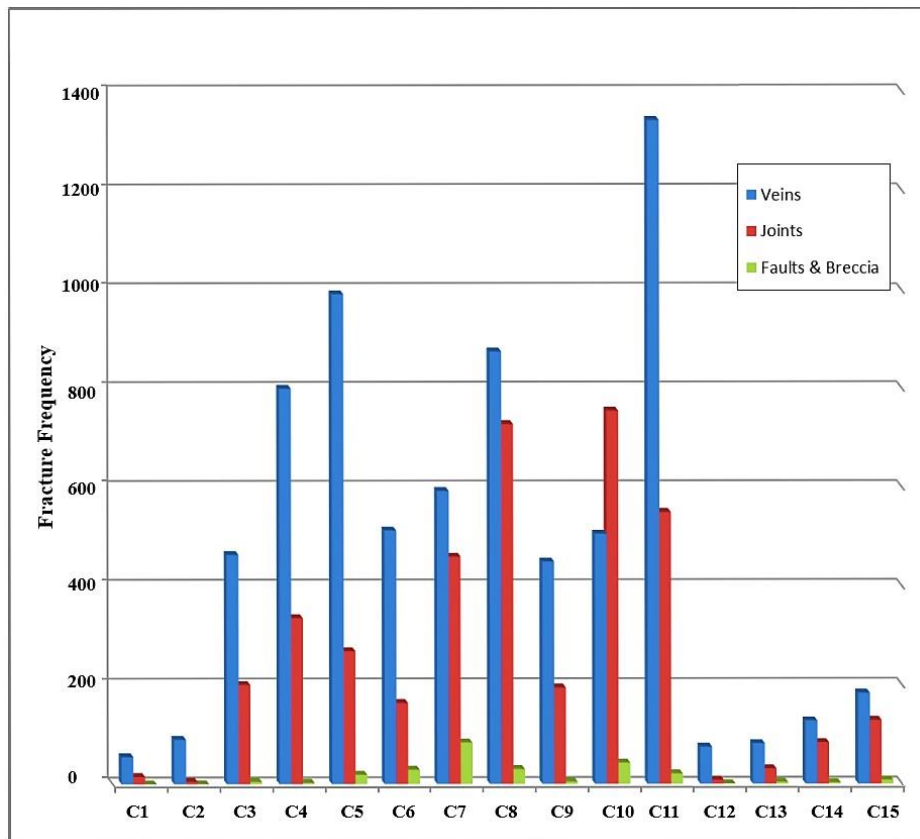


Figure 3.17 Fracture counts of opening- and shear-mode fractures in all cores.

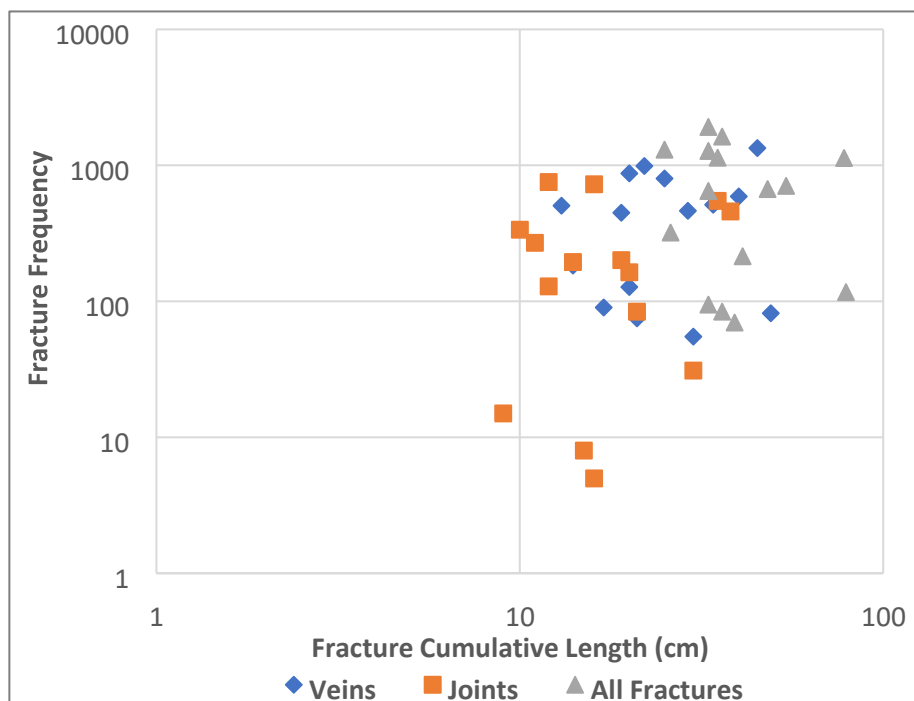


Figure 3.18 Logarithmic plot of the average values of fracture cumulative length versus the average values of fracture frequency for all cores.

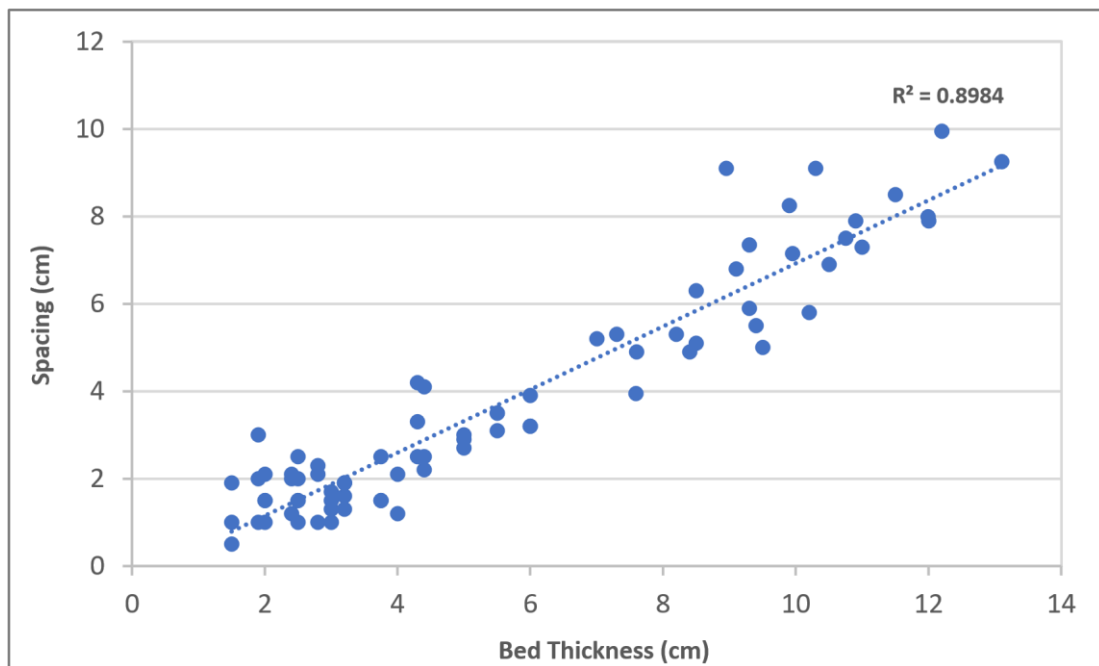


Figure 3.19 Plot of fracture spacing against bed thickness of opening-mode fractures.

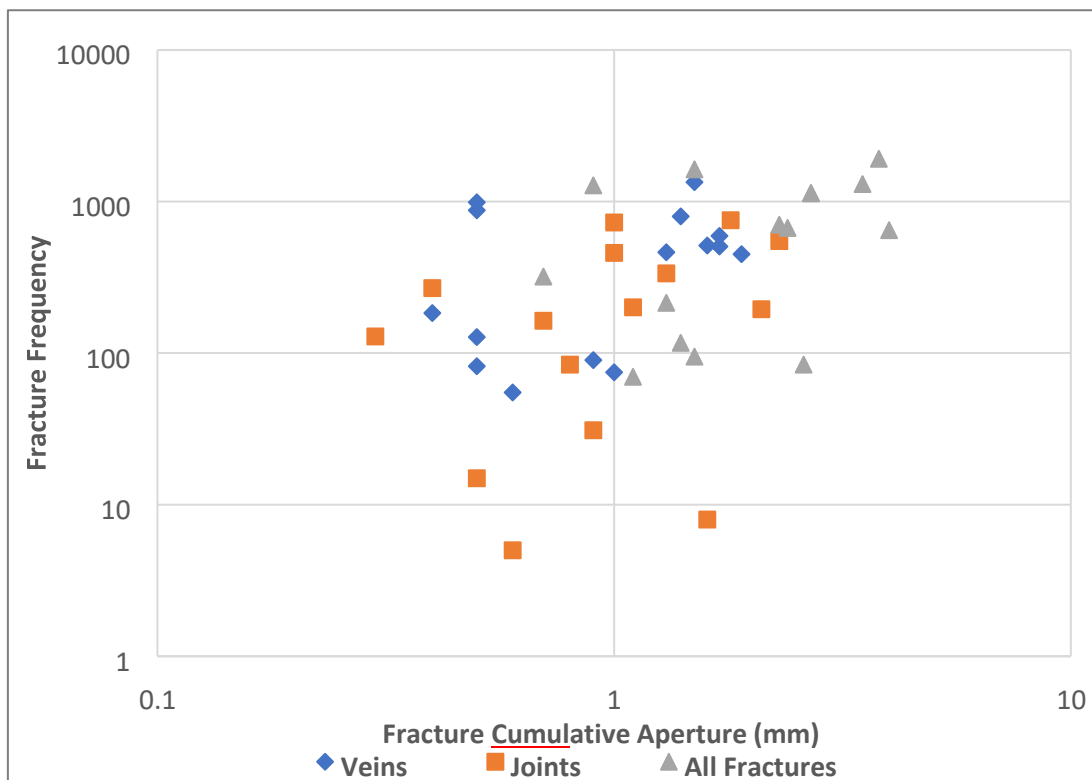


Figure 3.20 Logarithmic plot of the average values of fracture cumulative apertures versus the average values of fracture frequency for all cores.

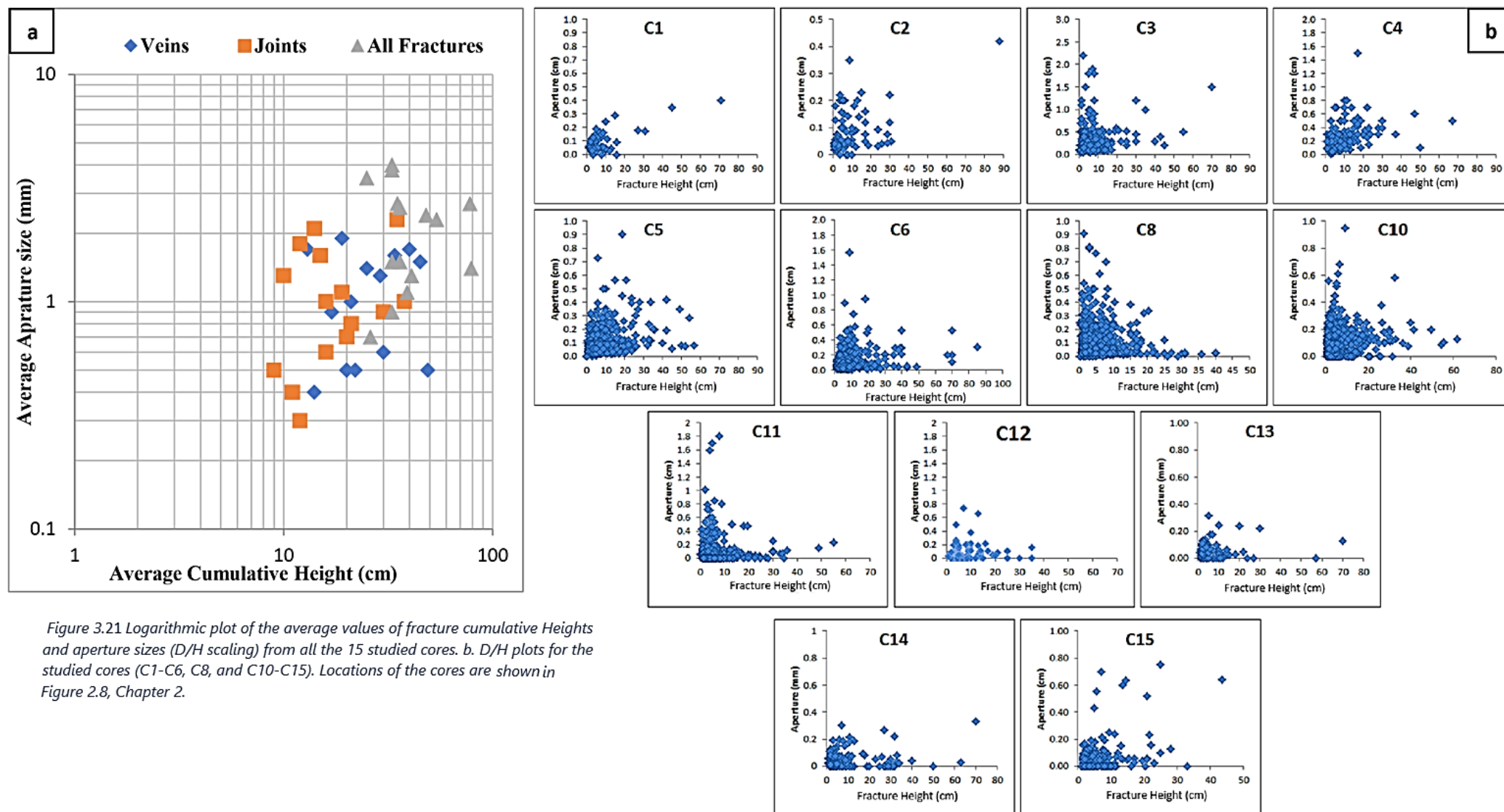


Figure 3.21 Logarithmic plot of the average values of fracture cumulative Heights and aperture sizes (D/H scaling) from all the 15 studied cores. b. D/H plots for the studied cores (C1-C6, C8, and C10-C15). Locations of the cores are shown in Figure 2.8, Chapter 2.

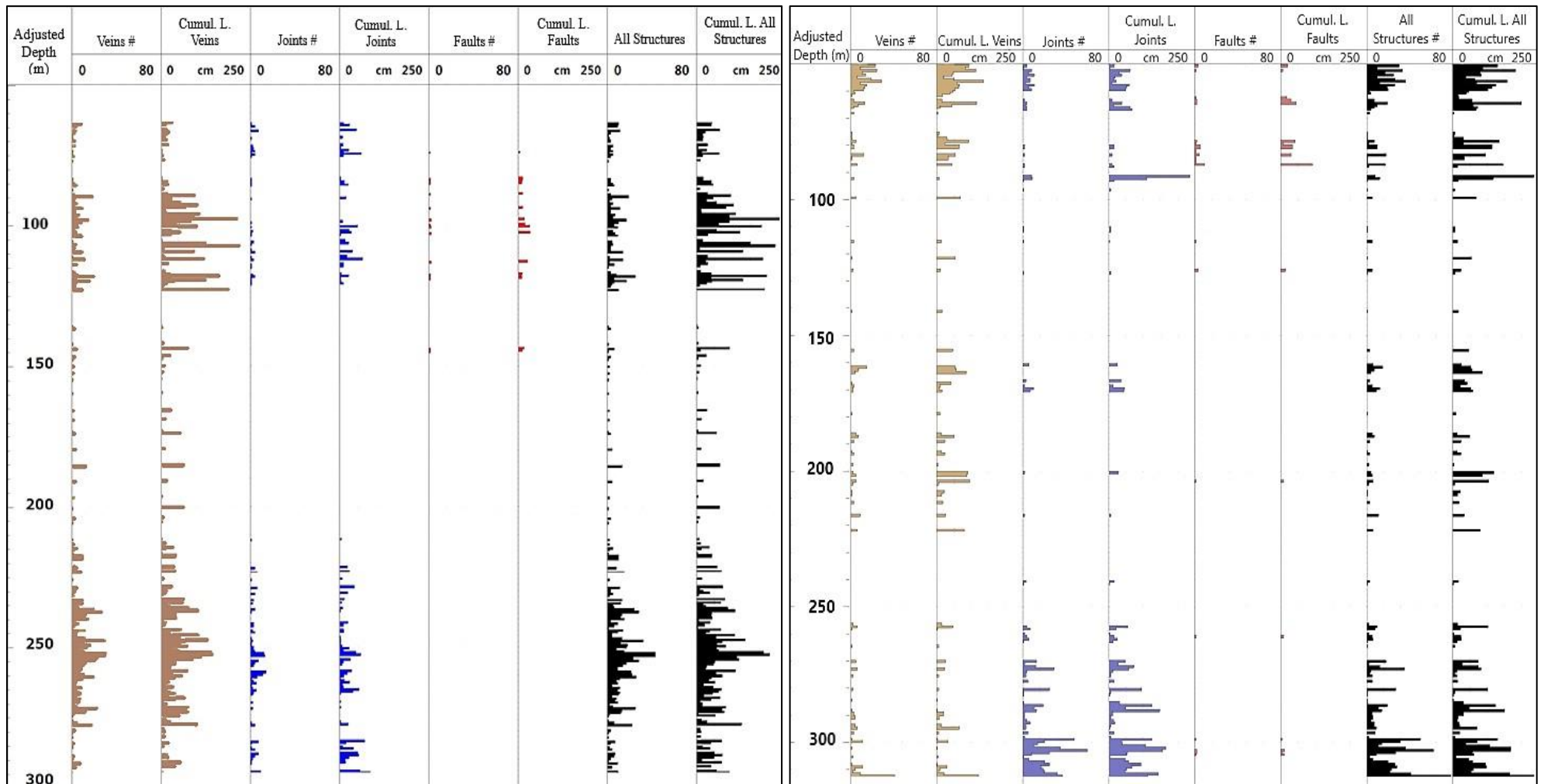


Figure 3.22 Examples of fracture logs constructed from fracture data for Core C5 from the Harrana area in the central west (on the left) and Core C10 from the Azraq area in central east (on the right).

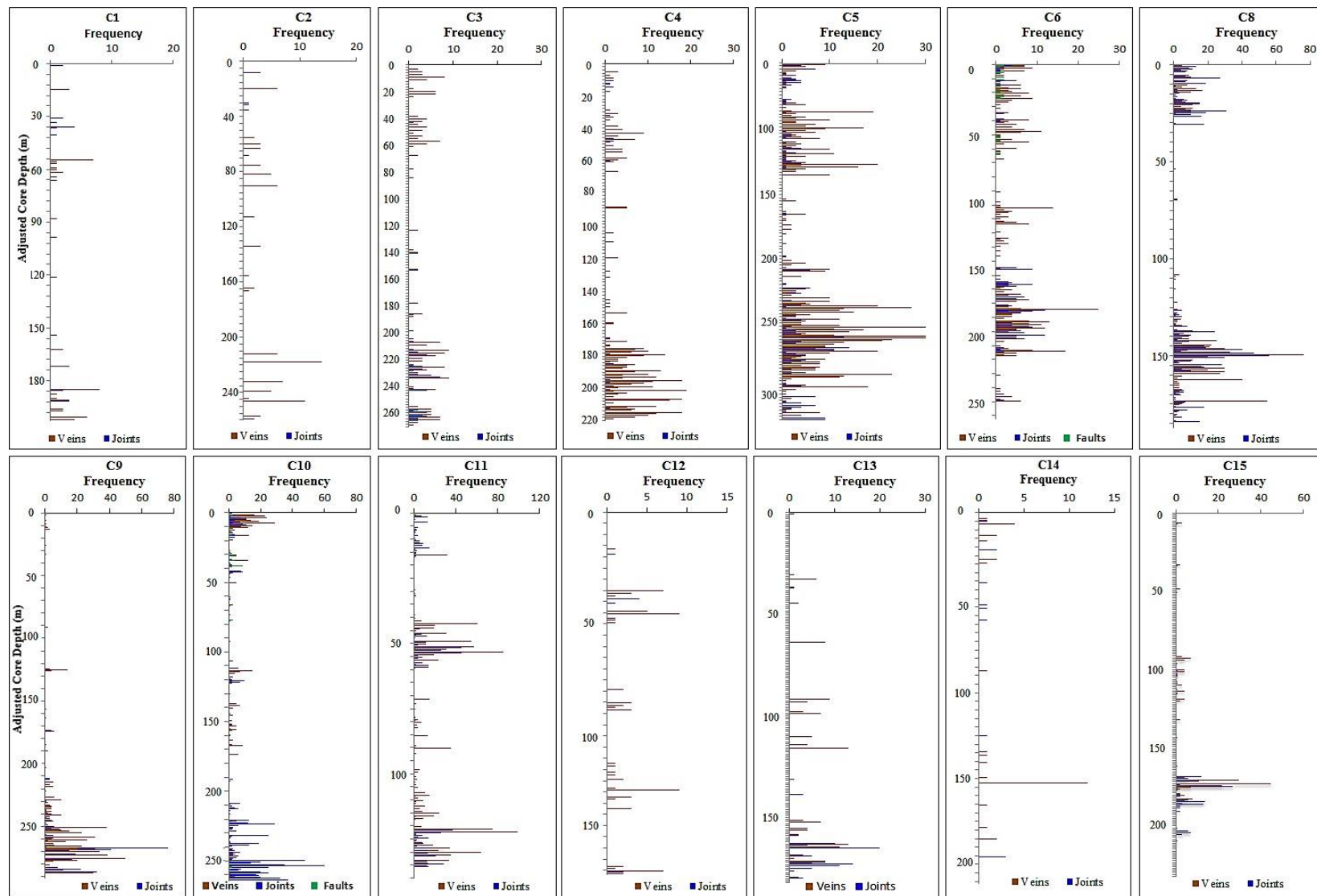


Figure 3.23 Plots of fracture frequency versus core depth for the studied cores.

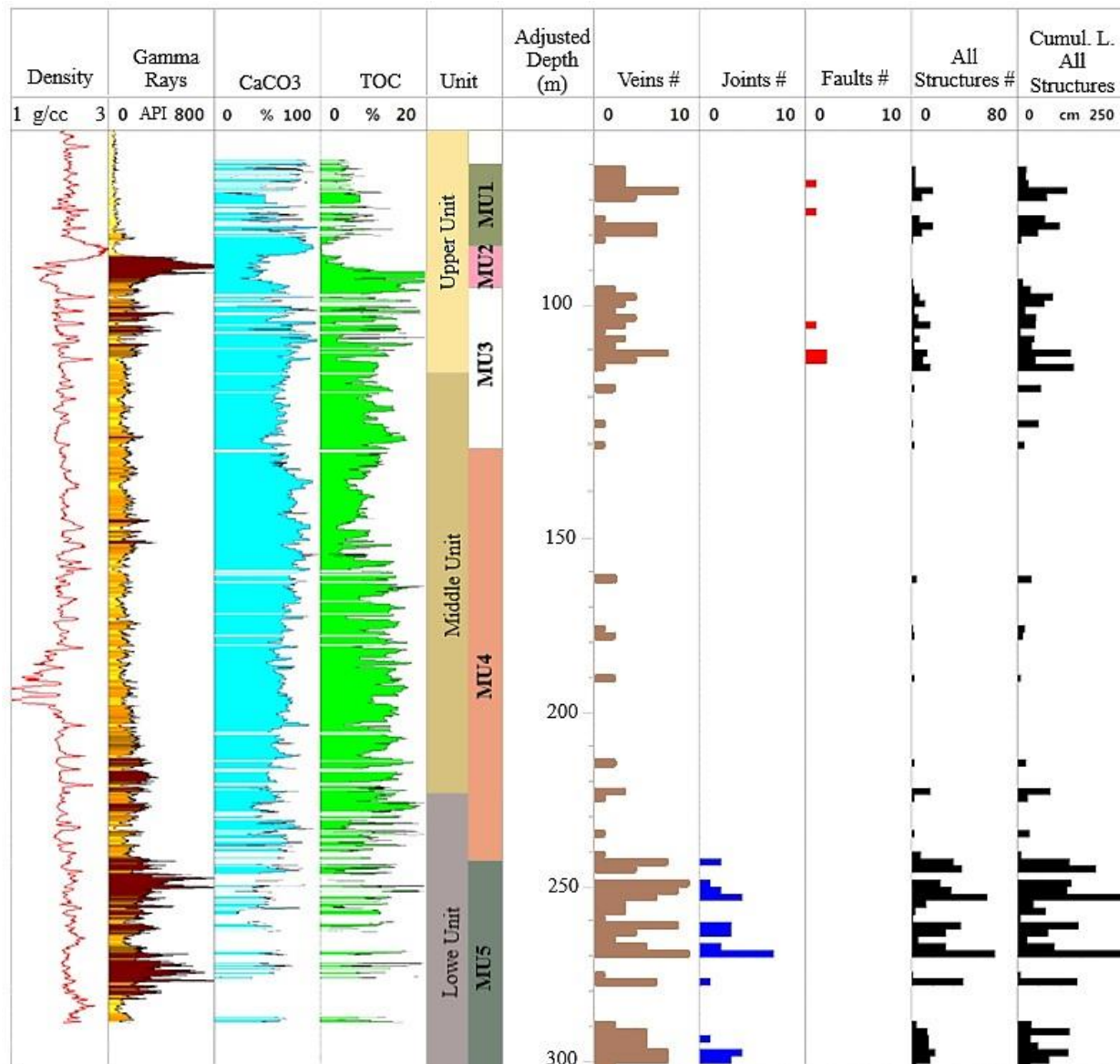


Figure 3.24 Fracture, petrophysical and geochemical data plotted versus depth for Core C3, Harrana Area (central-west Jordan).

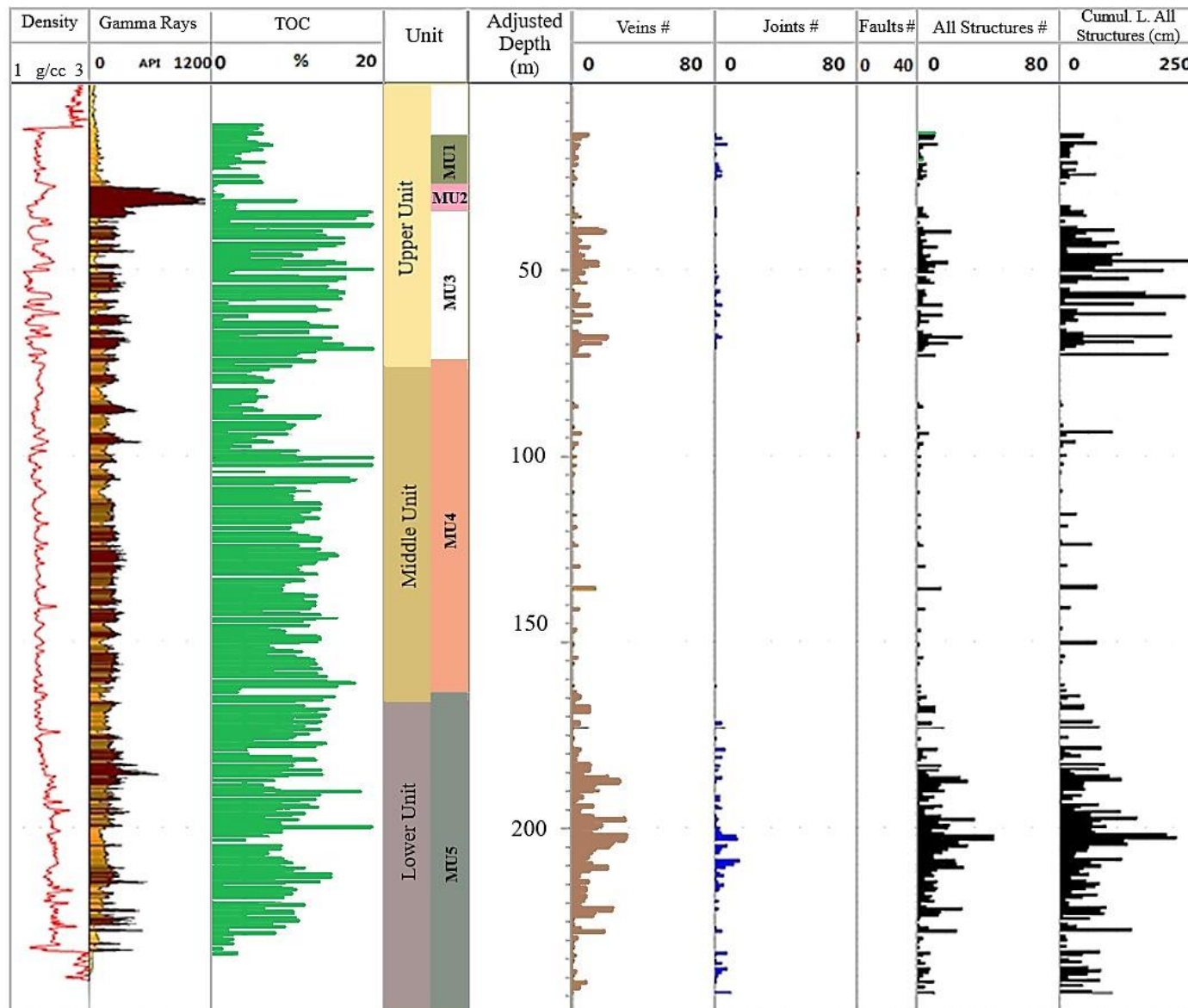


Figure 3.25 Fracture, petrophysical and geochemical data plotted versus depth for Core C4, Harrana Area (central-west Jordan).

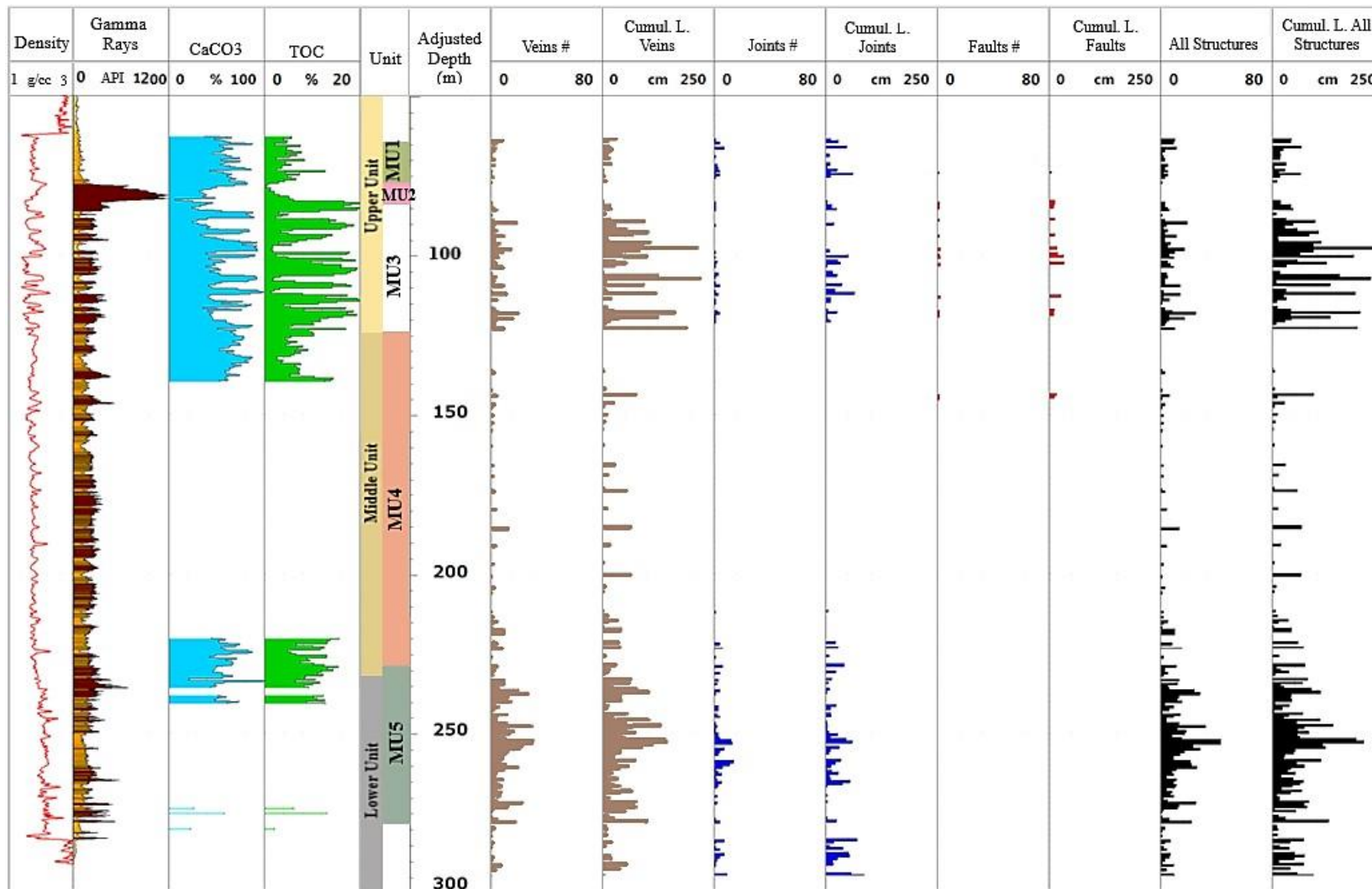


Figure 3.26 Fracture, petrophysical and geochemical data plotted versus depth for Core C5, Harrana Area (central-west Jordan). Note the high intensity of fractures in the Lower and Upper units and the low intensity in the Middle Unit.

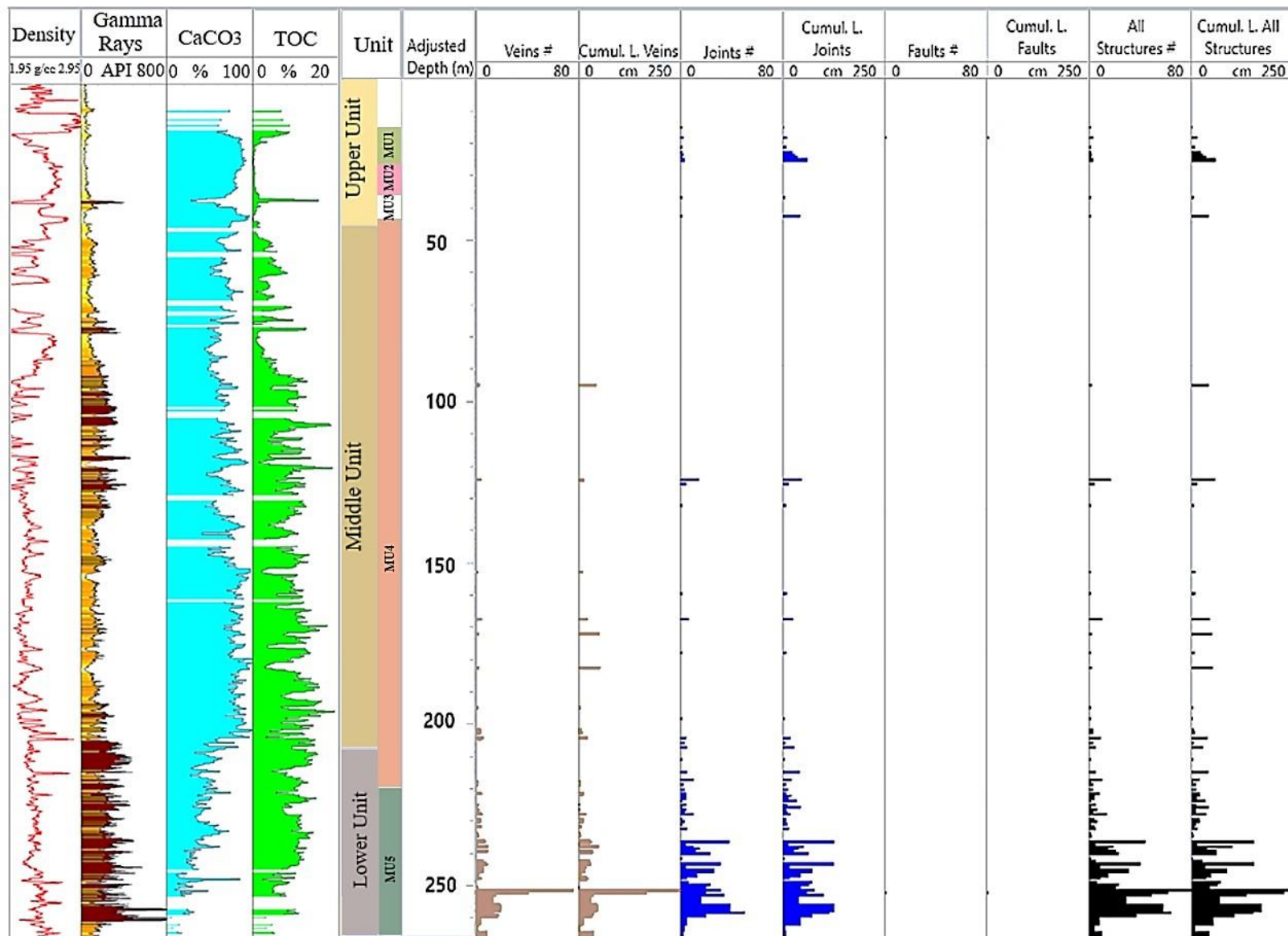


Figure 3.27 Fracture, petrophysical and geochemical data plotted versus depth for Core C9, Azraq Area (central-east Jordan).

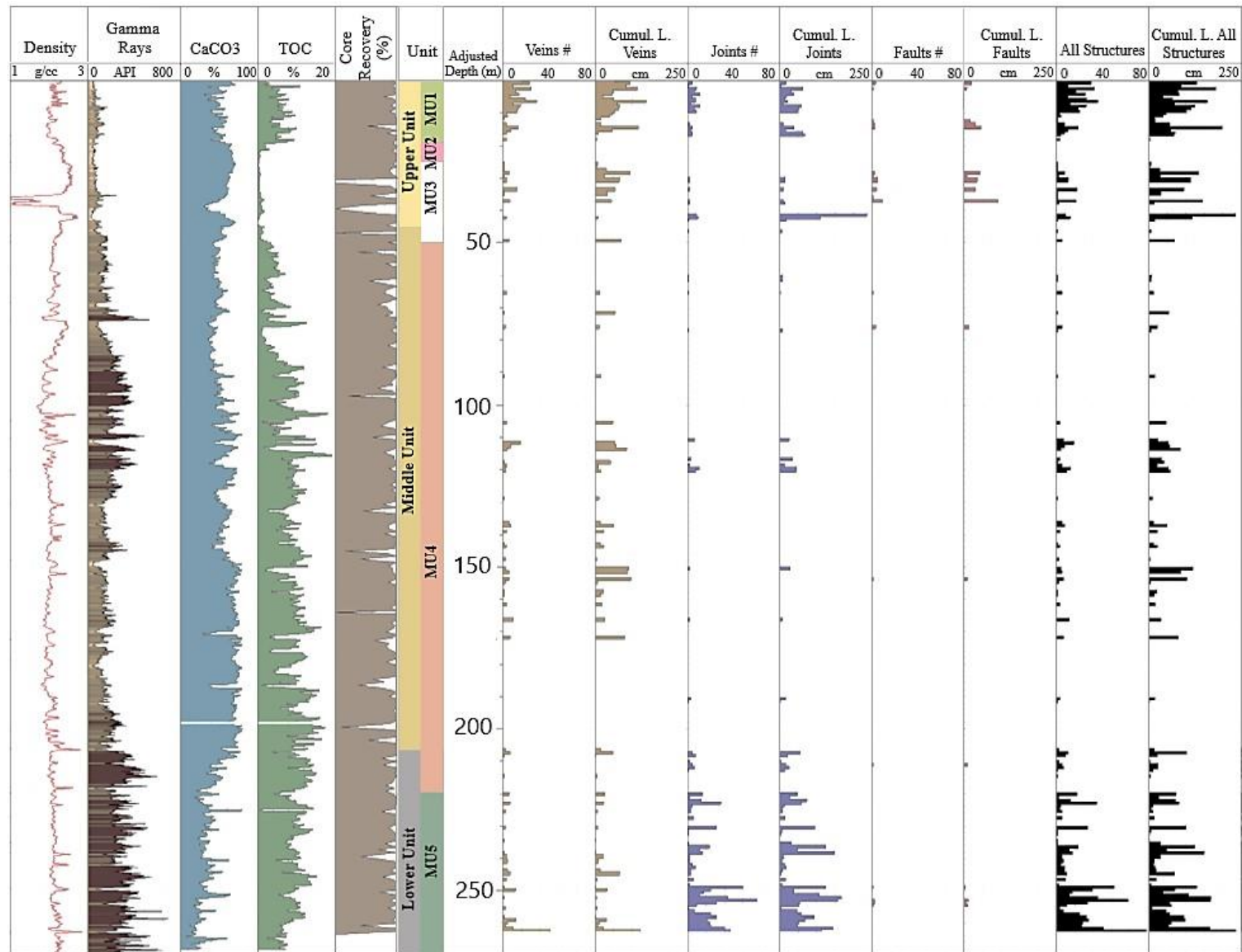


Figure 3.28 Fracture, petrophysical and geochemical data plotted versus depth for Core C10, Azraq Area (central-east Jordan). Note the relatively high intensity of fractures in the Middle Unit compared to the Middle Unit in the other cores.

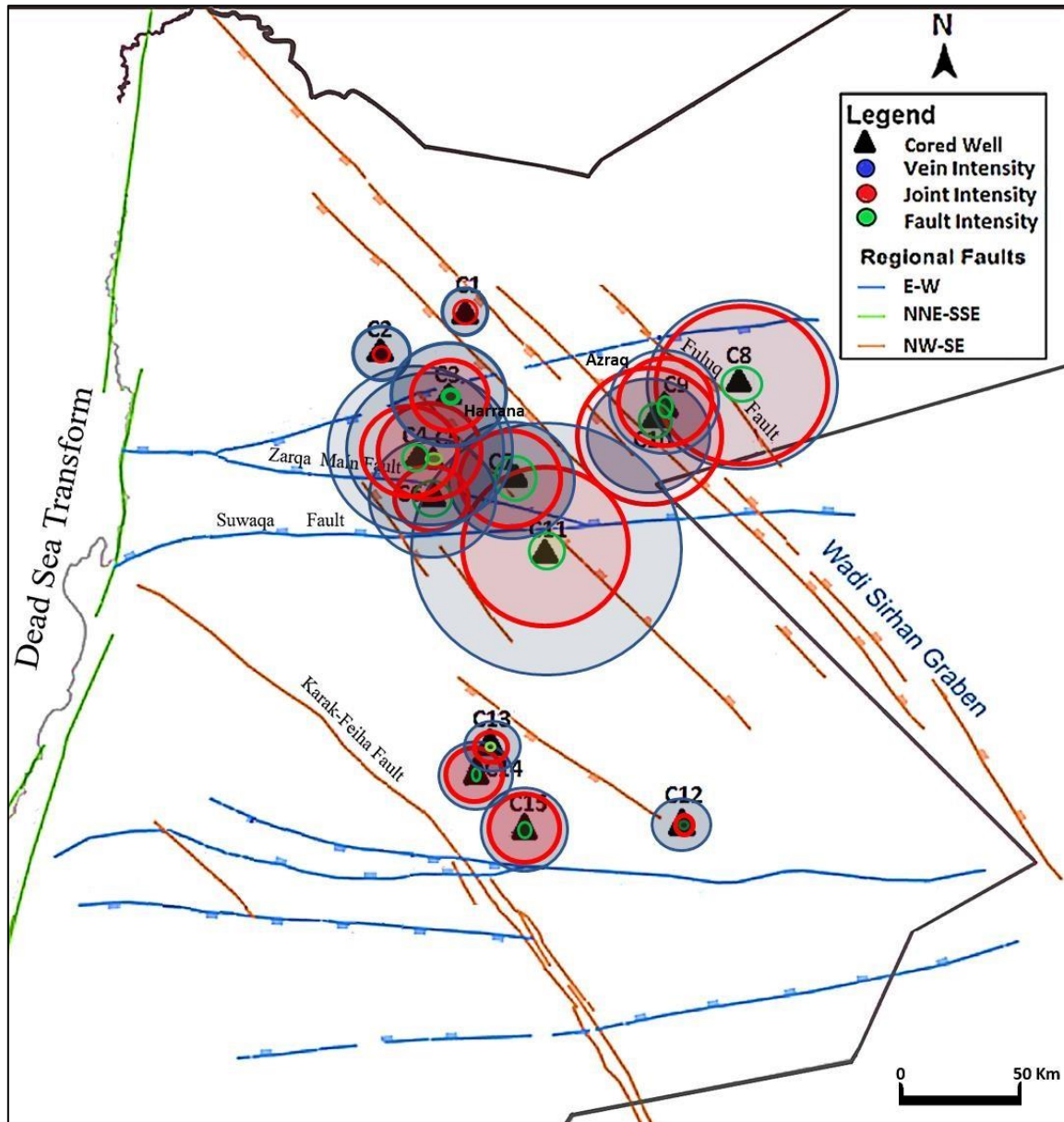


Figure 3.29 Bubble chart showing the intensity of the different fracture sets (veins, joints and faults) described from the fifteen studied cores. Fracture intensity was calculated by dividing the number of fractures by the number of fractured layers (layers with any kind of fracture). The total number of veins in the core C11, represented by the biggest blue circle, is 1343 veins.

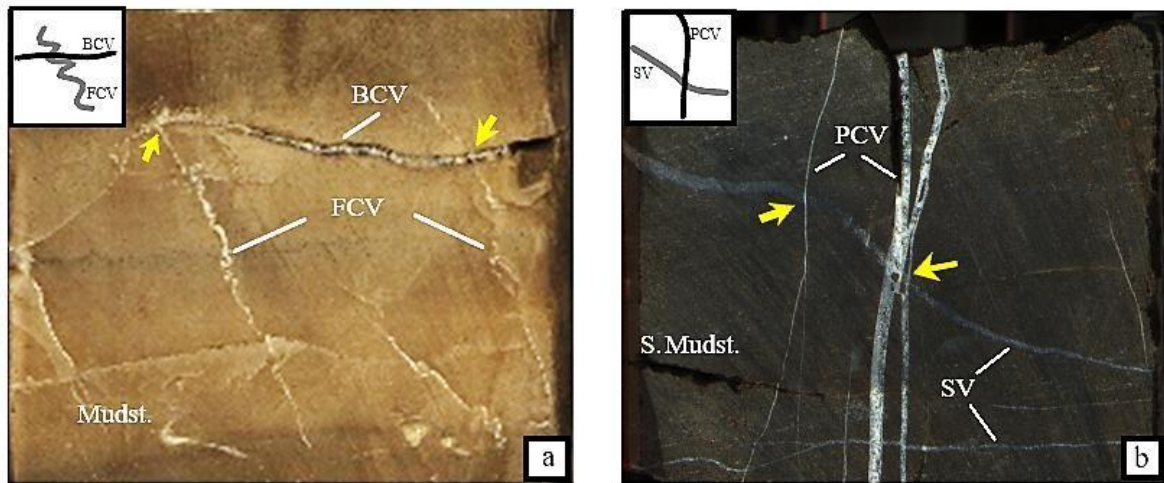


Figure 3.30 Core photographs showing the crosscutting relationship (yellow arrows) between fractures. PCV, Planar Calcite Vein; SV, Silica vein; BCV, Bedding-Parallel Calcite Vein. FCV, Folded Calcite Vein; S. Mudst., Silicified Mudstone. Core photograph width is 10 cm.

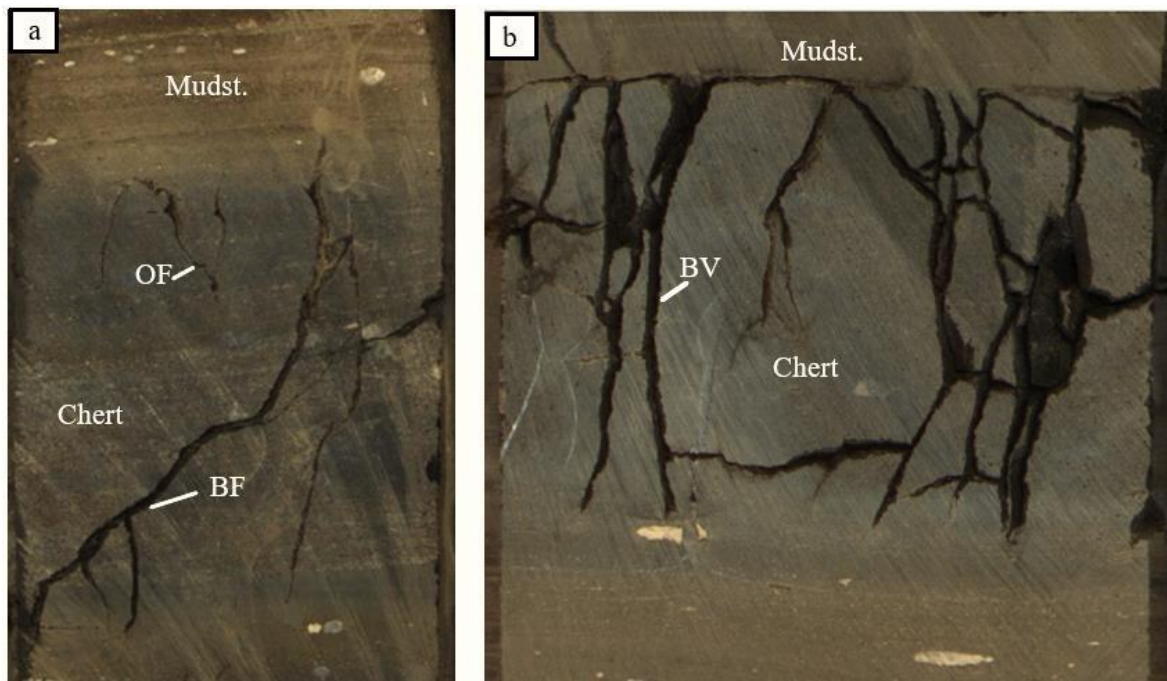


Figure 3.31 Core photographs showing the fracture interaction and connectivity. OF, Open Fracture; BF, Partially-filled bitumen fracture; BV, Bitumen Vein, PCV, Planar Calcite Vein; FCV, Folded Calcite Vein. Note that fractures in both (a) and (b) are confined to beds. Core photograph width is 10 cm.

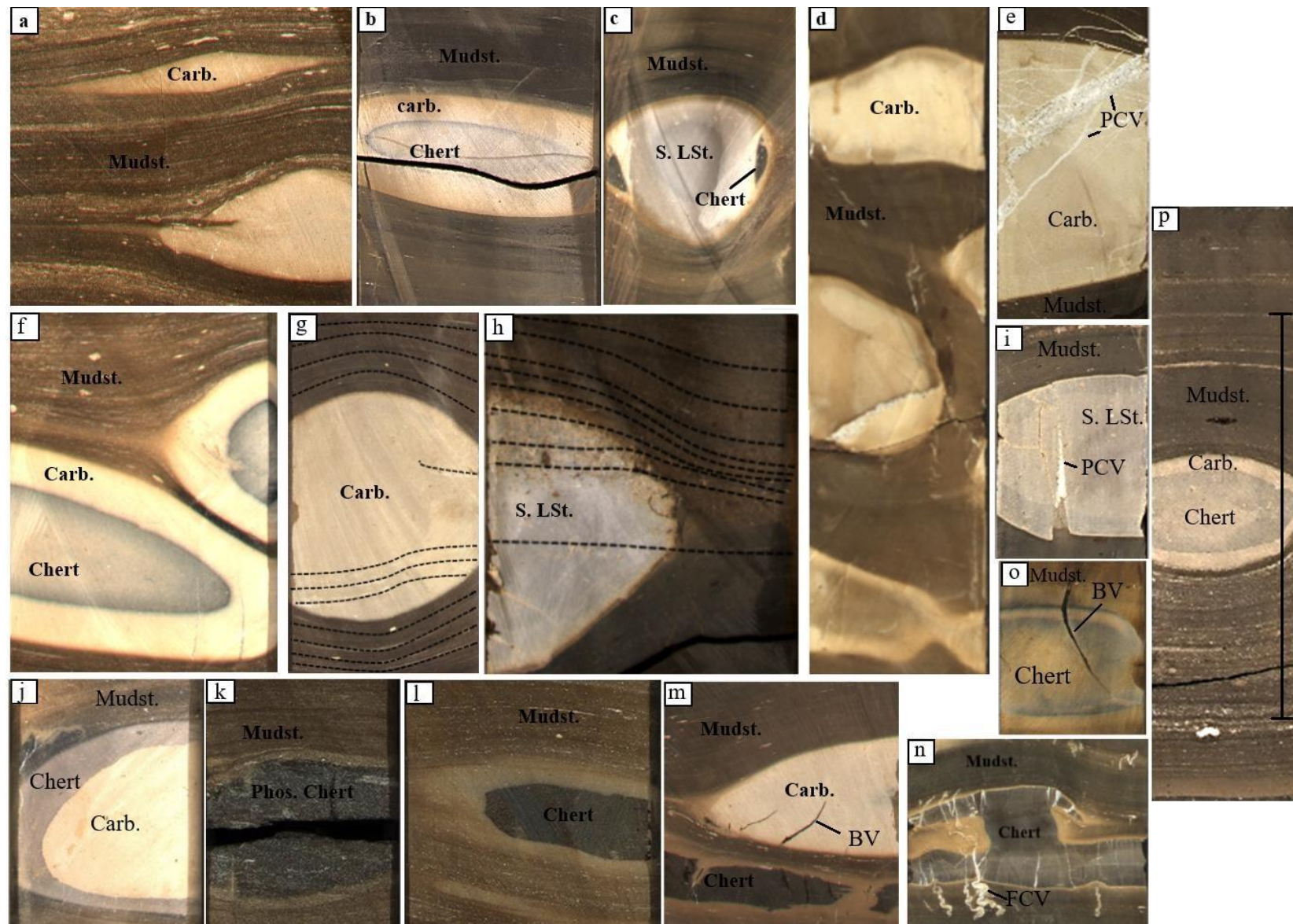


Figure 3.32 Examples from the cores of the different types and shapes of carbonate and chert concretions and their relationships with fractures. Carb. Carbonate; S. LSt., Silicified Limestone; BV, Bitumen Vein; PCV, Planar Calcite Vein; FCV, Folded Calcite Vein. a. Lensoidal carbonate concretion in mudstone. b. Tabular composite chert-carbonate concretion. c. Semi-spherical silicified limestone concretions with chert filling triangle spaces at the edges. d. A cluster of irregular carbonate concretion within a single bed. e. Carbonate concretion with calcite veins inside and around the concretion. f. Elliptical composite chert-carbonate concretion. g. Spherical carbonate concretion preserving the sedimentary laminae. h. Elliptical silicified limestone concretion preserving the sedimentary laminae. Note the bending of laminae around concretions in (g-h). i. A planar calcite vein in a silicified limestone concretion. Note that the vein tapering towards the concretion centre. j. A composite carbonate-chert concretion hosted by mudstone. k. Elliptical phosphatic chert concretion in mudstone. l. Tabular chert concretion in mudstone. m. Elliptical carbonate concretion and lensoidal chert concretion hosted by mudstone. Note the bitumen veins in the carbonate concretion. n. Irregular chert concretion crosscut by folded calcite veins and hosted by mudstone. Bitumen vein crosscut chert concretion and tapering towards the concretion centre. p. A clear example of the laminae bending (strain field) around concretions. Black line represents the strain field. Core width is 10 cm.

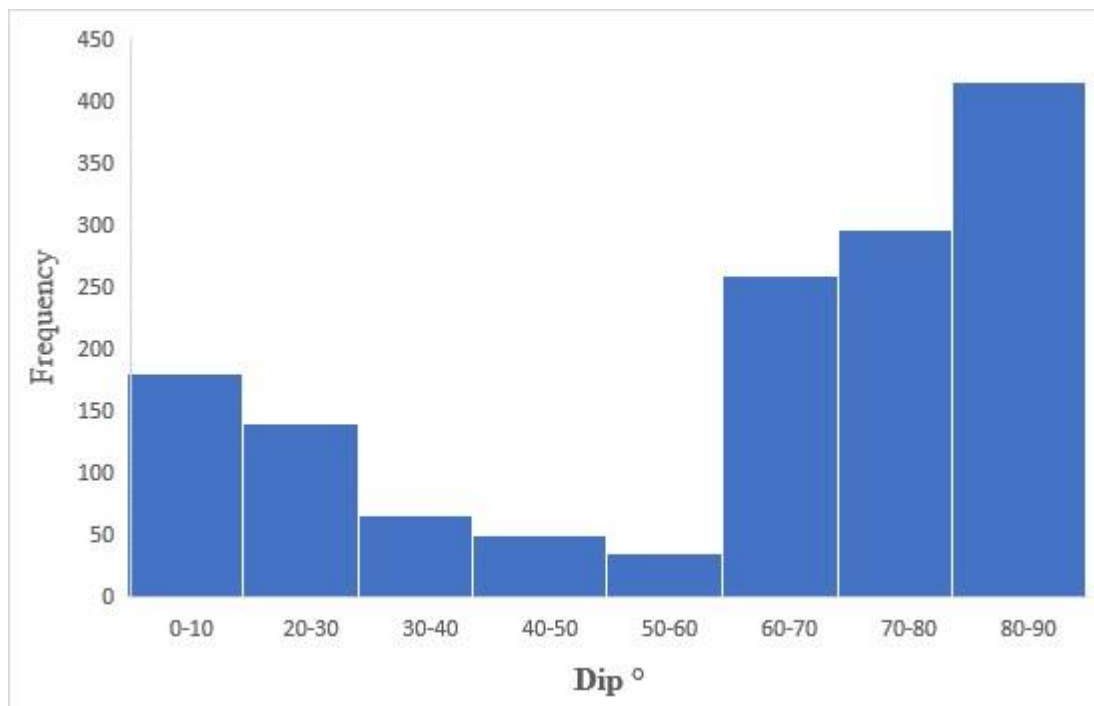


Figure 3.33 Histogram showing the dip angles of opening-mode fractures measured from the fifteen-studied core. N=1439

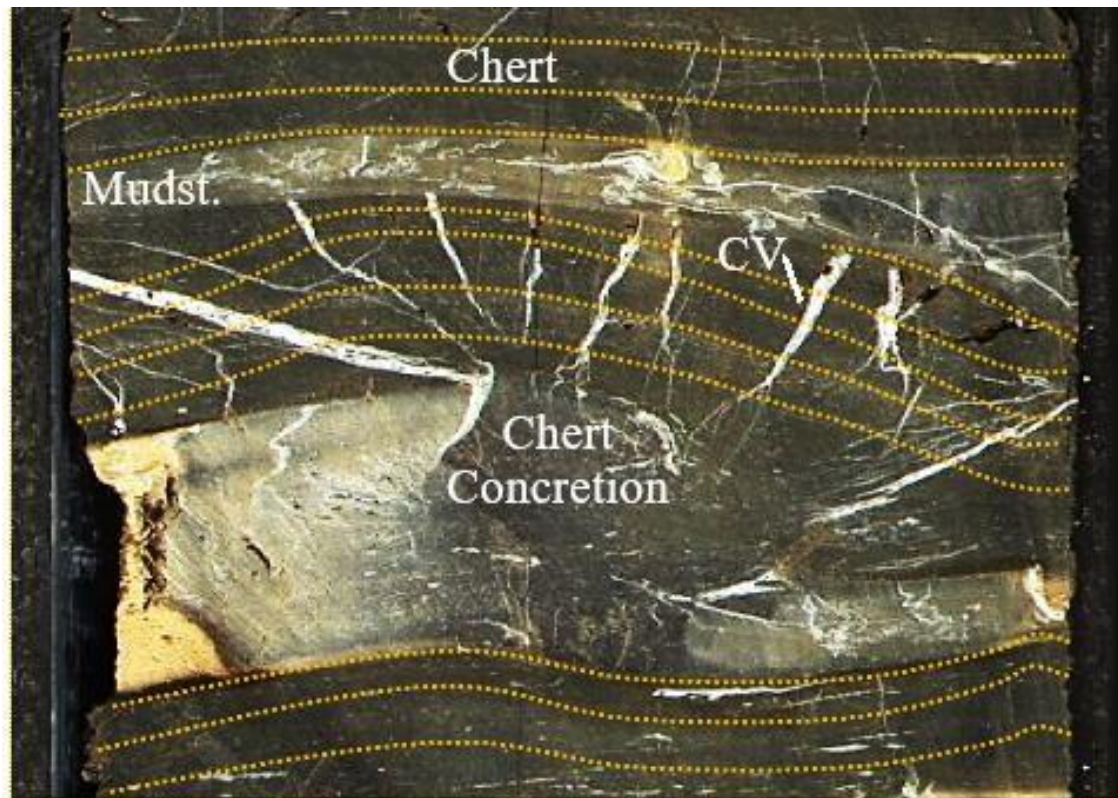


Figure 3.34 Core photograph showing the radial orientation of calcite veins inside a chert concretion. CV, Calcite Vein. Orange-dotted lines are the lamina boundary. Core width is 10 cm.

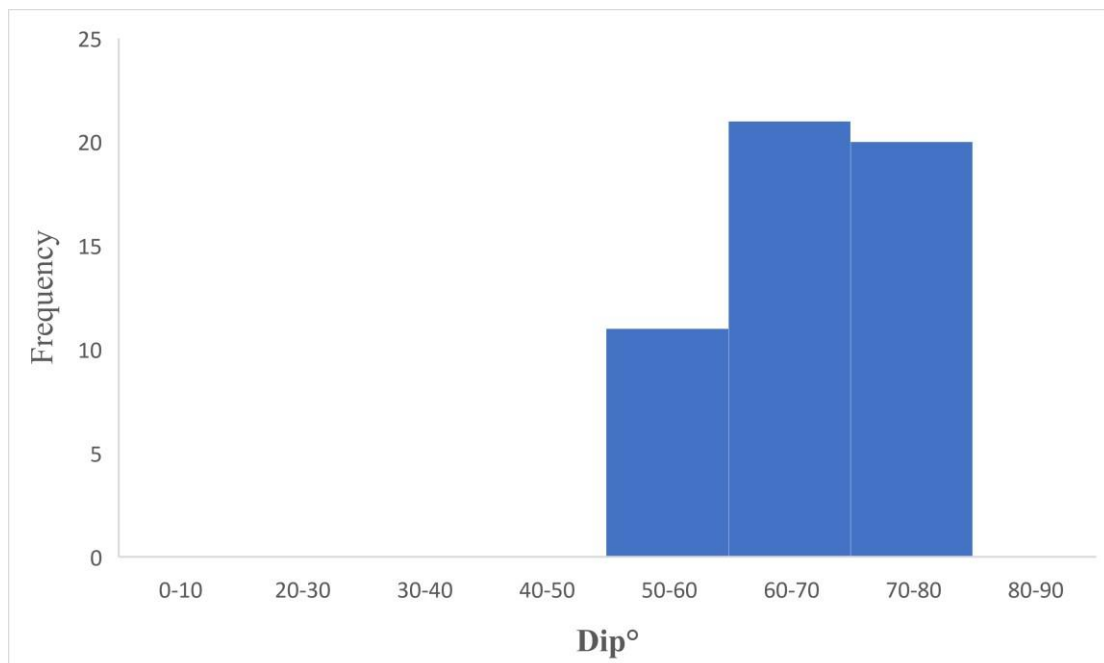


Figure 3.35 Histogram showing the dip angles of shear fractures measured from the fifteen-studied core. N=52.

Core Name	MU#	Adjusted Top Depth(m)	Adjusted Bottom Depth(m)	Thickness (m)	Core Name	MU#	Adjusted Top Depth(m)	Adjusted Bottom Depth (m)	Thickness (m)
C1	MU1	0	11	11	C9	MU1	0	13	13
	MU2	11	26	15		MU2	13	24	11
	MU3	26	55	29		MU3	24	39	15
	MU4	55	135	80		MU4	39	225	186
	MU5	135	169	34		MU5	225	286	61
C2	MU1	0	34	34	C10	MU1	0	20	20
	MU2	34	55	21		MU2	20	28	8
	MU3	55	78	23		MU3	28	51	23
	MU4	78	187	109		MU4	700	221	170
	MU5	187	200	13		MU5	221	264	43
C3	MU1	0	21	21	C11	MU1	0	17	17
	MU2	21	29	8		MU2	17	40	23
	MU3	92	81	52		MU3	40	59	19
	MU4	81	195	114		MU4	59	227	48
	MU5	195	232	37		MU5	227	107	28
C4	MU1	0	14	14	C12	MU1	0	3	3
	MU2	14	21	7		MU2	3	20	17
	MU3	21	62	41		MU3	20	34	14
	MU4	62	103	43		MU4	34	97	63
	MU5	103	146	115		MU5	97	145	48
C5	MU1	0	12	12	C13	MU1	0	7	7
	MU2	12	16	4		MU2	7	14	7
	MU3	16	56	40		MU3	14	30	16
	MU4	56	162	106		MU4	30	130	100
	MU5	162	227	65		MU5	130	202	72
C6	MU1	0	34	34	C14	MU1	0	1	1
	MU2	34	37	3		MU2	1	27	26
	MU3	37	63	26		MU3	27	41	14
	MU4	63	144	81		MU4	41	135	94
	MU5	144	240	96		MU5	135	182	47
C7	MU1	0	23	23	C15	MU1	0	46	46
	MU2	23	40	17		MU2	46	73	27
	MU3	40	85	45		MU3	73	89	16
	MU4	85	223	138		MU4	89	125	36
	MU5	223	258	35		MU5	125	191	66
C8	MU1	0	15	15					
	MU2	15	19	4					
	MU3	19	27	8					
	MU4	27	132	105					
	MU5	132	187	55					

Table 3.3 Mechanical unit tops and thicknesses determined for the fifteen cores C1-C15

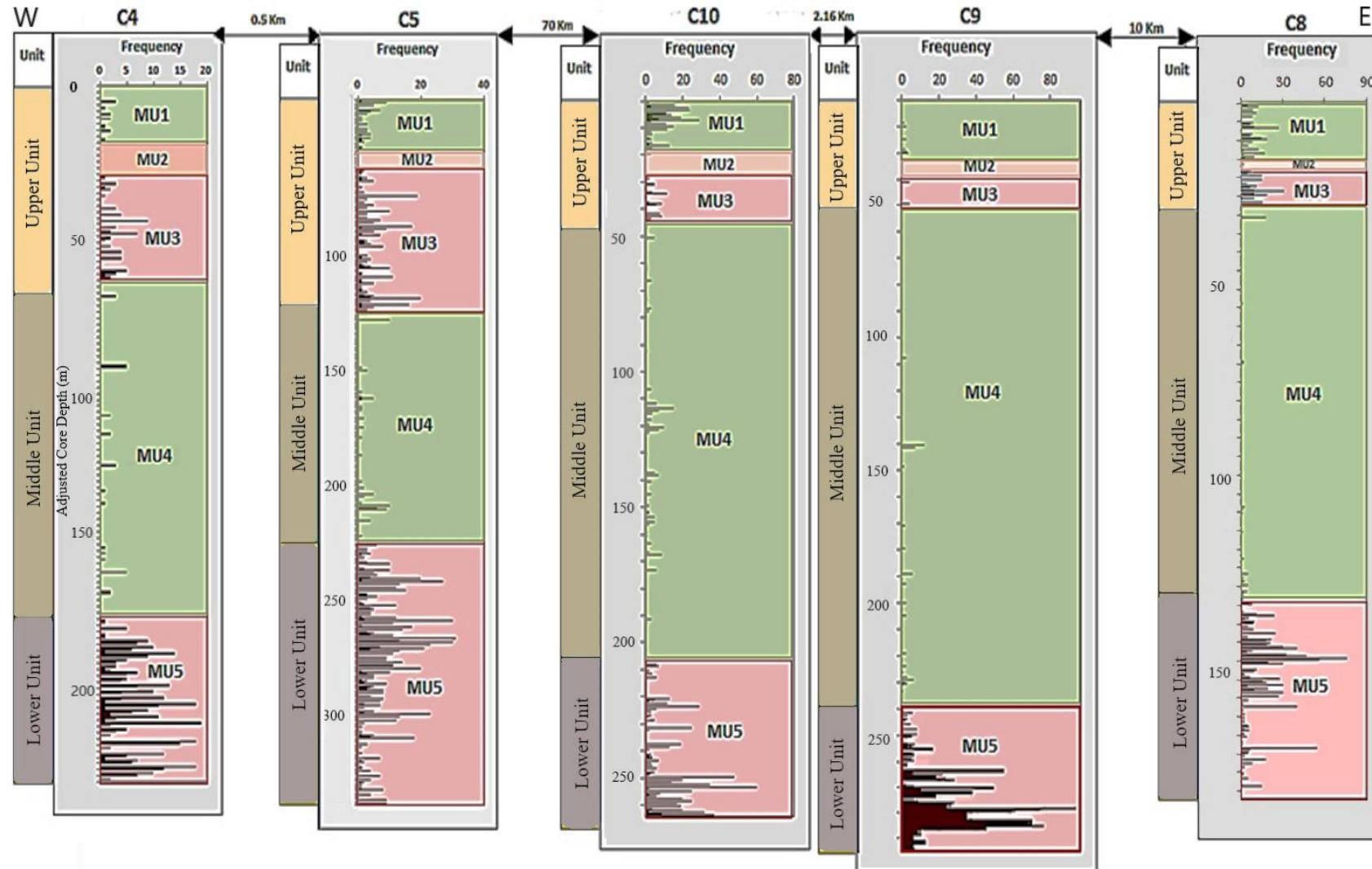


Figure 3.36 A regional correlation panel of fracture distribution and mechanical units (Lower, Middle and Upper) and subunits (MU1-MU5) for the cores C4, C5, C8, C9 and C10. Datum line is the top of the Upper Unit.

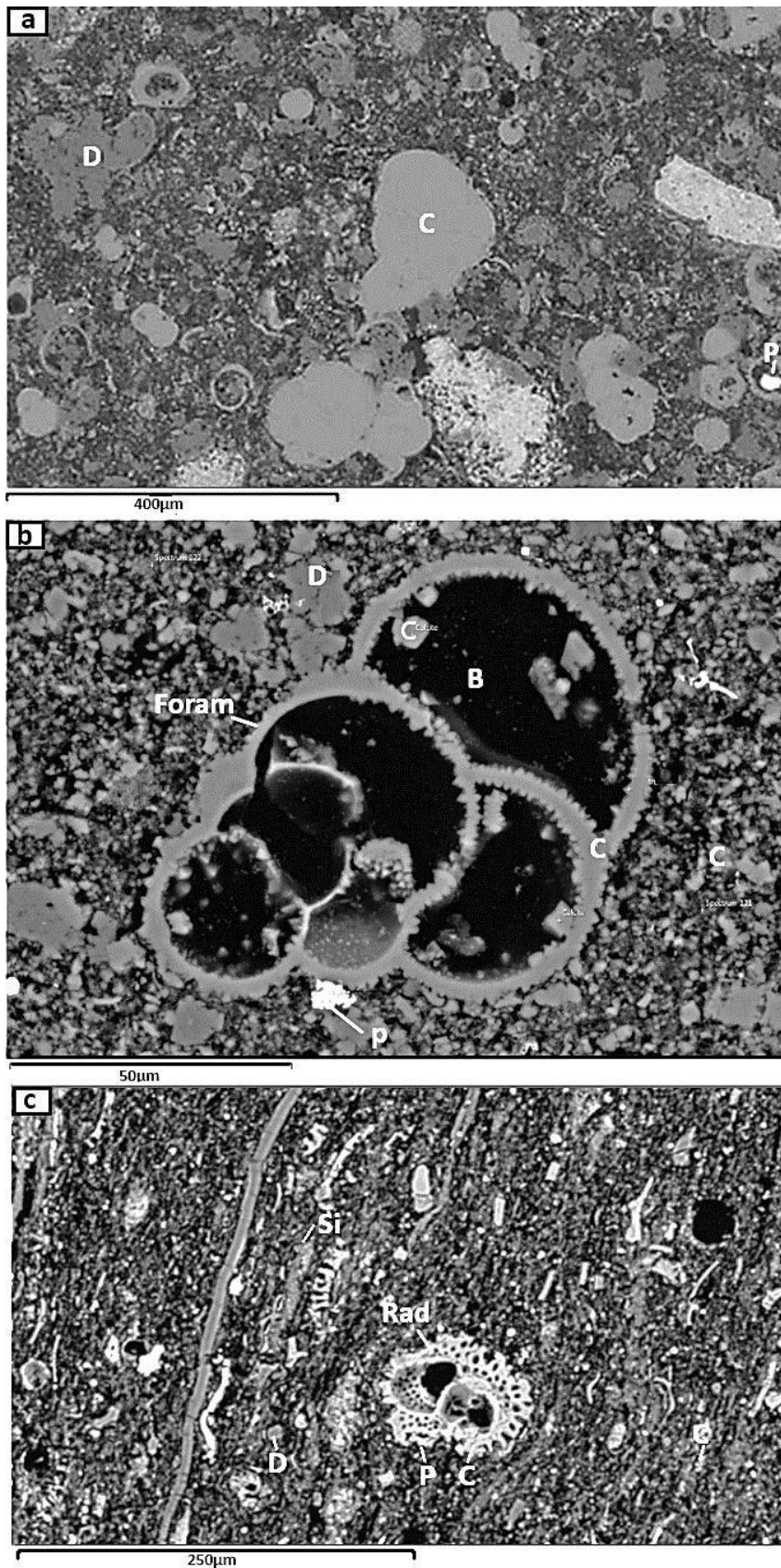


Figure 3.37 BSE images showing the composition of the mudstone intervals. C, Calcite; D, Dolomite; P, Pyrite; B, Bitumen; Si, Silica; Rad, Radiolarian. a. BSE image of marl interval from the Lower Unit, Core C4. Note that the calcite cement entirely fills microfossils and pervasively cements the microporous matrix. b. BSE image of an organic-rich Mudstone from the base of the Middle Unit, Core C9. Note that bitumen is present in both the microfossils (Foraminifera) and the matrix. c. BSE image of phosphatic-rich marl interval from the Lower Unit, Core C13. Note the abundance of fish fragments and bioclasts. The calcified and pyritized radiolarian test in (c) demonstrates that the original silica was dissolved and replaced by calcite.

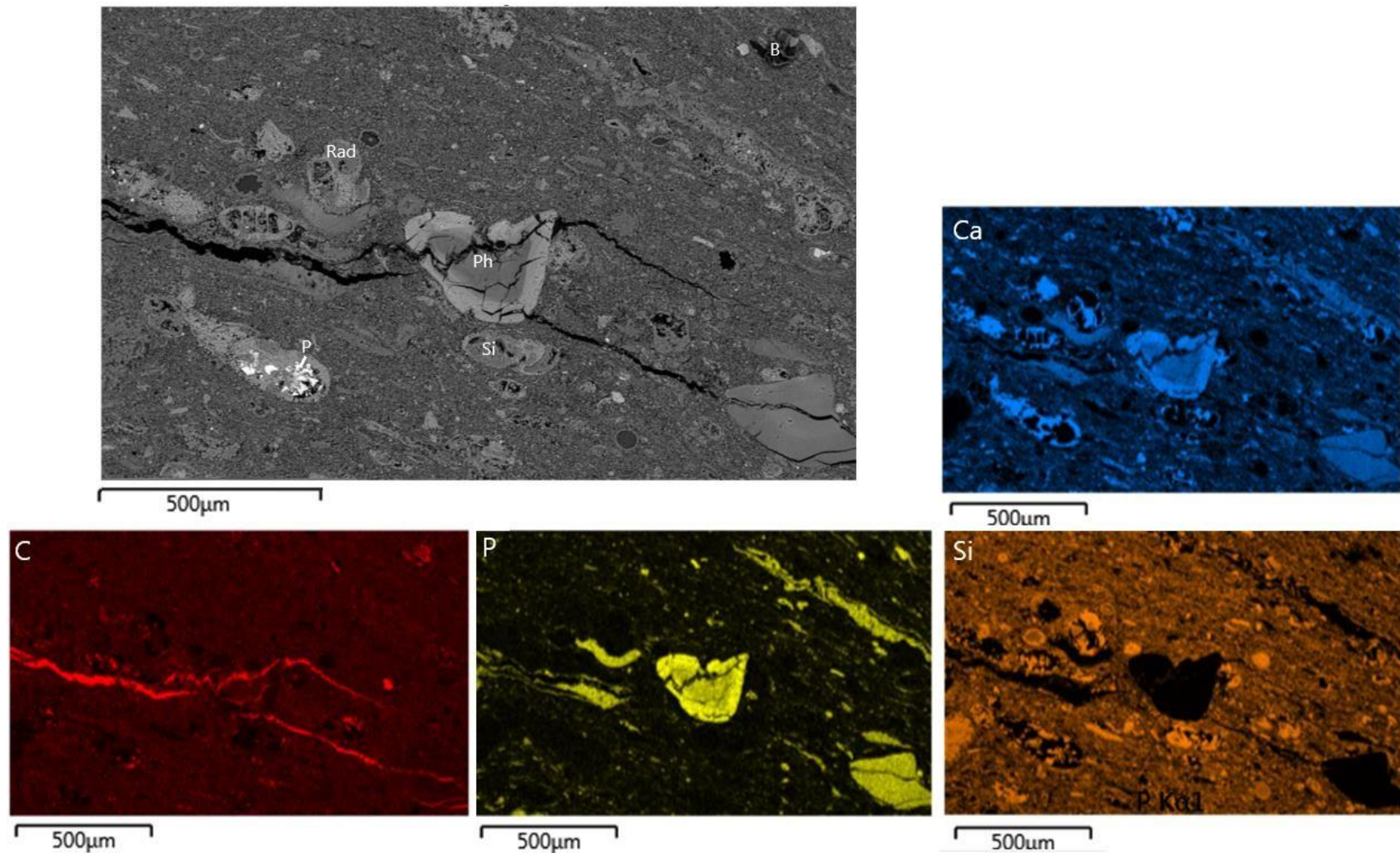


Figure 3.38 BSE image showing the host mudstone in the Lower Unit. Rad, Radiolarian; Ph, Phosphorite; P, Pyrite; Si, Silica; B, Bitumen. Note the bitumen vein transecting the grains and microfossil tests. EDS maps showing the chemical composition (Ca, C, P and Si) of the host rock. Note that the radiolarian is filled by both calcite and silica.

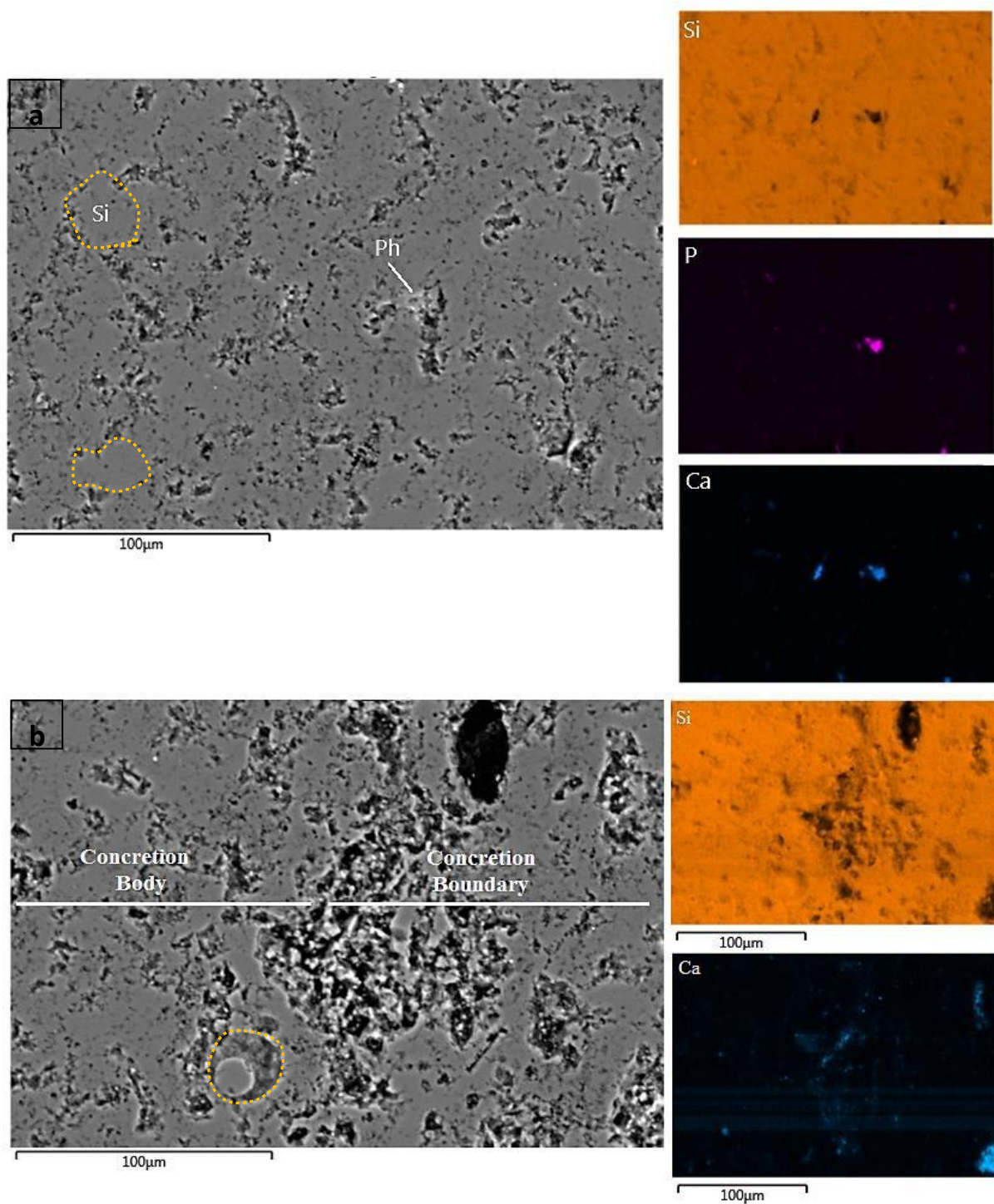


Figure 3.39 a. BSE image of chert concretion from the Lower Unit, Core 13. Note that Silica represents >90%. On the right, EDS images showing mineral composition of the chert concretion on the left. Si, Silica; Ph, Phosphorite. b. BSE image showing the gradational boundary of the chert concretion. On the right, EDS maps showing mineral composition of the boundary. Note the change from silica to calcite towards the concretion boundary. Orange-dotted shapes show the "ghost" of radiolaria and foraminifera.

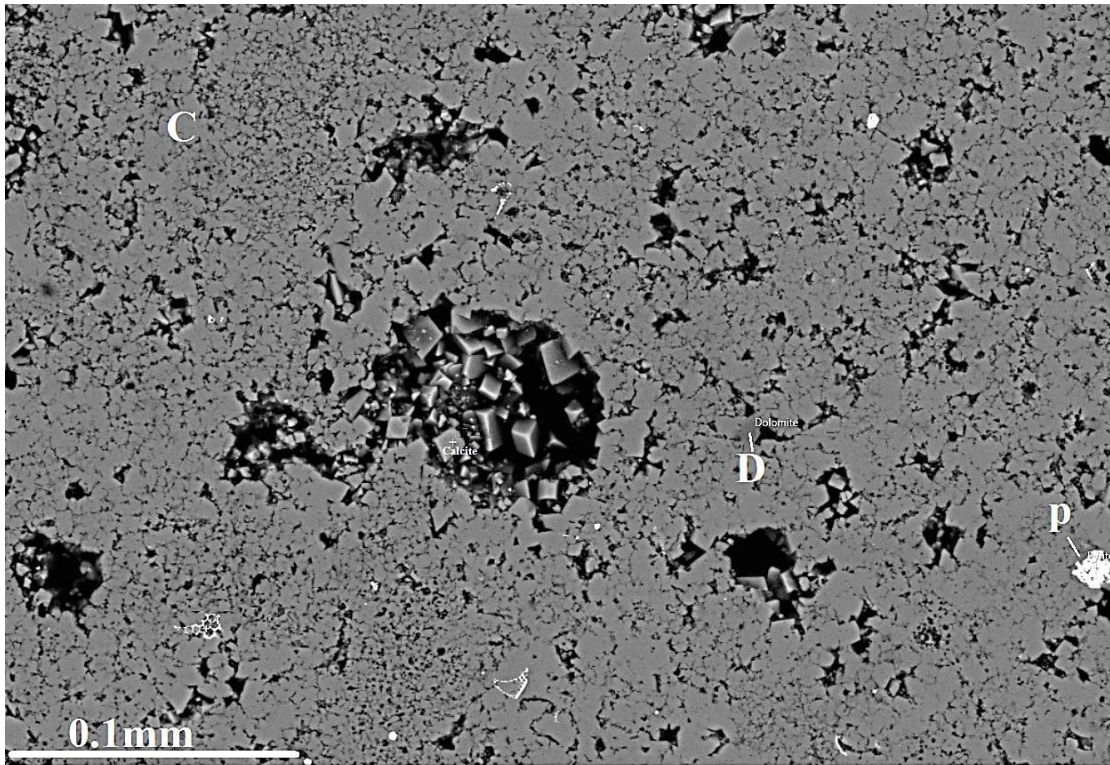


Figure 3.40 BSE image of carbonate concretion from the Lower Unit, Core C9. C, Calcite; D, Dolomite; P, Pyrite. Note that >90% of the concretion is calcite.

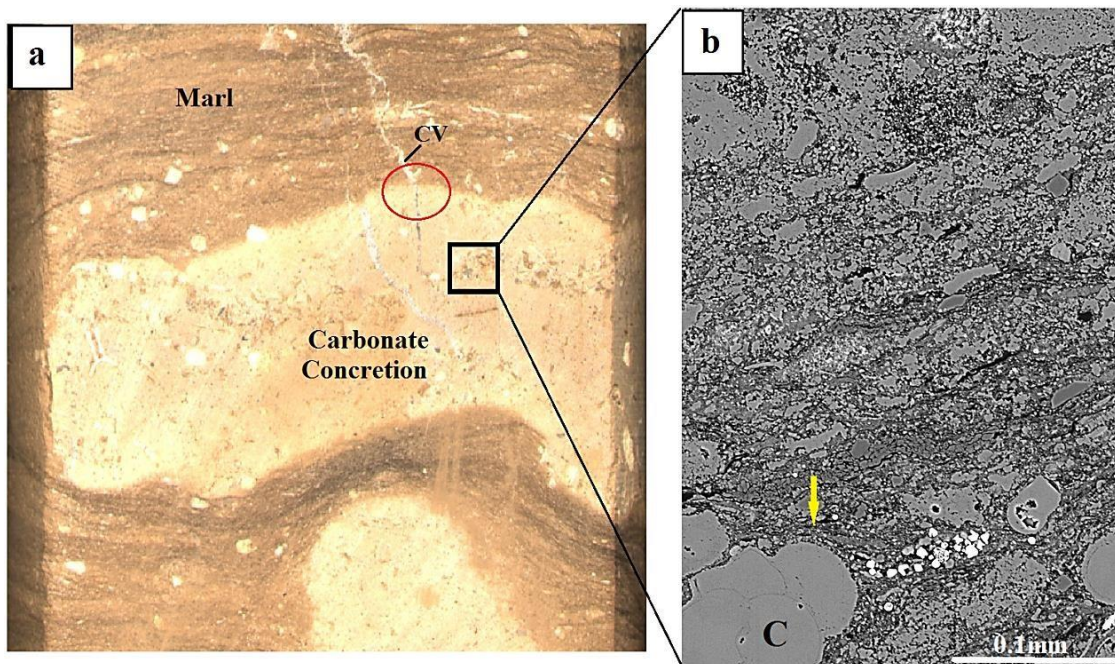


Figure 3.41 a. Core photograph of carbonate concretion in phosphatic marl interval from the Lower Unit, Core C5. CV, Calcite Vein. Core width is 10 cm. Note the change of vein direction when crossing the carbonate concretion (red circle). b. BSE image of inset in (a) showing the bending of sediment around microfossils (yellow arrow).

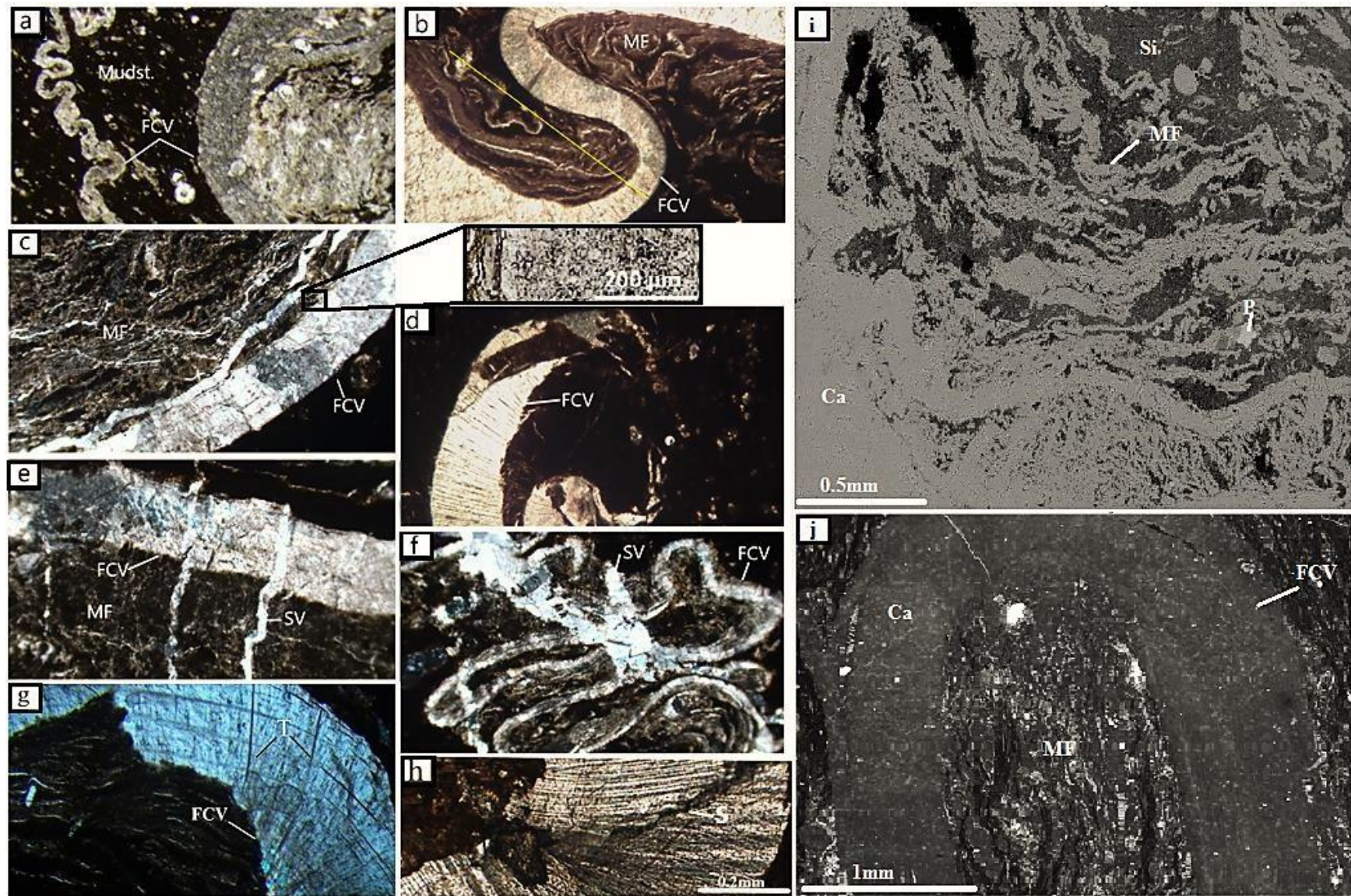


Figure 3.42 Microphotograph (CL) of folded calcite veins. FCV, Folded Calcite Vein; MF, Microfractures; SV, Silica Vein; T, Twin plane; S, Stylolite; Ca, Calcite; Si, Silica; P, Pyrite. Field of View from (a) to (g) 4mm. a. Coexistence of thick and thin folded veins in mudstone interval in the Lower Unit, Core C5. b. Microfractures within the inner arcs of a folded vein. Note the unidirectional folding axis of both the folded vein and its associated microfractures (yellow line). c. Folded vein showing crack-seal texture and microfractures in the inner arc. d. Branching tip of folded calcite vein. e. Silica veins crosscutting folded calcite vein. f. Excessive folding of a thin, folded calcite vein that is crosscut by silica vein. g. Twin planes on the hinge of the folded calcite vein. Note the folding of the twin planes with the vein folding. h. Stylolite in the folded calcite vein. i. and j. BSE images showing the microfractures folding in the inner arc of the folded veins.

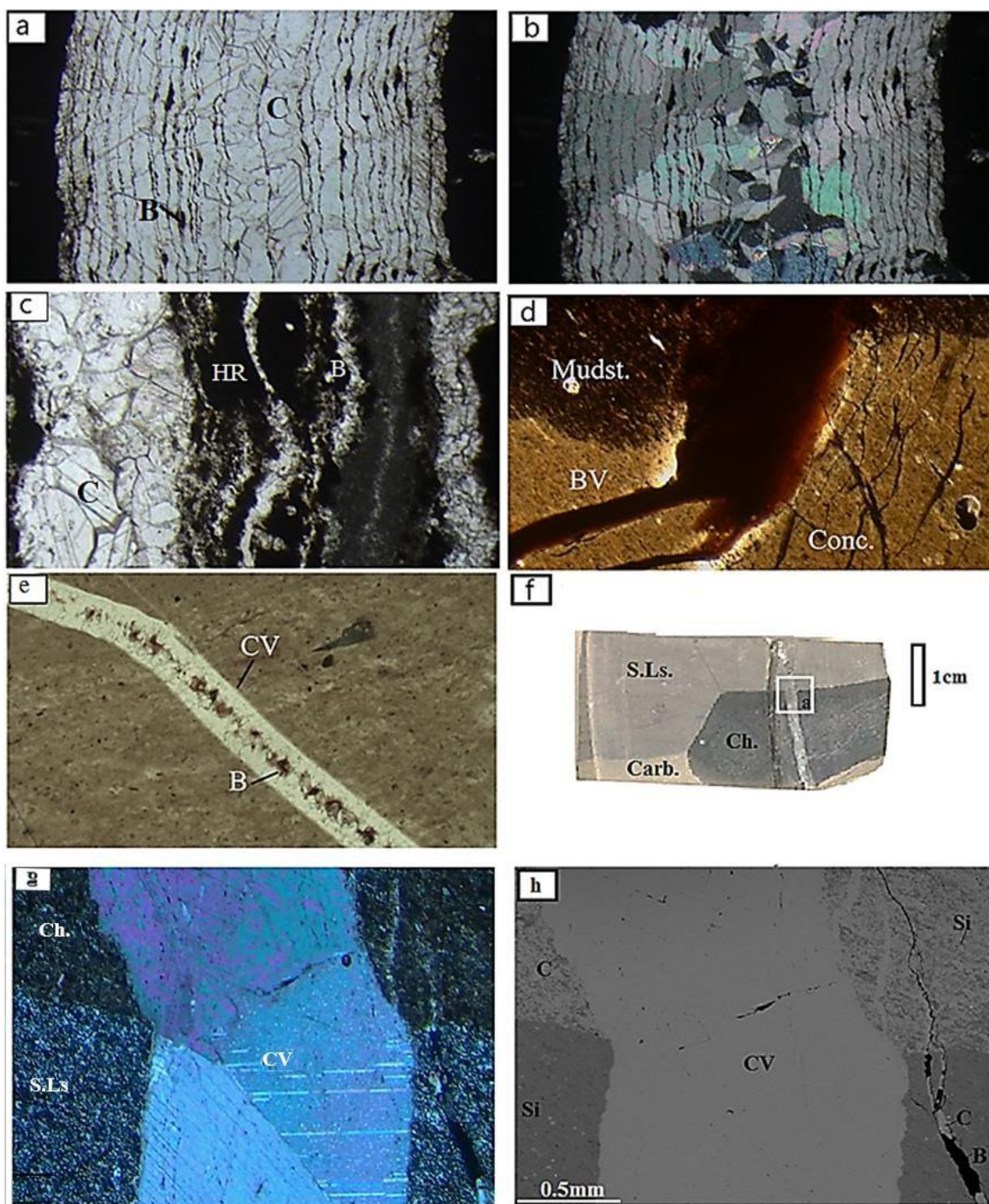


Figure 3.43 Planar calcite veins microphotographs. Field of view is 2.6 mm. C, Calcite; B, Bitumen; BV, Bitumen Vein; Conc., Carbonate Concretion; CV, Calcite Vein; S.Ls., Silicified Limestone; Ch, Chert; Si, Silica. a. Planar calcite vein showing the crack-seal texture in PPL. b. same as (a), but in CL. c. PPL photo micrograph of planar calcite veins showing the host rock and bitumen inclusions within the crack-seal texture. Note the sparry calcite crystals. d. Bitumen veins crossing a carbonate concretion. e. Planar calcite veins exhibiting syntaxial texture, with bitumen include at the median line. f. Hand sample of planar calcite vein crossing silicified limestone and chert concretion. g. microphotograph in CL of inset in (f) showing a bending in the calcite vein when it crosses the chert concretion. h. BSE image of (g).

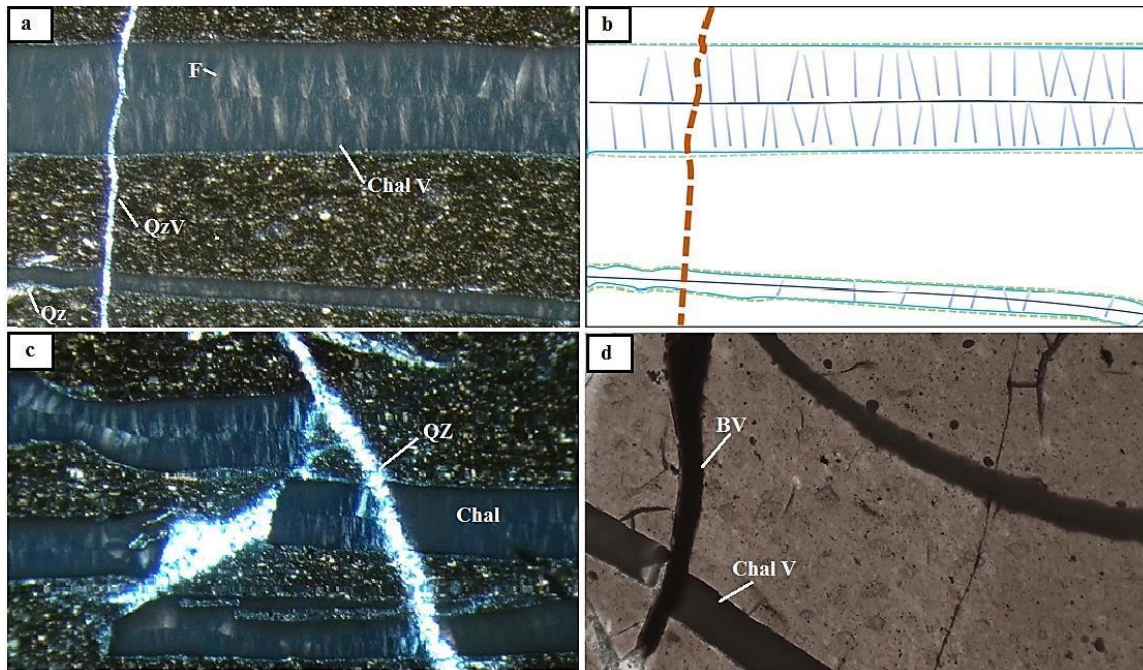


Figure 3.44 Microphotograph showing silica veins. F, Fibre; Chal V, Chalcedony Vein; Qz, Quartz, QzV, Quartz Vein; BV, Bitumen Vein. Field of view is 2.6mm. a. Horizontal chalcedony veins and vertical microcrystalline quartz vein in chert interval from the Lower Unit, Core C9. b. Schematic drawing of (a). Note that the vertical quartz vein (brown-dashed line) crosscutting fibrous chalcedony veins (blue lines), and thinner quartz veins (green-dashed, horizontal lines), which are aligned parallel to the walls of the chalcedony veins, are also crosscut by the vertical quartz vein, implying multiple stages of silica veins growth. c. Sub-vertical quartz veins crosscutting horizontal chalcedony veins. d. Vertical bitumen vein crosscutting sub-horizontal chalcedony veins in silicified limestone interval in the Lower Unit of the core.

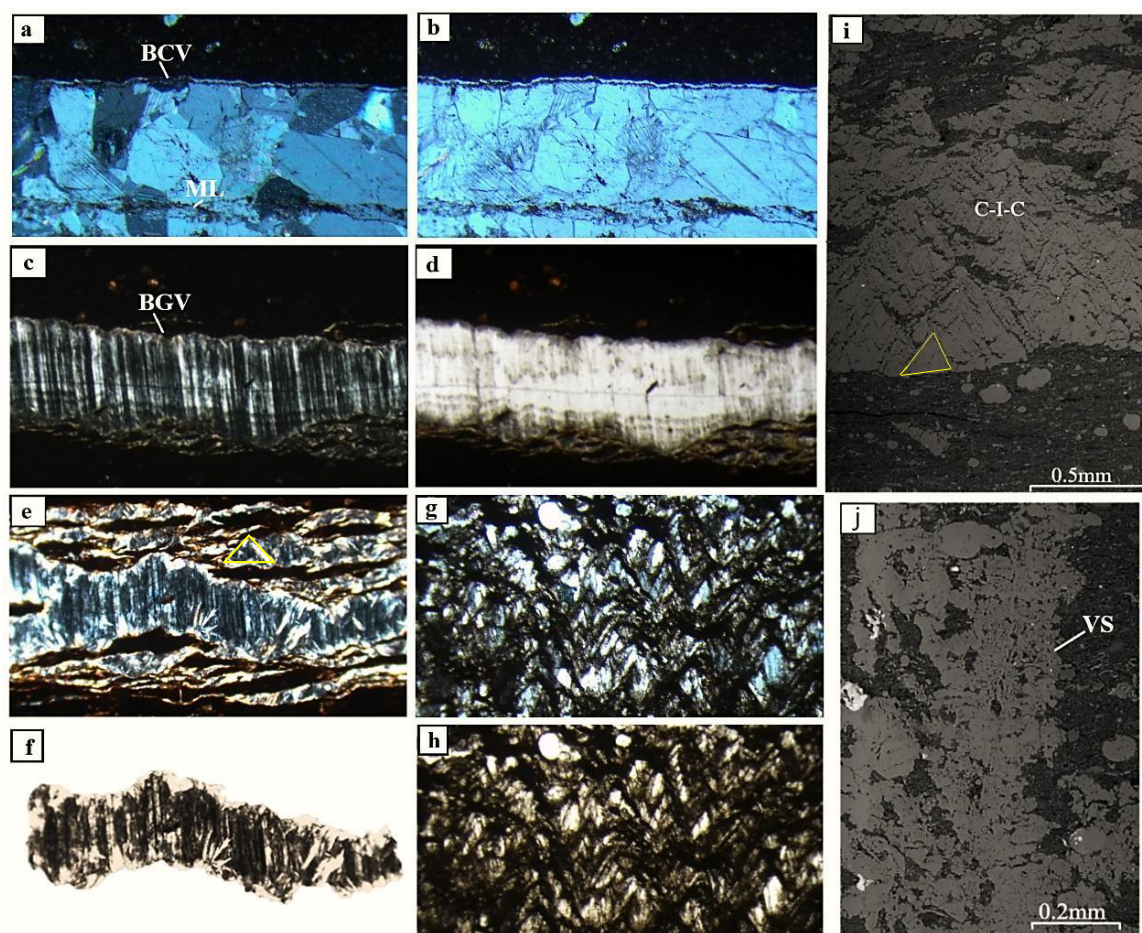


Figure 3.45 Examples of bedding-parallel calcite veins. Field of view from (a) to (g) is 2.6 mm. BCV, Bedding-Parallel Calcite Vein; BGV, Bedding-Parallel Gypsum Vein; ML, Median Line; C-I-C, Cone-in-Cone; VS, Vertical Strand. a. CL microphotograph of bedding-parallel calcite vein showing elongate-blocky texture. Note the Hostrock inclusions along the median line. b. Same as (a), but in PPL. c. CL microphotograph of bedding-parallel gypsum vein showing fibrous uniaxial texture. d. Same as (c), but in PPL. e. CL microphotograph of horizontal aggregates of fibrous calcite cone-in-cone. Note the thick fibrous calcite zone in the middle. f. Illustration of the fibrous zone in (f). g. CL microphotograph of calcite cone-in-cone. h. Same as (g), but in PPL. i. BSE image of calcite cone-in-cone structure. Note the conical shapes (yellow triangle) and the lower sharp contact with the host rock. j. BSE image of a vertical strand of cone-in-cone.

Sample ID	Core Name	Hostrock Lithology	Type of Calcite Structure	Vein $\delta^{13}\text{C}$	Vein $\delta^{18}\text{O}$	HR $\delta^{13}\text{C}$	HR $\delta^{18}\text{O}$	Sample ID	Core Name	Hostrock Lithology	Type of Calcite Structure	Vein $\delta^{13}\text{C}$	Vein $\delta^{18}\text{O}$	HR $\delta^{13}\text{C}$	HR $\delta^{18}\text{O}$
1	C13	Marl	Planar	-11.21	-10.81	-7.18	-5.59	23	C2	Marl	Folded	-9.43	-2.76	-9.00	-3.25
2	C6	Marl	Planar	-11.93	-7.82	-8.71	-8.43	24	C5	Shale	Folded	-10.67	-7.88	-6.54	-3.63
3	C3	Marl	Planar	-10.66	-6.68	-5.22	-3.18	25	C6	Marble	Planar	-11.58	-11.75	-17.1	-11.52
4	C3	Marl	Planar	-10.03	-12.61	-6.01	-3.34	26	C6	Marble	Planar	-16.78	-10.31	-16.29	-10.69
5	C7	Marl	Planar	-10.35	-12.76	-1.40	-4.97	27	C6	Marble	Planar	-10.70	-10.79	-20.12	-9.96
6	C10	SM	Planar	-7.27	-11.85	-7.64	-6.27	28	C6	L.St	Planar	-12.95	-9.16	-9.33	-7.68
7	C7	L.St	Planar	-6.95	-14.53	-8.08	-7.90	29	C8	SL	Planar	-6.85	-6.24	-6.48	-3.60
8	C13	SL	Planar	-13.20	-8.65	-12.91	-8.80	30	C8	SM	Planar	-8.67	-4.82	-8.51	-3.13
9	C13	SL	Planar	-12.58	-8.18	-12.91	-8.81	31	C9	SL	Planar	-6.47	-4.79	-6.10	-6.04
10	C15	Marl	Planar	-10.15	-11.79	-11.78	-6.88	32	C4	Shale	Fault	-1.15	-13.77	-2.53	-3.47
11	C1	Marl	Planar	-9.15	-10.22	-9.74	-6.19	33	C4	Marl	Fault	-1.13	-15.62	-4.95	-4.42
12	C13	Shale	Planar	-4.33	-9.43	-1.07	-3.34	34	C12	Shale	Fault	-5.80	-14.00	-1.95	-2.71
13	C13	SL	Planar	-8.72	-11.43	-3.67	-2.26	35	C13	marl	cone-in cone	-7.86	-9.56	-10.88	-3.82
14	C6	SL	Planar	-15.65	-6.33	-13.64	-5.91	36	C2	marl	cone-in cone	-5.99	-6.48	-4.21	-3.60
15	C6	PM	Planar	-12.99	-11.34	-4.11	-3.14	37	C12	marl	cone-in cone	-7.77	-2.94	-7.30	-5.05
16	C13	SL	Planar	-14.16	-9.16	-12.91	-8.81	38	C2	marl	cone-in cone	-6.72	-4.56	-6.12	-4.50
17	C6	PM	Planar	-22.83	-10.68	-4.96	-3.87	39	C14	ML	cone-in cone	-8.47	-9.51	-5.74	-4.67
18	C13	Marl	Folded	-9.22	-11.65	-7.18	-5.59	40	C12	ML	H. blocky vein	-5.12	-13.59	-4.79	-10.37
19	C2	Marl	Folded	-8.12	-9.41	-5.61	-4.80	41	C6	Shale	H. fibrous vein	-9.10	-7.20	-6.46	-5.37
20	C5	Marl	Folded	-14.91	-6.67	-9.22	-5.16	42	C6	L.st	H. fibrous vein	-4.46	-12.88	-6.40	-6.75
21	C1	Marl	Folded	-7.48	-11.06	-5.02	-6.21								
22	C6	Marl	Folded	-9.95	-8.05	-9.23	-8.40								

L.St, Limestone; SM, Silicified Marl; SL, Silicified Limestone, PM, Phosphatic Marl; HR, Host Rock; SM, Silicified Marl; ML, Marly Limestone; H, Horizontal.

Table 3.4 $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ compositions of calcite structures and their host rocks

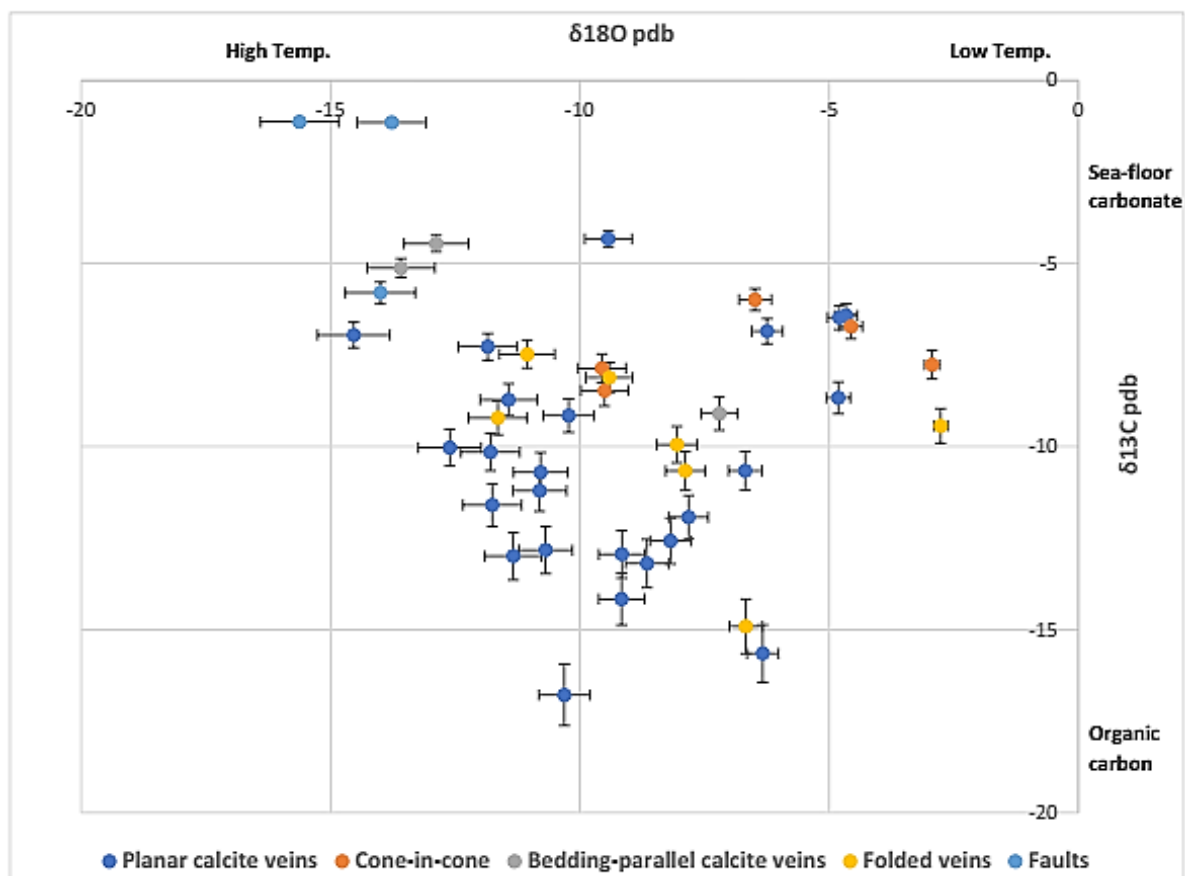


Figure 3.46 Plot of $\delta^{13}\text{C}$ against $\delta^{18}\text{O}$ compositions of all calcite structures analysed.

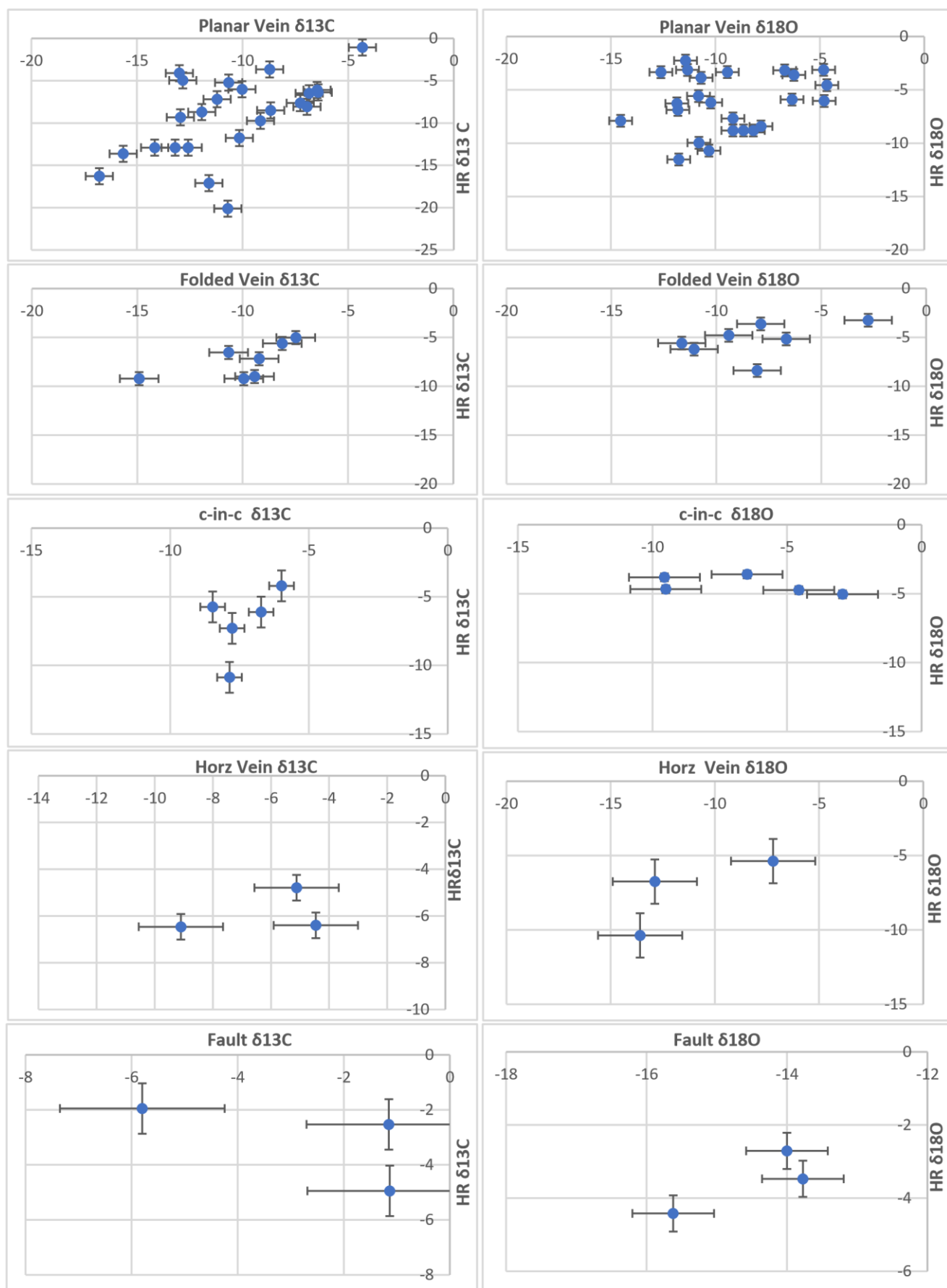


Figure 3.47 Plot of $\delta^{13}\text{C}$ against $\delta^{18}\text{O}$ compositions of all samples analysed.

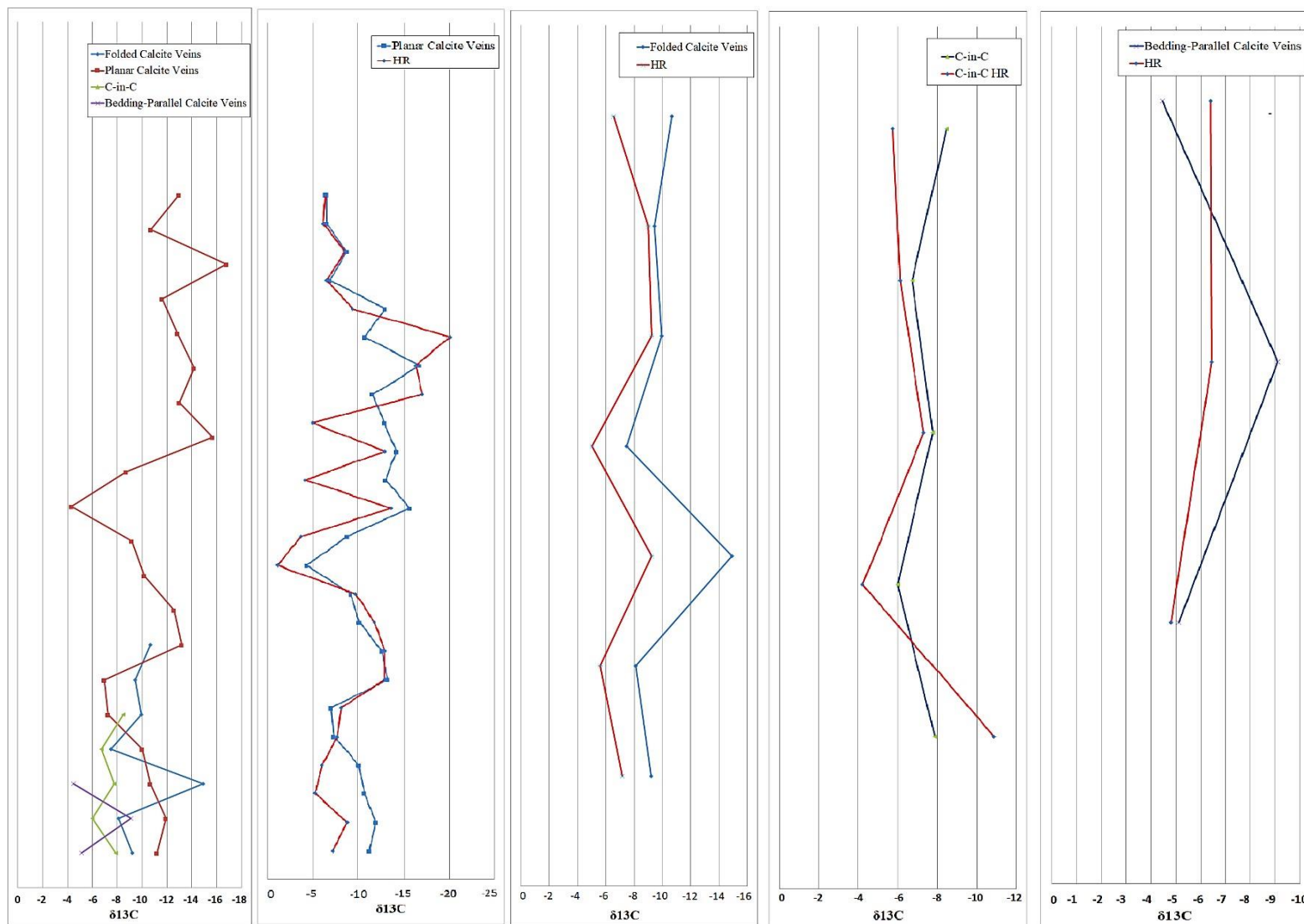


Figure 3.48 Plots of $\delta^{13}\text{C}$ of the different calcite structures analysed and their host rocks.

Sample Number	Core Name	HR Lithology	Host Rock $^{87}\text{Sr}/^{86}\text{Sr}$	Calcite Structure $^{87}\text{Sr}/^{86}\text{Sr}$	Calcite Structure Type	Sample Number	Core Name	HR Lithology	Host Rock $^{87}\text{Sr}/^{86}\text{Sr}$	Calcite Structure $^{87}\text{Sr}/^{86}\text{Sr}$	Calcite Structure Type
1	C2	Marl	0.7078	0.7078	Folded Vein	17	C1	Marl	0.7079	0.7078	Planar Vein
2	C13	Marl	0.7079	0.7077	Folded Vein	18	C15	Marl	0.7080	0.7079	Planar Vein
3	C2	Marl	0.7078	0.7079	Folded Vein	19	C8	Marl	0.7078	0.7079	Planar Vein
4	C1	Marl	0.7078	0.7079	Folded Vein	20	C4	Marl	0.7078	0.7079	Planar Vein
5	C3	Marl	0.7078	0.7078	Planar Vein	21	C9	Marl	0.7079	0.7078	Planar Vein
6	C3	Marl	0.7077	0.7079	Planar Vein	22	C6	Shale	0.7079	0.7078	Horz Vein
7	C7	Shale	0.7078	0.7078	Planar Vein	23	C6	Limestone	0.7079	0.7078	Horz Vein
8	C7	Marl	0.7078	0.7078	Planar Vein	24	C12	Shale	0.7078	0.7079	Horz Vein
9	C10	Marl	0.7078	0.7078	Planar Vein	25	C13	Limestone	0.7078	0.7077	C-in-C
10	C13	Shale	0.7078	0.7078	Planar Vein	26	C1	Marl	0.7078	0.7078	C-in-C
11	C12	Marl	0.7078	0.7078	Planar Vein	27	C12	Marl	0.7079	0.7078	C-in-C
12	C6	Marl	0.7077	0.7078	Planar Vein	28	C2	Marl	0.7078	0.7078	C-in-C
13	C6	Marble	0.7079	0.7078	Planar Vein	29	C2	Marl	0.7078	0.7077	C-in-C
14	C5	Marl	0.7079	0.7080	Planar Vein						
15	C6	Marble	0.7079	0.7078	Planar Vein						
16	C6	Marl	0.7078	0.7080	Planar Vein						

Table 3.5 $^{87}\text{Sr}/^{86}\text{Sr}$ data of the calcite structures and their host rocks

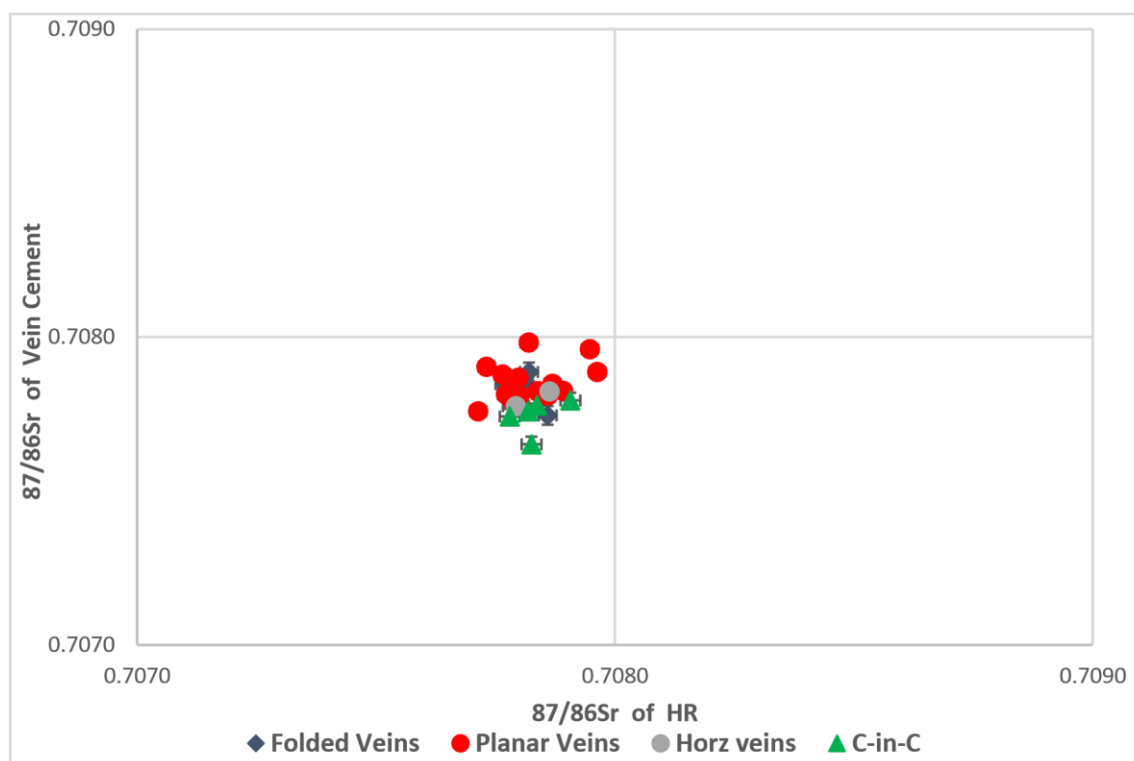


Figure 3.49 Plot of the $^{87}\text{Sr}/^{86}\text{Sr}$ data of the calcite structures against their host rocks.

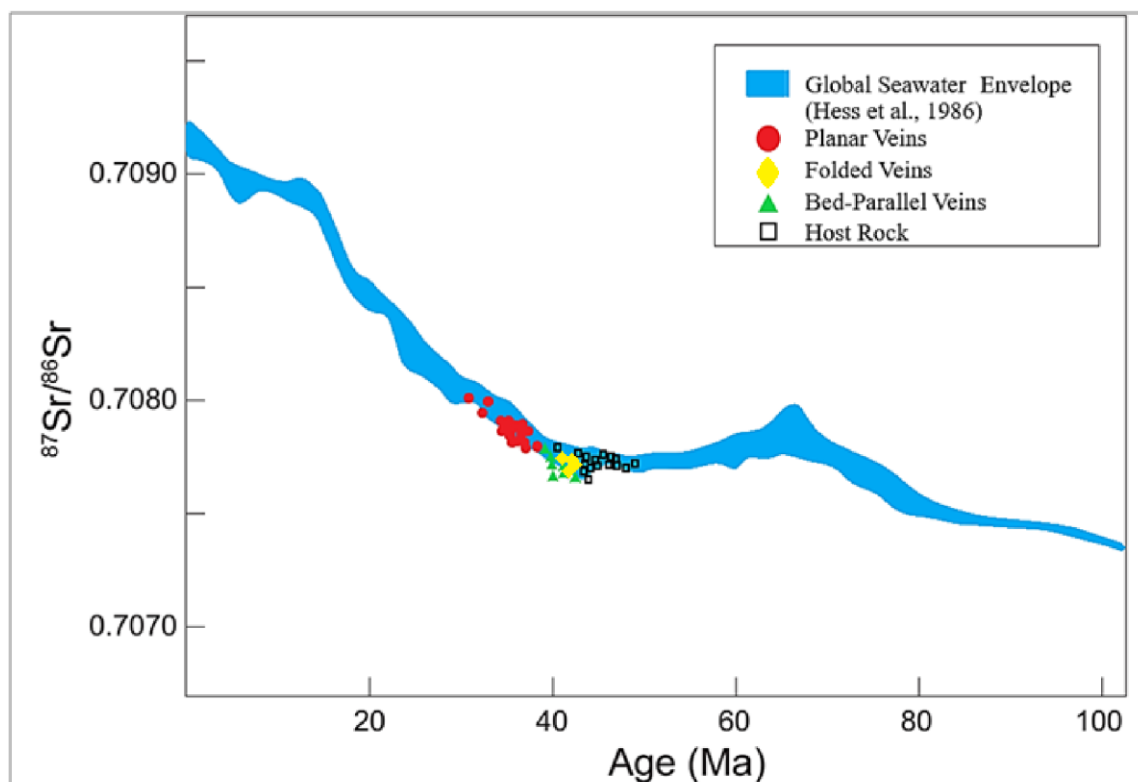


Figure 3.50 Plot of the $^{87}\text{Sr}/^{86}\text{Sr}$ data of the sample analysed on the global seawater envelope (modified from Hess et al., 1986). The age of the calcite samples was also correlated to the stable isotope and radiometric (U/Pb) data.

Sample Number	Sample Type	$^{238}\text{U}/^{206}\text{Pb}$	$^{208}\text{Pb}/^{206}\text{Pb}$
1	Planar Vein	130.704	1.113
2	Planar Vein	139.429	1.086
3	Planar Vein	137.336	1.034
4	Planar Vein	119.877	1.190
5	Planar Vein	113.485	1.277
6	Planar Vein	131.662	1.144
7	Planar Vein	117.319	1.304
8	Fault	307.100	0.145
9	Fault	325.562	0.085
10	Fault	324.271	0.061
11	Fault	321.509	0.072
12	Folded Vein	140.971	0.062
13	Folded Vein	139.161	0.070
14	Bed-Parallel Vein	79.958	1.000
15	Bed-Parallel Vein	93.235	1.360

Table 3.6 The data obtained from the U-Pb dating analysis

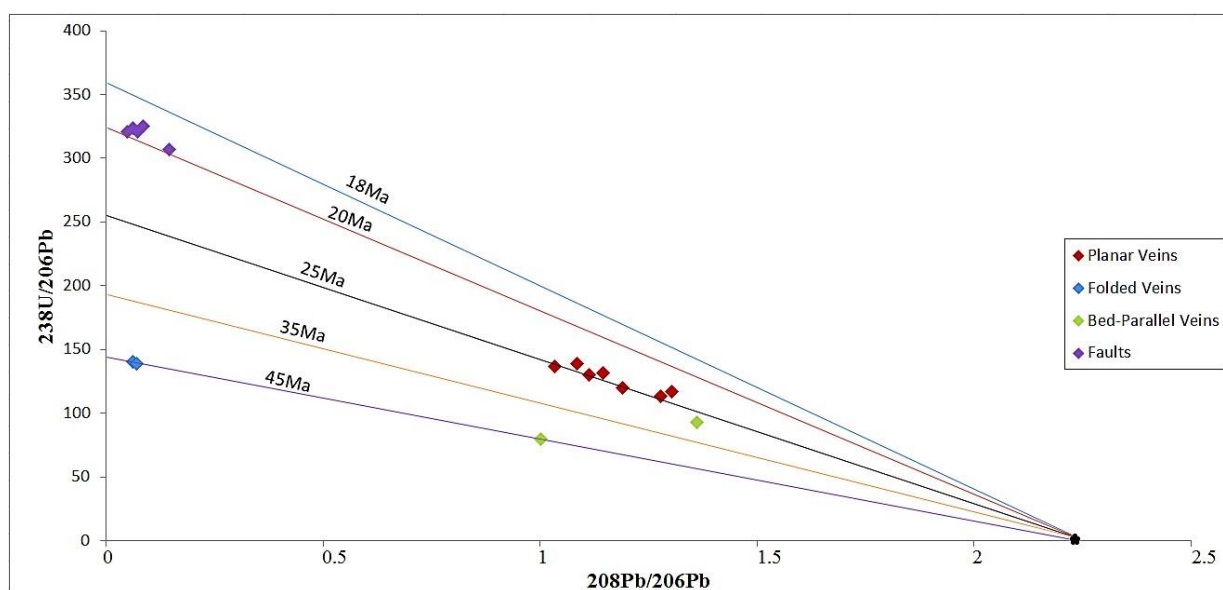


Figure 3.51 Plot of the $^{238}\text{U}/^{206}\text{Pb}$ against $^{208}\text{Pb}/^{206}\text{Pb}$ of the different calcite structures. Lines represent the age reference. Note the consistency in age between the two folded vein samples (blue) collected from two different cores (C4 and C5) from central Jordan.

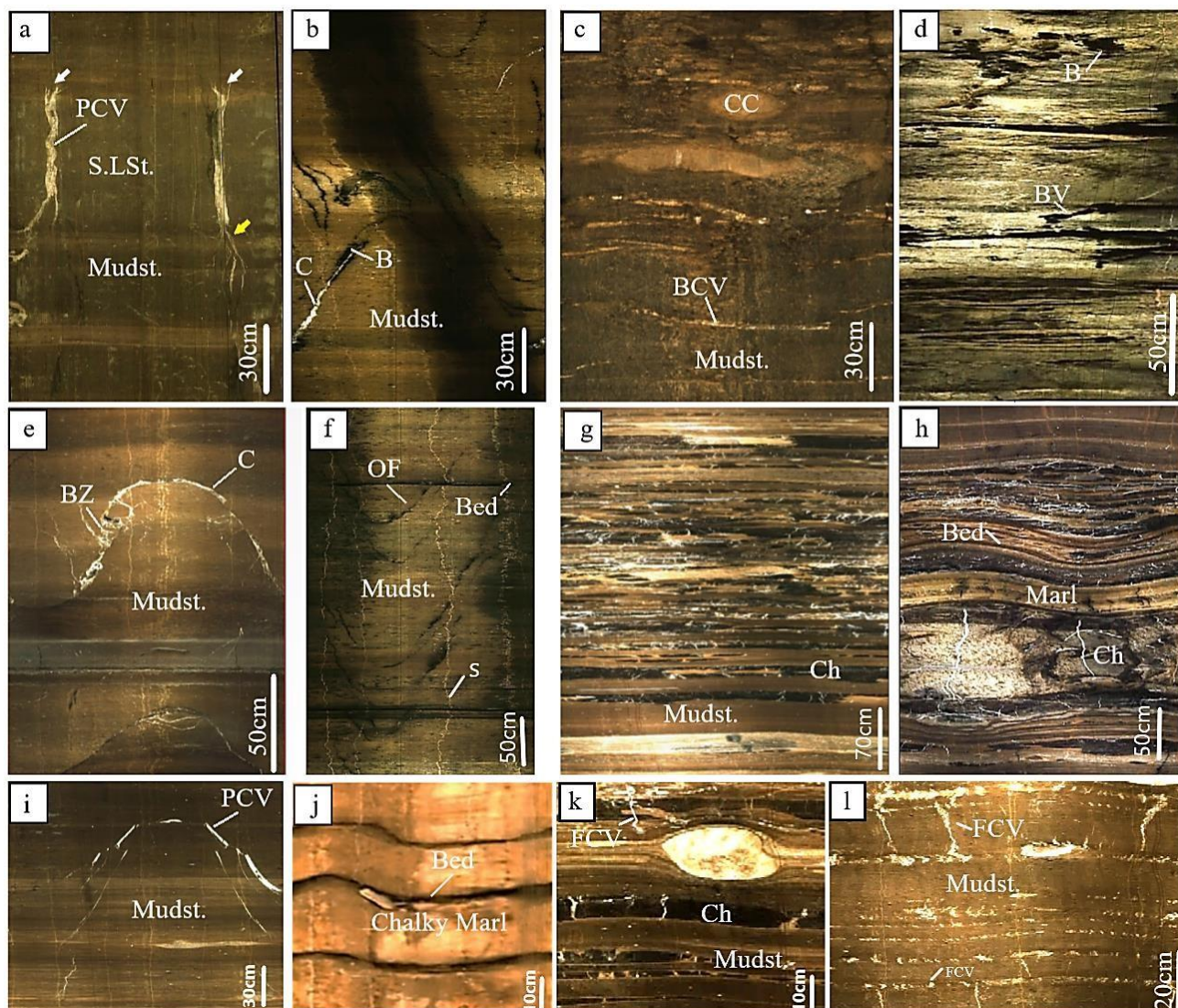


Figure 3.52 Borehole images of structures observed in the Jordan oil shale. PCV, Planar Calcite Vein; S.LSt., Silicified Limestone; C, Calcite; B, Bitumen; CC, Carbonate Concretion; BCV, Bed-Parallel Calcite Vein; BV, Bitumen Vein; BZ, Breccia Zone; OF, Open Fracture; Bed, Bedding Plane; S, Striation Ch, Chert; FCV, Folded Calcite Vein. a. Vertical calcite vein. Note the branching tip of the vein (white arrows) and the vein splaying and stepping (yellow arrow). b. Veins cemented by calcite and bitumen. c. Bedding-parallel calcite veins in mudstone. d. Bitumen in vugs and veins in marl interval. e. Dilatational breccia zone. f. Open fractures (joints) in mudstone. g and h. Bed-confined veins. i. Subvertical calcite vein in mudstone. j. Bedding planes in marl interval. k and l. Folded calcite veins in the Lower Unit.

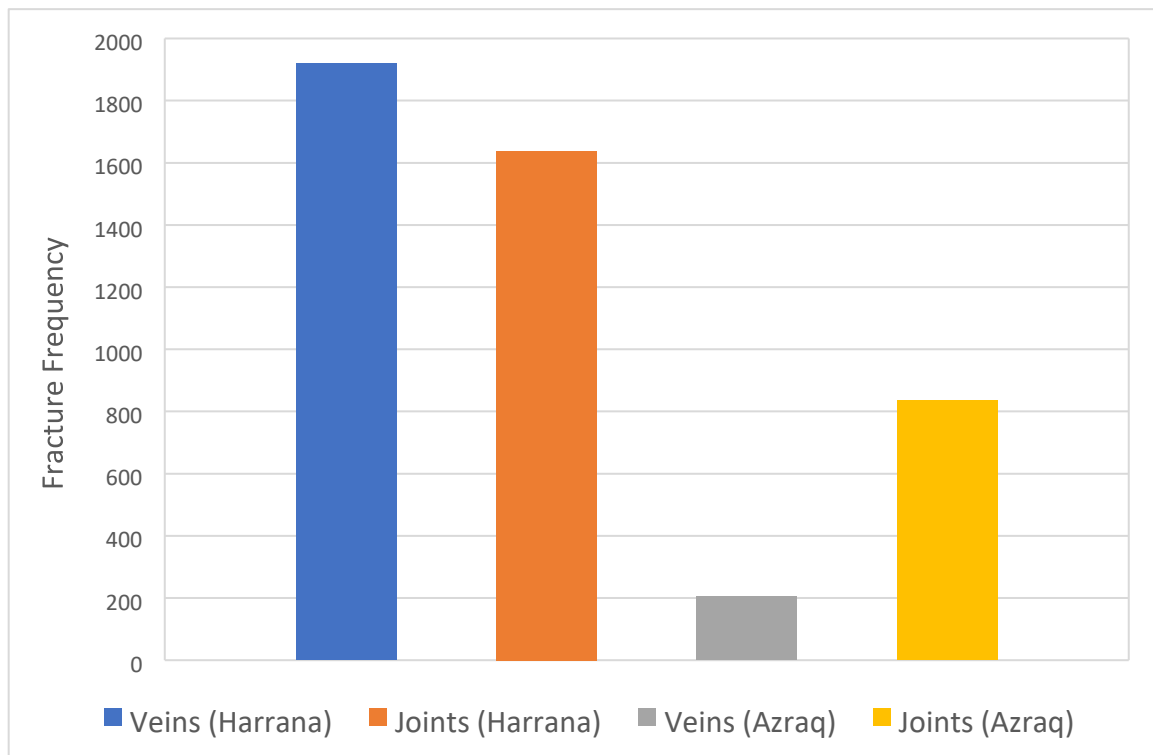


Figure 3.53 Totals of fractures interpreted from the BHIs of wells from both Harrana and Azraq area.

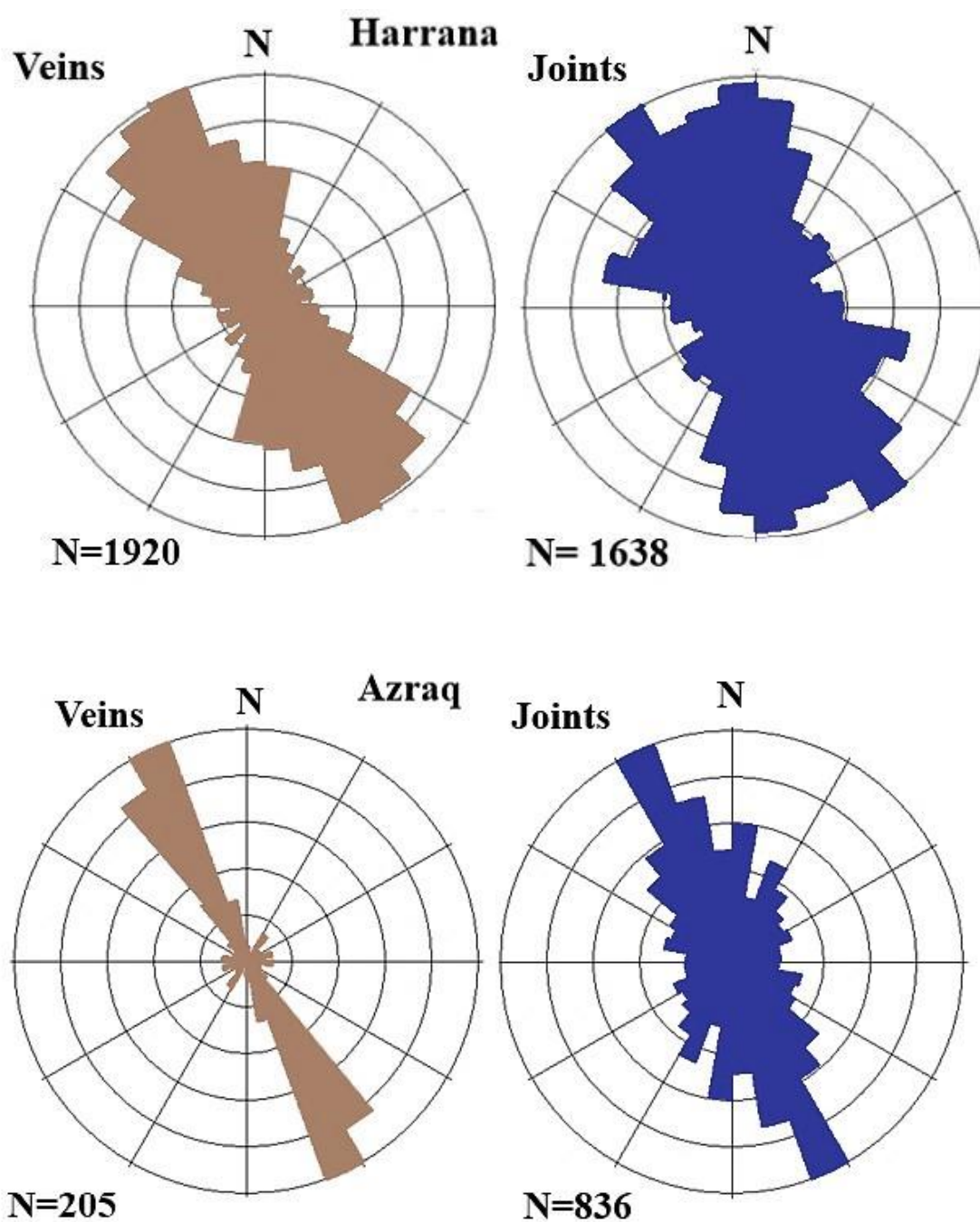


Figure 3.54 Rose diagrams for opening-mode fractures described from BHIs in Harrana and Azraq. The Harrana dataset is collected from 14 wells whereas the Azraq dataset comprises 7 wells. The locations of the wells are shown in Figure 2.8, Chapter 2.

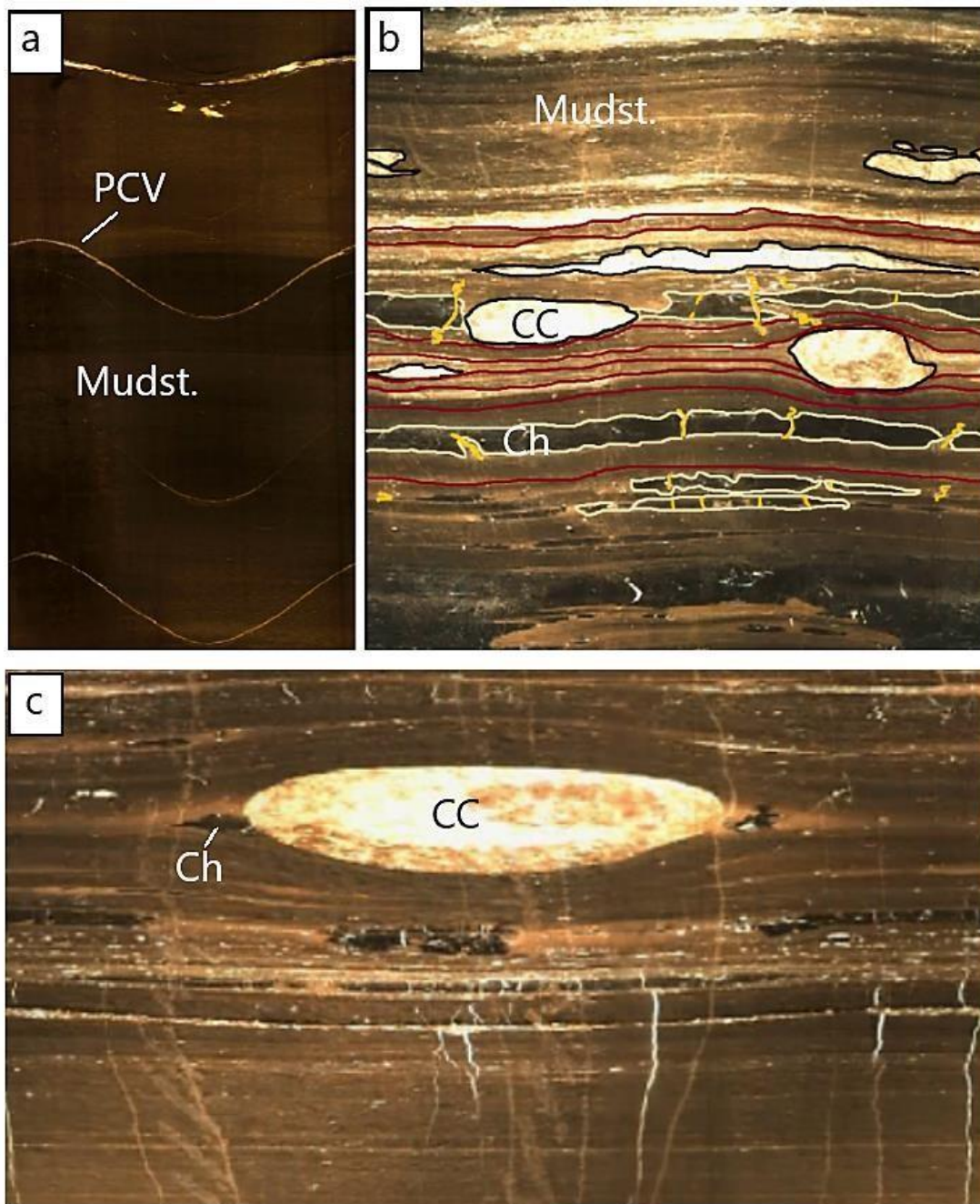


Figure 3.55 Example of fractures and concretions from the BHIs. PCV, Planar Calcite Vein; CC, Carbonate Concretion; Ch, Chert. a. Planar Calcite veins aligned regularly. b. Folded veins in non-parallel arrangement (Yellow lines) Note the different shapes and geometries of chert and carbonate concretions. Red lines represent the lamina boundaries. c. Elliptical carbonate concretion with two triangle spaces filled by chert on the edges. Field of view is equivalent to the borehole diameter.

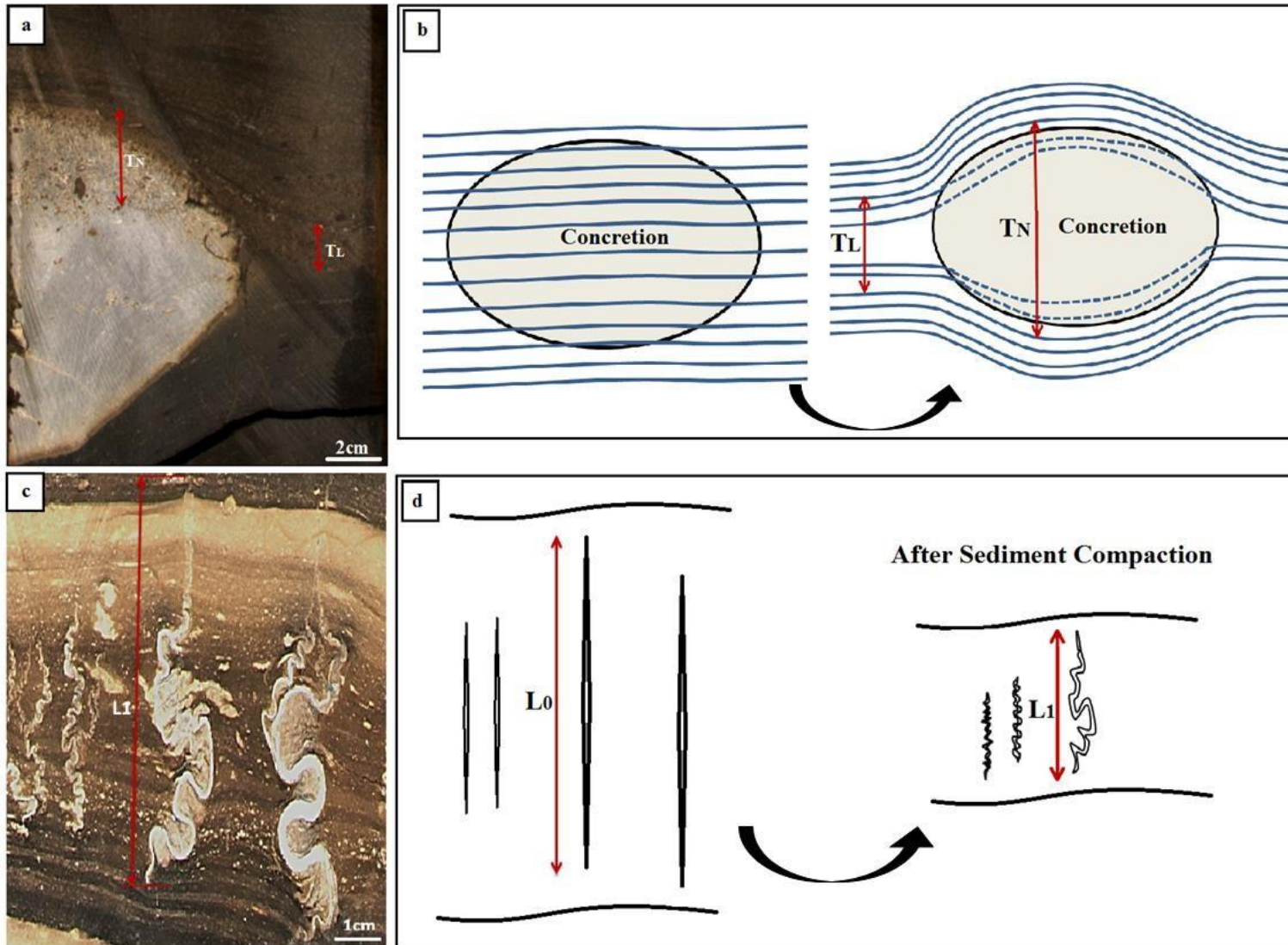


Figure 3.56 Quantification of compaction around concretions and folded calcite veins. T_N , Un-compacted bedding thickness (thickness between the beds immediately above and below the concretion); T_L , compacted bedding thickness (thickness between same horizons, lateral to the concretion); L_0 , meandering length of folded veins; L_1 , straight distance between tips of folded vein (height of compacted vein). a. Core Photograph of carbonate concretion showing differential compaction around concretion. b. Illustration of laminae before and after compaction. Compaction ratio is (T_N/T_L) . c. Core Photograph of folded calcite veins. Note that thinner veins are more folded than thicker folded veins. d. Illustration of vein before and after compaction. Compaction ratio is (L_0/L_1) .

Sample number	Location of the host core	L ₀ (cm)	L ₁ (cm)	Compaction ratio of folded veins (Cv)	Min Cv	Max Cv	Average Cv
1	Harrana, CW	13	7	1.86	1.25	2.25	1.74
2	Harrana, CW	9	4	2.25			
3	Harrana, CW	22	11	2.00			
4	Harrana, CW	20	10.5	1.90			
5	Harrana, CW	15	8.7	1.72			
6	Harrana, CW	17	12.5	1.36			
7	Harrana, CW	5.5	2.5	2.20			
8	Harrana, CW	5	3.2	1.56			
9	Harrana, CW	7	4	1.75			
10	Harrana, CW	15.2	9.2	1.65			
11	Harrana, CW	11.5	7.5	1.53			
12	Harrana, CW	7.5	4	1.88			
13	Harrana, CW	10	6.3	1.59			
14	Harrana, CW	5	4	1.25			
15	Harrana, CW	2	1.2	1.67			
16	Azraq, CE	1.5	1	1.50	1.25	2.57	1.87
17	Azraq, CE	1.3	0.6	2.17			
18	Azraq, CE	3	1.56	1.92			
19	Azraq, CE	1.8	0.7	2.57			
20	Azraq, CE	2	1.3	1.54			
21	Azraq, CE	2.5	2	1.25			
22	Azraq, CE	2	1.1	1.82			
23	Azraq, CE	7	3.65	1.92			
24	Azraq, CE	5	2.53	1.98			
25	Azraq, CE	5	3	1.67			
26	Azraq, CE	3.8	1.75	2.17			
27	Azraq, CE	9	5.1	1.76			
28	Azraq, CE	2	1	2.00			
29	Azraq, CE	3	1.6	1.88			
30	Azraq, CE	2.5	1.29	1.94			
31	Azraq, CE	2	1.1	1.82			
32	Azraq, CE	4	2.15	1.86			
33	Bayer, S	5	3.5	1.43	1.24	1.97	1.53
34	Bayer, S	9.5	6	1.58			
35	Bayer, S	2	1.5	1.33			
36	Bayer, S	7	4.2	1.67			
37	Bayer, S	5.6	3.6	1.56			
38	Bayer, S	6.1	4.1	1.49			
39	Bayer, S	2.5	1.7	1.47			
40	Bayer, S	7.5	3.8	1.97			
41	Bayer, S	2.1	1.7	1.24			
	All folded veins				1.24	2.57	1.75

Table 3.7 Compaction Ratio of folded calcite veins measured from the core

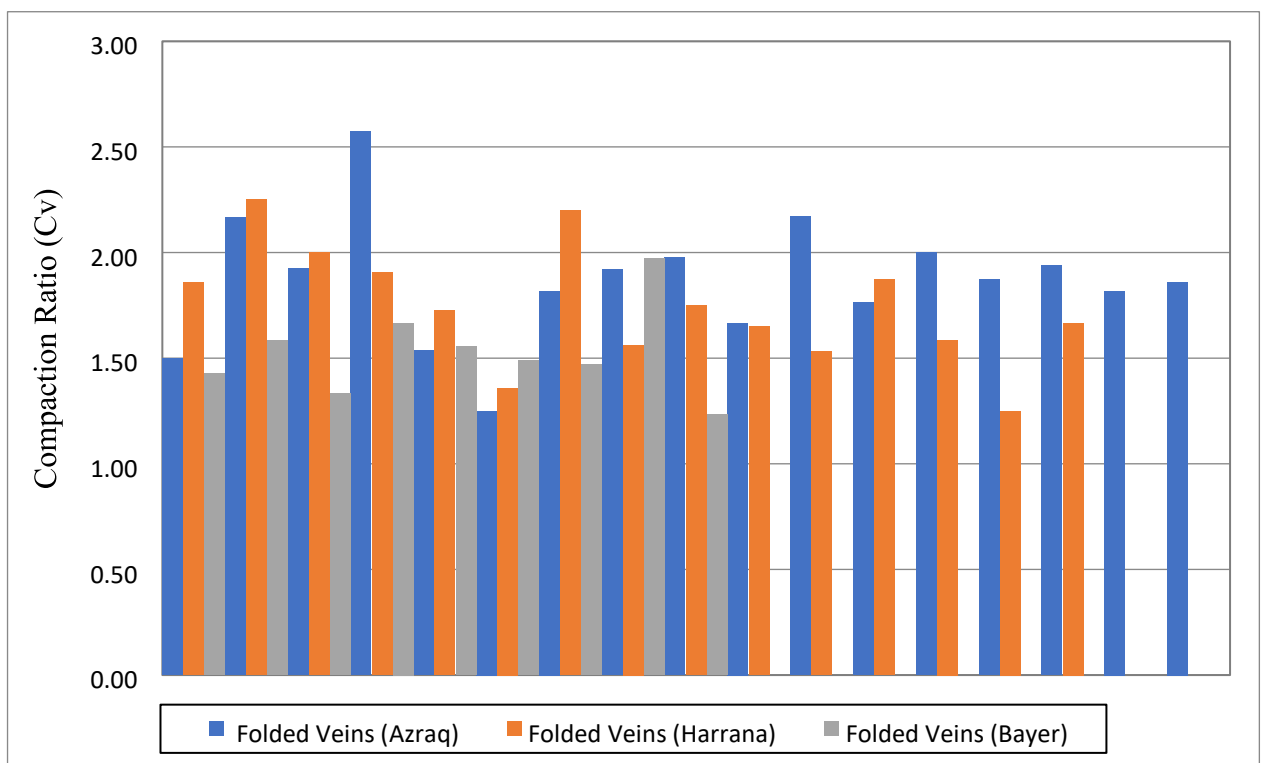


Figure 3.57 A histogram showing the compaction ratios for folded calcite veins measured from all cores. Note the interesting lower compaction ratio for the folded veins in the Bayer cores.

Chert Concretion				Carbonate Concretion			
Sample Number	T _N	T _L	Compaction Ratio (C _N)	Sample Number	T _N	T _L	Compaction Ratio (C _N)
1	5	3	1.67	1	20	5.5	3.64
2	3	2	1.5	2	12	7	1.71
3	3.5	2.5	1.4	3	20	3.2	6.25
4	2	1.5	1.33	4	15	4	3.75
5	5.5	2	2.75	5	25	3.8	6.58
6	6	3.5	1.71	6	14	2.5	5.6
7	1.8	1	1.8	7	15.5	5	3.1
8	4	2.5	1.6	8	14	4.5	3.11
9	4	1.5	2.67	9	20	7	2.86
10	15.5	5	3.1	10	19	5	3.8
11	11	3	3.67	11	22	7.5	2.93
12	14	6	2.33	12	20	6	3.33
13	5.5	2	2.75	13	13	5	2.6
14	4	2	2	14	12	3	2.67
15	12	5	2.4	15	5	1.5	3.33
				16	5	2	2.5
				17	4.5	2.5	1.8
				18	4	2	2
				19	2.5	1	1.74
Average	6.45	2.83	2.18	10.00	13.82	4.11	3.33

Table 3.8 Compaction measurements around chert and carbonate concretions in the cores

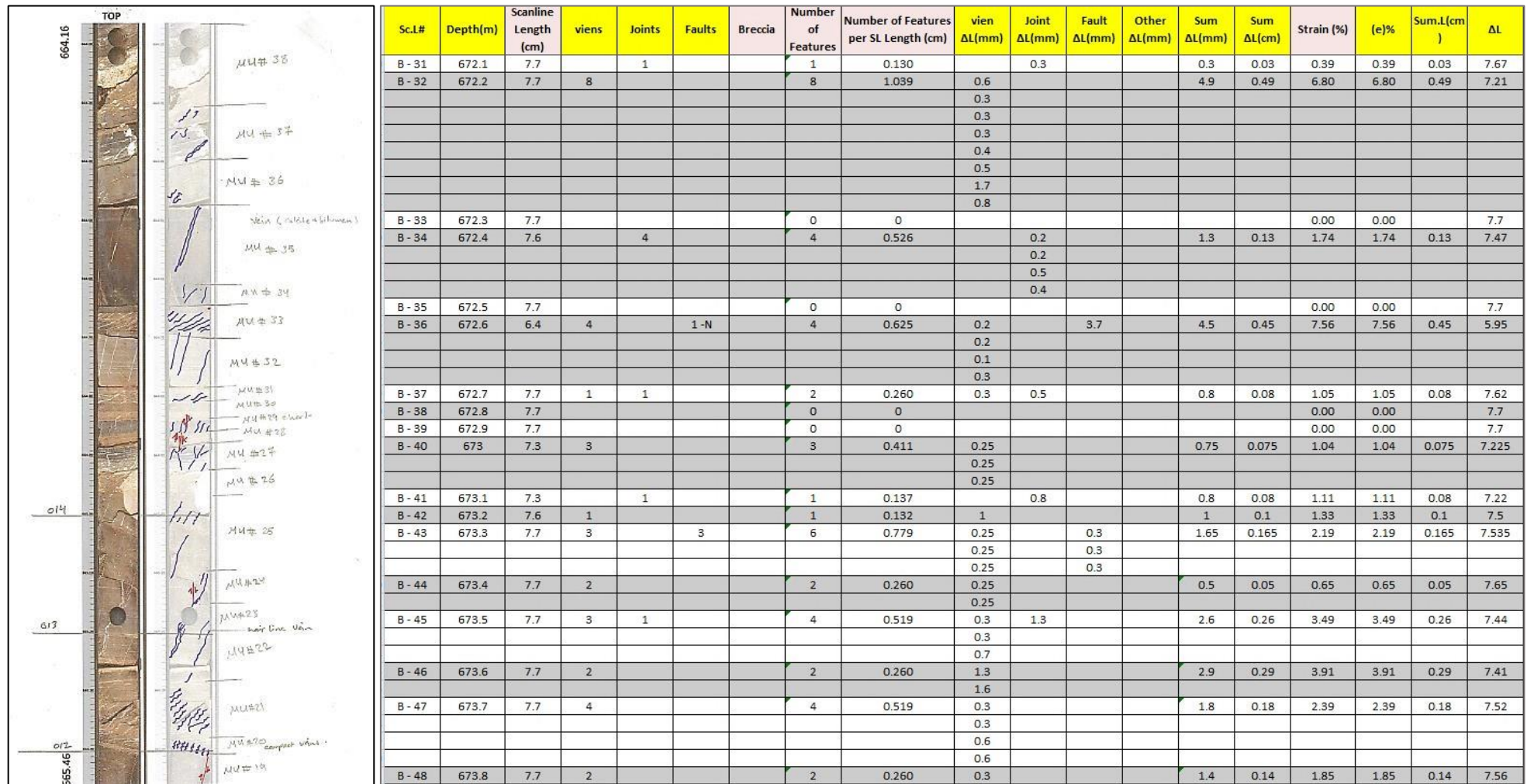


Figure 3.58 An example of the scanline measurements made from the cores.

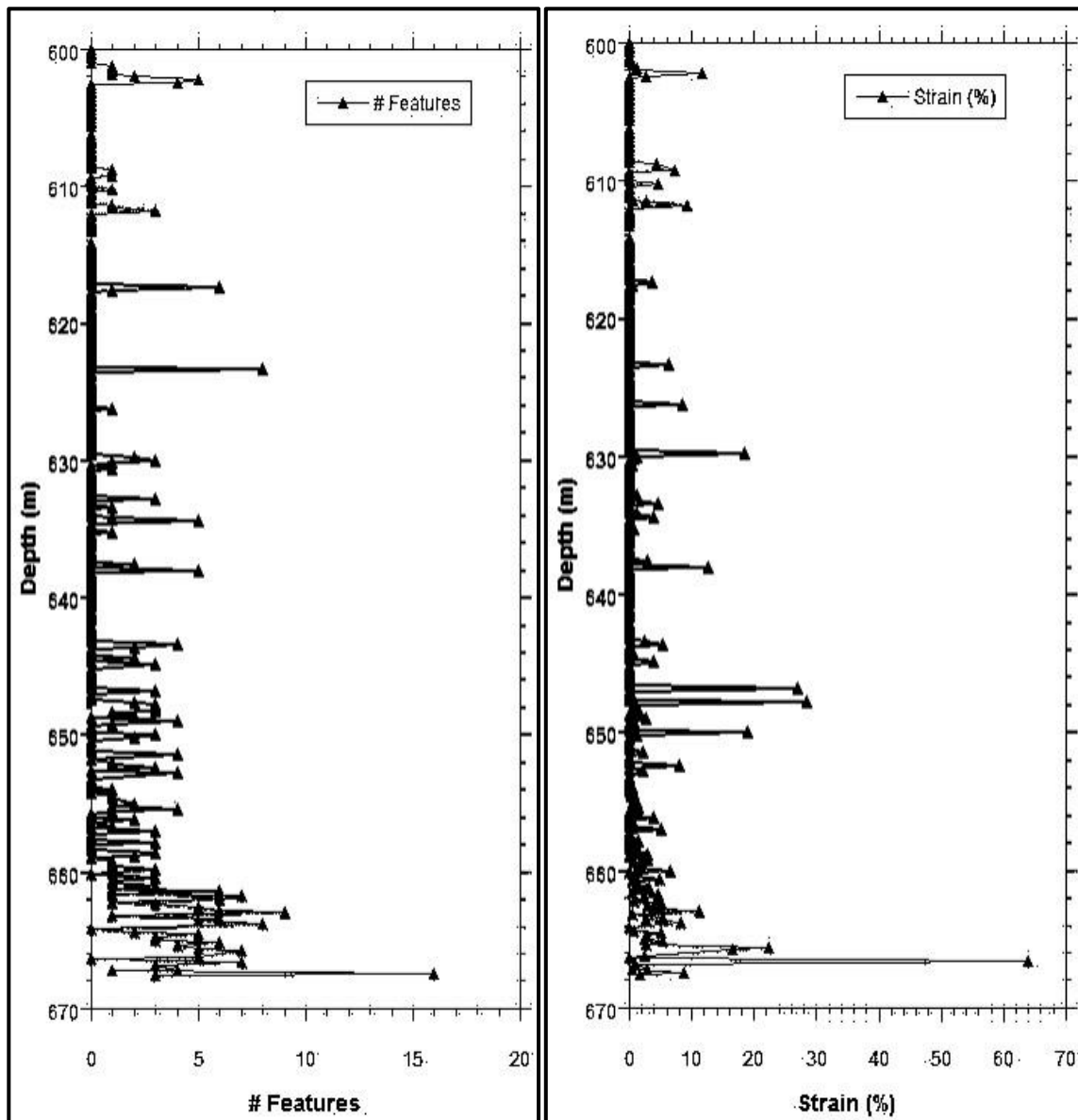


Figure 3.59 Number of features (fractures and breccias) and strain analysis for a fault hanging wall observed in the core from well C7.

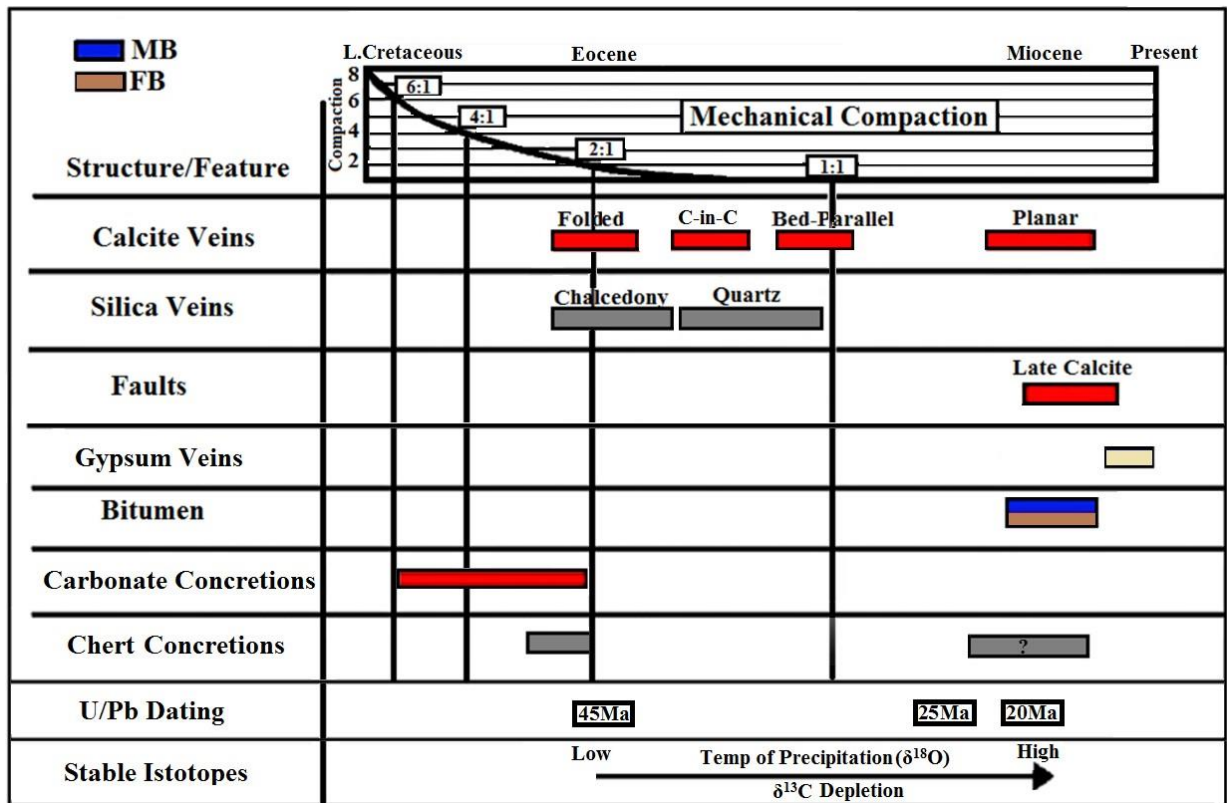


Figure 3.60 A paragenetic sequence of the different fracture cements observed in the cores. MB, Matrix Bitumen; FB, Fracture Bitumen. Compaction ratio curve is modified from Hooker et al., 2017a).

Chapter 4: Characterization of Outcrop Natural Fracture Systems and Constraining the Timing and Mechanisms of their Development in the JOS

4.1 Introduction

The study of outcrop analogues can be used to complement the analysis of subsurface fractures (Casini et al., 2016), particularly if the outcrops are selected carefully (Laubach et al., 2009), and therefore allows the construction of a complete structural framework for the rocks of interest. Different studies have combined subsurface data obtained from different sources such as core, borehole image and wireline logs and seismic reflections with data from outcrops for the sedimentary and structural characterisation of rocks (e.g., Levell et al., 1988, Ameen, 2014; Casini et al., 2016; Morris et al., 2016; Alfahmi, 2018; Lai et al., 2018). The differences in burial/uplift history, however, should be considered when undertaking surface to subsurface fracture correlations, because outcrop fractures may not be usually representative of the subsurface fracture. Carefully selected outcrops can provide important, 3D geological and structural information that can help obtain a better understanding of the different fracture characteristics and their spatial distributions (Bertotti et al., 2007).

The previous chapter described the subsurface fracture systems, their vertical distribution, timing, and genetic mechanisms in the Upper Cretaceous-Eocene strata of Jordan from analysis of core and borehole image logs. The present chapter provides a wider picture of the fracture spatial distribution and correlates the outcrop fractures to the subsurface fractures.

This chapter mainly aims to describe fractures from outcrops of the Upper Cretaceous-Eocene rocks, and to understand their propagation and spacing properties, that may be insufficiently represented in the cores due to the limited sampling or limited core width.

The chapter is organised in five major sections. The first section is devoted to introducing the methods and datasets used to characterise and measure fracture systems from the outcrop. The second section provides an overview of the geology of the study outcrops. The third section presents the findings of the analysis of fractures from the study of outcrops in three main subsections (1) the macroscopic description (field scale) of fracture types, fracture spacing, and fracture distribution and orientation, (2) the microscopic description of fracture properties and (3) the geochemical analysis of fracture cements and host rocks. The fourth section provides the results of the strain analysis around concretions. The fifth section discusses the main findings including the timings, distributions and formation mechanisms of the different fracture systems observed in the JOS outcrops.

4.2 Methods and Datasets

4.2.1 Methods of Fracture Description from Outcrop

Fieldwork was carried out in order to describe and measure the different natural fracture systems from outcrops, analyse their lateral distribution, understand their relationships with concretions, and analyse the strain field around concretions. Fractures were described, measured and sampled from five carefully-selected outcrops representing road cuts, quarries and trenches exposing the Upper Cretaceous-Eocene rocks that are equivalent to the cored intervals. The exposures studied are distributed in central and southern Jordan (see Figure 2.8; chapter 2). The first exposure is located in the Harrana area, 50 km east of Amman and very close to the cored wells drilled in this area (C1-C7), offering a good

opportunity to correlate the subsurface fractures to the outcrop fractures. Three outcrops are located in the Karak Governorate in central Jordan, about 110 km south of Amman. The fifth exposure is located in Wadi Al-Bustan in the Al-Shoubak area, south Jordan (Figure 2.8; Chapter 2).

Two of the studied outcrops are sections of freshly mined quarries. The quarry sections offer a 3D view that allows the analysis of the fracture geometries, orientations and horizontal distribution. Fracture characteristics, including fracture height, aperture size, spacing, and mineralogy, were described and measured from each study outcrop. Diluted hydrochloric acid (HCl) and scratch tests were used to distinguish the different cements filling the fractures. A hand lens (10X) was used to examine small fractures and fracture tips. Fracture spacing was measured using the scanline method, by measuring fracture spacing in any single bed along a horizontal scanline (e.g., Priest & Hudson, 1981; Priest 1993; Bons et al., 2004; Zeeb et al., 2013) (Figure 4.1).

4.2.2 Petrographic Analysis Methods

The different fractures observed in the outcrops, fracture cements, concretions and host rocks were sampled from all the studied localities. Fracture and host-rock samples were selected from freshly cut rocks. 4 cm-long core plugs were sampled from carbonate concretions from centre-to-edge, using a portable Makita drill machine, horizontally and vertically across the concretion, to systematically examine the composition and the internal structure of the concretion body.

Thin sections with a thickness of 30 μm were then prepared from the collected fragmented and plugged samples, to study the mineralogy, internal structure and microscopic properties of the different fracture sets, concretions and their host rocks. Samples were studied using cross-polarized light, using a Nikon Optophot Microscope linked to a high

resolution digital camera. Selected samples of fractures, concretions and host rocks were investigated under the SEM using a FEI Quanta 650 FEG scanning electron microscope. BSE images were created to study mineral compositions and textures of these samples. Elemental mapping was carried out using an EDS attached to the SEM. The SEM analysis was performed at a working distance of about 10 mm.

4.2.3 Geochemical Analysis Methods

Powdered samples were drilled from the samples collected from cemented fractures, and their host rocks to study the isotopic compositions of veins and host rocks. Samples of the previously sampled core plugs from the carbonate concretions (center-to-edge) and their host rocks were pulverized using a microdrill, to investigate the $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ stable isotopes. The samples were later run at the isotope laboratory of the Open University using a MAT 253 mass spectrometer. The methodology of the stable isotopic analysis is presented in detail in Chapter 3, Section 3.2.1.3.1.

X-ray diffraction (XRD) analysis was performed on selected vein samples in order to identify their compositions. Samples were drilled from vein cement using a microdrill. Powdered samples were then run using a PANalytical Empyrean diffractometer at the University of Oxford.

4.3 Locations and Geology of the Study Outcrops

This section provides an overview of the locations and geology of the study outcrops. A detailed geological and structural context of the study succession is provided in Chapter 2. The first outcrop, in the Harrana area (31° 51' 54" Latitude and 36° 37' 55" Longitude), marks the upper part of the study succession, representing a large portion of the URC and its lower boundary with the MCM (most of the Upper Unit) (Figure 4.2). The second exposure, the El-Lajjun Quarry, is located in the El-Lajjun area, central Jordan, midway

between the Karak City and the Qatrana intersection with the Desert Highway (Amman-Aqaba highway): 31° 12' 53" Latitude and 35° 57' 00" Longitude. The quarry is a part of an oil shale deposit that covers an area of approximately 20 km². Both the Lower Unit and the Middle Unit are exposed at the El-Lajjun area (Figure 4.2). The quarry section mainly shows the oil shale unit and the overlying chalky marl unit (Figure 4.3). Two horizons of carbonate concretions are observed in this quarry, with the thickness of the quarry section being about 40 m. These quarry rocks have been previously described in many studies (e.g., Shawabkeh, 1991; Abed et al., 2005; Hakimi et al., 2016, 2018). The main structural feature in the El-Lajjun area is the El-Lajjoun Graben, which controls a morphological depression surrounded from the East and West by N-S striking faults (NRA, 1990) (see Section 2.3 and Figure 2.3a, Chapter 2 for more information).

The third outcrop is represented by the Qatrana Roadcut, which is exposed in different places along the highway that links the Qatrana and the Karak cities (31° 13' 06" Latitude and 35° 59' 52" Longitude). Most of the lower portion of the study succession is exposed along this roadcut (Figure 4.2). These strata mark the Qatrana Phosphorite Member (the upper member of the AHP formation) and the lowermost part of the MCM (Figure 4.2). The roadcut varies in thickness between 30 and 40 m along a distance of about 20 km. The rocks are mainly composed of phosphatic marl interbedded with phosphatic chert and thin oil shale beds. Three main concretion horizons are observed in this roadcut section (Figure 4.2).

The fourth locality is the Sultani open-pit Quarry (see Figure 2.8, Chapter 2). It is located in the Sultani area, about 115 km south of Amman, and 10 km southeast of the third outcrop (Qatrana Roadcut): 31° 06' 37" Latitude and 35° 00' 43" Longitude. The quarry extends over an area of 0.25 km². The total thickness of the studied section is

approximately 45 m. This quarry exposes the strata of the Middle Unit (Figure 4.2). The quarry is a part of an oil shale deposit confined to a NW-SE trending graben structure, with an area of 32 km².

The fifth outcrop is located in Wadi Al-Bustan (30° 21' 46" Latitude and 35° 26' 52" Longitude), south Jordan (see Figure 2.8, Chapter 2 for the outcrop location). At this location, the entire Upper Cretaceous-Eocene succession including the three stratigraphic units (Lower, Middle and Upper) is exposed (Figure 4.2). The outcrop shows the full succession along an enormous valley (see Figure 2.16a, Chapter 2), offering an opportunity to study fracture distribution and orientation along a complete section, and then correlating them with the core fracture data. This site also afforded a 3D investigation of fracture sets and concretions on a massive platform (Figure 4.4).

At the area of central Jordan, including the Harrana, El-Lajjun, Qatrana, and Sultani outcrops, the structure is controlled by normal high-angle faults trending mostly NNW-SSE, and for the most part concave to the NE (NRA, 1985). Some of the faults display a reversal of throw along strike, and have intervening blocks of strata which, in consequence, tend in the present day to be tilted to the NNE with scarp slopes facing SSW. Block faulting along curved faults with opposite downthrows has led to the formation of the Sultani depression. Gentle folding is noticeable across the area about fold axes that generally trend NNW-SSE. Drag folds adjacent to the faults have led to local high dips (NRA, 1985). The faulted blocks are the principal morphological control.

4.4 Results of Outcrop Fracture Characterization

4.4.1 Natural Fractures Observed in Outcrop

4.4.1.1 Fracture Description and Mineralogy

Natural fractures, including both opening-mode and shear-mode fractures, are observed in all of the studied outcrops in different types and geometries (Figure 4.5). These include veins (mineralised fractures), joints (open fractures), faults, and breccia zones (Figure 4.5). Based on their morphology, two main types of veins are found. These are planar veins and folded veins (Figure 4.5a).

Similar to the core, fracture networks in the studied succession exhibit a wide variety of fracture fills. Filling materials observed are calcite, silica, gypsum or bitumen (Figure 4.5). Vein minerals are usually compatible with the lithologies of their host rocks. Calcite veins are usually observed to be present in limestone and chalky marl, whereas silica veins are commonly observed in silicified and chert beds and concretions (Figure 4.5). This is also a characteristic of the fracture fills of the fracture systems described from the cores in Chapter 3, and it suggests that the sealing of fracture networks has occurred in a chemically-closed system (Czerniakowski et al., 1984; Blyth et al., 2004; Hooker et al., 2017b).

4.4.1.1.1 *Folded Veins*

Folded veins are observed in outcrops from both central and south Jordan; in the Qatrana Roadcut and Wadi AL-Bustan (Figure 4.6a-d). The veins are observed in chalky marl beds that are adjacent to chert beds or concretions (Figure 4.6a-c). Folded veins are usually observed to be confined to their host beds (Figure 4.5a). The veins are highly folded in the marl and they continue straight when they transect chert or silicified intervals (Figure 4.6a-c and e-g).

Folded veins are observed as being filled only by calcite (Figures 4.5a and 4.6a-d). These veins show different geometries; they are present either vertical or inclined (Figure 4.6a-d). However, the vein tips usually align on a single plane (Figure 4.6a-d). Folded veins are present as either thick veins or as thin veins (Figure 4.6a-d). Thin folded veins exhibit more folding and pointed tips compared to thick veins (Figure 4.6a and e). Folded calcite veins are much shorter than other vein types observed in the outcrops, with a maximum present-day vein height less than 10 cm (Figures 4.5a and 4.6a-d).

As interpreted in Chapter 3, these veins were folded by sediment compaction that led to ~50% reduction in the vein height. Interestingly, folded veins are found in the Qatrana Roadcut in a marl bed that also hosts planar calcite veins parallel to these folded calcite veins (Figure 4.6c). This provides evidence from the field for the two different timings of development of these veins; planar veins postdate the development of the folded calcite veins. Such an observation was not documented from the cores; this highlights the importance of studying fractures from outcrop analogues.

4.4.1.1.2 Planar Veins

Planar veins are abundant in the outcrops and are found in all rock types, including mudstone, chert and carbonate rocks in the Harrana, Qatrana and Wadi Al-Bustan outcrops (Figures 4.5a-b and 4.6f). They are observed to be cemented by calcite or silica (Figure 4.5a-b and i.). Their aperture sizes vary from hairline veins to cm-scale veins (Figure 4.5a-b).

Planar calcite veins are observed in parallel arrangements (Figure 4.5b). They are recognised by their straight geometry, and they exhibit uniform aperture along the vein trace with a decrease towards the tips, particularly in thin veins (Figure 4.5a-b and c). Thick planar calcite veins show branching ends (Figure 4.6f and i). Thick veins exhibit a crack-seal texture (Figure 4.6f).

Planar calcite veins were observed in two outcrop localities and are mostly orthogonal to bedding (Figure 4.5a-b). They range from several centimetres to several metres in height. The veins are usually observed confined to beds (Figure 4.5b), although in the Qatrana Roadcut, most of these veins are observed to vertically transect different beds and show high vein heights (up to 10 m) (Figure 4.5a). This indicates that these veins have developed at a late stage.

Planar silica veins are usually present in chert and silicified intervals (Figures 4.5i and 4.7). They are observed in different geometries; horizontal veins, vertical veins or a subvertical veins among all the studied localities. In the Qatrana Roadcut, silica veins are always orthogonal to bedding, even if they are present in chert beds that are bent around concretions (Figure 4.7). This suggests that silica veins have formed very early, prior to the chert lithification and formation of carbonate concretions. If they had formed after the bedded chert was lithified, they would be expected to concentrate in the areas close to the greatest curvature of the beds.

4.4.1.1.3 Bedding-Parallel Veins

4.4.1.1.3.1 Bedding-Parallel Calcite Veins

Bedding-parallel calcite veins are observed in the Qatrana Roadcut and Wadi Al-Bustan localities. They are usually observed in chalky marl, marly limestone and shale along the bedding boundaries (Figure 4.8a-b). The veins have a uniform thickness (aperture size), and they are laterally continuous over tens of metres.

The veins are totally cemented by calcite, and exhibit three different types of fibrous texture; antitaxial, elongate-blocky and cone-in-cone (Figure 4.8). The calcite crystals show an antitaxial texture in the oil shale interval (i.e., in the Middle Unit) in Wadi Al-Bustan (Figure 4.8a-b). In this type of vein, the fibres are orthogonal to the vein walls

(Figure 4.8a-b). Elongate-blocky bedding-parallel veins were observed only in the Qatrana Roadcut (Figure 4.8c). Bedding-parallel calcite veins can be also found to be bent around carbonate concretions (Figure 4.8d-f). In such a case, the veins exhibit the third type of texture, which is the cone-in-cone (Figure 4.8 d-f).

4.4.1.1.3.2 Bedding-Parallel Gypsum Veins

Bedding-parallel gypsum veins are only observed in the Sultani Quarry in the organic-rich oil shale interval and the overlying massive bituminous marl (Figure 4.9). They extend along the bedding planes for >10 m.

In the oil shale interval, the veins are only found as fibrous bedding-parallel gypsum veins similar to those observed in the core (described in Chapter 3). These veins are composed of well-developed gypsum fibres and a narrow median line that extends along the vein centre (Figure 4.9d). The thickness of these veins ranges from a few millimetres to 2 cm. The fibres show antitaxial growth, in which the fibres grow from the median line outwards (towards the wall rock) following the opening routes (Durney & Ramsay, 1973; Hilgers & Urai, 2002a; Bons & Montenari, 2005; Bons et al., 2012). In such veins, the gypsum fibres are orthogonal to the vein walls, independently of the vein dip (Figure 4.9d-e).

Bedding-parallel gypsum veins are present in the bituminous marl unit as a massive network of bed-parallel gypsum veins, with inclined sigmoidal gypsum veins interlinking the bedding-parallel (Figure 4.9a-c). In this unit, the bedding-parallel veins are composite, and exhibit a flat morphology. Some of these veins display lenticular shapes that pinch out laterally, particularly if they are <1 m long (Figure 4.9a). They typically comprise two parts; a white powdery vein and an inner brown, fibrous vein (Figure 4.9b). The fibrous gypsum vein is usually contained within the white powdery vein, forming a median horizon in the white powdery veins (Figure 4.9b). The white part exhibits a rough surface

appears as poorly-developed fibres that are seen as vertical striations (Figure 4.9b). The inner fibrous vein shows antitaxial growth from a median suture line (Figure 4.9e).

In the bituminous marl unit, gypsum veins vary in height from a few centimetres to a few metres (Figure 4.9a), and their thickness ranges from a few millimetres up to 30 cm. The oblique gypsum veins are composed of white powdery gypsum and a thin fibrous gypsum vein (Figure 4.9c). However, some of these veins lack the presence of the median fibrous vein. The oblique gypsum veins have small branches along the vein body, and these branches lack any median fibrous veins (Figure 4.9a and c). The absence of the median fibrous zone in the small branches indicates that these branches were formed later, possibly by redissolution and precipitation of gypsum (see Discussion, Section 4.8.1.3). The oblique sigmoidal gypsum veins are observed to merge into the thick bed-parallel veins (Figure 4.9a). Fibres are observed to be vertical in all fibrous veins regardless of the vein orientation (Figure 4.10). This may suggest that the veins have opened by fluid injection against vertical σ_3 (Rodrigues et. al., 2009).

Field observations of gypsum veins in the marl unit, also provides evidence of late deformation in many parts of the gypsum vein networks observed in the marl unit. This includes shearing, horizontal shortening, and minor faulting crossing both the fibrous and powdery veins (Figure 4.11).

‘Overstepping’ of the gypsum veins is also observed in the bituminous marl interval (Figure 4.11e-f). The term ‘overstepping fractures’ is used to define the linkage occurring between sub-parallel isolated fractures, where the tips are adjacent (Biddle & Christie-Blick 1985). The veins overstep by a bridging process between vein segments at the vein tips. At the first stage, isolated veins develop horizontally. Fractures then start to open and the spacing between their tips decreases. The tips propagate towards each other, leading to

the inclusion of host rock present around the tips in the newly forming vein. Finally, the bridge breaks down, and a single vein forms by the linkage of the two segments (Figure 4.10f).

4.4.1.1.4 Bitumen-Bearing Fractures

Bitumen-bearing fractures are only observed in the El-Lajjun Quarry as entirely bitumen-filled veins (Figure 4.5c). These bitumen veins are observed in an organic rich shale interval, and are usually observed to be confined to the bed boundaries (Figure 4.5c). They are present in vertical geometry, and they are always in parallel arrangements (Figure 4.5c). The veins exhibit mm-scale aperture sizes, and up to 15 cm in height. They show an average value of 5 cm for fracture spacing. No observations of any other fracture types were documented in the beds hosting bitumen veins. These veins are discussed in detail in the next chapter (Chapter 5).

4.4.1.1.5 Joints

Joints are the most common fracture type observed in the five study localities. They are typically present in the chert and silicified intervals, and are usually confined to beds (Figure 4.5h). They are normally highly clustered where they are associated with concretions or faults (Figure 4.5h). They can be found in the beds that are bending around concretions or in the damage zones of faults. They vary in their aperture sizes from mm- to cm-scale. These joints have fracture heights that range between 3 and 60 cm.

4.4.1.1.6 Shear Fractures

Faults and breccia zones are commonly observed in the outcrops (Figure 4.5h). The faults are mainly observed in the Upper Unit of the studied succession, which is exposed in the Harrana and Wadi Al-Bustan localities. They are occasionally found in the Lower Unit

exposed along the Qatrana Roadcut. No observations of faults were reported from the El-Lajjun and Sultani quarries that expose the Middle Unit.

Faults are easily distinguished in the outcrop by the lamina offset, and they usually show normal displacement (Figure 4.5h). They are observed in different scales, ranging from cm- to m-scale. Faults are observed either as simple faults (Figure 4.12) or as complex faults showing fault core and two damage zones (Figure 4.5h). The fault core exhibits gouge, breccia and cataclasites that are usually cemented by calcite (Figure 4.5h). Late opening-mode fractures, including joints and planar calcite veins, consistent with the core observations, are present in association with faults (Figure 4.5h). They usually cluster in the fault damage zones. The outcrops, however, allowed a better understanding of the abundance and spatial distribution of opening-mode fractures within the fault damage zones. The opening-mode fractures show a general decrease in intensity with increasing distance outwards from the fault core (Figure 4.13).

4.4.2 Concretions and Concretion-associated Structures

This section will be devoted to describing the different concretion types observed in the outcrops and their associated structures.

4.4.2.1 Description of Concretions

Multiple styles of carbonate and chert concretions are observed in all outcrops investigated in this study. Concretions are usually observed to be stratigraphically confined. Each type of both the carbonate and chert concretions is present in different horizons; a single horizon has only one type of concretions (Figure 4.14a). Both the carbonate and chert concretions were seen in different lithologies, particularly in the lower and upper units of the study succession. At the Qatrana Roadcut, chert beds are observed to be bent around carbonate concretions (Figure 4.14b). In such cases, thin chert beds show higher amounts

of folding than thicker chert beds. For example, above and below the large carbonate concretions in Figure 4.14b the chert beds exhibit a contrasting degree of folding. Above the concretion the thick beds are not folded, whereas below, a set of the beds are folded to a horizon of small carbonate concretions below.

Both symmetrical and asymmetrical concretions are observed (Figure 4.15). Symmetrical concretions exhibit varied shapes. They can be found in spherical, elliptical, tabular or lenticular shapes (Figure 4.15). In the Sultani Quarry, only two main shapes of the carbonate concretions are present, spherical carbonate concretions and elliptical carbonate concretions. The other four localities show examples of all concretion shapes, and each concretion shape is in turn confined to a single stratigraphic horizon. This may suggest that the precipitation rate was different from one horizon to another (Huggett, 1994).

In the Qatrana Roadcut, tabular carbonate concretions are observed to preserve parts of chert laminae inside the concretion body (Figure 4.15c). This may suggest replacive growth of the concretion (see Discussion, Section 4.8.4 for more detail).

Interestingly, in both the Qatrana Roadcut and the Sultani Quarry, some carbonate concretions are observed to have a chert zone at the concretion edge (i.e., zoned carbonate concretions) (Figure 4.15d), which are also observed in a single horizon. This type of concretion can be observed in either spherical or elliptical shapes. Other concretions in the Qatrana Roadcut were found to have triangular shapes filled by chert at the edges (Figure 4.15e). This type of concretion is only found in lenticular shapes.

The concretion diameter ranges from 15 cm to more than 3 m for carbonate concretions and from 10 cm to 60 cm for chert concretions in all the studied localities (Figure 4.16). The concretion spacing ranges from several centimetres to several meters, with average spacing of 2.5 m for carbonate concretions and 20 cm for chert concretions in all localities.

Several remarkable examples of carbonate concretions, with the laminae preserved entirely, are seen in the Qatrana Roadcut (see Figure 4.8d). In such examples, the bedding can be easily traced through the concretion, but with an enlarged bed thickness compared to the surrounding beds. This indicates a replacive concretion growth (Selles-Martinez, 1996; Gaines & Vorhies, 2016).

Other concretions are observed to push the laminae apart in asymmetrical geometries, suggesting a displacive concretion growth (Figure 4.14b). Such concretions have no sedimentary structures preserved inside the concretion body, and they usually show asymmetrical bending of the beds over-and underlying the concretion (Figure 4.14b).

4.4.2.2 Concretion-associated Fractures

Different fracture types are observed to be associated with concretions in the outcrops (Figures 4.7, 4.8d, 4.14b, 4.17, 4.18, and 4.19). These fractures display different relationships to concretions. They are observed to be highly developed inside and immediately around carbonate and chert concretions.

In the Sultani Quarry, some carbonate concretions in the lower concretion horizon in the bituminous marl unit, immediately above the oil shale unit, exhibit a ‘septarian’ structure (Figure 4.17), with cemented fractures cutting across the concretion core. Such fractures usually tend to cluster in the concretion centre (Raiswell, 1971; Astin, 1986), and appear to taper towards the concretion margins (Figure 4.17). They can be present as a single fracture or as a network of interconnected fractures (Figure 4.17).

Fibrous bedding-parallel veins filled by calcite (in the Qatrana Roadcut) and others filled by gypsum (in the Sultani Quarry) were observed to bend around carbonate and chert concretions (Figures 4.8d and 4.18a-b). Other examples from the Sultani Quarry show bedding-parallel gypsum veins crosscutting carbonate concretions (Figure 4.18c-d). This

indicates a late development of these gypsum veins, postdating the concretion (see Discussion; Section 4.8). A critical observation regarding the relative timing question is that the gypsum fibres in the bedding-parallel gypsum veins at the Sultani Quarry exhibit orthogonal directions to the vein wall in the veins that are bent around concretions (Figure 4.18a-b).

In the Qatrana Roadcut, silica veins were observed in the chert beds that are folded around concretions (Figures 4.7 and 4.14). In this case, the veins were seen to be confined to the chert beds. They are consistently oriented orthogonally to the bed boundary, and they fan in dip with the folding of the chert beds around concretions, displaying a ‘radial’ vein orientation (Figures 4.7 and 4.14). The veins exhibited aperture sizes that are remarkably large in the areas bending around concretions, compared with those of the veins present in the other parts of the beds that preserve the horizontal alignment, and without any bending (Figure 4.7).

Vertical planar calcite veins were observed at the Qatrana Roadcut to crosscut carbonate concretions, as well as the host rocks above and below the concretion (Figure 4.7). This suggests that these veins postdate the growth of the concretions (see Discussion for more detail).

When open fractures (joints) are seen, they are usually confined to the beds that show a stiff character (i.e., silicified or chert beds) and host concretions (Figure 4.19). In such cases, the joints show a distribution that is consistent with the deformation of the folded laminae around the concretion.

4.4.3 Fracture Measurements

4.4.3.1 Fracture Geometry, Frequency and Distribution

The results of fracture frequencies, heights, aperture sizes, and spacing of all fractures collected from the five outcrops along the vertical and horizontal scanlines are presented in Table 4.1 and Figures 4.20-4.22. Fractures have a wide range of fracture heights. Both opening-mode fractures and faults show a range from a few centimetres to a few meters in height (Figures 4.5 and 4.20a). The highest fracture heights were measured from the Sultani Quarry, Qatrana Roadcut and Wadi Al-Bustan for the bedding-parallel veins (Figure 4.20a). In the Qatrana Roadcut, bedding-parallel calcite veins range from 1 to 5 cm in aperture sizes and from 1 to 5 m in Length, whereas planar calcite veins show aperture sizes range from few millimetres up to 3 cm and heights from 2 cm to 15m (Figure 4.20). Folded calcite veins have heights up to 10 cm.

Fractures are usually cemented in mudstones (Figure 4.5), whereas they can be found cemented and uncemented in chert/silicified units (Figures 4.7 and 4.19). Aperture sizes of fractures in the outcrop vary from mm- to cm-scale (Figure 4.20b). The largest aperture sizes of all vertical fractures are usually observed in chert and silicified units. In addition, aperture size is large for fractures that are bent around concretions in the places that have a greater curvature (see Figure 4.7) and for gypsum veins in the Sultani Quarry (up to 30 cm).

Fracture spacing is clearly observed to be controlled by the bed thickness in the outcrops; closely-spaced fractures are documented in thin beds (Figure 4.5b), whereas widely spaced fractures are observed in the thick beds (Figure 4.5a). The lateral extent of beds that the outcrops have effectively offered, have allowed good opportunities to take accurate measurements of fracture spacing over statistically valid distances along the bedding.

Fractures have high intensity in the chert and silicified beds (Figures 4.14b and 4.19). They usually show a higher intensity in the beds that are bending around concretions and in the fault zones. In the Qatrana Roadcut, vertical planar calcite veins crosscut the entire section, including carbonate and chert concretions (Figure 4.7). Folded calcite veins and silica veins are only seen to be confined to their host beds (Figures 4.5 a, d and 4.7).

Similar to that observed in the core, opening-mode fractures are the most common type observed in the outcrop (Table 4.1). Fracture distribution was examined vertically and horizontally in the studied outcrops. The vertical distribution of all fracture types is, to some extent, consistent with fracture distribution in the cores. All types of fractures are mostly clustered in the upper and the lower units that are composed primarily of chert and silicified units, whereas low fracturing is in general observed within the Middle Unit (Figure 4.22). However, the Middle Unit, representing the Sultani Quarry, has high bedding parallel fracturing (bedding-parallel gypsum veins) compared to that in the cores. Planar calcite veins are observed in the three stratigraphic units (Lower, Middle and Upper) of the studied succession, with a high cluster only in the lower and upper units (Figure 4.22). Folded veins are only present in the Lower Unit, as observed in the Qatrana Roadcut (Figure 4.22). Calcite bed-parallel veins are observed in the Qatrana Roadcut, representing the Lower Unit, while gypsum bed-parallel veins are only observed in the Sultani Quarry which marks the Middle Unit of the section (Figure 4.22). Horizontally, fractures have high horizontal assemblages within the chert beds, particularly if the beds are folded around concretions (Figures 4.7 and 4.19).

4.4.3.2 Fracture Orientation and Inclination

Opening-mode fractures vary in orientation from vertical and sub-vertical fractures (e.g., calcite veins, silica veins and joints) to horizontal veins (e.g., bedding-parallel calcite veins and bedding-parallel gypsum veins) (see Figure 4.5). Fracture strike data are shown in

Figure 4.23a. Fractures show orientations that range between 90° and 180° (Figure 4.23a). The results of fracture dips measured from all the study outcrops are displayed in Figure 4.23b. Fractures have a wide range of dip values. Most fractures in the Qatrana Roadcut have dip values range from 70° to 90° (Figure 4.23b). Fractures in the Sultani Quarry are mostly 0-10° in dip (Figure 4.23b).

In the Qatrana Roadcut, bedding-parallel calcite veins were observed to be crosscut by long, vertical planar calcite veins. Planar calcite veins were also observed to crosscut silica veins. This indicates that both the bedding-parallel calcite veins and silica veins predate planar calcite veins.

4.5 Microscopic Description of Outcrop Fractures and Concretions

This section presents a description of the microscopic properties of the different types of fractures, and the carbonate and chert concretions observed in the studied succession. The microscopic properties examined include mineralogy, internal structure, textural relationships, cross-cutting relationships between fractures, and relationships between fractures and concretions. It is worth mentioning that folded calcite veins observed in the outcrop are not described here because they are not easy to sample, due to the friable nature of the weathered host rocks.

4.5.1 Planar Calcite Veins

Vertical planar calcite veins show multiple-stage crack-seal texture, represented by multiple bands of calcite cement (Figure 4.24a-b). The calcite cement mainly shows syntaxial texture, with elongate-blocky crystals. Late sparry calcite crystals are seen within the vein cement (Figure 4.24a). Inclusions from the host rock are observed within the cement of planar calcite veins aligned along the boundaries of the crack-seal bands (Figure 4.24a-b). The veins exhibit a sharp contact with the host rock (Figure 4.24a-b).

4.5.2 Planar Silica Veins

Two types of silica veins are observed in the thin sections, fibrous chalcedony veins and microcrystalline quartz veins (Figure 3.24c-d). Both types exhibit fibrous textures (Figure 4.24c-d). Vertical and subvertical silica veins exhibit elongate-blocky texture, whereas horizontal silica veins display antitaxial texture (Figure 4.24c-d). Micro-silica veins are mostly planar. They are usually present as a complex network in silicified and chert beds or concretions.

Silica cement within the silica veins is entirely free of bitumen inclusions. Multiple examples of horizontal silica veins are documented crosscut by bitumen veins (Figure 4.24e). This observation is consistent to those documented from the cores and it indicates that the development of silica vein occurred at early stage of burial evolution, prior to the formation of the bitumen veins.

4.5.3 Bedding-Parallel Calcite Veins

Bedding-parallel veins are usually fibrous (Figure 4.24f). The veins display an elongate-blocky texture; calcite crystals show crystal growth from the vein wall towards the vein centre (Figure 4.24f). Host rock inclusions are observed within the vein, with a highest concentration to the vein walls (Figure 4.24f).

4.5.4 Bedding-Parallel Gypsum Veins

At the microscopic scale, the white powdery gypsum vein exhibit microcrystalline gypsum, with low negative relief (Figure 4.25). The median gypsum vein is composed of well-developed fibrous gypsum, with antitaxial texture (Figure 4.25c, f).

The microcrystalline gypsum reveals granoblastic gypsum (Figure 4.25e, h). This indicates secondary formation of these gypsum crystals. These gypsum microfacies are

characterized by the limited presence of anhydrite relics. The disordered texture may reflect fast crystal growth rates and supersaturation.

The contact between the fibrous vein and the surrounding white fabric is gradual (Figure 4.25). The fibres of the fibrous vein show an orthogonal direction to the vein walls, even while the vein is bending around concretions (Figure 4.25).

The bedding-parallel gypsum veins in the SEM show some gypsum fibres dispersed along the gradual contact between the fibrous vein and the white powdery vein (Figure 4.25b). This may suggest two possibilities; the white powdery vein is secondary gypsum that was presumably originally composed of fibrous gypsum or anhydrite before it transforms to micro crystals, or that the dispersed fibres present in the white powdery part have started to crystallise from the surrounding microcrystalline gypsum.

4.5.5 Bitumen-Bearing Fractures

Bitumen-bearing fractures are observed in the thin sections either in the form of bitumen veins, in which bitumen entirely fills the fracture (Figure 4.24e), or as inclusions in the planar calcite veins. Only vertical and subvertical bitumen veins are observed in the thin sections prepared from all outcrop localities. Bitumen filling the bitumen veins show an amorphous appearance. Bitumen veins are observed to crosscut horizontal silica veins that show no presence of bitumen inclusions (Figure 4.23e).

4.5.6 Chert Concretions

Chert concretions consist mainly of microcrystalline quartz (Figure 4.26). Radiolaria and foraminifera tests are rarely observed inside the concretions. When they are present, they are usually observed to be cemented by chalcedony (Figure 3.26). Calcite is hardly observed in these concretions.

4.5.7 Carbonate Concretions

Carbonate concretions are composed of >90 % calcite in all concretions analysed. They are mainly of micritic nature (Figure 4.27). Apatite fills the cavities of microfossils. The calcite microcrystals are distributed in a very fine-grained matrix (Figures 4.27 and 4.28).

In most of the samples collected from concretion that show symmetrical strain field around the concretion (see next section), the laminae can be traced as alternating dark and light bands or through the aligned grains (Figure 4.27d). Systematic microscopic examination of carbonate concretions, from centre-to-edge (Figure 4.27), shows that the crystal size, porosity and density of cement change with distance. The size of calcite crystals increases towards the concretion margins. The amount of cement in the samples from the centres of the concretions is usually more than those from the samples of the concretion margins (Figure 4.27). Microfossils and bone remains are observed to be preserved in relatively undeformed shapes inside the concretions (Figure 4.27a, c).

In the concretions that show asymmetrical shapes associated with asymmetrical strain field, the concretions show no internal structures or laminae (Figure 4.28). The calcite crystal in these concretions are of fibrous origin; the crystals are in the form of micro-spars (Figure 4.28a-b) like in the previous case (mentioned above). They are present in unaligned pattern. This may suggest displacive growth (see Discussion, Section 4.8.4)

4.6 Geochemistry

4.6.1 Stable Isotopic Analysis of Carbonate Concretions

The stable isotopes data of carbonate concretions and calcite veins are shown in Table 4.1. The isotopic ratios of carbonate concretions show a relatively narrow range of samples analysed from the centre to edge of each individual type of concretion (Table 4.1; Figure

4.29). However, the results show that the edges are more depleted in carbon than the centres (Figure 4.29). The carbon isotopic values of the centres of the concretions are closer to those of the host rock samples. The oxygen isotopic values range from -9.35 to 4.38‰ of all the analysed carbonate concretions. This suggests that the concretion formation commenced at shallow burial depth, but continued during progressive burial (Woo & Khim, 2006).

The carbon isotopic values of planar calcite veins transecting the carbonate concretions in the Qatrana Roadcut (range from -11.1 to -9.13 ‰) show that the veins are more depleted in carbon than the concretions and the host rocks (Figure 4.29). This suggests that the planar calcite veins have developed at late stage (see Discussion).

4.6.2 XRD

XRD analysis was carried out on samples from bed-parallel veins collected from the Sultani Quarry, to identify the mineralogy of vein material. The results show that the cement of the white powdery veins is mainly composed of gypsum (Figure 4. 30). The central fibrous vein is also gypsum, but with minor quantities of calcite and silica detected in samples collected from the median line of the vein (Figure 4.30).

4.7 Strain Analysis around Concretions

Strain was analysed around both the carbonate and chert concretions observed in the outcrops. The complete exposure of the beds hosting concretions in the outcrops afforded a unique opportunity to measure the degree of deformation around concretions (Figure 4.31). This analysis was performed by measuring the strain field around concretions, which is represented by the thinning of beds around the concretion (Figure 4.31). The strain shadow disappears within a distance up to 2 m above and below the concretion. The compaction

degree of the beds above and below the concretions, the concretion strain field (Figure 4.31) was measured and calculated according to Hooker et al. (2017) as follows:

$$C_N = \frac{T_N}{T_L}$$

Where,

T_N : the original thickness of a stratigraphic package across the concretion vertically,

T_L : the compacted thickness of the same package lateral to the concretion.

The results of this analysis are presented in Table 4.3. They show that the compaction ratios of the chert concretions are significantly lower (average value 2.5) than those for the carbonate concretions (average value 3).

4.8 Discussion

Field work has principally facilitated compiling geographic and geological data, helped in the description of the fractures and examination of their abundance in the Upper Cretaceous-Eocene rocks. It has allowed the correlation of the outcrop fractures with the subsurface fractures (described in Chapter 3), and allowed an opportunity to understand their spatial distribution across the area of study. In addition, it has also offered the opportunity to sample fracture characteristics and understand their relationships to different diagenetic features such as concretions.

4.8.1 Development of Outcrop Fractures

4.8.1.1 Fracture Distribution

Both the vertical and lateral fracture distributions were measurable in the outcrops due to the availability of freshly cut and quarried rocks in the carefully selected exposures, including two quarries (e.g., Sultani and El-Lajjun quarries), road cuts (e.g., Qatrana Roadcut). Wadi Al-Bustan, in the South, has also allowed a full exposure of the study

succession that in turn, helped in examining the vertical distribution of fractures (Figure 4.22).

There is a strong correlation between subsurface and outcrop fractures based on their types, geometries, general intensities, and vertical distribution. Fractures are mainly clustered in the lower (the Qatrana Roadcut) and upper units (the Harrana outcrop) (Figure 4.22). However, outcrops have revealed that the Middle Unit has relatively higher fracture intensity than that reported from the core (Figure 4.22), with a wider fracture spacing that was not measurable from the cores due to the limited core width.

The distribution analysis reveals that fracturing is generally controlled by the mechanical properties of the host rocks. It is interpreted in a number of studies that the difference in the mechanical rock properties (e.g., contrast in elastic properties of rocks) can create anisotropy in fracture intensity (e.g., Bourne, 2003; Laubach et al., 2009). This is clearly shown by the observation that most of the fractures (mainly opening-mode fractures) observed in the individual study exposures, are seen to be confined to their host beds (Figure 4.5a-c, h). In addition, the clustering of fractures in the rocks that show a high stiffness property (e.g., chert and silicified intervals) indicates that fracture development has been also affected by the physical and mechanical properties of the rock.

The abundance of the different types of opening-mode fractures in three localities (the Qatrana and Harrana exposures and the Lower Unit in Wadi Al-Bustan), in the chert intervals and in the beds that host chert and carbonate concretions (see Figures 4.14b, 4.19 and 4.22), reflects that both silica and carbonate diagenesis have controlled fracturing. This is clear at least, in the parts of the sections that show a high proportion of chert beds and other diagenetic features (e.g., concretions).

Bed thickness is also considered as an important factor that controls fracture spacing (Ladeira & Price, 1981; Pollard & Aydin, 1988; Wu & Pollard, 1995; Bai et al., 2000; Schöpfer et al., 2011). The large lateral extent of the exposure of the rocks of interest in the studied outcrops reveals that fractures are typically closely spaced in thin beds, whereas they have higher spacing values in thick beds. The positive linear relationship between fracture spacing and bed thickness (Figure 4.21) for both planar veins and joints reflects the control of mechanical properties of rocks on fracture distribution (Bai et al., 2000) and an isotropic stress field, indicating tectonically-imposed fracturing. Spacing of folded calcite veins, in contrast, have no clear relationship with bed thickness (Figure 4.21), indicating that these veins have developed by a mechanism and in a stress field that are different from that for other planar fractures (planar veins and joints).

The final Discussion Chapter (Chapter 6) discusses all the possible controls on fracture distribution and shows the differences and similarities in fracture distribution between the subsurface fracture systems and the outcrop fracture systems examined in this study.

4.8.1.2 Fracture Timing

The timing of the different types of outcrop fractures was constrained through crosscutting relationships, relationships between fractures and host beds and concretions, petrographic examination, and isotope geochemistry. These different techniques have been used in different studies separately, or in combination, to constrain the timing of fracture development (e.g., Uysal et al., 2011; Kolodny et al., 2014; Beaudoin et al., 2015; Hansman et al., 2018; Hooker et al., 2018).

The transecting geometry of the thorough-going, planar calcite veins of the beds that host bed-confined, folded calcite veins in the Qatrana Roadcut (Figures 4.5 and 4.6c, g) strongly supports the early timing of the folded calcite veins and a late formation of the

planar calcite veins (see Chapter 3, Section 3.6.2 for detailed discussion on the timing of folded and planar calcite veins).

The presence of the planar calcite veins together with folded calcite veins in the same bed indicates that planar veins developed at a later stage of sediment compaction. This is consistent with the analysis of the timing of development of the subsurface planar calcite veins observed in the core. This specific observation was not documented from the core alone, highlighting the significance of employing the outcrop analogues to complement the study of the distribution of subsurface fracture systems.

Another piece of evidence on the late development of the through-going, planar calcite veins comes from the crosscutting relationship observed between planar calcite veins the beds and concretions that host both the bedding-parallel calcite veins and the silica veins in the Qatrana Roadcut (Figure 4.7), which indicates that both the bedding-parallel calcite veins and silica veins predate planar calcite veins.

The occurrence of silica veins in the chert beds that are folded around concretions (Figures 4.7 and 4.14), with consistently orthogonal orientations to the bed boundary, fanning in dip with the folding of the chert beds around concretions, and that have aperture sizes that are remarkably large in the areas bending around concretions (Figure 4.7), indicates that silica veins have developed at very early stage, prior to the lithification of chert bed. If the silica veins had formed after the lithification of chert beds, they would be expected to concentrate only in the areas close to the greatest curvature of the beds but not in all parts of the chert beds (both the folded and non-folded chert beds).

At the microscale, the presence of bitumen inclusions within the crack-seal texture of the planar calcite veins suggests that bitumen was present during the formation (during the opening and sealing) of these veins. The crosscutting relationships between silica veins

and bitumen veins in samples collected from the Lower Unit (from the Qatrana Roadcut) (Figure 4.24e) strongly supports the interpretation that silica veins formed at early stage of basin evolution prior to the formation of bitumen veins.

Bedding-parallel gypsum veins were observed in the cores to be well developed in the oil shale interval (at present day depth ranges between 143 m and 156 m) and are interpreted to have formed early based on petrographic and geochemical analyses (see Chapter 3 for more information). Similar veins were also observed in the equivalent oil shale interval in the Sultani Quarry (Figure 4.9d), with internal structure (antitaxial) and relationships to host rocks that are similar to those observed in the cores. This indicates that fibrous bedding-parallel gypsum veins observed in the oil shale interval in the Sultani Quarry have also developed at the same time as those in the cores.

In contrast, the composite gypsum veins in the overlying marl unit in the Sultani Quarry at least the white powdery gypsum veins (Figure 4.9a), suggest late formation of veins that may have resulted by late diagenetic alteration (see next section for more information on the formation mechanism). The presence of these late veins limited to 10 m thick beds suggests that the veins have developed at late stage of basin history in this specific interval.

The clustering of joints (unsealed fractures) in the deformed beds around carbonate concretions and in the fault damage zones implies that this type of fracture has formed at different stages during the burial and uplift history (Laubach, 1988; Engelder, 1993).

4.8.1.3 Fracture Formation Mechanisms

The different genetic mechanisms of fracture formation of the folded and planar calcite veins, the silica veins, and the bedding parallel calcite veins are discussed in Chapter 3, Section 3.6.3. To avoid repetition, the discussion in this section is devoted towards the

formation mechanism of bedding-parallel gypsum veins that have not been discussed in detail in Chapter 3 due to their limited abundance in the core. The other fracture types observed in the outcrops (silica veins, planar calcite veins and septarian fractures) are also discussed in relation to the concretions in the next section. The formation mechanisms of the different structures observed in the studied succession in both the subsurface (core and image logs) and the outcrops are compiled together and discussed in the final Discussion Chapter, Chapter 6.

Bedding-parallel fibrous veins are very common in sedimentary basins (Cobbold et al., 2013), especially within rock units of low permeability. Fibrous gypsum is the most common type of gypsum that usually precipitates within fractures forming veins (Raman & Ramdas, 1954; Machel, 1985; Warren, 2006; Rustichelli et al., 2016; Meng, 2017). Fibrous veins, in general, are considered as an important tool to understand the deformation history of their host rocks (Ramsay, 1980).

To date, most of the studies on fibrous veins deal with syntaxial veins (Hilgers & Urai, 2002b; Hilgers et al., 2004; Nüchter & Stöckhert, 2007). Less attention has been paid towards understanding antitaxial veins (Bons et al., 2012; Meng, 2017). The fibrous bedding-parallel gypsum veins observed in both the oil shale and the overlying marl units in the Sultani Quarry, exhibit an antitaxial texture (Figures 4.9d-e and 4.25c, f). This suggests that the veins developed under dilation during gypsum crystallization (Bons et al., 2012; Gasparrini et al., 2013). The gypsum in the veins is interpreted to have been formed from sulphate and calcium ions derived locally from the sulphur-rich calcareous oil shale with the presence of near-surface groundwater.

The presence of gypsum veins in the oil shale unit is consistent with the very low permeability and fissility nature of the shale that resulted in the formation of the only set of

unidirectional veins (the bed-parallel fibrous veins), compared to the overlying marl unit that has a multi-directional gypsum vein set.

The main mechanism that controlled the formation of the gypsum veins in general in the JOS is argued here to be the fluid overpressuring in which the pore pressure of the fluids exceeds the vertical overburden stress:

$$pF > \sigma v + T$$

Where,

pF : is the fluid pressure; and

σv : is the vertical stress.

T : the tensile strength of the host rock

Overpressuring driven by fluid expulsion during diagenetic processes can generate bedding-parallel veins. Such processes have been widely discussed in the literature for different types of fluids; the smectite to illite transformation, and primary oil migration (e.g., Bruce, 1984; Freed & Peacor, 1989; Parnell & Carey, 1995; Katahara, 2006, Parnell et al., 2017a, b).

Another mechanism that is believed at least to have contributed to the formation of these fibrous gypsum veins is the force of crystallization. Force of crystallisation can lead or assist the formation of bedding-parallel fibrous veins in the rocks (Maliva & Siever, 1988). The shape of fibres and their vertical orientation (against gravity) suggests that force of crystallisation may have contributed to the development of the fibrous gypsum veins in both the oil shale and the marl intervals. The internal crystallization force could uplift the overburden if the supersaturation of rocks with fluids is high enough. The fibrous veins,

therefore, could expand and deform the host rocks, if the force of crystallization exceeded the sum of the overburden stress and the tensile strength of the JOS:

$$Pc \geq T + \rho gh$$

Where,

Pc: is the force of crystallization;

T: is the tensile strength of the rocks;

ρ : is the density of the rocks; and

h: is the maximum depth.

Force of crystallization was first argued by Weyl (1959) as a process contrary to pressure solution. The force of crystallization has recently been linked to explain the growth of antitaxial gypsum veins in Somerset, UK (Meng, 2017).

Gypsum is a typical weathering product of Cretaceous marine shales in many semiarid regions around the world (Hattin, 1965; Retallack, 1983). The white powdery gypsum veins that have been observed only in the outcrop (Sultani Quarry) in the marl unit are interpreted here to have formed as a product of a diagenetic alteration of previously formed gypsum. This is evident from the microcrystalline nature of gypsum within the white powdery vein (Figure 4.25b, e, h), which suggests transformation of gypsum/anhydrite fibres into microcrystalline gypsum crystals (Testa & Lugli, 2000). The white powdery veins indicate that the veins may have formed by direct, precipitation of gypsum during the vein development or by fast recrystallization of gypsum fibres, from the median fibrous vein, as microcrystalline gypsum (Figure 4.25). The latter is supported by the presence of discrete gypsum fibres or fibre relics in the microcrystalline part (Figure 4.25b, e). In both possibilities gypsum could have been sourced from local evaporitic fluids containing sulphur that is abundant in the host sulphur-rich mudstone rocks.

The overstepping or bridging between bedding-parallel gypsum veins (Figure 4.11e-f) indicates reactivation of the veins that occurred after the formation of the veins (Christie-Blick & Biddle, 1985).

4.8.4 Development of Concretions and their Relationships with Fractures

The vast majority of concretions are documented in mudrocks (e.g., Dix & Mullins, 1987; Astin & Scotchman, 1988; Calver & Baillie, 1990). The occurrence of carbonate and chert concretions in mudrocks is one of the most clearly diagenetic characteristics of the hostrock formation (Astin & Scotchman, 1988) that offer information about the paleo-biogeochemical development in sediments (Raiswell, 1971, 1976, 1988; Huggett, 1994; Selles-Martinez, 1996; Hesse, 2011; Dale et al., 2014; Gaines & Vorhies, 2016). The variety in the concretion-forming minerals from carbonate, quartz, to a zonation of carbonate and silica (Raiswell, 1971; Nelson & Lawrence, 1984) reflects a range of diagenetic environments of concretion formation, which in turn, is controlled by the composition of host rock and the depositional environment (Woo & Khim, 2006). The nature of the concretion cement, texture and porosity can help to constrain the mode and depth of the concretion growth. For a concretion to form in sediment, the pore water has to be supersaturated with respect to the mineral that will start to precipitate. For example, in the case of the carbonate concretion formation, the water has to be supersaturated with Ca^{+2} or Mg^{+2} and CO_3^- ions.

Carbonate concretions, which are abundant in the outcrop sections, typically form in organic-rich mudstones, where a significant proportion of the bicarbonate needed for carbonate precipitation is provided from the decomposition of organic matter in the host sediments (Gaines & Vorhies, 2016). The size of the crystals forming the concretions can unravel the growth history with regards to the growth rate of concretions (Woo & Khim,

2006). In the case of the concretions observed in the JOS, the increase in the calcite crystal size shown by the petrographic examination of systematically sampled (centre-to-edge) carbonate concretions, reflects a decrease in the concretion growth rate.

Many studies have explained the different scenarios of concretion-growth mechanisms (e.g., Selles-Martinez, 1996; Gaines & Vorhies, 2016). Two main types have been reported in most of these studies; (1) replacive concretion growth and (2) displacive concretion growth. The high abundance of concretions in the outcrop sections allowed a unique opportunity to examine the development of the different types of concretions, and to raise questions about concretions abundance within the cored intervals.

Replacive concretions are usually recognised by the preserved sedimentary structures and laminae in the concretion body, which in the JOS concretions are clearly seen, for examples, in specific horizons in the Qatrana Roadcut (Figure 4.8d-f). In addition, the symmetrical shapes of these concretions in the study succession that resulted in symmetrical strain fields around concretions also indicate that the concretions have grown replacively. In the replacive growth, the cements are thought to replace the host material or passively fill the pore space (Selles-Martinez, 1996; Marshall & Pirrie, 2013).

Displacive concretions, in contrast, can be recognised by their massive nature with no preserved internal sedimentary structures or laminae. A major indicator of the displacive growth of concretions is the fracturing of their host rocks (Saigal & Walton, 1988). This type of concretion is believed to form under shallow burial conditions and against the overburden load, and can be developed by force of crystallization (Selles-Martinez, 1996; Dewers & Ortoleva, 1990). It can grow when oversaturation of parent brines is associated with un-cemented, highly porous hosts, which is only possible under shallow burial conditions or by early overpressuring of sealed beds. Displacive growth also requires the

shear strength of the host to be very low and a hydrostatic stress field within the host (Selles-Martinez, 1996).

Displacive concretions were reported in the study localities in many examples, particularly in the Qatrana Roadcut. The evidence on displacive growth of the JOS carbonate concretions lies in the following key observations documented from the field, some of which have not previously been discussed in the literature; (1) the absence of any internal structures and the non-uniform pattern of crystals inside the concretions (Figure 4.28a-c), (2) the preferred asymmetrical shapes of concretions that resulted in folding in the chert beds, particularly thin chert beds around concretions (see Figure 4.14), (3) the differences in the strain fields (asymmetrical strain field) around the concretions compared to those measured around replacive concretions that reflect a uniform differential compaction (Figure 4.31), and (4) the formation of the triangular spaces that are filled by chert at the concretion edges (Figure 4.15e).

The two new observations that have not been discussed before, namely: the formation of the triangular spaces at the concretion edges and the folding of the chert beds around carbonate concretions, are used here to explain the displacive growth of concretions. Displacive growth may occur when the host is not completely lithified and this can help in the relative dating of some types of concretion-associated fractures. This leads the question what has controlled this bending of chert beds around concretions?

From outcrop observations, it was inferred that the thickness of chert beds has played a key role in controlling the folding of chert beds around concretions (Figure 4.14b). Chert beds were fractured by the combined effect of regional and local stresses. Original fractures (silica veins) in the bent chert beds above and below concretions have also been bent, but remain orthogonal to the host-bed (Figure 4.7). New, local. New, local fractures

(joints) were generated in the chert beds as a result of the deformation (Figure 4.7). The chert beds seem to have been not completely lithified at the time of the concretion growth. Silica veins may also have developed during early diagenesis, presumably when the lithification of the chert beds commenced. The concretion has forcibly folded the fractured, semi-lithified chert beds upwards and downwards (with respect to the centre of the concretion) through its growth history. This forced folding should be distinguished from simple differential compaction processes by the extremely localised strain patterns. In other words, the folding of chert beds occurred during the formation of the displacive concretions. Consistently, the same type of fractures in other parts of the beds (not bent; not associated with concretions) at the same section, remained in their original vertical attitude (Figure 4.7). This interpretation corresponds to the idea that displacive concretionary bodies usually tend to make room for themselves by pushing away their host sediments (Selles-Martines, 1996).

The other new observation that is explained here in relation to the displacive growth of concretions is the formation of the triangular spaces around some carbonate concretions (Figure 4.15e). Such spaces are argued to form at the concretion edges during its displacive growth within the host sediment, taking the shape of triangles. After the formation of these triangular voids, the chert seems to have filled the spaces. This observation was not only documented in the outcrop, but also it has been seen in the core and in the borehole image logs in equivalent intervals (see Figure 6.1, Chapter 6).

Despite difficulties in identifying the mechanisms that control concretion growth, the growth model has not always been stated by previous authors (Raiswell & Fisher, 2000; Marshall & Pirrie, 2013), the timing of the concretion growth, can be relatively constrained by analysis of the degree of compaction in the concretion compared with the adjacent host rocks (Astin & Scotchman, 1988; Marshall & Pirrie, 2013) (Figure 4.32), or

by stable isotopes and the growth rate of crystals (Woo & Khim, 2006). Field observations and laboratory analyses can offer information on how, where and when individual concretions formed (Marshall & Pirrie, 2013). For instance, newly quarried sections afford a fresh surface that can help in tracing the sedimentary laminae and sedimentary structures through and around concretions and examine the different concretion characteristics that help in the relative dating of concretion formation (Figure 4.32).

In the study succession, both the replacive and displacive concretions seem to have formed prior to any significant compaction or at least continued to grow during burial and compaction (Figure 4.32a-b). This is evident from the enhanced compaction of the host mudstone leading to the bending of shale and marl laminae around concretions (Marshall & Pirrie, 2013). The bending of chert beds is, however, excluded from this interpretation. If the chert beds have been bent after lithification, they would show high degree of fracturing and deformation, which is not the case of the chert beds that are folded around concretions in the Qatrana Roadcut. The chert beds seem to have been folded during the concretion growth.

The presence of the non-deformed fossils and the preserved sedimentary structures inside some carbonate concretions (Figure 4.27) indicates that the concretions formed after the deposition of the sediment (Marshall & Pirrie, 2013). Therefore, concretions are always 'diagenetic' features. Diagenesis can occur immediately after deposition and continue to depths of several kilometres. The formation of carbonate concretions in the JOS have been mostly related in previous studies to very early diagenesis at shallow burial depths (Abed et al., 2005; Makhoulf & El-Haddad, 2006). In the present study, it is also concluded (based on the petrographic and isotopic data) that the concretions were formed very early at shallow burial depth. The complete shapes of fossils preserved within the concretions (Figure 4.27a-c) and the presence of septarian fractures inside the concretions (Figure

4.17) are strong indicators of early formation of concretions at shallow burial depths (Astin, 1986; Astin & Scotchman, 1988; Raiswell, 1971; Hesslebo & Palmer, 1992).

Septarian fractures have been reported in many localities such as those found Lias Group (Lower Jurassic) and Kimmeridge Clay Formation (Upper Jurassic) in Dorset, UK (e.g., Wolff et al., 1992; Hesslebo & Palmer, 1992; Kiriakoulakis et al., 2000). Hesslebo & Palmer (1992) have argued that septarian cracks formed within concretions probably at maximum burial depths of <30 m prior to exhumation. Some authors have characterized septarian fractures as tensile fractures that usually form during burial and compaction of the host sediments (Astin, 1986; Maltman, 1994; Selles-Martinez, 1996; Hounslow, 1997). Astin (1986) argued that the septarian fractures are tensile fractures that form by dehydration during diagenesis. Hounslow (1997) suggested that they may form over a wide depth range. However, he suggests that a condition for their formation is that the pore pressure is higher inside the concretion than in the host sediment, generating local overpressuring that led to fracturing. It was suggested that this may have occurred because early cementation reduced the permeability of the concretion, confining the fluids in the concretion. Selles-Martinez (1996) suggests that it is possible for septarian cracks to develop as a result of the increase of stress around a stiff concretion in plastic sediment, which is a case that overpressuring is not essential for septarian fracturing (Raiswell & Fisher, 2000). Wetzel (1992) argued that septarian fracture formation initiates from both plastic deformation and dewatering of the concretion body itself. All these models are consistent with the idea that the formation of concretions was followed by the development septarian fractures.

The folding of chert beds around concretions is another line of evidence on the early timing of displacive concretion growth. Although the deflection of bedding around the concretions is a primary feature associated with different types of concretions (Raiswell,

1971; Mozley, 1996; Raiswell & Fisher, 2000, Rodrigues et al., 2009), the chert beds were observed to show high and asymmetrical deflection above and below displacive concretions (Figures 4.14b, 4.19), whereas in the case of replacive concretions, they show slight symmetrical deflection and in many cases, keep the normal horizontal bedding above and below concretions (Figures 4.8d and 4.15b-c). The latter is usually common in the replacive concretions that show tabular or elliptical morphologies (Figures 4.8d and 4.15c-d). It appears that the thickness of the chert beds played a key role in controlling the degree of folding of chert beds around concretions. This is supported by the observation that thin chert beds are extremely tightly folded around concretions that show asymmetrical morphology (Figure 4.14b), while thicker chert beds are only slightly folded or in some cases, not affected at all (Figures 4.14b and 4.19). In Figure 4.14b, it is interpreted that the lower chert bed has been pushed down by the effect of the concretion growth; it is difficult to see how this geometry could have resulted from differential compaction above. The orientation of the fracture set (silica veins) in the thin chert bed that underlies the carbonate concretion implies that growth of the carbonate concretion postdates silica fracturing, and is coeval with folding of the chert beds. A horizon of small carbonate concretions in the lower part of the section (highlighted in black ellipsoids) seems to have also bent by the growth influence of the main concretion. This bending in the lower chert bed and the nodule horizon below the main concretion indicates that the growth mechanism of the carbonate concretion was displacive growth, principally in a downward direction. Additionally, the fracturing (jointing) in the lower boundary of the overlying thick chert bed indicates that the fracturing has occurred in this bed, possibly due to the progressive growth of the concretion when the chert started to become more brittle during shallow burial diagenesis.

The presence of silica veins in the chert beds that are folded around carbonate concretions in the Qatrana Roadcut reflects fracturing that occurred prior to both the concretion growth and the folding of chert beds. This is evident from the orthogonal direction of these veins to the chert beds (Figure 4.7). The progressive concretion growth appears also to have contributed to the increase of the vein aperture by reactivation of these fractures when the chert was still not completely lithified; the veins have large aperture sizes in the parts of the chert beds that have greatest curvature (Figure 4.7), whereas they exhibit narrow apertures in the parts of the same chert beds that preserved their horizontal setting due to the absence of concretions.

Vertical planar calcite veins, in contrast, require a different explanation. They were observed in many examples in the Qatrana Roadcut to transect the carbonate concretions and the beds above and below the concretions (Figure 4.7), indicating that these veins postdate the growth of the concretions. This is supported by the stable isotopic data that show that the veins are more depleted in carbon than the concretions and the host sediments.

The immediate occurrence of fibrous gypsum veins in the Sultani Quarry above and below carbonate concretions, with different relationships with concretions, either bending around concretions or transecting the concretion bodies, offer an additional opportunity to examine the development of both the concretions and the associated gypsum veins and constrain their timing. Fibrous veins are a major focus of many structural studies in which the geometry of fracture is used to understand the type of deformation (Ramsay & Huber, 1983; Rodrigues et al., 2009) or the state of stress (Pollard et al., 1982, Rickard & Rixon, 1983) in the host rock at the time of fracturing. The direction of fibres in these veins at the Sultani quarry has been investigated in the areas where the vein is wrapping around the concretion. The veins appear to follow the deflections in bedding around the concretions

(Figure 4.18). Rodrigues et al. (2009) have argued that fibrous segments of concretion-associated fibrous veins can have radial distribution that coincides with the deflections of the beds around concretions. The orthogonal fibre direction in the gypsum vein segments, regardless their dips, indicates that the veins have developed at a late stage that postdates the concretion growth; the veins have been opened along the deflected beds around the concretions. The gypsum veins were observed in the Sultani Quarry to transect the carbonate concretions (Figure 4.18). This strongly demonstrates the late formation of these veins and provides additional evidence that the veins postdate the formation of the concretions.

As mentioned above, the formation of concretions occurs by cementation of the grains in the sediment; the mineral precipitates from the pore water and simply takes the place of the water in the rock. If a cementation occurs at early stage in the compaction and burial history, there will be spaces between the grains and the cement will occupy a large volume of the concretion (Marshall & Pirrie, 2013). In contrast, if the minerals form at late stage of the burial history, the remaining space for cement to pervade will be much lesser. At the microscopic scale, the former case is evident in which the concretion shows high quantities of cement, implying an early timing of development of these concretions (Figure 4.27).

The systematic stable isotopic analysis of both types of carbonate concretions shows a narrow range of samples analysed from the centre to edge of each individual type of concretion (Table 4.1; Figure 4.29). However, the isotopic results show that the edges are more depleted in carbon than the centres (Figure 4.29). This supports that the concretion formation commenced at shallow burial depth, but may have continued during progressive burial (Woo & Khim, 2006).

4.9 Conclusions

- 1) Natural fractures are abundant in the Upper Cretaceous-Paleogene outcrops, with a dominance of opening-mode fractures. Fractures observed include veins, cone-in-cone structures, joints, bitumen-filled fractures, and faults. Opening-mode fractures are the most common observed fractures.
- 2) Rock properties and diagenesis are major factors that controlled fracture distribution and development in the succession.
- 3) Concretions have different relationships with fractures ranging from fracturing resulting from concretion growth (displacive growth) to late fracturing enhanced by the pre-existence of these rigid diagenetic features (differential compaction).
- 4) The timing of fracture development in the Upper Cretaceous-Eocene rocks can be divided into two main constraints: (a) early fractures including folded calcite veins, bedding-parallel veins and silica veins and (b) late fractures (including late tectonic fractures including planar calcite veins, joints and faults; and late diagenetic fractures such as the white powdery gypsum veins).
- 5) Different formation mechanisms have controlled fracturing in the Upper Cretaceous-Eocene succession. These include fluid pressure (e.g., bed-parallel calcite veins and gypsum veins), early diagenesis (e.g., folded calcite veins, planar silica vein), late diagenetic process (e.g., white powdery gypsum veins), force of crystallization (e.g., fibrous gypsum veins), and late tectonic processes (e.g., planar veins and joints within damage zones associated with faulting).

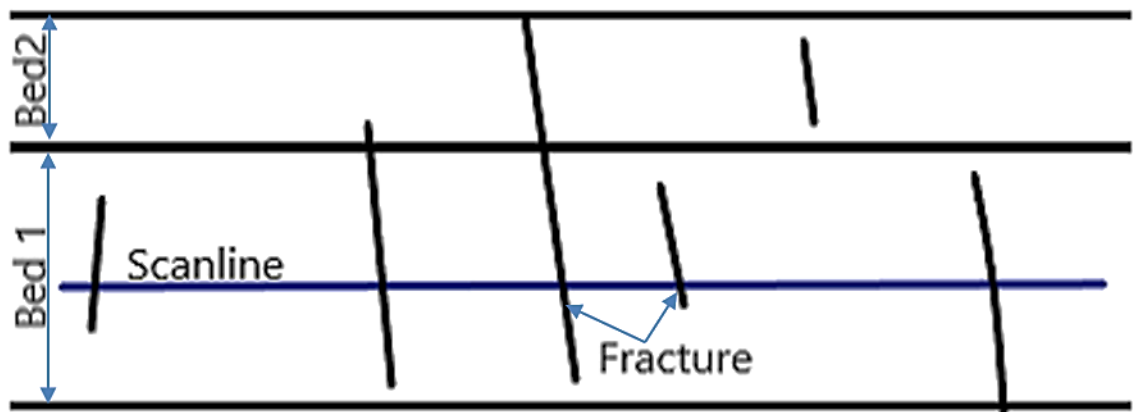


Figure 4.1 Illustration of the scanline method used to measure fracture properties (height, width and spacing) from the outcrop.

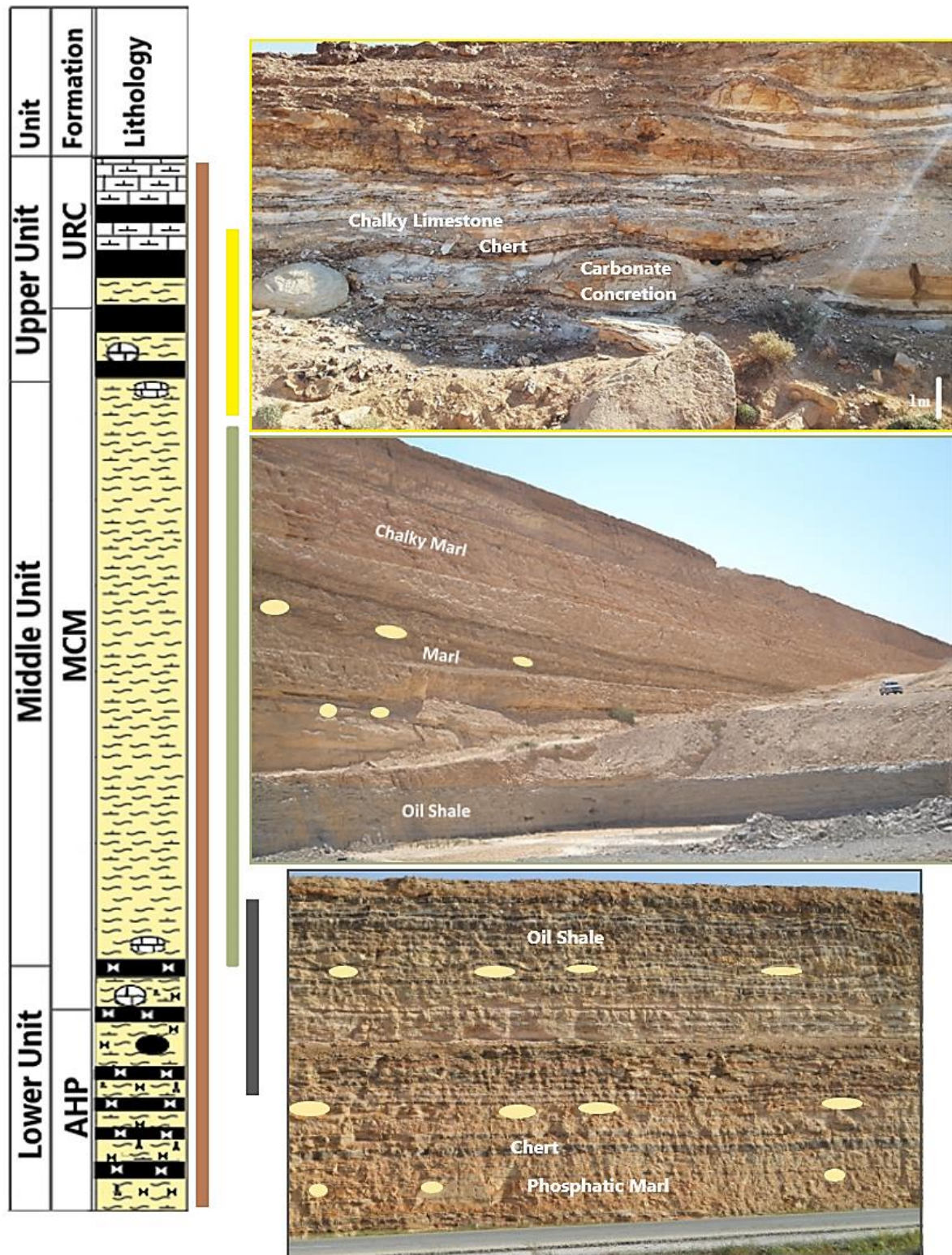


Figure 4.2 Representation of the study succession from the different studied outcrops. Brown line represents the whole succession in Wadi Al-Bustan; grey line represents the Qatrana Roadcut; green line represents the Sultani and El-Lajjun Quarries; yellow line represents the Harrana Outcrop. Grey-framed photo from the Qatrana Roadcut; green-framed photo from the Sultani Quarry; yellow-framed photo from the Harrana Outcrop. Note the different concretion horizons (beige ellipsoids).

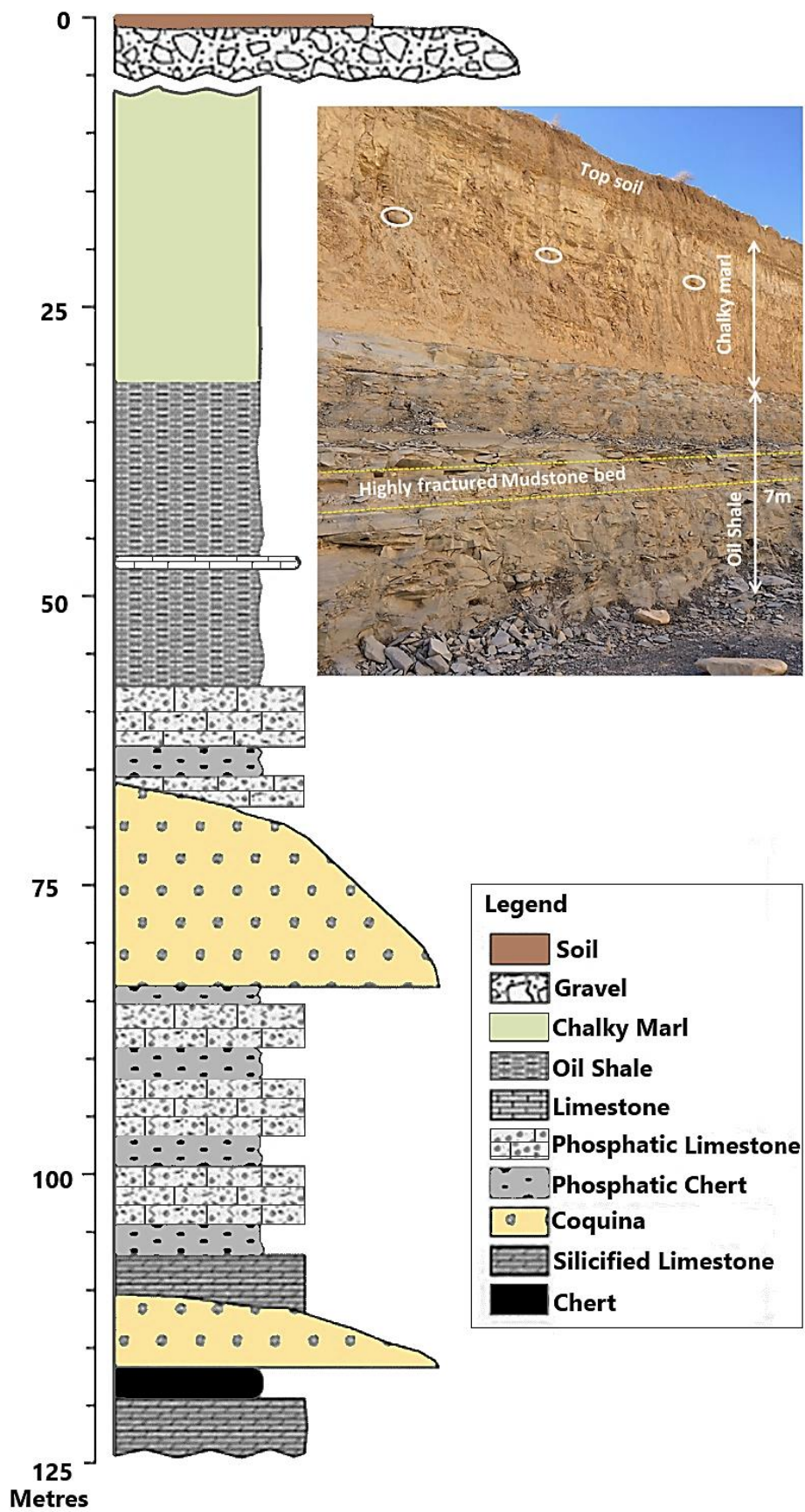


Figure 4.3 Columnar section showing the lithology of the El-Lajjun area (modified from Shawabkeh, 1991). Field photograph on the right is from the El-Lajjun Quarry.



Figure 4.4 Field photographs showing examples of the outcrop platform in Wadi AL-Bustan.

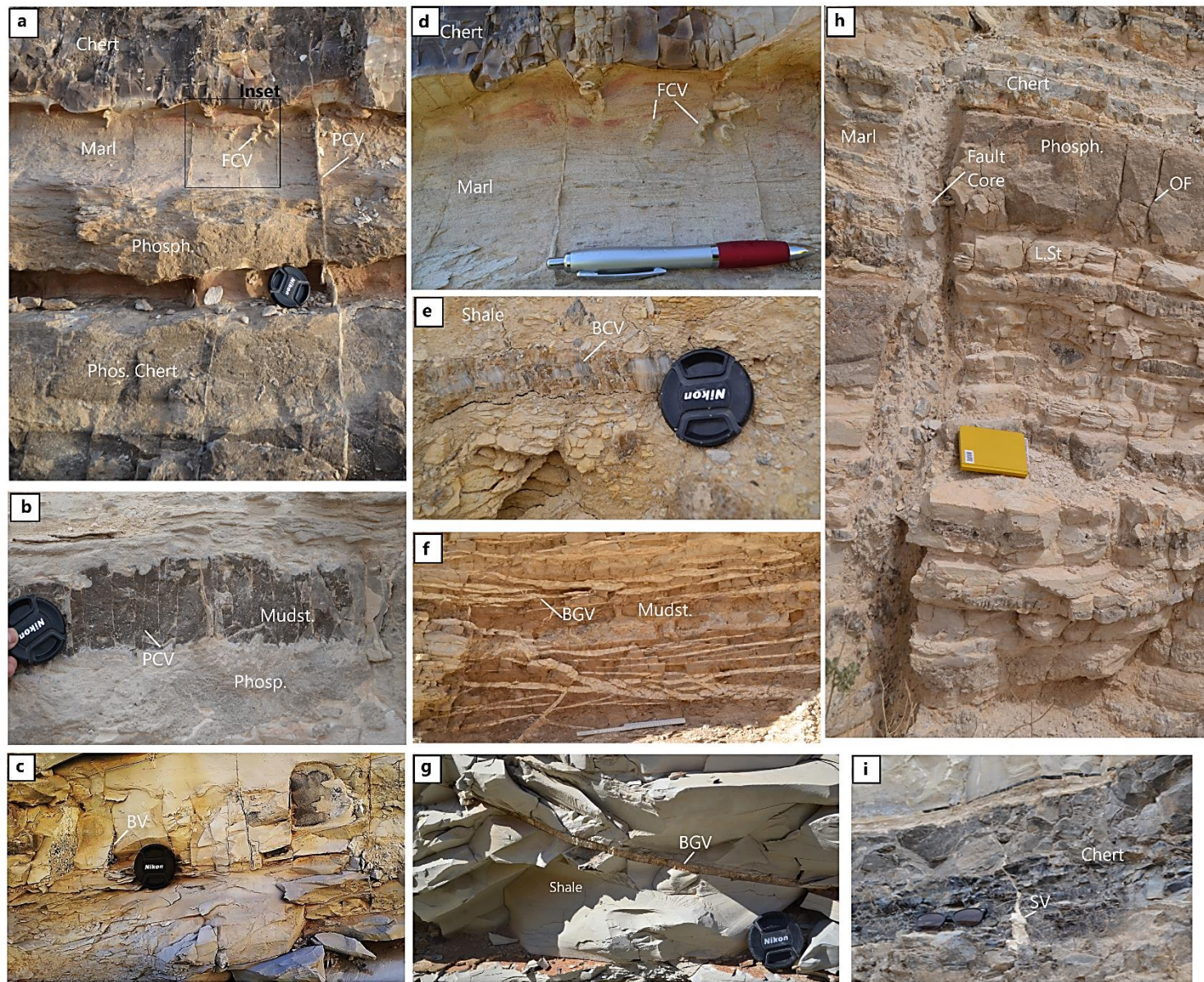


Figure 4.5 Field photographs showing examples of the different fracture types observed in the study outcrops. PCV, Planar Calcite Vein; FCV, Folded Calcite Vein; BGV, Bedding-Parallel Calcite Vein; BGV, Bedding-Parallel Gypsum Vein, OF, Open Fracture; Phosph., Phosphatic Chert; L.St, Limestone; SV, Silica Vein. a. Thoroughgoing PCVs and bed-confined FCVs in the Qatrana roadcut. Note the coexistence of planar and folded veins in the same bed. b. PCVs confined to mudstone bed in the Qatrana roadcut. The veins are closely spaced. c. BVs in oil shale interval in the El-Lajjun Quarry. d. Enlargement of the inset in (a). e. BGV in oil shale interval in the Middle Unit in Wadi Al-Bustan. f. BGVs in mudstone in the Sultani Quarry. g. BGVs in oil shale interval in the Sultani Quarry. h. Fault zone and open fractures in the Qatrana Roadcut. Note the fault core including breccias. i. SV in chert interval in the Qatrana Roadcut.

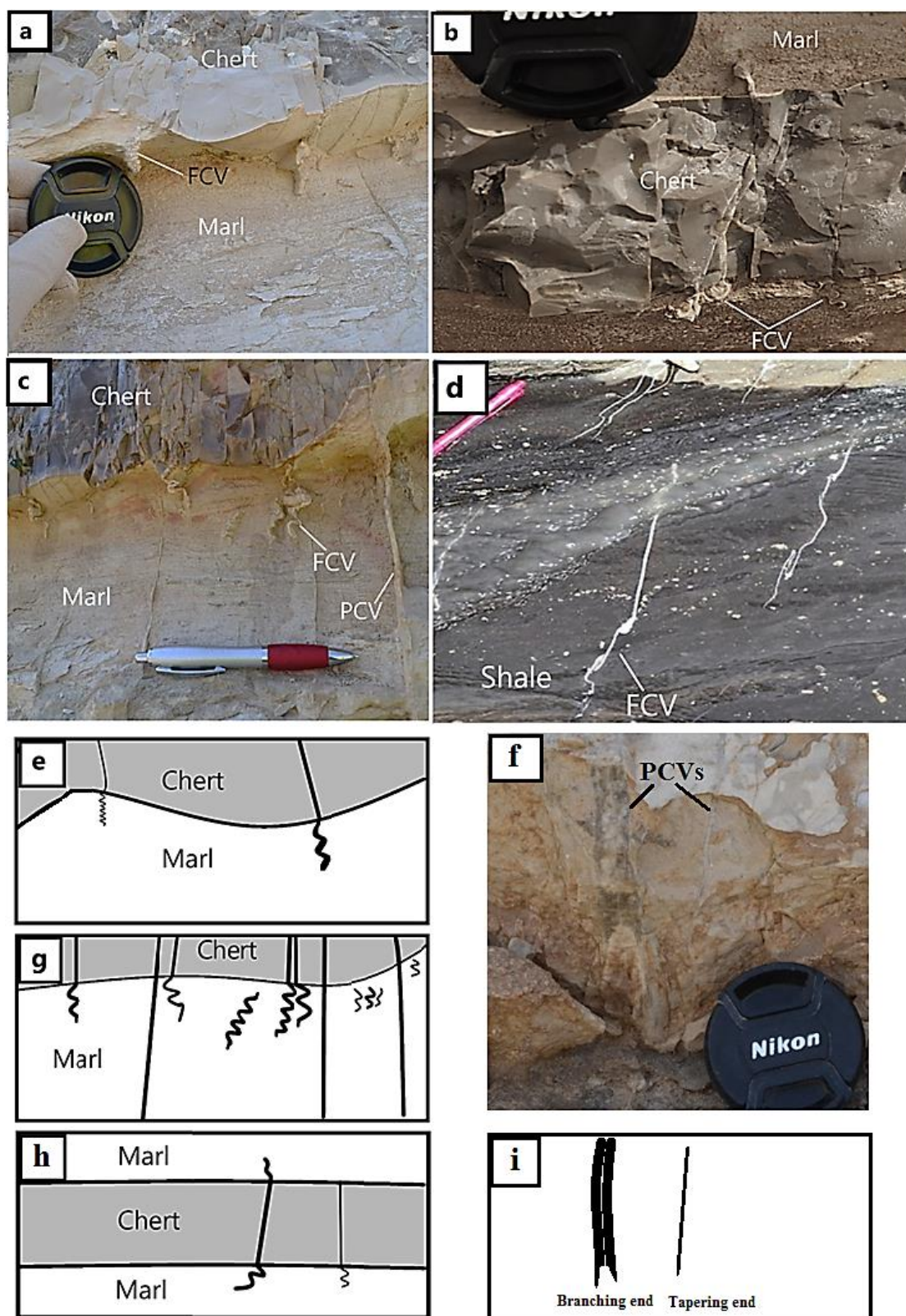


Figure 4.6 Field photographs of folded and planar calcite veins. FCV, Folded Calcite Vein; PCV, Planar Calcite Vein. a-c. Thick and thin folded calcite veins in marl interval the Qatrana Roadcut. d. Folded calcite vein on a shale platform in Wadi Al-Bustan. e. Drawing of (a). Note that the folded veins continue straight when transect chert interval. f. Planar calcite veins with different thicknesses and ends. Note the branching end of the thick vein and the tapering end of the thin vein. g. Drawing of (c). h. Drawing of (b). i. Drawing of (f).

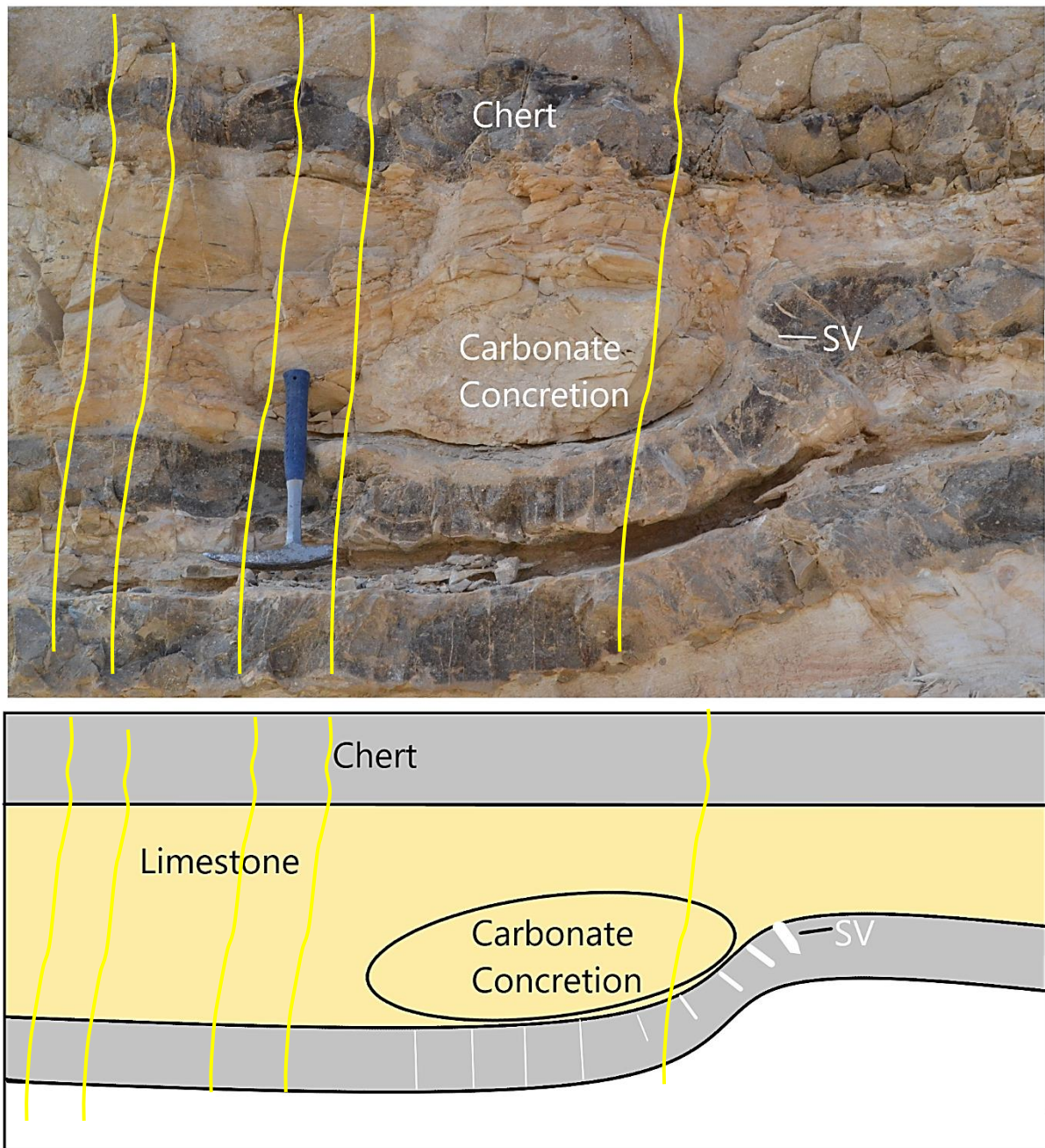


Figure 4.7 Field photograph and drawing showing silica veins (white lines) in a chert bed bending around concretion. Note the maximum aperture size in the areas that show high degree of bending. Note the throughgoing calcite veins (yellow lines) that transect the concretion and the host rocks.

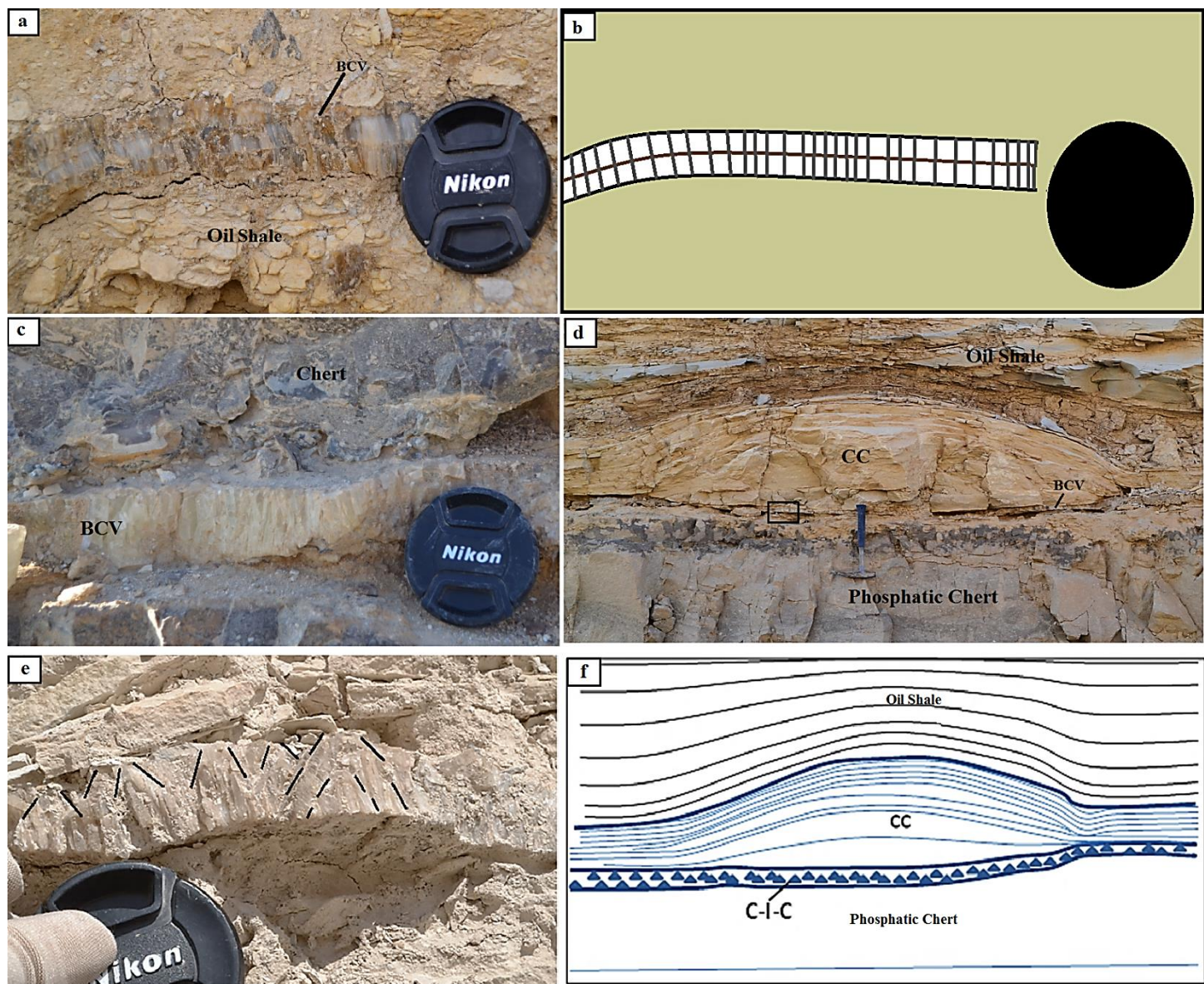


Figure 4.8 Field photographs and drawing of bedding-parallel calcite veins observed in the study outcrops. BCV, Bedding-parallel calcite vein; CC, Carbonate Concretion. a. BCV in oil shale interval in the Middle Unit in Wadi Al-Bustan. b. drawing of (a). Note the antitaxial texture where fibres growing from the median line (brown line) outwards perpendicular to the vein wall. c. BCV in the Lower Unit in the Qatrana Roadcut. d. Cone-in-cone bending at the margin of carbonate concretion in the Qatrana Roadcut. e. Enlargement of the inset in (d). f. drawing of (d). Note the laminae preserved inside the concretion.

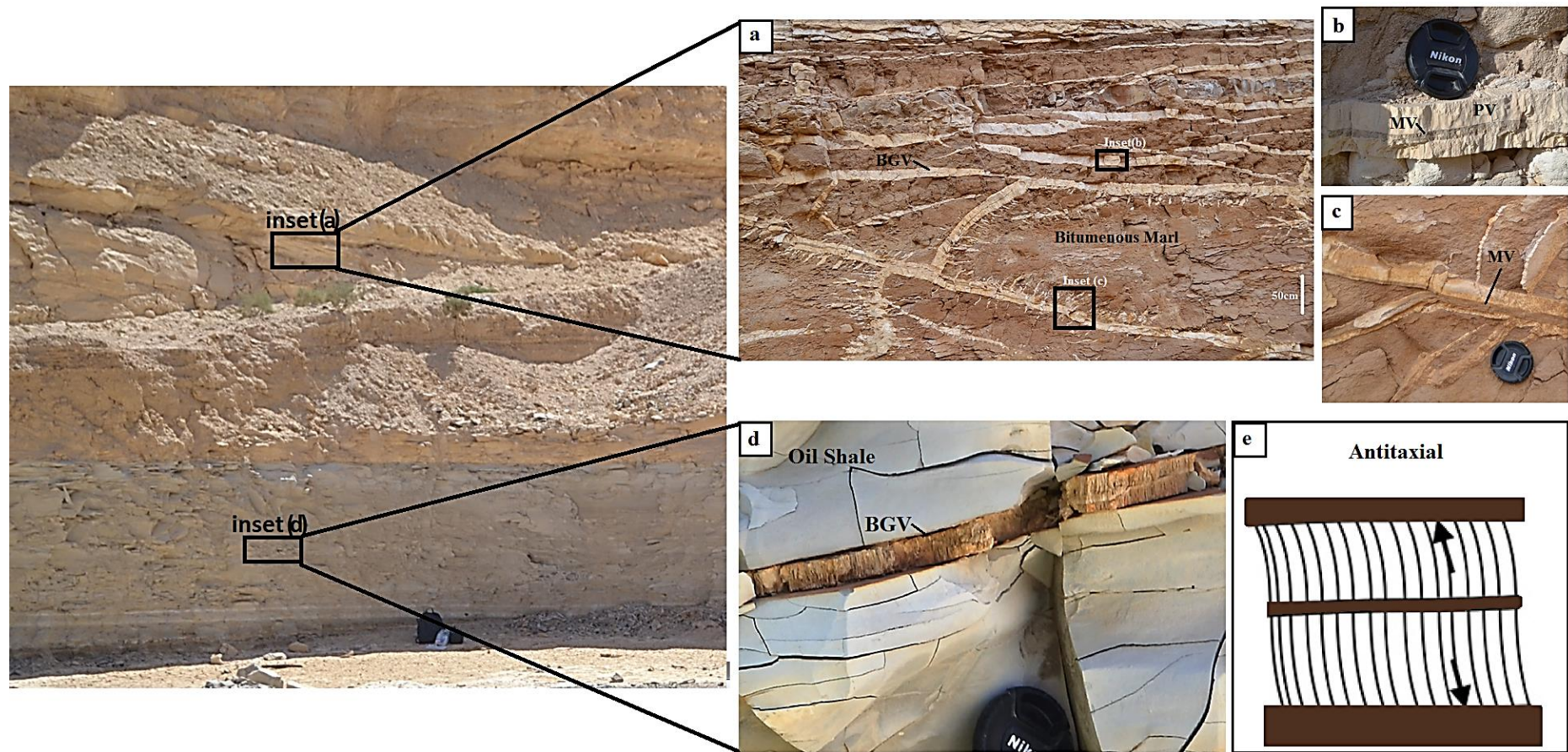


Figure 4.9 Bedding-parallel gypsum veins in the Sultani Quarry. BGV, Bedding-Parallel Gypsum Vein; MV, Median Vein; PV, Powdery Vein. a. A network of bedding-parallel and inclined gypsum veins in the bituminous marl interval. b. Enlargement of inset (b) in Figure (4.9a). Note the two vein horizons. c. Enlargement of inset (c) in Figure (4.9a). Note the presence of the median line in the inclined vein and the absence of this line in the branches. d. Bedding-parallel gypsum vein in the oil shale interval. e. Drawing of the antitaxial texture observed in the fibrous gypsum veins.

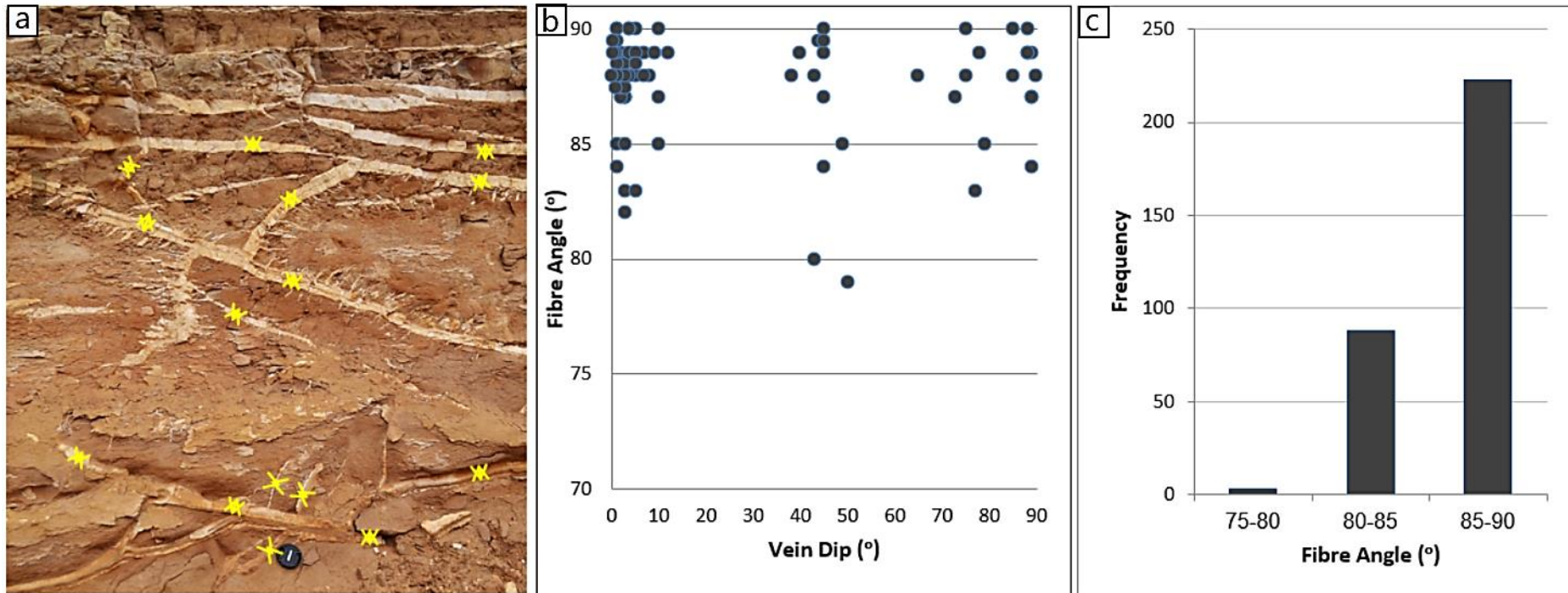


Figure 4.10 a. Field photograph showing the orthogonal directions of the gypsum fibres to the vein wall. b. Plot of vein dips against the angle between fibers and vein wall. c. Histogram showing the distribution of angles between fibres and vein walls. $N = 314$

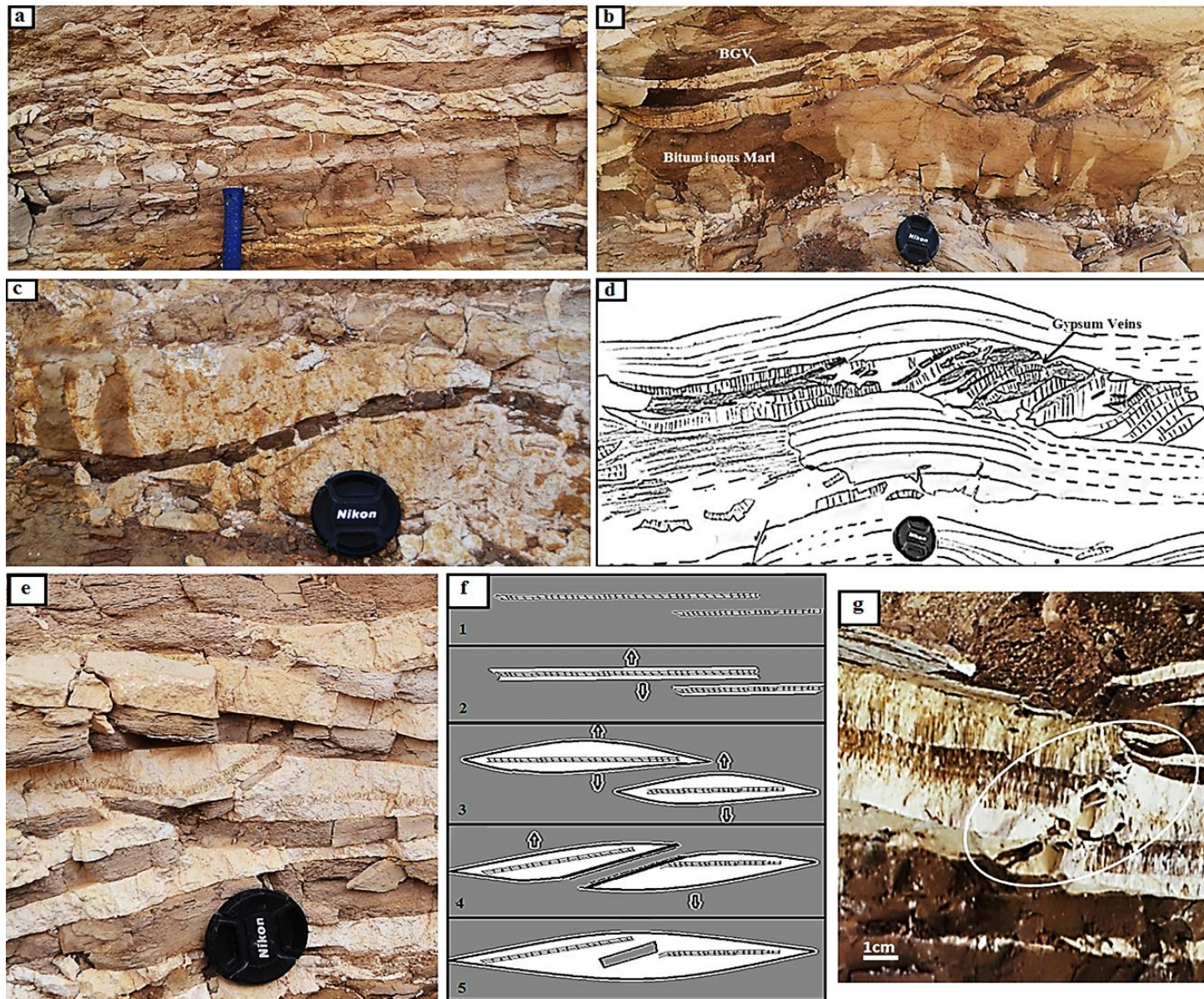


Figure 4.11 Examples of different deformation features in gypsum veins and host rocks in the Sultani Quarry. a. and b. Field photographs showing deformation related to horizontal shortening. c. Field photograph showing bridging between two lenticular gypsum bodies. d. Drawing of (b). e. Bridging/overstepping in gypsum veins. f. Drawing of the overstepping stages. g. Micro faulting.

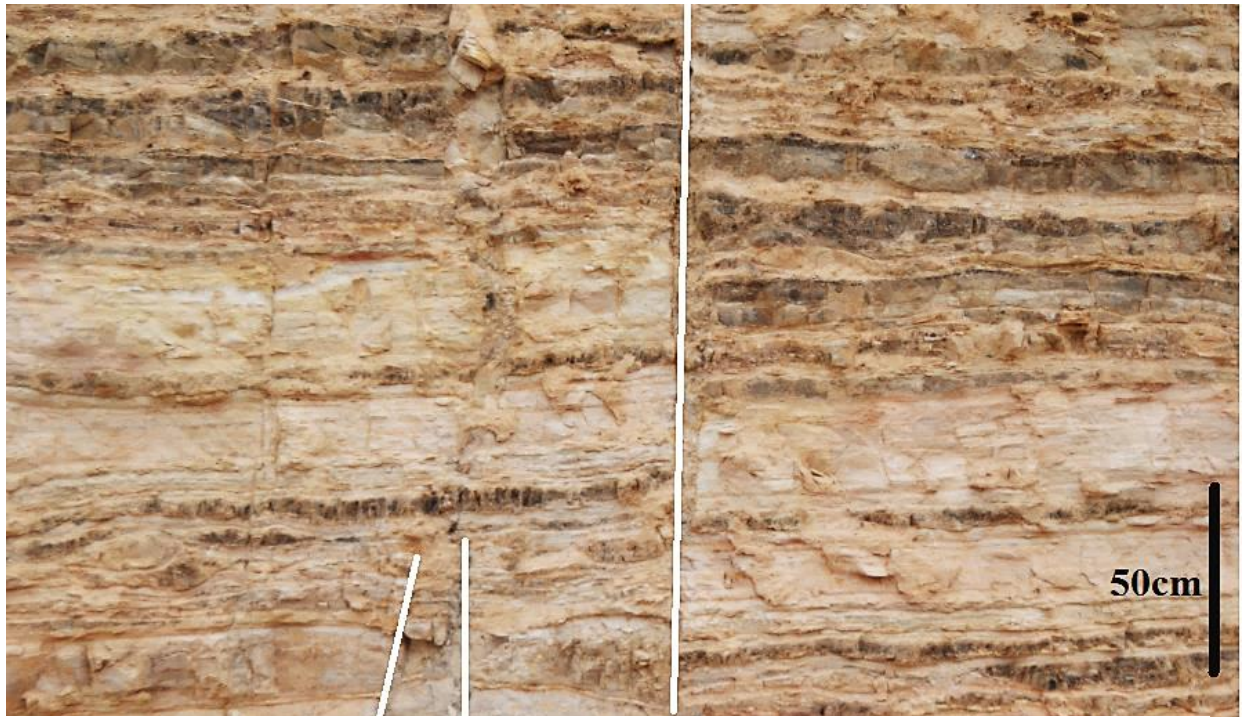


Figure 4.12 Examples of simple faulting in the Qatrana Roadcut. White lines are fault planes.

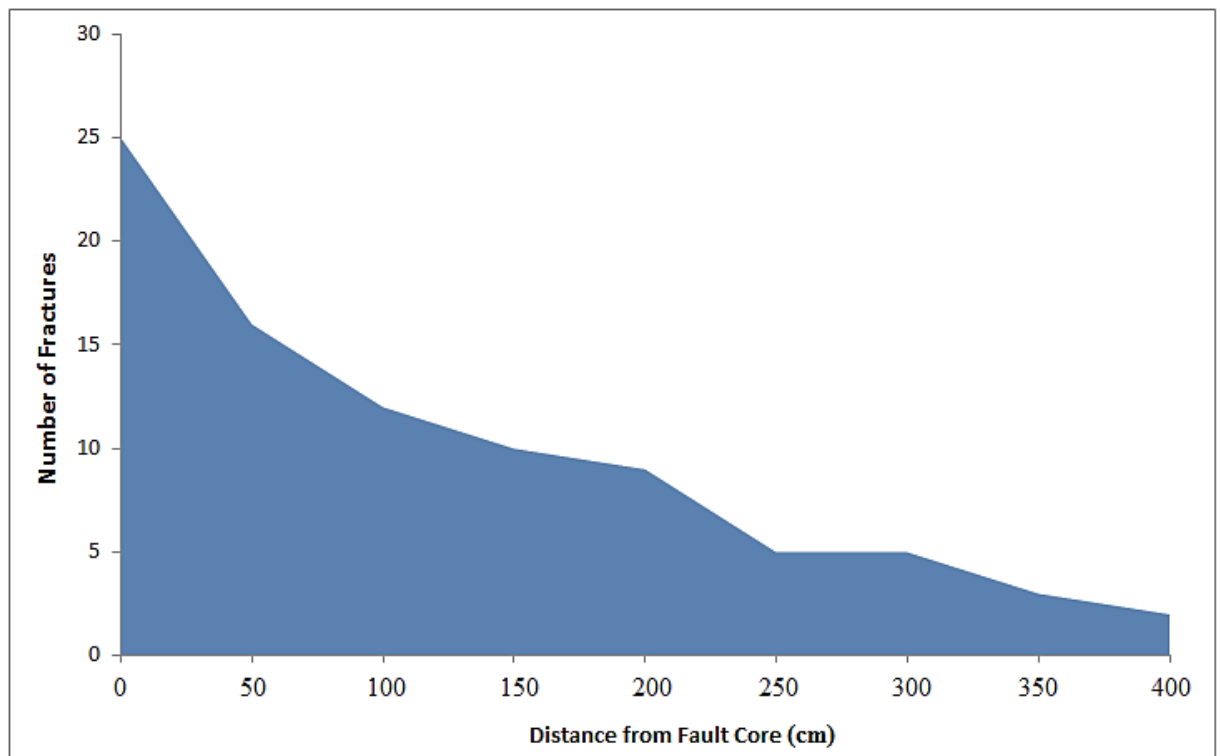


Figure 4.13 The intensity of opening-mode fractures associated with faults in the Qatrana roadcut.



Figure 4.14 Field photographs showing the different horizons of concretions in the Qatrana Outcrop (a) and the bending of thin chert beds around carbonate concretions (b). Note the direction of silica veins (white lines) in the chert beds in (b) and the bending of the small-concretion horizon (black elliptical shapes). Ch, Chert; CC, Carbonate concretion.

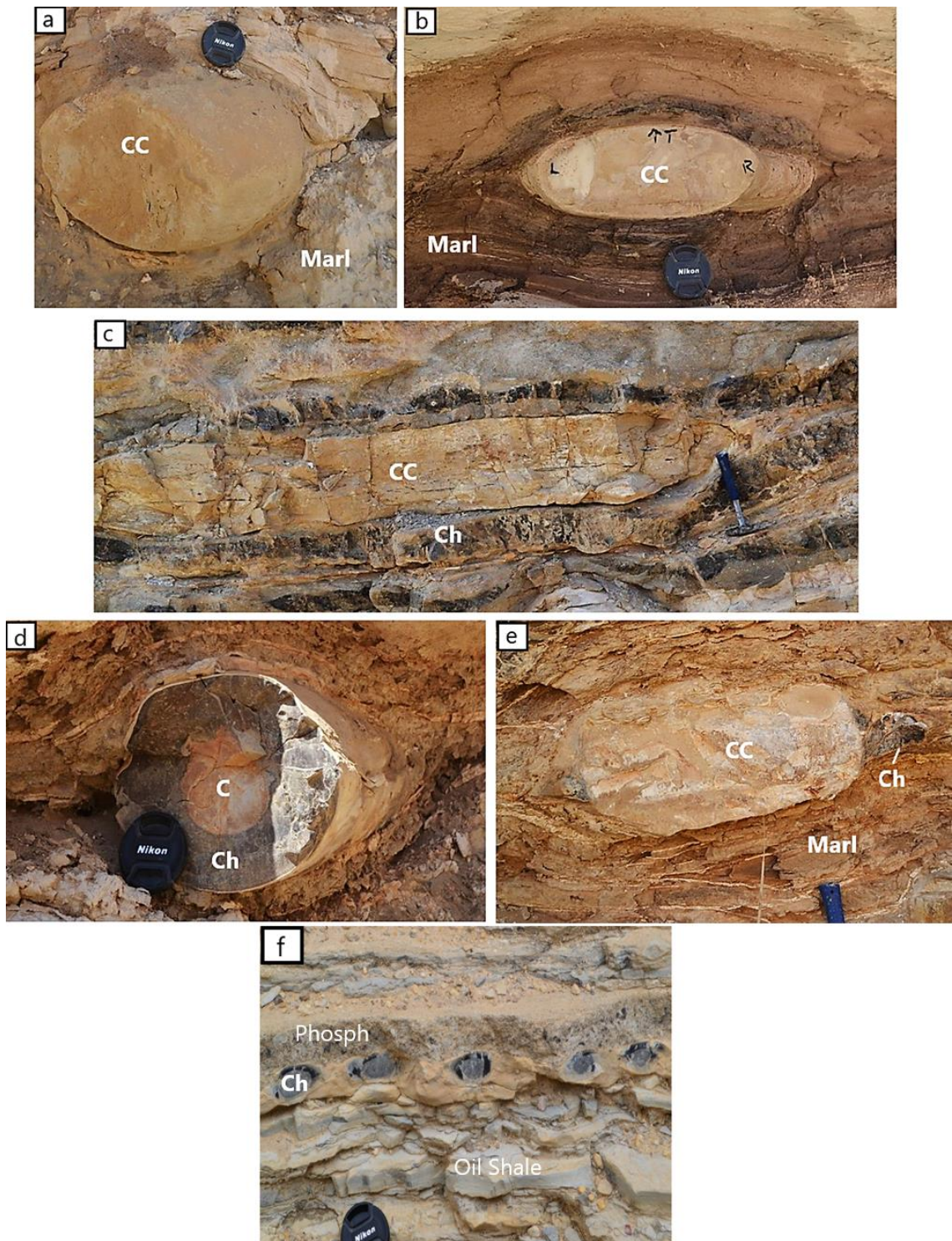


Figure 4.15 Field photographs showing examples the different types of carbonate and chert concretions observed in the study outcrops. CC, Carbonate Concretion; Ch, Chert; C, Carbonate; Phosph; Phosphatic marl. a. Spherical carbonate concretion in the Sultani Quarry. b. Elliptical carbonate concretion in the Sultan i Quarry. c. Tabular carbonate concretion in the Qatrana roadcut. Note the chert laminae inside the concretion. d. Zoned carbonate-chert concretion in the Sultani Quarry. e. Carbonate concretion with triangular shapes filled by chert at the edges in the Sultani Quarry. f. Horizon of chert concretions in Wadi Al-Bustan.

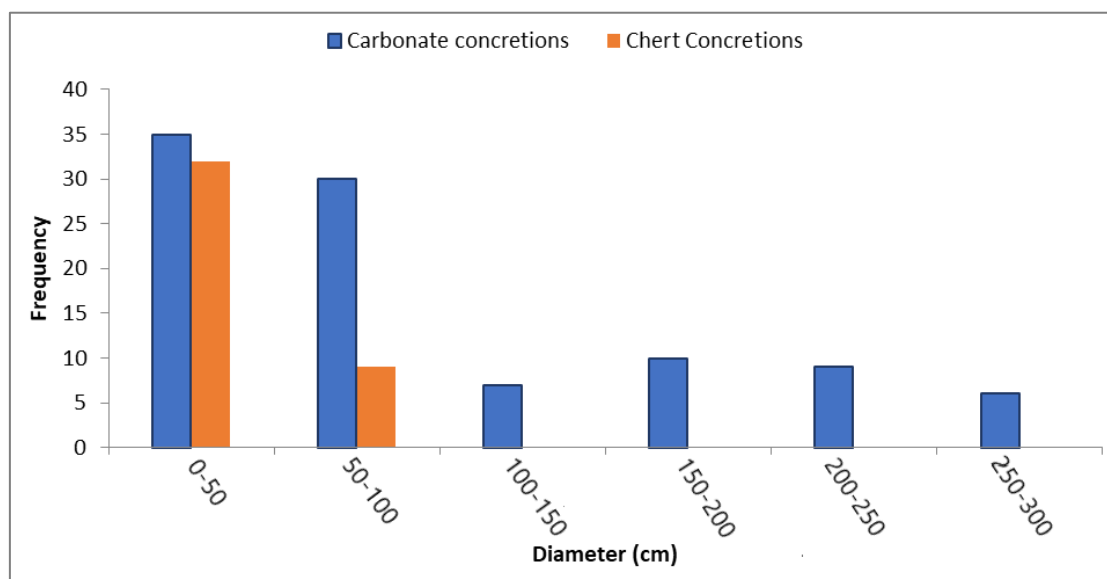


Figure 4.16 Histogram showing the diameter values of carbonate and chert concretions in all the study outcrops.

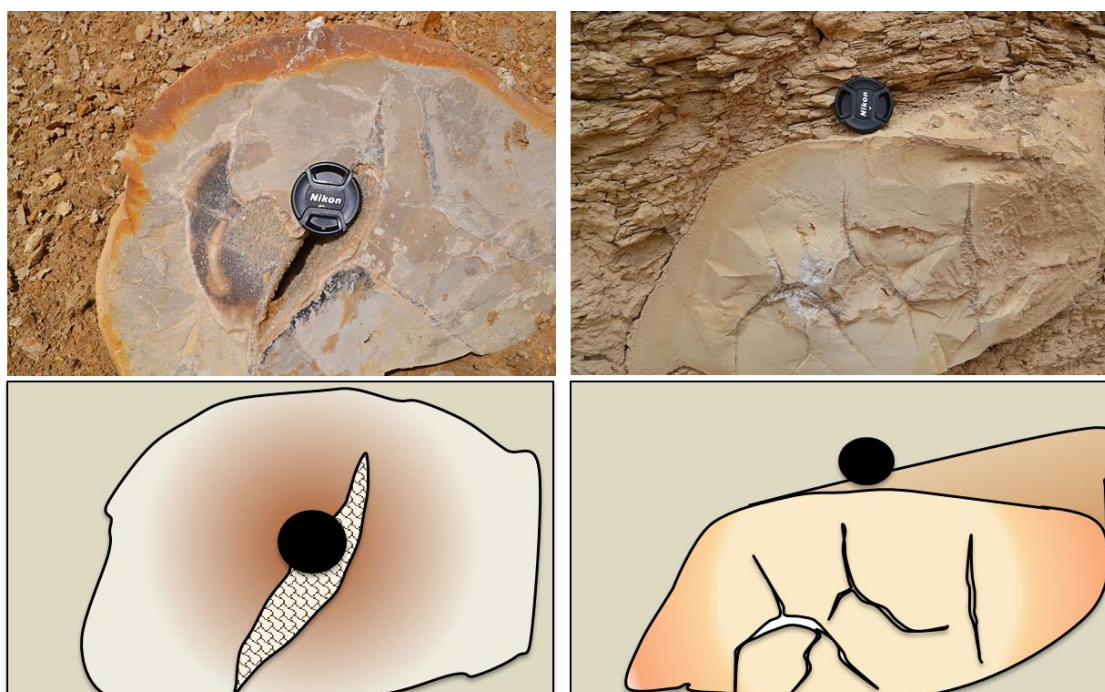


Figure 4.17 Examples of septarian fractures in carbonate concretion in the Sultani Quarry.

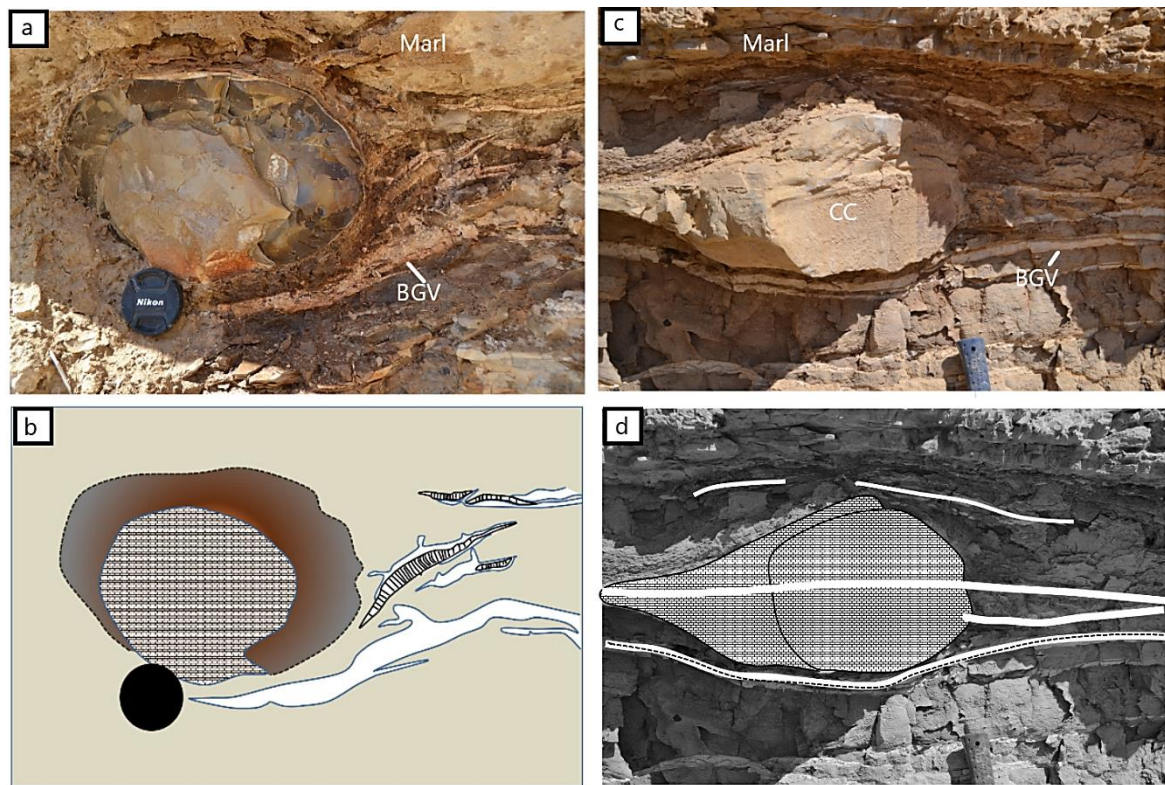


Figure 4.18 a. Bedding-parallel gypsum vein bending around a concretion. b. Drawing of (a). c. Bedding-parallel gypsum veins transecting a carbonate concretion. d. Drawing of (c).

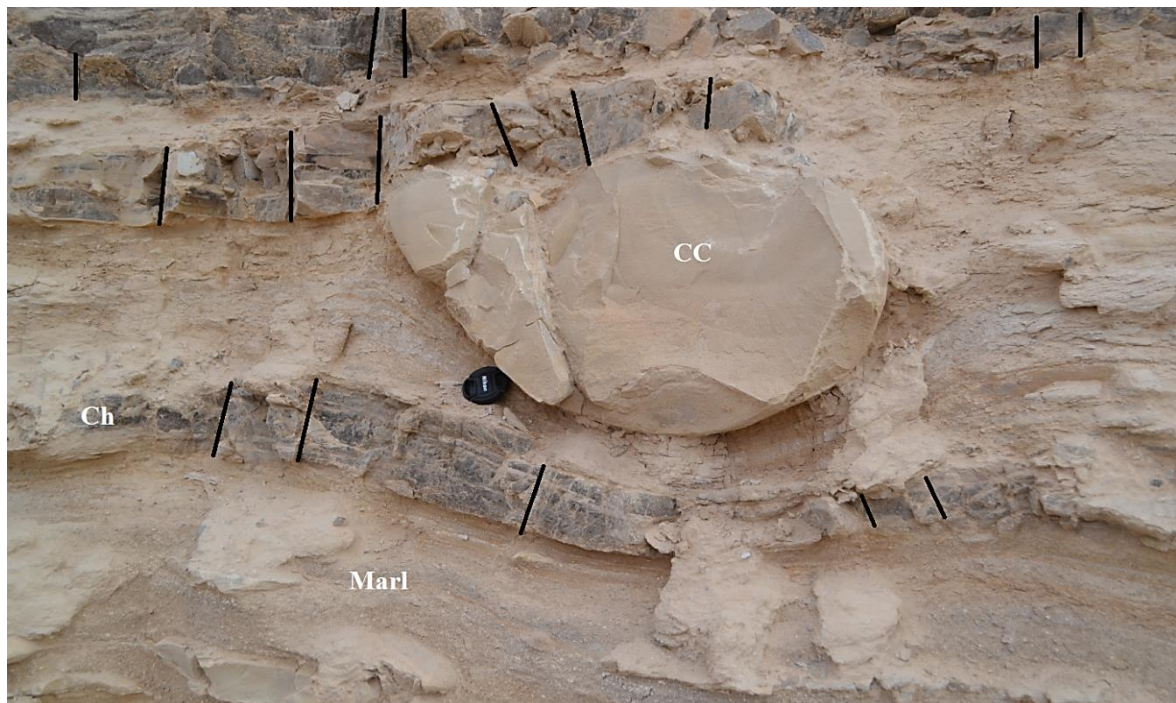


Figure 4.19 Field photograph showing joints (black lines) in chert beds bending around a carbonate concretion. CC, Carbonate Concretion; Ch, Chert.

Outcrop Name	Thickness of the Studied Section (m)	Location	Representative Unit	Total Number of Fractures	Fracture Type
Harrana	35	Central Jordan	Upper	85	Planar calcite, silica veins and joints
El-Lajjun Quarry	40	Central Jordan	Lower	30	Bitumen veins
Qatrana Roadcut	40	Central Jordan	Lower	152	Calcite veins (planar, folded and bedding-parallel), silica veins and joints
Sultani Quarry	45	Central Jordan	Middle	40	Bedding-parallel gypsum veins
Wadi Al-Bustan Exposure	150	South Jordan	Lower, Middle and Upper	250	Calcite veins (planar, folded and bedding-parallel), silica veins and joints
Total	310			557	

Table 4.1 Fracture dataset from all of the study outcrops

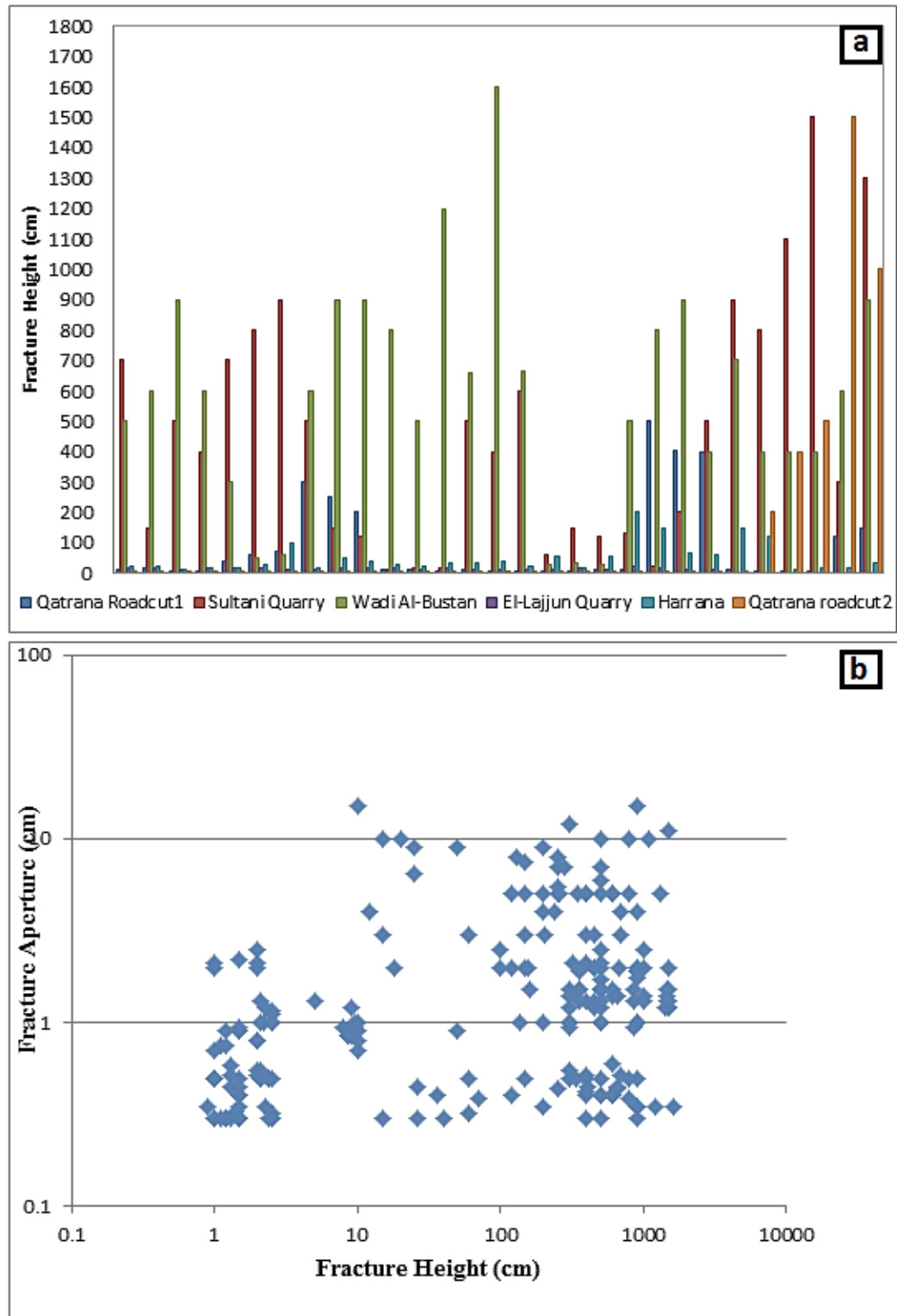


Figure 4.20 a. Fracture height of all fractures from all the study outcrops. Note that the Qatrana Roadcut (1 and 2) represents the Lower Unit; Sultani Quarry represents the Middle Unit; Harrana Outcrop represents the Upper Unit; and Wadi Al-Bustan Outcrop represents the entire section (see Figure 4.22 for more information). b. Aperture/height data of all opening-mode fractures observed in the study outcrops. Large apertures represent the apertures of the bedding-parallel gypsum veins in the Sultani Quarry.

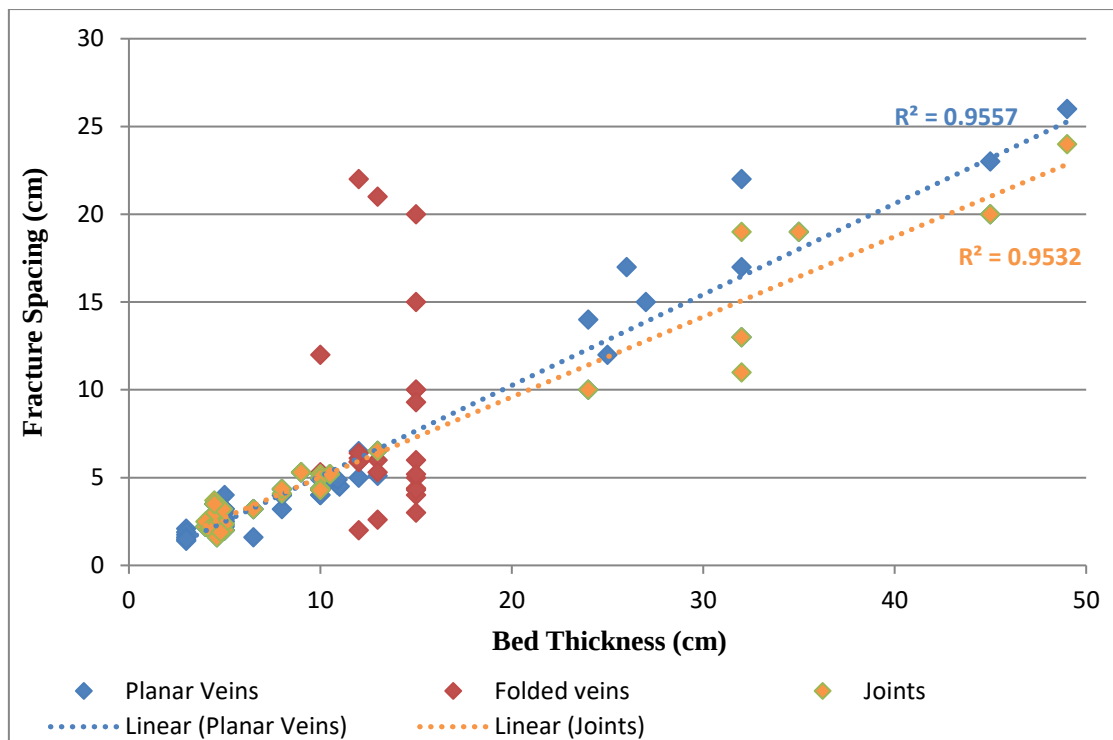


Figure 4.21 Plot of fracture spacing against bed thickness of planar and folded calcite veins. $N = 154$

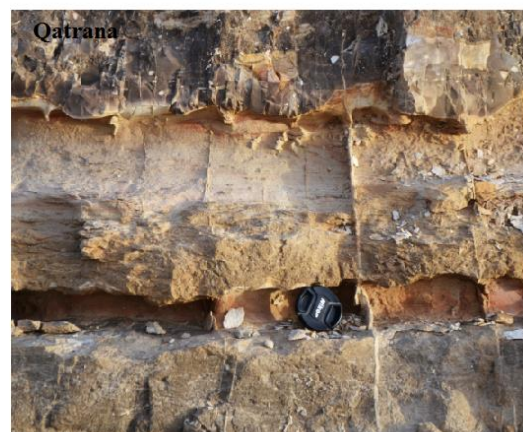
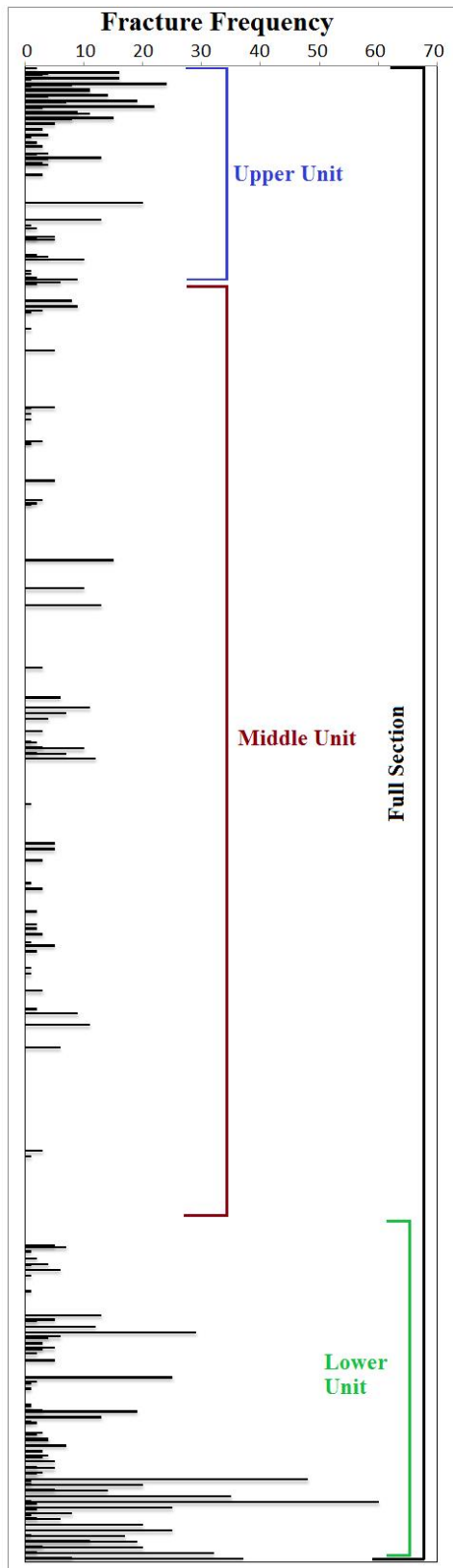


Figure 4.22 The distribution of fracture along the study succession. Note the clustering of fractures in the lower (Qatrana Roadcut) and upper (Harrana Outcrop) units. The Middle Unit (in the Lajjun and Sultani quarries) show relatively low fracture intensity. The full section was available in Wadi Al-Bustan.

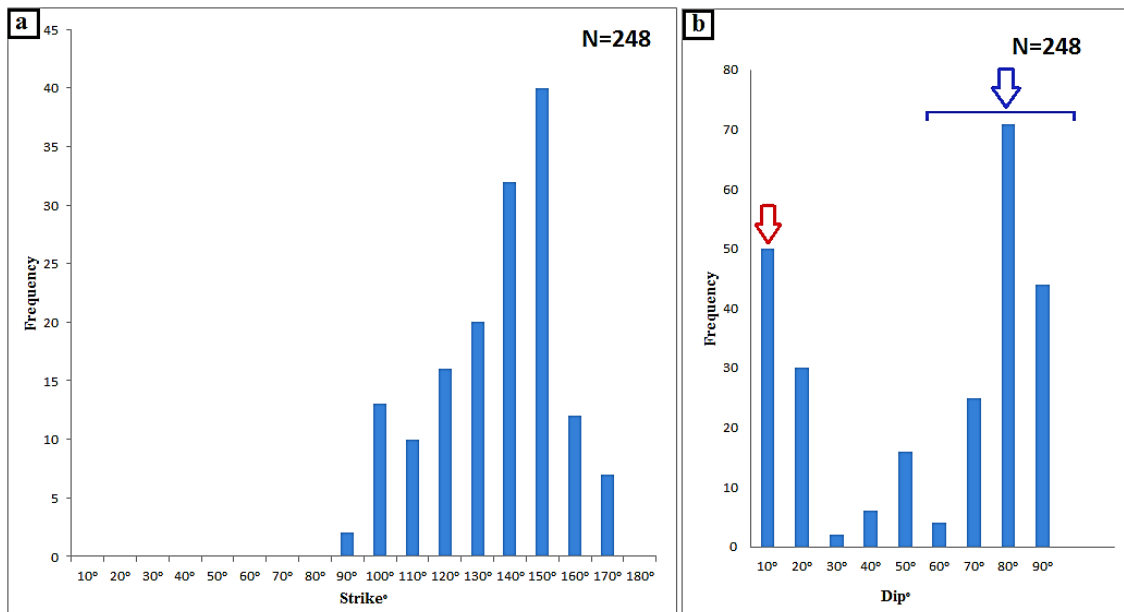


Figure 4.23 Orientation measurements of fractures in the outcrops for strike (a) and dip (b). Red arrow represents dip values of bedding-parallel veins in the Sultani quarry and blue arrow points to the fracture dip values in the Qatrana roadcut.

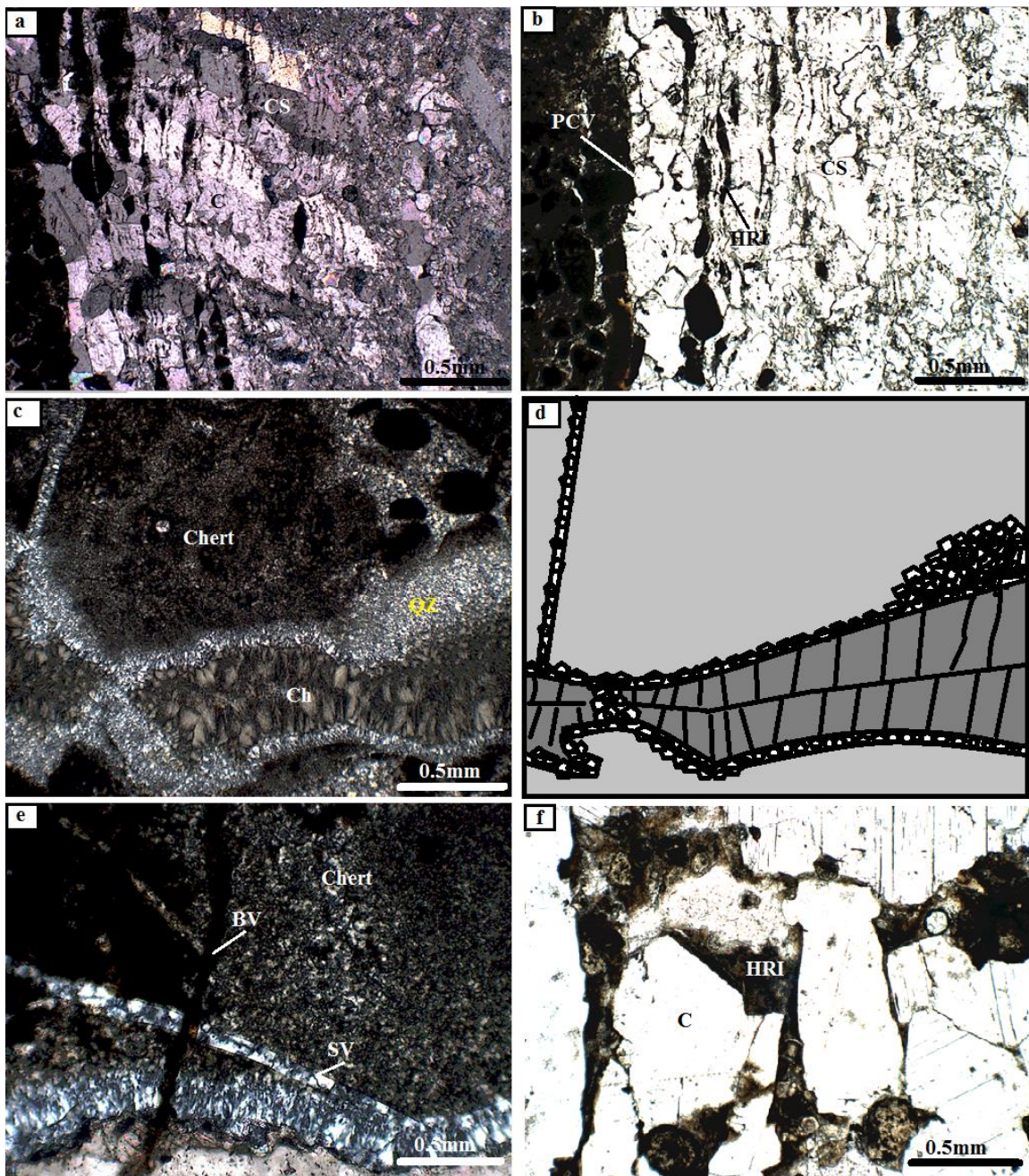


Figure 4.24 Microphotographs showing the texture and composition of the fractures observed in the outcrops. C, calcite; CS, Crack Seal; PCV, Planar Calcite Vein; HRI, Hostrock Inclusion; Qz, Microcrystalline Quartz; Ch, Chalcedony; BV, Bitumen Vein; SV Silica Vein. a. Vertical planar calcite vein showing a crack-seal texture in CL. b. Same as (a) but in PPL. c. Fibrous chalcedony vein enclosed by microcrystalline quartz vein at both walls in a folded chert bed in the Qatrana Roadcut. d. Drawing of (c). e. Bitumen vein crosscutting silica veins in chert interval from the Qatrana Roadcut. f. Bedding-parallel calcite vein showing elongate-blocky texture. Note the host rock inclusions within the vein.

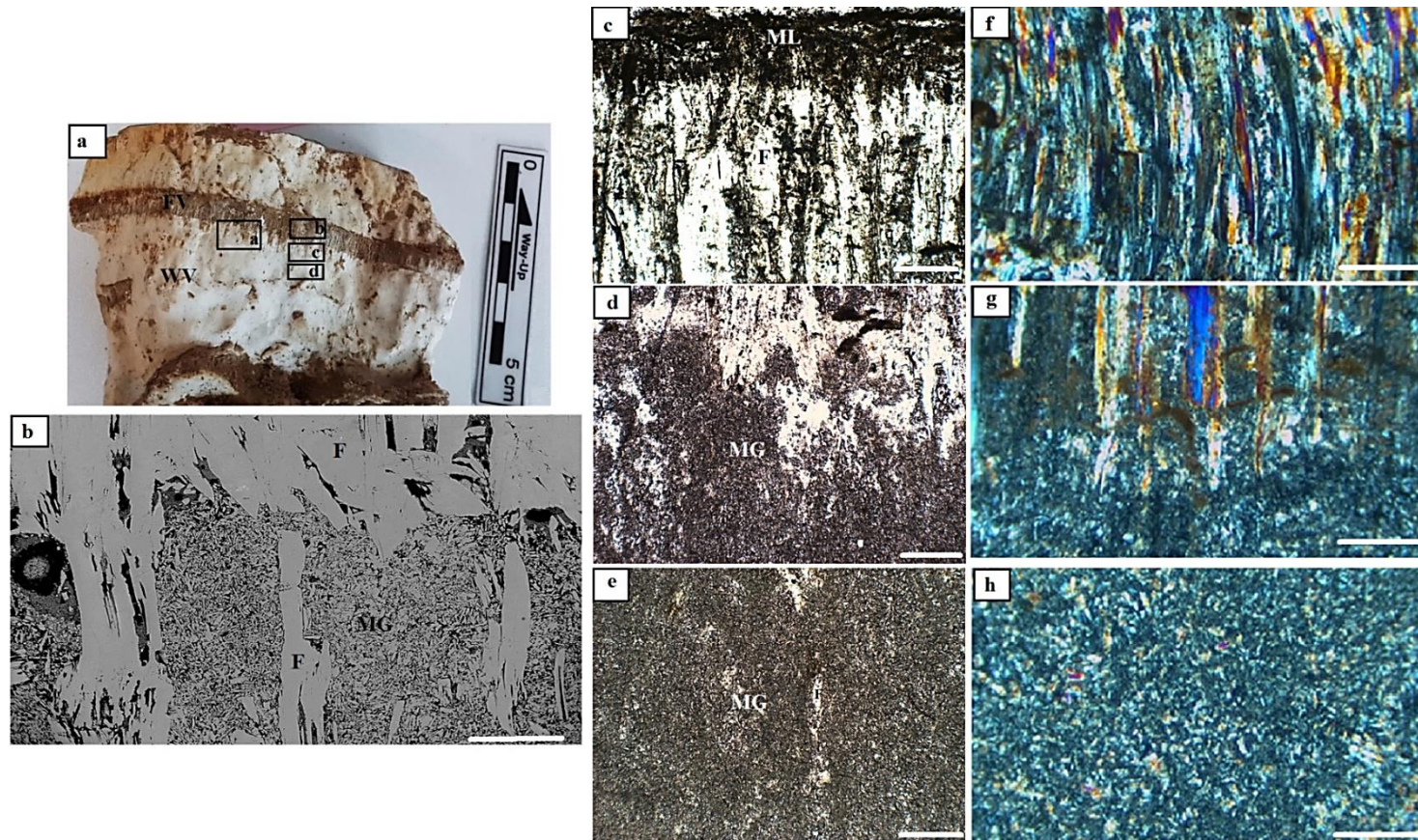


Figure 4.25 Bedding-parallel gypsum veins. FV, Fibrous Vein; WV, White Powdery Vein; F, Fibre, MG, Microcrystalline Gypsum; ML, Median line. Scale bar from (c) to (h) is 0.5mm. a. Hand specimen showing the median fibrous vein and the white powdery vein. b. BSE image of the inset (a) in Figure 4.23a. c-e. Microphotographs of the insets (b-d) in Figure 4.23a in PPL. f-g. Microphotographs of the insets (b-d) in Figure 4.23a in CL. Note the change in texture from fibrous gypsum to microcrystalline gypsum with the move from the median vein in (c, f) towards the white powdery vein (e, h).

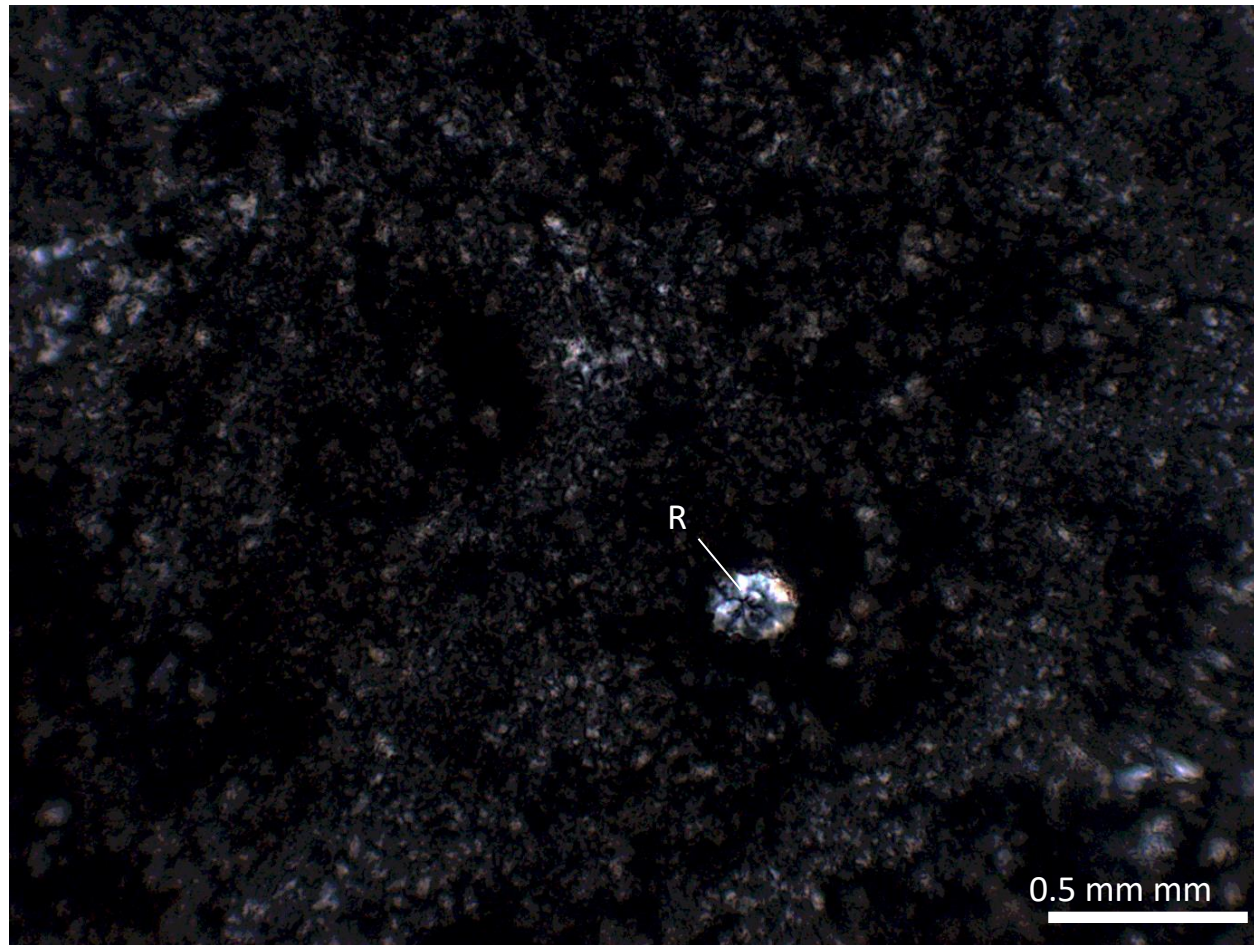


Figure 4.26 Microphotograph in CL showing a chalcedony-filled radiolarian.

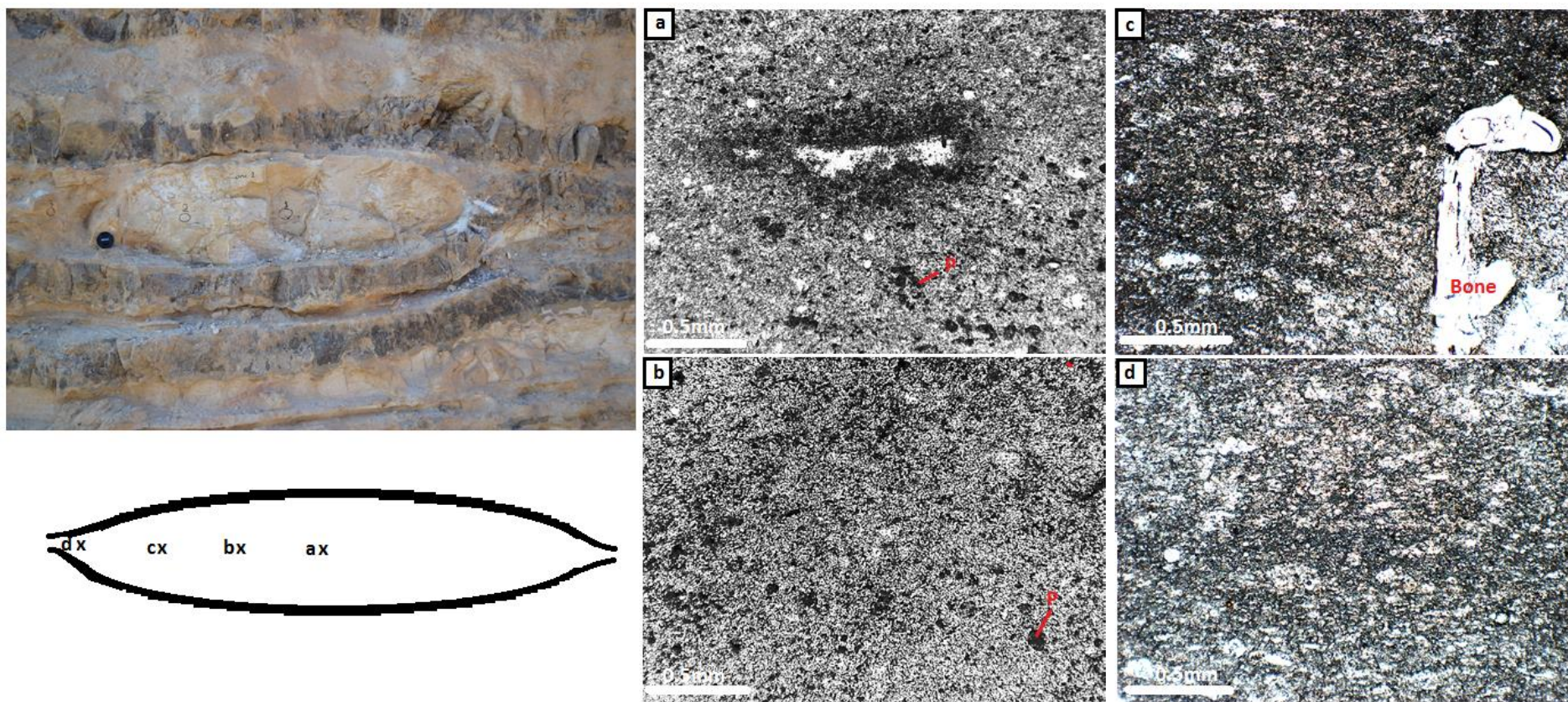


Figure 4.27 Microphotographs of systematically sampled (drawing: a, centre and d, edge) elliptical carbonate concretion showing the change in porosity grain size and amount of cement from the centre towards the margin of the concretion. P, Pore Space. Note the decrease in pore spaces from the centre towards the edge. Note the preserved bone fragment in (c) and the aligned grains in laminae patterns in (d).

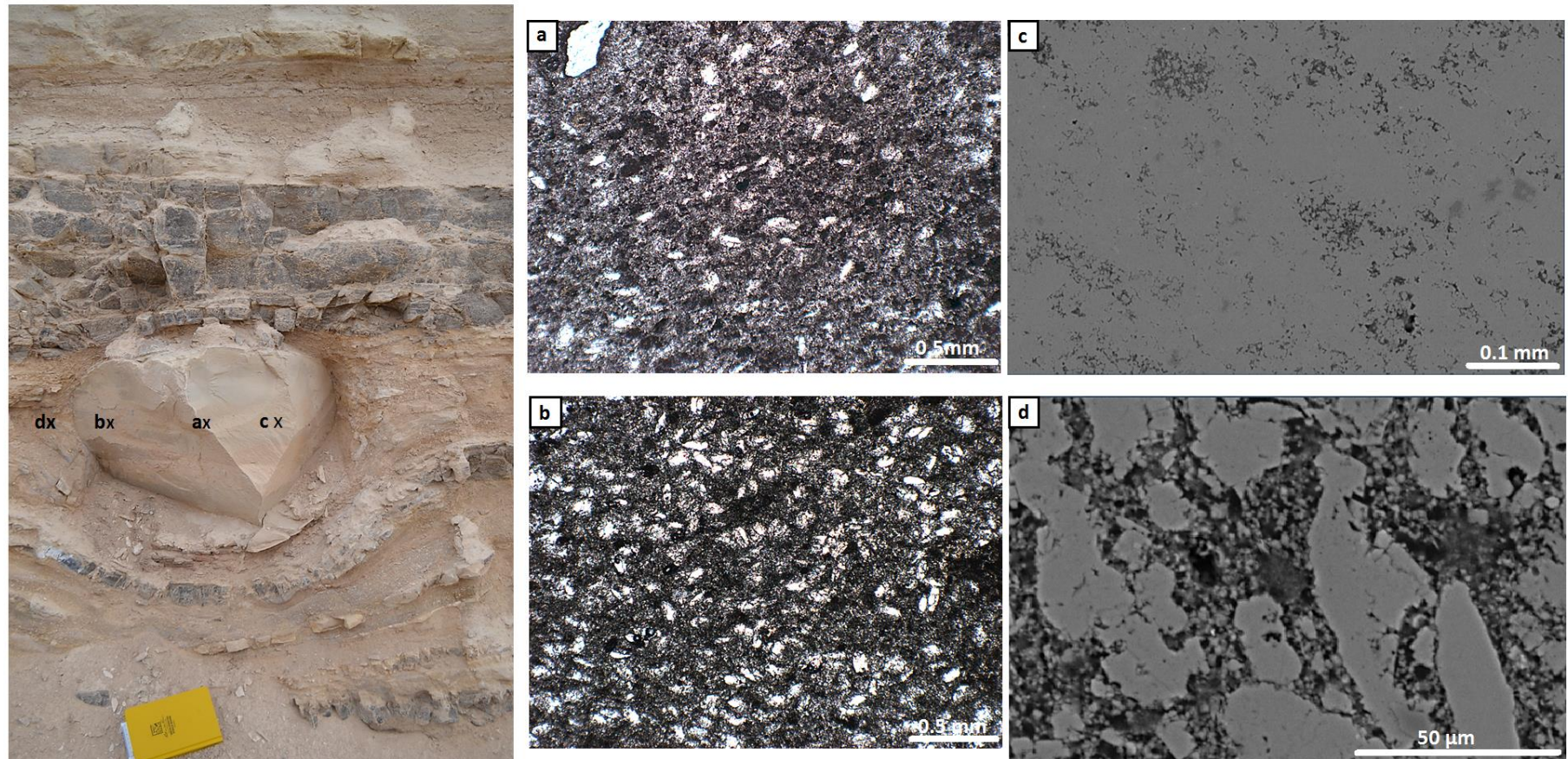


Figure 4.28 a-c. Microphotographs showing the composition of systematically sampled, asymmetrical carbonate concretion. Note the multiple directions of the randomly aligned semi fibrous calcite crystals. c. BSE image showing the texture of the concretion. d. BSE image showing the composition of the host rock that abuts the concretion.

Sample	$\delta^{13}\text{C}$ vpdb ‰	$\delta^{18}\text{O}$ vpdb ‰	Growth
1	-8.145	-9.348	Displacive?
2	-9.577	-8.299	
3	-10.582	-8.881	
Host Rock	-9.402	-10.239	
1	-6.923	-4.920	Replacive?
2	-7.032	-4.379	
3	-7.133	-4.678	
4	-6.072	-4.972	
Host Rock	-7.171	-5.057	
1	-6.831	-5.320	Replacive?
2	-7.122	-6.379	
3	-6.931	-7.678	
Host Rock	-7.250	-6.170	
Planar Calcite Vein	-10.180	-13.550	
Planar Calcite Vein	-9.131	-12.120	
Planar Calcite Vein	-11.102	-10.210	

Table 4.2 Isotope data of carbonate concretions and planar calcite veins transecting the concretions in the Qatrana roadcut. 1to 4: centre (1)-to-edge of the concretion

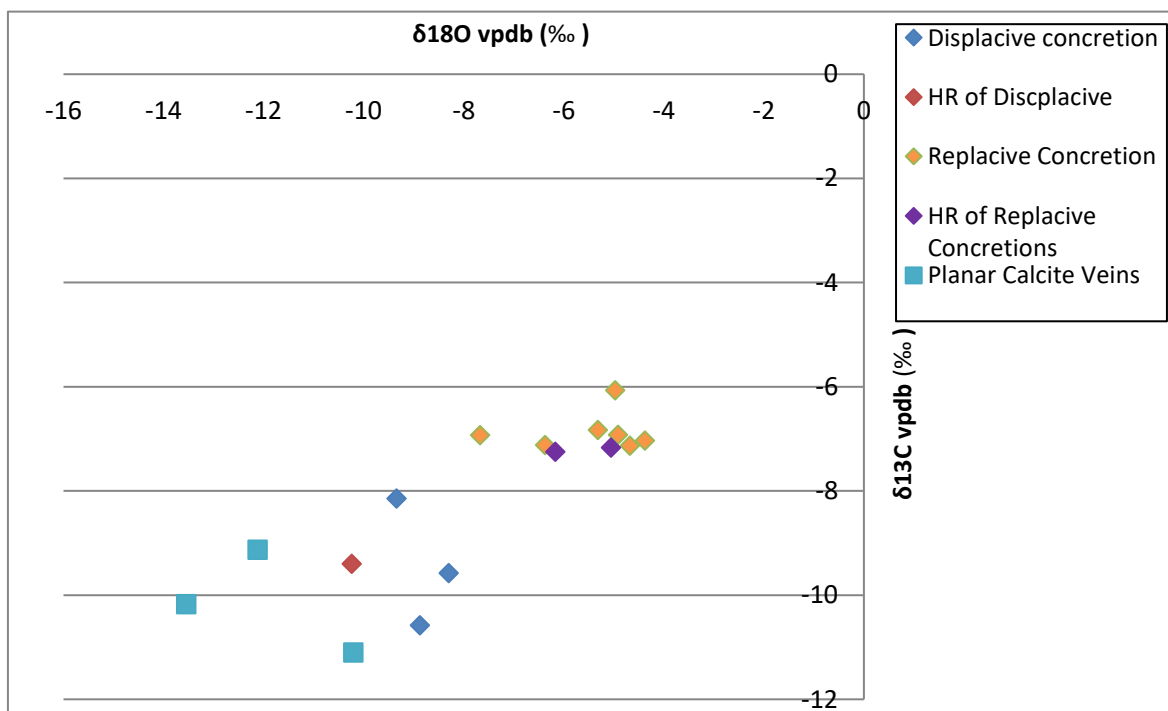
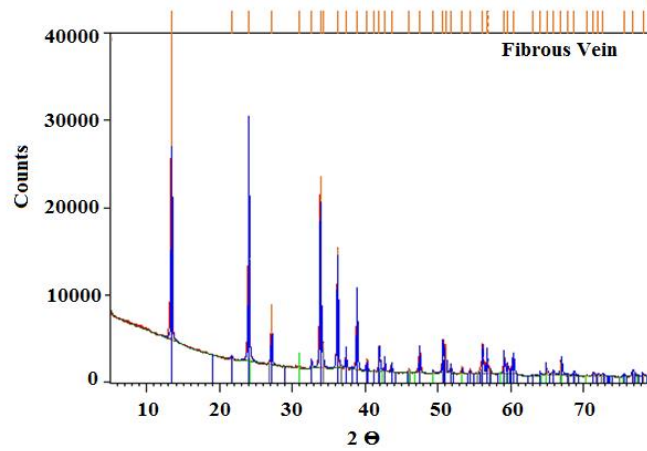
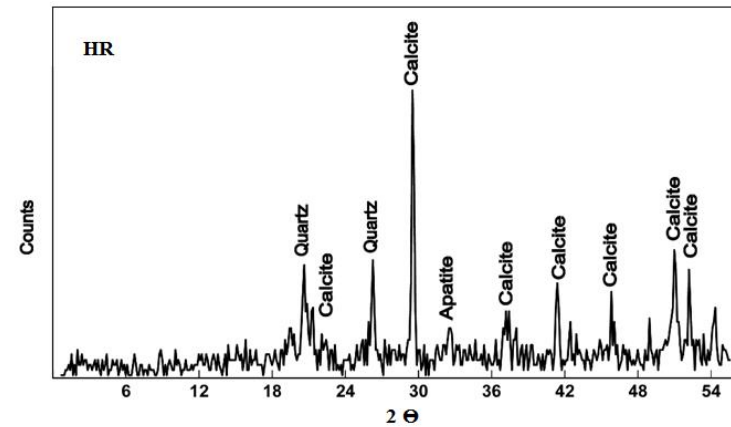


Figure 4.29 Plot of stable isotopes data of carbonate concretions and planar calcite veins transecting the concretions.



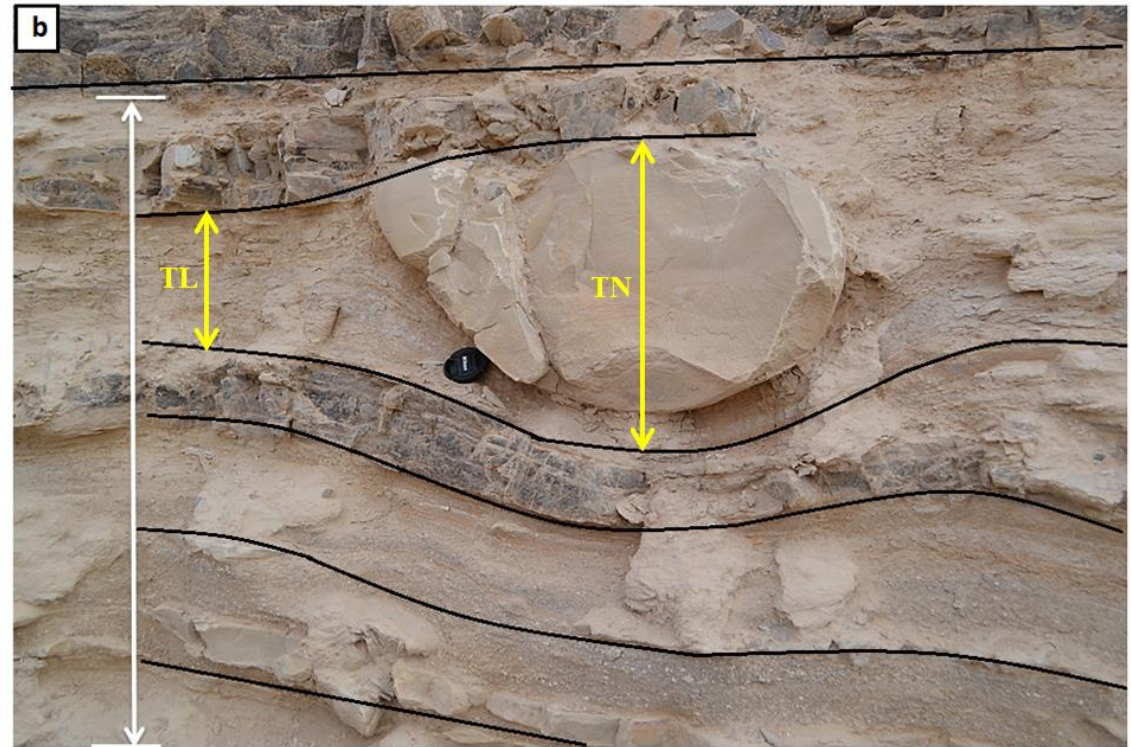
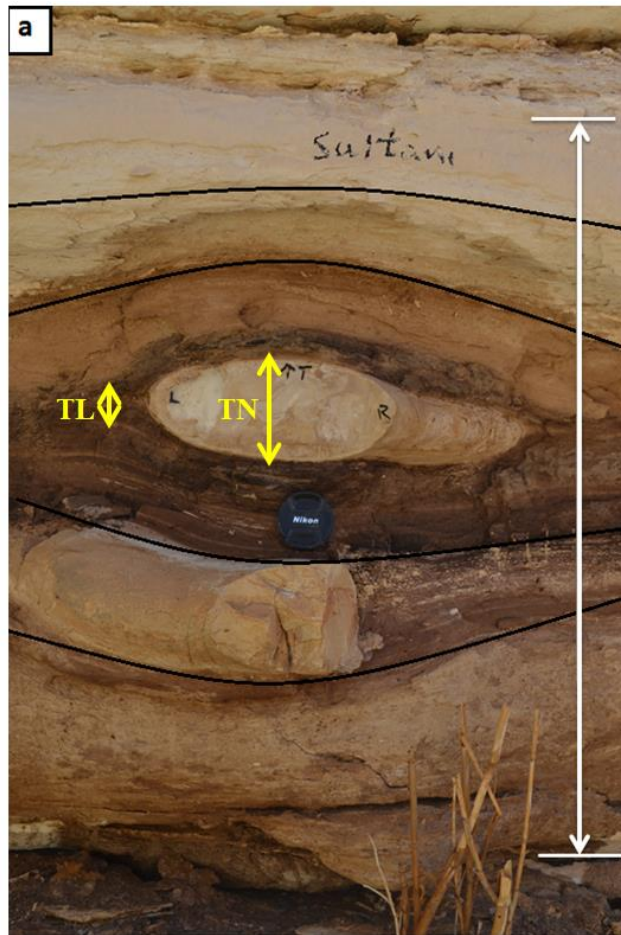


Figure 4.31 Measurement of the compaction ratio around concretions in a symmetrical strain field (a) and an asymmetrical strain field (b) in the outcrops.

Concretion Number	T _N (Chert Conc.)	T _L (Chert Conc.)	Compaction Ratio (C _N)	Concretion Number	T _N (Carbonate Conc.)	T _L (Carbonate Conc.)	Compaction Ratio (C _N)
1	14	5	2.80	1	45	15	3.00
2	17	6	2.83	2	48	17	2.82
3	3.75	2.5	1.50	3	50	14	3.57
4	15	6	2.50	4	55	18	3.06
5	14	5	2.80	5	58	20.5	2.83
6	18	7	2.57	6	46	15	3.07
7	14	5	2.80	7	48	13	3.69
8	10	3.5	2.86	8	45	13.5	3.33
9	6	2.5	2.40	9	55	19	2.89
10	5	2	2.50	10	58	19	3.05
11	5	2.5	2.00	11	46	16	2.88
				12	50	14	3.57
				13	74	25	2.96
				14	210	60	3.50
				15	71	24	2.96
				16	80	25	3.20
				17	75.5	24	3.15
				18	60.5	14	4.32
				19	40	13	3.08
				20	39	15	2.60

Table 4.3 Compaction Ratios measured around chert and carbonate concretions

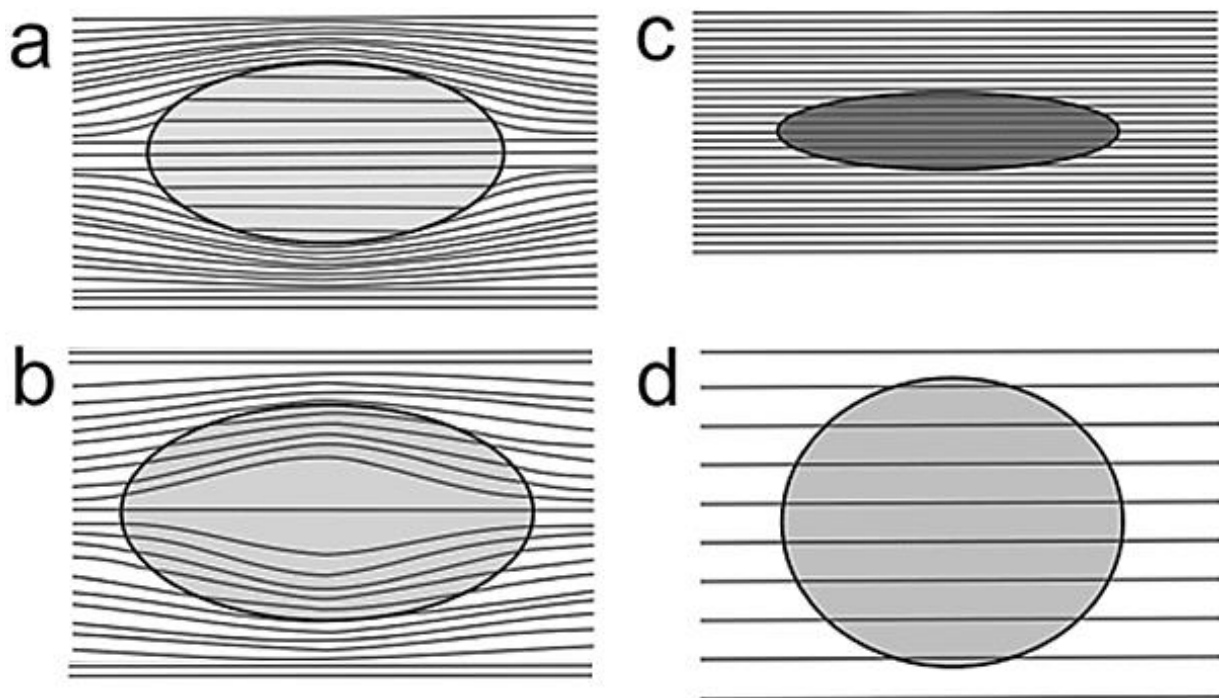


Figure 4.32 Diagram showing the different possible occurrences of concretions and their relationships to the host sediment that can allow us to work out the timing of concretion growth relative to sediment compaction. Ellipses represent the outer margin of the concretions. a. early diagenetic concretion formed before compaction. b. early diagenetic concretion which continued to grow during compaction. Note that the beds are deformed inside and outside the concretion. c. and d. represent concretions that formed after compaction. Modified from Marshall & Pirrie, 2013.

Chapter 5: The Development of Bitumen-Bearing fractures in the JOS

5.1 Introduction

Natural fractures play an important role in hydrocarbon migration and accumulation (Nelson, 1985; Mandl & Harkness, 1987; Jones et al., 1998; Aydin, 2000; Childs et al., 2009; Wang et al., 2016). Hydrocarbons in the form of solid bitumen have been documented within fractures and pores in reservoirs and overlying seal rocks. These occurrences occur on a range of scales, including bitumen in large-scale structures such as Gilsonite dikes (Verbeek & Grout, 1983, 1992), in mineralized veins (Parnell et al., 2003, 2017a, 2017b; Fink, 2016) and in microfractures (Ryder et al., 2013).

Most of the studies have focused on investigating bitumen occurrences that are hosted in stratigraphic units other than the source rock from which the bitumen was derived, and have used a variety of research methodologies including experimental work, field-based investigations and conceptual modelling (e.g., Parnell et al., 1991, 1994, 1998; Peters et al., 1996; Wilhelms et al., 2001; Wang et al., 2015). There is, however, a paucity of studies that focus on solid bitumen formation and migration within the confines of the source rock interval, and as a result, there are many unresolved questions concerning this earliest phase of hydrocarbon mobilisation (Pepper, 2017).

The prolific oil shale deposits of Late Cretaceous to Paleogene age in Jordan provide an excellent opportunity to study the earliest stages of catagenesis, and in particular the factors leading to the formation of bitumen veins, and bitumen inclusions in mineralized veins. They are identified both in outcrop and across a large area in the subsurface through drilling operations in the past 20 years. The oil shales have been studied previously for their organic matter enrichment, maturation, hydrocarbon potential and sedimentological

and palaeo-environmental implications (Abed et al., 2005, 2009; Abed & Arouri, 1994; Ali Hussein et al., 2014a, 2014b; Hakimi et al., 2016, 2018; März et al., 2016). The organic geochemical characteristics of bitumen were only studied for bitumen present within the oil shale matrix (e.g., Abed et al., 2005; Hakimi et al., 2016). Fracture-filling bitumen present in these rocks has not been previously characterized for its thermal maturity and origin.

The scope of this chapter is to examine the origin of the bitumen present within fractures and vugs and its relationships to fracturing in the Late Cretaceous-Paleogene oil shale of Jordan. The chapter has three main aims: (1) to define the source and emplacement history of the bitumen and test whether the bitumen is derived from a local source (i.e., JOS) or has migrated from a deeper petroleum source by studying its organic-geochemical characteristics; (2) to better understand the transport mechanism of the bitumen into fractures and vugs; and (3) to investigate whether bitumen maturation and migration coincide with fracturing, and if so, establish the nature of relationships between bitumen mobilisation and fracturing.

Recent studies of the natural fracture systems within the JOS have shown that diagenetic processes play an important role in fracture formation and distribution in the Cretaceous-Palaeogene strata of Jordan (Hooker et al., 2017a; Huggett et al., 2017). Hooker et al. (2017a) presented evidence of fracture formation controlled by silica diagenesis whereby fractures formed due to volume redistribution and loss associated with the opal-quartz transformation during early burial. Fractures were simultaneously filled by calcite. Their study demonstrated that chemical reactions can lead to open fracture pore space and that the created space can be filled by diagenetic mineral phases.

Bitumen is present within the oil shale matrix, natural fractures and adjacent vugs in the JOS, either in macroscopically visible deposits or as inclusions in the fracture cement. The

geostatistical analysis of fracture distribution, petrographic investigation and geochemical analyses have been carried out on samples collected from 15 continuously cored sections through the JOS and the El-Lajjun Quarry.

The contents of this chapter are organised under four major sections. The first section presents the sampling techniques, the methods used to describe bitumen-bearing fractures and the geochemical methods used to examine the fracture and matrix bitumens. The second section reports the macroscopic, microscopic and geochemical results obtained from the study of the bitumen-bearing fractures. The third section provides discussions of the source of bitumen present in the fractures and in the matrix, the bitumen generation and migration in the JOS and the development of the different type of bitumen-bearing fractures. The final conclusions of this chapter are provided in the fifth section.

5.2 Methodology

5.1 Sample Selection and Description

Bitumen was investigated within fractures, vugs and the matrix of host rocks in both core and outcrop located in central and south Jordan (see Figure 2.8, Chapter 2). The fifteen examined cores include five cores in the Harrana area, three in the Azraq area, two in the central area, two in the Sirhan area, and three in the Bayer area (see Section 3.2.1.1, Chapter 3), with a total cored interval of all cores of 3212 m. The cored intervals range in length from 137 m to 293 m per well (see Table 3.1, Chapter 3), with a core diameter of 10 cm. The frequency, aperture and length of the bitumen-filled fractures were described and measured in each core.

The quarry section provides an opportunity to analyse the horizontal distribution of bitumen-bearing fractures. Bitumen-bearing fractures and their spacings were measured

along horizontal scan lines (Priest & Hudson, 1981; Priest, 1993; Bons et al., 2004; Zeeb et al., 2013).

Thirty-eight representative samples were collected from core and outcrop (32 samples from the fifteen studied cores and 6 from the El-Lajjun Quarry) and submitted for organic-geochemical and petrographic analyses. The samples were taken from both bitumen-bearing fractures and their host matrix. At least one fracture bitumen (FB) sample from each core was selected along with a matrix bitumen (MB) sample from the host organic-rich mudrock. If bitumen-bearing fractures occurred in chert or carbonate beds or concretions, matrix samples were taken from the adjacent mudstone beds (a few centimeters away from concretions). The average vertical distance between fracture and matrix bitumen samples is ~2 cm (see Appendices).

Thin sections with a thickness of 30 µm were prepared from the fracture and matrix samples to study the bitumen present within the fractures and host rock. Optical analyses of thin sections of selected samples were made in cross-polarized light with a Nikon Optophot Microscope linked to a digital camera. The thin sections were also examined using a FEI Quanta 650 FEG scanning electron microscope. BSE images were created to study mineral compositions and textures of the samples. Elemental mapping was performed using an EDS attachment to the SEM. All SEM analyses were performed at 20 kV and a working distance of about 10 mm.

5.2 Organic Geochemistry

Geochemical analyses of the matrix samples, including TOC and Total Sulphur (TS), were performed using a LECO elemental analyser. Rock Eval pyrolysis was used to characterize the type and maturity of the organic matter using a Vinci Rock-Eval 6 Analyser.

Twenty-four samples (22 from cores and 2 from the Quarry) were selected for extract analyses. Special care was taken during preparing the samples to avoid any contamination. The Soxhlet-extractor glassware was put in the oven at 450 °C to be sterilised before using. Rock samples were prepared by rinsing with distilled water and dichloromethane (DCM). The samples were drilled out of fractures/vugs and rock matrix using a microdrill. Pulverized samples were then weighed and loaded into thimbles. The thimbles were previously cleaned by DCM and left to dry. The extractable organic matter (bitumen) was then extracted from samples using a standard Soxhlet extraction technique (Mueller & Philp 1998). A 3:1 mixture of DCM and Methanol (CH₃OH; abbreviated here MeOH) was prepared in 250 ml volumetric flasks that were then fixed into the base of the Soxhlet Extractors (Figure 5.1a). The thimble containing the pulverized and weighed sample were loaded into an extraction chamber fitted by adaptors to a distillation flask from the bottom and to a condenser from the top (Figure 5.1b). The Soxhlet Extractors were run for 24 hrs. The extracts were then blown down under N₂ gas. Asphaltenes, which represent a substantial fraction of the extract, were precipitated from the whole extract using an excess of hexane (Aske et al., 2001). The samples were then centrifuged at 5000 rpm, and the supernatant solution of each sample, representing the maltene fraction, was carefully transferred into a separate glass vial and blown down to about 0.5 ml by N₂ gas. Elemental Sulphur was removed using activated copper prior to column chromatography.

The extracts were passed through a column containing silica gel to remove any remaining water, before being separated by alumina (Al₂O₃) column chromatography (Figure 5.1c). The extract sample was eluted sequentially by hexane/DCM (9:1, v/v) and DCM/MeOH (1:1, v/v) to separate aliphatic and aromatic fractions respectively.

Biomarkers from the maltene fraction were analysed using an Agilent 5977B GCMS. The GC oven program was set to 80 °C for 1 min, then increased to 130 °C at 20 °C/min, and

then to a maximum temperature of 325 °C at 10 °C/min which was held for 30 min. Triterpane and sterane biomarkers were identified from the maltene fraction with the standard methods using m/z 191 and m/z 217 fragmentograms run in SIM mode. The biomarker results were compared to an internal standard (SRM2266). Identification of biomarkers was carried out based on the retention time.

The complete Soxhlet extraction procedure is shown in Figure 5.2.

5.3 Results

5.3.1 Fracture Description

5.3.1.1 Fracture Types and Mineralization

Bitumen is either present within the mineralized veins, or entirely fills fractures and vugs. In the mineralized veins, bitumen was observed to be only present within vertical planar calcite veins and bedding-parallel calcite veins (termed here BCVs) (Figure 5.3a and f); no observations of bitumen were documented within folded calcite veins. Fractures entirely filled by bitumen are referred to here as bitumen veins (BVs) (Figure 5.3b-e and g-h). Bitumen in unfolded calcite veins (planar veins) is observed either as solid inclusions within the cement or as fillings of the remaining porosity and inter-crystalline pore space (Figure 5.3a and f). No crosscutting relationships between calcite veins, that lack the presence of bitumen, and bitumen veins were documented.

Bitumen is not found in silica-cemented veins, but BVs are observed to crosscut silica veins (Figures 5.3h and 5.4a-b), and some BVs exploit the preexisting silica veins by running immediately adjacent and parallel to silica veins (Figure 5.4a-b). These observations establish that the timing of BVs post-dates the silica veins.

There are many examples in the cores of horizontal BCVs crosscutting early folded calcite veins (Figure 5.4f-g), but BCVs and BVs have not been observed crosscutting each other.

Importantly, no systematic crosscutting relationships are observed between horizontal BVs and vertical BVs. This suggests that all geometries and types of bitumen-bearing fractures are coeval.

There are only a few observations of bitumen inclusions associated with shear fractures (faults), such as observed in the lower unit of the core from well C7 at a depth of 667 m. These few cases mainly represent bitumen that occurs within the fault cement or in opening-mode fractures that are associated with fault damage zones.

Fractured beds are predominantly those cemented by carbonate and silicate minerals (Huggett et al., 2017). Fracture cements generally reflect the lithology of the host unit such that calcite veins are observed in calcareous mudstones and limestones whereas chalcedony and microcrystalline veins are present in silicified and chert beds.

In the quarry section, only BVs are observed, and they are confined to their host oil shale beds. No mineralized veins containing any form of bitumen were observed in the quarry (i.e., no BCVs).

5.3.1.2 Fracture Geometry

The majority (>70%) of the bitumen-bearing fractures (BVs and BCVs) observed in the cores are sub-vertical to vertical (Figures 5.3a, d-f and 5.4c-e). The dip angles of most vertical bitumen-bearing fractures observed in the core vary with the change in lithology (Figures 5.3a and 5.4e). Due to the unknown orientation of the core, the true fracture strike was not constrained in this study. Horizontal fractures (bedding-parallel) and a combination of horizontal and vertical fractures forming complex fracture networks also occur (Figures 5.3c, h and 5.4f-g).

Vertical bitumen-bearing fractures in the cores range from 1 to 120 cm in height (parallel to the core axis), whereas the length of bed-parallel fractures (orthogonal to core axis) is

not measurable due to the limited core width. Vertical core fractures measured within the homogenous organic-rich marl unit (Middle Unit), have smaller apertures and greater heights than fractures within the heterogeneous units (Lower and Upper), which are commonly restricted vertically by chert and silicified bands. BVs transect their host beds or concretions entirely (Figures 5.3d-e, g-h, and 5.4a-e).

BVs are less strictly planar (more irregular) than other mineralized veins (i.e., calcite veins and silica veins) and bitumen-bearing fractures (i.e., BCVs) (Figures 5.3 and 5.4d-e). Bitumen-bearing fractures show a range of tip morphologies varying from single tapering ends to branching ends (Figures 5.3 and 5.3). Individual thick BVs commonly branch, particularly within mudstone beds (Figure 5.4e). BVs tend to have larger apertures in the chert, silicified and carbonate beds or concretions than in the mudstone (Figures 5.3d-e, g-h and 5.4d-e). Both BCVs and BVs tend to deflect in their trajectory when crossing different lithofacies (Figure 5.1a, d, and 5.2e).

Generally, BVs have larger aperture sizes than other bitumen-bearing fractures (Figure 5.4d-e). The maximum aperture size of BVs averages 0.90 cm while it averages 0.40 cm for BCVs. BVs display apertures that are variable along fracture traces (Figure 5.4d-e). In all types of bitumen-bearing fractures, fracture aperture decreases in a fairly regular taper towards the tips (Figure 5.3a-h and 5.4a-c).

BVs observed in the El-Lajjun quarry are planar and exclusively vertical/subvertical (Figure 5.4c). They range in height from 5 cm to 20 cm and have a mean maximum aperture of 0.20 cm. The veins in outcrop do not exhibit any branching features that are comparable to the BVs observed in the cores.

5.3.1.3 Fracture Abundance and Distribution

Bitumen-bearing fractures are found dominantly in the lower section of in all cores, which comprises organic-rich mudstones interbedded with silicified units or chert beds or

concretions (Figures 5.5 and 5.6). The majority of bitumen-bearing fractures in the Lower Unit are confined to silicified, chert and carbonate intervals positioned between unfractured, organic-rich mudstone beds, particularly those mudstone beds with higher contents of TOC and S (Figure 5.6; Table 5.1).

The frequencies of all bitumen-bearing fractures were normalized to the length of recovered core (Figures 5.5 and 5.6). The data reflect high intensities of bitumen-filled fractures and vugs with the progressive increase of the present-day depth (Figures 5.6, 5.7 and 5.8). A few BVs (5 % of total bitumen fractures) are observed in the Middle Unit in the homogenous organic-rich mudstone interval that lacks chert.

Bitumen-bearing fractures are more abundant in the Azraq cores (CE Jordan) than the Harrana cores (CW Jordan) (Figures 5.5, 5.7 and 5.8). This increase in abundance eastwards may reflect the greater maximum burial depth (Figure 5.8). The frequency of bitumen fractures in the Lower Unit of the core in the East is more than double the frequency in the same unit in the West. In the South, bitumen-bearing fractures are less abundant; this is reflected by the frequency of these fractures in the cores from the wells C12-C15 (Figure 5.5).

Generally, BVs are highly interconnected in the core; it is difficult to find individual BVs that do not intersect others (Figures 5.3b, d, g-h and 5.4e). Other bitumen-bearing fractures are commonly observed as singular fractures (Figures 5.3a, e and 5.4d). Both horizontal and vertical bitumen-bearing fractures are observed within the Lower Unit; however, horizontal fractures are less abundant than vertical fractures and 90% of them are only filled by bitumen (as BVs).

BVs are closely spaced in the core and tend to cluster. This clustering increases the chance of observing them within the limited core width and may explain the high intensity of BVs

compared to other bitumen-bearing fractures that might have a wider fracture spacing that exceeds the core width.

Bitumen fractures are also observed in the outcrop at the El-Lajjun quarry within the lower part of the MCM formation confined to a 20-cm thick organic-rich oil shale bed (Figure 5.4c). The quarry allowed a 3D visualization of bitumen-bearing fractures and their horizontal distribution. The fractures heights are mostly equal to the bed thickness. The average fracture spacing of the measured fractures is 12 cm.

5.3.1.4 Bitumen-Bearing Fractures Associated with Concretions

An interesting observation regarding the occurrence of the bitumen fractures in the JOS is that anomalously low organic content mudstone zones (see sections 5.3.2.3 and 5.3.3.3) are usually seen in a contact zone with those chert concretions that host bitumen-bearing fractures, particularly BVs (Figure 5.9). Such zones have never been previously described and are termed here informally as ‘halos’. Halos can be described here as discoloured mudstone zones with a halo-shaped like and are present at the rims of chert concretions. 90% of the observed halos are associated with chert concretions (Figure 5.9). No observations have been reported from the cores or outcrops of the presence of any halos associated with carbonate concretions. This is exemplified by the co-occurrence of chert and carbonate concretions in a single organic-rich mudstone bed, in which the chert concretion is surrounded by a mudstone halo, whereas the carbonate concretion has no such zone (i.e., halo) surrounding the concretion (Figure 5.3d). Halos are stratigraphically clustered in the Lower Unit of the study succession of Jordan, where the organic-rich mudstone interbedded with chert intervals.

Some halos are seen to be completely enclosed by irregular chert concretions, and separated from their host organic-rich mudstone (Figure 5.9c). In such cases, the halos normally take the shape of the adjacent chert concretions. In contrast, symmetrical chert

concretions are usually surrounded by halos with regular shapes (Figure 5.9a-b). This observation strongly suggests that the halos may have resulted from alteration caused by chert formation (i.e., reaction rims or haloes).

Halos are always observed to contain or to be associated with bitumen veins (BVs), but not with mineralised veins containing bitumen (i.e., BCVs). 94% of BVs are <4 mm away from the nearest halo (Figure 5.10). 60% of the BVs are present within the halos (Figure 5.10). BVs are well developed in these halos and are normally propagating into the adjacent chert concretions (Figure 5.9). A possible hypothesis is that chert formation has altered the local sediment chemistry and/or pore water temperature, thereby promoting early bitumen generation through mobilisation of interstitial organic matter (see Discussion, Section 5.4.3).

Bitumen veins propagate from the halos mainly towards the underlying chert concretions, but not upwards into the host mudrocks (Figure 5.9a). There is a strong correlation between the colour intensity of halos and the aperture of the hosted or adjacent bitumen veins; the halos are much lighter in colour wherever there is a bitumen vein with a large aperture (Figure 5.9a).

5.3.2 Microscopic Description

Petrographic examination revealed a wide variety of host rock lithologies, fracture cements, textures and crosscutting relationships. This demonstrates a complex history of deposition, fracturing and sealing. It also allowed characterisation of the microscopic occurrences of bitumen and its migration pathways.

5.3.2.1 Host Rock

Petrographic investigation of both the lower unit, which is the host of most of the bitumen-bearing fractures, and the middle unit shows that the mudstone mainly consists of calcite,

quartz, organic matter (bitumen) and apatite, with locally considerable amounts of chalcedony, dolomite, and occasionally pyrite (see Figure 3.37, Chapter 3). Clay is only rarely observed and quartz is commonly of microcrystalline nature. Calcite has cemented most microfossils and radiolaria are locally abundant.

The interbedding of chert and mudstone in the lower unit is believed to be the result of diagenetic alteration of locally abundant concentrations of radiolaria and diatoms (Hooker et al., 2017a). The chert is believed to have developed by replacing carbonate components in the original host material with silica from biogenic components (Hooker et al. 2017; Huggett et al. 2017). This is consistent with the low silica content of the adjacent mudstones.

Bitumen is observed as small amorphous particles in variable shapes and as flakes that are present in the matrix of the organic-rich mudstones (oil shale) in both the lower and middle units (Figure 5.11c-d).

5.3.2.2 Bitumen-Bearing Fractures

This section focuses on describing the petrographic properties of bitumen-bearing fractures (mainly BVs and BCVs). Detailed petrographic descriptions of all other fractures observed in the cores and outcrops, are provided in Chapter 3 and Chapter 4 (under the sections 3.4.1.6 and 4.5, respectively).

Fracture bitumen under the microscope is observed to either entirely fill the fracture (i.e., bitumen in BVs; Figure 5.11c and f), or to be texturally enclosed by calcite cements (i.e., in BCVs; Figure 5.11a). BCVs are characterized by a crack-seal texture (e.g., Ramsay 1980), representing multiple generations of calcite crystals with elongate-blocky morphology (Figure 5.11a). The veins appear under the microscope as multiple parallel bands of cements. The crack-seal texture indicates that calcite precipitation was synchronous with fracture opening. Micro-bitumen inclusions (sub-millimetre-scale) and

inclusions of organic-rich host-rock are present within the cement of these calcite veins (Figure 5.11a-b) which indicates that bitumen was present during formation of the veins (c.f. Parnell and Carey, 1995). Bitumen filling the vacant segments of partially-filled BCVs appears under the microscope to have similar petrographic properties to the bitumen within the calcite cement.

Bitumen micro-veins (BMVs) are observed in internally branching networks, resulting in incipient brecciation of the organic-rich mudstone (Figure 5.11f-g). Horizontal and sub-horizontal microfractures entirely filled by bitumen are also commonly observed within the organic-rich mudstone (Figure 5.11c). This microfracturing is observed particularly around or at the margin of calcite and pyrite crystals (Figure 5.11c-d).

Microscopic examination of folded calcite veins shows that these early veins do not contain any traces of bitumen but are entirely cemented by calcite. This indicates that bitumen generation and migration postdate the development of folded veins. It also strongly supports the macroscopic observations described above from the core.

BVs exhibit a clear crosscutting relationship with silica veins at the microscale but not with BCVs. Importantly, the silica cement in silica veins is entirely free of bitumen inclusions. BMVs with varied directions were observed, including vertical BMVs, horizontal BMVs and network BMVs. However, no crosscutting relationships between horizontal and vertical BMVs were detected.

5.3.2.3 Low-Organic Zones (Halos)

Microscopic examination of the halos, low-organic content zones around chert concretions, shows that the boundary between the halo and the host organic-rich mudstone is gradational (Figure 5.12). Bitumen is present as filaments within both the host mudstone and at the boundary with the halo (Figure 5.12).

Within the boundary zone, there is a progressive increase in the size of bitumen filaments, and they appear to connect towards the halo (Figure 5.12). BVs nucleate within the halos and develop towards the chert concretion (Figure 5.12). Well-developed BVs are only seen in the halo and in the chert concretion; no BVs were seen in the adjacent organic-rich mudstone hosting the halos and chert concretions.

EDS maps show high organic carbon content in the pores within the host mudstone, whereas it is more concentrated within micro veins in the halos (Figures 5.13 and 5.14). The halos therefore represent regions with anomalously low bitumen content in the matrix. Within the transitional zone, the zone between the halo and the host mudstone, carbon content is high. Silica content shows a gradual decrease from the chert concretion towards the halos (Figures 5.13 and 5.14), indicating that the progressive silicification is somehow related to the development of bitumen veins (see Discussion, Section 5.4.3.3).

5.3.3 Organic Geochemistry

5.3.3.1 Source Rock Characterization

Geochemical properties of the JOS samples are presented in Table 5.2. The JOS has high TOC values ranging from 6 to 30 wt% and S contents from 1.5 to 8.3 wt%. Hydrogen Index (HI) values range from 583 to 880 mg HC/g TOC, indicating excellent oil-prone kerogen. The Tmax values range from 389 to 424 °C (<0.6% VRE), and the Production Index (PI) has an average of about 0.03 mg/g TOC, both parameters indicating a thermally immature source rock (Figure 5.15). The low maturity of the JOS indicates that regional maximum paleo-burial depths were likely shallow (few hundred metres), or more accurately, paleo-burial temperatures were low (~ 30-50 °C).

5.3.3.2 Biomarkers of Fracture and Matrix Bitumens

Table 5.3 lists pertinent characteristics of the m/z 191 (Triterpenoids) and 217 (Steranes) fragmentograms from extracts of both fracture (FB) and matrix bitumen (MB). The most prominent feature of the triterpane fragmentogram is the high concentration of gammacerane (Figure 5.16a-b), which is an indicator of water column stratification possibly due to salinity variations (Mello et al., 1988; Sinninghe-Damste et al., 1995, Peters et al., 2005). C29:C30 $\alpha\beta$ hopane ratios range from 0.27-0.60 and gammacerane: C30 $\alpha\beta$ hopane ratios from 0.34-0.93. The 22S/(22S+R) C₃₁ $\alpha\beta$ extended-hopane ratio ranges from very immature values of 0.2 to the equilibrium value of 0.6 (Peters et al., 2005).

Sterane biomarkers are commonly used to indicate depositional environments and organic matter biomass input (Huang & Meinschein, 1979; Seifert & Moldowan, 1981; Volkman, 1986; Peters et al., 2005). Table 5.3 lists the relative amounts of the C27, C28 and C29 $\alpha\alpha\alpha$ R steranes, which are the predominant isomers in all the samples. The samples all show a predominance of C27 steranes (47-58%), with subequal amounts of C28 and C29 steranes comprising the rest (Figures 5.16c-d, and 5.17). All the samples, both matrix and bitumen, cluster in a relatively small area in Figure 5.17, indicating very similar biomass contributions to the kerogen of the JOS (Huang & Meinschein, 1979; Seifert & Moldowan, 1981; Volkman, 1986; Peters et al., 2005).

20S/(20S+20R) $\alpha\alpha\alpha$ and the $\alpha\beta\beta$ /($\alpha\beta\beta$ + $\alpha\alpha\alpha$) C28 sterane ratios range from 0.25-0.52 and 0.41-0.66 respectively (Table 5.3; Figure 5.18a). Most of the samples have yet to reach the equilibrium maturity values for these maturity parameters of 0.55 and 0.7, respectively, and attest to their low maturity.

5.3.3.3 Halos Geochemistry

The geochemical analysis shows that the halo matrix differs from the host mudstone matrix in the lower TOC values. RE data show low TOC values (average value 6.5 wt %) in the halos compared to the higher TOC values of the host mudstone (average value 14 wt %) (Table 5.4; Figure 5.19). This is consistent with the petrographic examination of the halos and the host organic-rich mudstone and supports the hypothesis that the halo matrix was originally saturated with organic matter before the formation of the chert concretion that forced the surrounding host mudstone (halo) to lose its organic-carbon content (Abu-Mahfouz et al., 2019b), leading to nucleate BVs (See Discussion, Section 5.4.2). The decreased TOC values in the halos indicates a lack of compaction due to silica cementation relative to the marls.

5.4 Discussion

The discussion focuses on the source and maturity of bitumen, bitumen generation and migration, and formation mechanism of bitumen fractures. It is based on the multi-scale observations, including field, core and petrographic observations and the geochemical data (TOC, S, RockEval, and biomarkers) of bitumen-bearing fractures and their host rocks.

5.4.1 Source and Maturity of Fracture Bitumens

The JOS bituminous and organic rich marls comprise an excellent oil prone source rock. The presence of Type I/II-S kerogens indicates high quality organic matter type that is able to generate low API gravity, asphaltene-rich bitumen at relatively low thermal stress because of the high sulphur (1.5-12 wt%) contents (Burnham, 2017).

The biomarker data obtained from the matrix and fracture bitumens from the JOS indicate that they are sourced from the same type of organic matter. Subtle variations in biomarker

parameters (e.g., C29/C30 hopane, gammacerane/hopane and the sterane carbon number distributions) between wells probably reflect slight paleogeographic differences in environment of deposition and biomass input. Although a detailed discussion of the formation of these source rocks is beyond the scope of this paper, the pertinent point is that there is a close similarity in the biomarker patterns of the matrix bitumen, and the solid bitumen found in fractures and vugs (Figures 5.16 and 5.18). Most compelling are the maturity-related hopane and sterane isomerization parameters (e.g., hopane 22S/(22S+22R), and sterane 20S/20S-20R) and $\alpha\beta\beta/(\alpha\beta\beta+\alpha\alpha\alpha)$ parameters) that are remarkably similar between matrix and fracture bitumen and indicate low thermal levels of stress, which is also consistent with bulk Tmax values that point to low thermal maturity (0.5-0.6% VRE). These lines of evidence strongly suggest that both matrix and fracture bitumen is locally derived and indigenous to the JOS and has migrated at most on the scale of a few beds, i.e., cm-scale to m-scale migration. This applies to both fractures and vugs hosted in the bituminous marls and to those that extend into adjacent chert and carbonate beds which inherently have little or no organic matter.

Samples from Core 6 and the quarry show the lowest levels of maturity for the matrix and fracture/vug bitumens (Figure 5.18a), suggesting the least amount of burial/thermal stress for these localities. This is consistent with their shallow present day burial depth, and likely maximum burial depth relative to their counterparts in the eastern part of the basin (Figure 5.7).

5.4.2 Bitumen Generation and Migration

The elevated TOC and sulphur contents of the JOS are considered here to be the prime factor in the relatively early mobilisation of a bitumen phase. Although the JOS is immature by conventional standards of thermal maturity, the high total content of sulphur

up to 8 wt% analysed in the present study (Table 5.2) and up to 12 wt% in other studies (e.g., Al Aljariri Alhesan, 2018), is interpreted to have contributed to early cracking of the Type I/II-S kerogen, resulting in the generation of bitumen (Burnham, 2017).

Hydrocarbon migration out of the source rock is usually considered as the explanation for the presence of solid bitumen in veins (Parnell & Carey, 1995; Fink et al., 2016). In the case of the JOS, the presence of solid bitumen within matrix, fractures and vugs within the main source interval is attributed to very short distance (mm- to m-scale) hydrocarbon migration through fractures and vugs. This is evidenced by the strongly correlated biomarker data of fracture bitumen and matrix bitumen.

The early-generated bitumen appears to have been squeezed out and mobilised not by buoyancy forces (as with some views of primary hydrocarbon migration), but under the vertical load imposed during peak burial. Following the recent model of Pepper (2017), it is suggested that the transformation from the kerogen to the bitumen phase generated an overpressure (i.e., by fluid volume increase) that was equivalent to the overburden stress and enabled the bitumen to migrate via newly propagating fracture pathways through the host organic-rich mudstone beds. This process is different from the conventional notion of primary migration where hydrocarbons saturate available pore spaces before being expelled into carrier beds. Instead, these rocks have extruded early-generated heavy bitumen under the overburden load and led to redistribution into vugs and BCVs or newly-forming fractures. The presence of early-generated bitumen in the pores of the low-permeability mudrocks has built up overpressure that triggered the development of new fractures (BVs).

The difference in physical and mechanical rock properties of the chert, carbonate and mudstone beds in the host source rock interval can be viewed as a form of mechanical stratigraphy (Gross et al., 1995). This interbedding of contrasting rock types may have

exerted a second order control on the bitumen migration in that local stress concentrations (e.g., around chert and carbonate concretions) may have promoted fracturing at bed boundaries that could be exploited by the mobilised bitumen (Bourne, 2003). It is also likely that the overpressure within the bitumen phase was instrumental in driving fracture propagation, as indeed implied by the prevalence of both vertical and horizontal bitumen veins (cf. Mandl & Harkness, 1987).

Following expulsion of bitumen by mechanical compaction when the load-bearing capacity of the source rocks was exceeded, it is suggested that the resulting volume decrease of the rocks caused overpressure that allowed the bitumen to migrate only short (cm-scale) distances from the saturated mudstone (oil shale) beds towards the adjacent silicified, chert and carbonate intervals via networks of newly propagating bitumen-bearing fractures (Figures 5.4d-f and 5.20). This pressure release mechanism is considered as an efficient process of bitumen expulsion resulting from mechanical compaction once the bitumen phase achieved load bearing capacity. It indicates that the compaction-related pressure has allowed the migration of the low-maturity bitumen generated from the local kerogen and squeezed it, creating fractures (Figure 5.20) (c.f. Parnell, 2017b).

Furthermore, the strongly comparable geochemical signatures of the fracture and matrix bitumen are a robust indication of local migration (cm-scale) of bitumen from the saturated mudstone (organic-rich oil shale) towards the brittle chert and carbonate intervals (Figures 5.15, 5.17 and 5.18). Petrographic observations (Figure 5.11f) show that the direction of fracture propagation is from the organic-rich mudstone beds towards the chert or carbonate concretions. The pathways taken by migrating bitumen from outside carbonate concretions directed inward to their interiors is also evident in microscopic observations (Figure 5.11f). The presence of the low-organic zone (halos) around some chert concretions in the Lower Unit of the section suggests that there was another important factor that affected bitumen

generation and mobilization. This factor is represented by the diagenetic alteration of the mudstone zones surrounding the chert concretions that are usually associated with the presence of BVs (Figure 5.9). These halos are interpreted here to have been formed as a result of the chemical reaction created by the formation of chert concretions (reaction rims) in the local organic-rich mudstone. The relatively low TOC values (average value 6.5 wt %) compared to the high TOC values of the host mudstone (average value 14 wt %) (Table 5.4; Figure 5.19), indicates that the halos are zones of transformation within the organic mudstone that hosts the chert concretions, where bitumen forms BVs. Supporting evidence comes from the petrographic investigation that shows that the halos have silica contents that are lower than that of the chert concretions and higher than that of the host organic-rich mudstone, and carbonate contents that are lower than that of the host mudstone and much higher than that in the chert concretion (Figures 5.13 and 5.14). This indicates that silica diagenesis during the basin history has almost certainly contributed to bitumen generation and mobilization.

To summarise, both mechanical and chemical processes are deduced to have controlled bitumen generation and migration in the JOS during the burial and uplift history. Most bitumen fracture fills occurred during peak burial, but some of the organic matter may have been mobilised during concretion formation at shallower depths (e.g., low-organic zones; halos).

5.4.3 Development of Bitumen-bearing Fractures

5.4.3.1 Distribution of Bitumen-Bearing Fractures

The vertical distribution of bitumen-bearing fractures in the cores is influenced by two main factors. In the Lower Unit of all cored successions, the fracture frequency shows a positive correlation with TOC and TS (Figure 5.6). This observation is consistent with the

model presented in the previous section for local bitumen generation and mobilisation during early catagenesis since the fracture frequency would be expected to be greatest where the concentration of kerogen is highest. It is interesting to observe that fracture frequency also correlates with TS, which is consistent with the arguments presented above that it is the high sulphur contents that allows very early bitumen generation.

The second factor is the gross burial history and thermal stress that the source interval experienced. The high frequency of bitumen-bearing fractures in the Lower Unit of the cores when compared to the overlying two units (Middle and Upper units) indicates that burial depth plays a role in the development of these fractures. The overall increase in fracture intensity of the lower fractured interval (Lower Unit) from west to east (Figures 5.5, 5.7 and 5.8) is another indicator that increased maximum burial in the east and enhanced thermal stress and peak temperatures play a role in fracture formation.

At the bed scale, bitumen-bearing fractures are mainly observed clustering in the chert and silicified beds, positioned between or beneath organic-rich beds in the lower unit. This predominance of bitumen-bearing fractures in layers with high rock stiffness indicates that the development of bitumen fractures has been primarily controlled by the physical and mechanical properties of the rock, represented by the small-scale mechanical stratigraphy. Rapid vertical changes in Poisson's ratio have been suggested as a means for localising opening-mode fracture development (Teufel & Clark, 1984; McDermott et al., 2013), and such a mechanism could apply here. The increased fracture intensity in the Lower Unit from west to east can be also related to lithofacies changes with a higher percentage of chert-rich beds (i.e. greater mechanical fracturing where chert is predominant).

Fracture analysis of the outcrop analogue provided some constraints on possible horizontal distribution of BVs in the middle unit, which in the core has very low fracture abundance. The average fracture spacing at outcrop was c. 12 cm which is almost double the typical

core diameter, and thus fracture frequency in the middle unit from core observation is certainly significantly underestimated (cf. Alfahmi, 2018).

5.4.3.2 Timing of Bitumen-Bearing Fractures

Cross-cutting relations are critical for constraining the timing of fractures (Beaudoin et al., 2015; Hooker et al., 2018). The lack of systematic cross-cutting relationships between the different types of bitumen-bearing fractures leads to the conclusion that these fractures were developing coevally during the period of active bitumen generation. BCVs and BVs seem to have crosscut all other type of fractures except for late calcite veins; suggesting that BCVs and BVs postdate all opening-mode fractures (i.e., silica veins and folded calcite veins) present in the studied succession and are coeval with late calcite veins (Figures 5.3h and 5.4a-b, f-g). The crosscutting relationship between vertical BVs and silica veins indicates that both the BCVs and BVs postdate silica diagenesis. The absence of any crosscutting relationships between vertical and horizontal BVs in the core and at the microscale implies a synchronicity in the development of both the horizontal and vertical BVs, and this in turn points to the development of a lithostatic overpressure in the bitumen phase (Mandl & Harkness, 1987).

At the microscale, the presence of bitumen inclusions within the cement of the planar calcite veins (i.e., BCVs), indicates that bitumen was present during the opening and sealing of the vein. This suggests that bitumen migration coincides with the formation of calcite veins. The crack-seal pattern of BCVs further indicates that calcite precipitation was synchronous with opening of the fracture. Similar arguments were advanced for the co-existence of bitumen and saturated mineralising pore fluids by Parnell and Carey (1995).

The late development of bitumen-bearing fractures (Hooker et al., 2017a) is consistent with the model presented above for bitumen mobilisation at the peak of thermal stress, i.e., at a late stage in the burial history prior to any uplift and cooling.

5.4.3.3 Formation Mechanisms of Bitumen-Bearing Fractures

Generally, opening-mode fractures propagate along a plane perpendicular to the minimum-compressive principal stress, normal to the plane of the crack (Secor, 1965). One mechanism for their formation is hydraulic fracturing, where a pore pressure exceeds the least compressive total normal stress, but they may also form at pore pressures significantly less than the overburden stress and by minor extensional or volumetric strains, or thermal stresses, particularly under shallow burial conditions (Eichhubl et al., 2001; Eichhubl & Aydin, 2003; Olson et al., 2009).

Veins are considered as important structural elements since they reflect stages of geological activity that occurred after formation of the host unit (Bons et al., 2012). According to Hooker et al. (2017a), calcite cement in the different types of calcite veins present in the JOS has been formed at many different phases during both burial and uplift of the host units. The high percentage of calcite veins in the calcareous, organic-rich oil shale relative to all opening-mode fractures suggests the circulation of calcium carbonate-rich fluids in the fracture networks.

Calcite veins are considered here to be formed by three possible fracture-forming mechanisms. These include diagenetic processes, tectonic stresses or fluid overpressure. The investigated bitumen-bearing fractures are coeval with late planar calcite veins which are believed to be tectonic-related fractures (Hooker et al., 2017a). The BCVs identified in the present study have elongate-blocky textures which reveal a moderate crack opening rate according to Prajapati et al. (2018). Elongate-blocky crystals demonstrate preferred

orientations at the final phase of growth, indicating the presence of growth competition during vein development (Hilgers & Urai, 2002a; Prajapati et al., 2018).

The sealing of BCVs shows overlapping filling stages. This is evidenced by the occurrence of bitumen within the crack-seal structure which suggests that the bitumen occurrence is syngenetic with the cementation of calcite veins (see Figure 5.11a-b) (cf. Parnell & Carey, 1995). The crack-seal texture implies incremental opening and sealing. This internal structure of these veins indicates that the calcite has started to precipitate in the fracture, and the bitumen filled the remaining porosity and the vacant segments before the fractures were entirely sealed. A process similar to this was proposed in a study of the Woodford Shale by Comer (2007).

The folded calcite veins are free of any bitumen traces and are thought to have formed during early burial in response to silica diagenesis and were folded by sediment compaction (the detailed formation mechanism of folded calcite veins is discussed in Chapter 3, Section 3.6.3.1). Planar silica veins are also thought to have formed early due to silica diagenesis. The sequence of events of all these fractures cements is summarized in Figure 3.60, Chapter 3.

Low permeability and high organic content play a key role in fracturing (Mandl & Harkness, 1987). The development of bitumen-bearing fractures appears to have been controlled (for BVs) or at least have been assisted (for BVCs) by hydraulic fracturing, driven by lithostatic fluid overpressure in the organic material. Brecciation in the organic-rich mudstone, particularly mosaic brecciation (see Figures 5.3b, h and 5.11g) can be indicative of hydraulically-fractured rock due to fluid overpressure within this low-permeability unit (Parnell & Carey, 1995; Passchier, 2000). The bitumen-supported breccia in the mudstone reflects a pressure regime that has been hydrostatic and then

developed to highly elevated overpressured led to fracture (i.e., brecciation) the host mudstone (Parnell & Carey, 1995).

The bedding-parallel BVs and BVCs present within the organic-rich oil shale and adjacent chert intervals (see Figures 5.3c and 5.4f-g) show a good example of hydraulic-fracturing produced by bitumen overpressuring in which the bitumen pore pressure exceeds the vertical stress generated by the overburden load.

Hydraulic fracturing usually occurs when the Mohr circle intersects the failure envelope in the tensile quadrant of Mohr space. The size of Mohr circle is governed by effective stress, which is represented by the difference between total stress and fluid pressure.

$$\sigma^- = \sigma - pF$$

Where,

σ^- is the effective stress,

σ is the total stress, and

pF is the fluid pressure.

At the microscopic scale, the development of bitumen filaments into well-developed microfractures (BMVs) within the organic-rich mudstone beds also indicates fracturing due to high pore pressure (Figure 5.5c). These BMVs that propagate within and around calcite and pyrite crystals are interpreted to have been generated by fluid pressure in the bitumen phase (**pB**) that exceeded the vertical stress (**σv**) and the tensile strength of the host rocks (**T**):

$$pB > \sigma v + T$$

Fracturing driven by fluid overpressure relating to maturation of organic matter has been documented in a range of studies (e.g., Jochum et al., 1995; Zhang et al., 2016; Meng et al., 2017). It is often suggested that volume expansion of kerogen during thermal maturation is the main driver for fracturing (Michels et al., 1994; Figueroa Pilz et al., 2017). In the case of the JOS, the generated bitumen is of early maturity. However, the yield of this early expelled bitumen seems to have been high enough to cause a volume change that resulted in fracturing.

The mechanical stiffness of chert and silicified beds is argued to have directed the fracturing to be straighter and with bigger apertures than in the ductile mudstone thus emphasising the role of the rock properties in fracturing and aperture development.

The aperture irregularity of BVs in silicified and chert beds (Figure 5.4d-e), the deflection of veins when they propagate in different beds (Figures 5.3a, and 5.4e) and the branching and splaying of BVs in mudstone (Figure 5.4e) all indicate that the fracturing generated by fluid overpressure was controlled by the rock properties (Bons et al., 2012). The veins seem to have been able to branch or change direction due to the ductility of the mud rocks, presenting further evidence of fracture formation due to fluid overpressure. Therefore, we argue that in the low-permeability mudstone beds, the bitumen overpressure has forced the mudstone to fracture or brecciate. This is consistent with the study of Parnell & Carey (1995).

An interesting piece of evidence for fracturing due to bitumen overpressure is that a single BV can propagate through three types (mudstone, chert and carbonate) of rocks with different mechanical properties (Figure 5.3d). This is also supported by microscopic observations where BVs and MBVs propagate from an organic-rich mudstone bed towards a carbonate concretion (see Figure 5.11f). The microscopic occurrence of bitumen within the matrix in both the lower and middle units of the core indicates that the bitumen has

been subjected to an external pressure that forced this *in-situ* bitumen to find a way to release this pressure by generating fractures in the low permeability mudstone matrix.

A final line of evidence for overpressured bitumen acting as a fracture driver comes from the observation that a low-organic content mudstone zones (halos) present around chert concretions and reflect a concentration of BVs. The presence of BVs in the low-organic content zones (halos) and adjacent chert concretions in some parts of the study succession (in the Lower Unit) indicates that volume change and mechanical properties of rocks are not the only factors that caused bitumen overpressuring, leading to fracturing. It is argued here that the halos have developed as a result of the formation of chert concretions. This has led to the development of BVs within these alteration rims from the local bitumen generated from the original organic-rich mudstone (Abu-Mahfouz et al., 2019b). The nucleation of BVs appears to occur in the halos with propagation into the chert concretions (Figure 5.12). A huge reduction in the disseminated organic matter seems to have occurred in the halos, resulting in the formation of these halos. This is evidenced by the low TOC content compared to the surrounding mudstone (Figure 5.19). The high intensity of BVs inside and around the chert concretions (in the halos) in this specific unit reflects the association chert formation and bitumen fracturing. The strong relationship between the aperture size of the BVs and the colour intensity of the halo also demonstrates hydraulic fracturing created by the bitumen overpressure.

5.5 Conclusions

- 1) The biomarker and petrographic results of fracture and matrix bitumen from 15 cores and one outcrop of the JOS show that the source of bitumen within fractures and adjacent vugs are very low-mature hydrocarbons expelled from the JOS matrix, and have not migrated from another petroleum source.

- 2) Maturity indicators imply that the oil shale is everywhere in the very early-oil generation stage, but with a regional variation from west to east resulting from greater maximum burial depth.
- 3) The present study has revealed that bitumen fractures developed by hydraulic fracturing as the result of fluid overpressure in the organic matter due to permeability destruction in the organic mudstones which prevents fluid loss. Overpressured bitumen has acted as a fracture driver, generating bitumen veins in both the organic-rich mudstones and the adjacent chert and silicified intervals.
- 4) Diagenetic processes (e.g., silica diagenesis) have contributed to the bitumen mobilization and the development of bitumen veins in the Lower Unit of the succession.
- 5) Rock properties played an important role in the distribution of bitumen-bearing fractures with high intensities of bitumen-bearing fractures occurring in chert and silicified beds.
- 6) Late calcite veins contain bitumen inclusion (BCVs) which shows that bitumen occurrence is syngenetic with the cementation of calcite veins, and was generated relatively late in the paragenetic sequence.

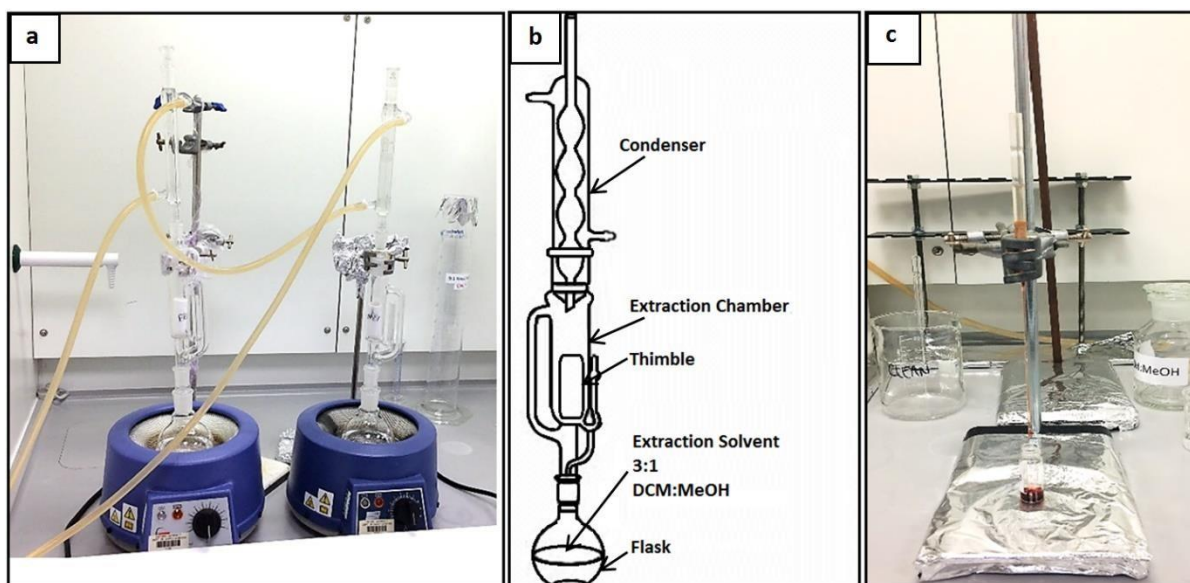


Figure 5.1 a. and b. A photograph showing the Soxhlet extractor set used to extract bitumen from the studied samples. b. A diagram illustrating the parts of the Soxhlet extractor. c. A photograph of the column chromatography apparatus.

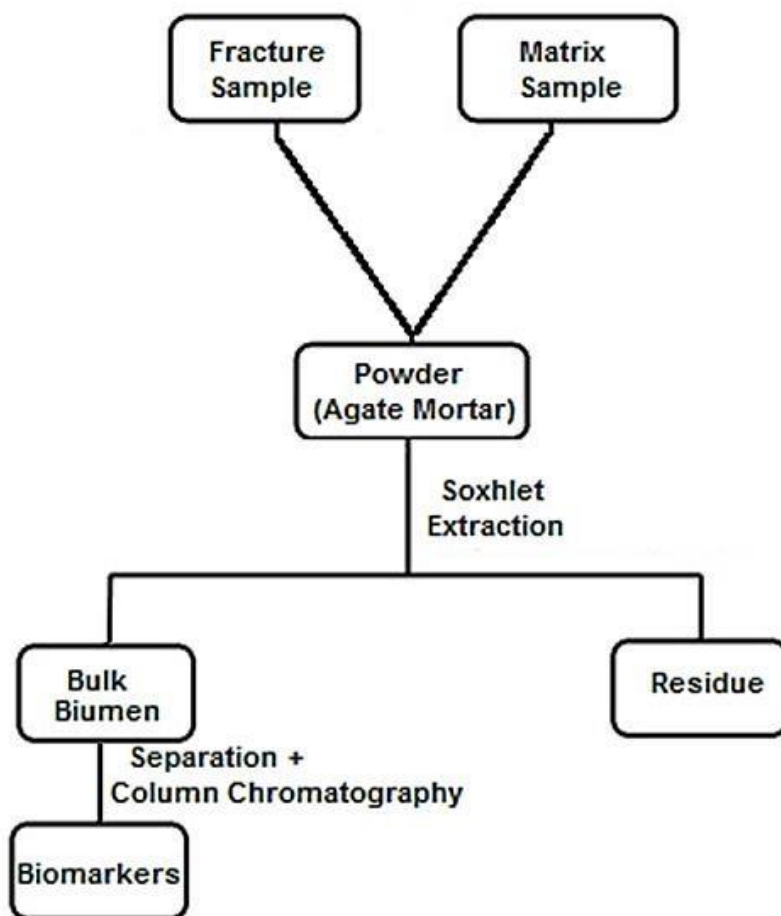


Figure 5.2 Flowchart showing the procedure of Soxhlet extraction.

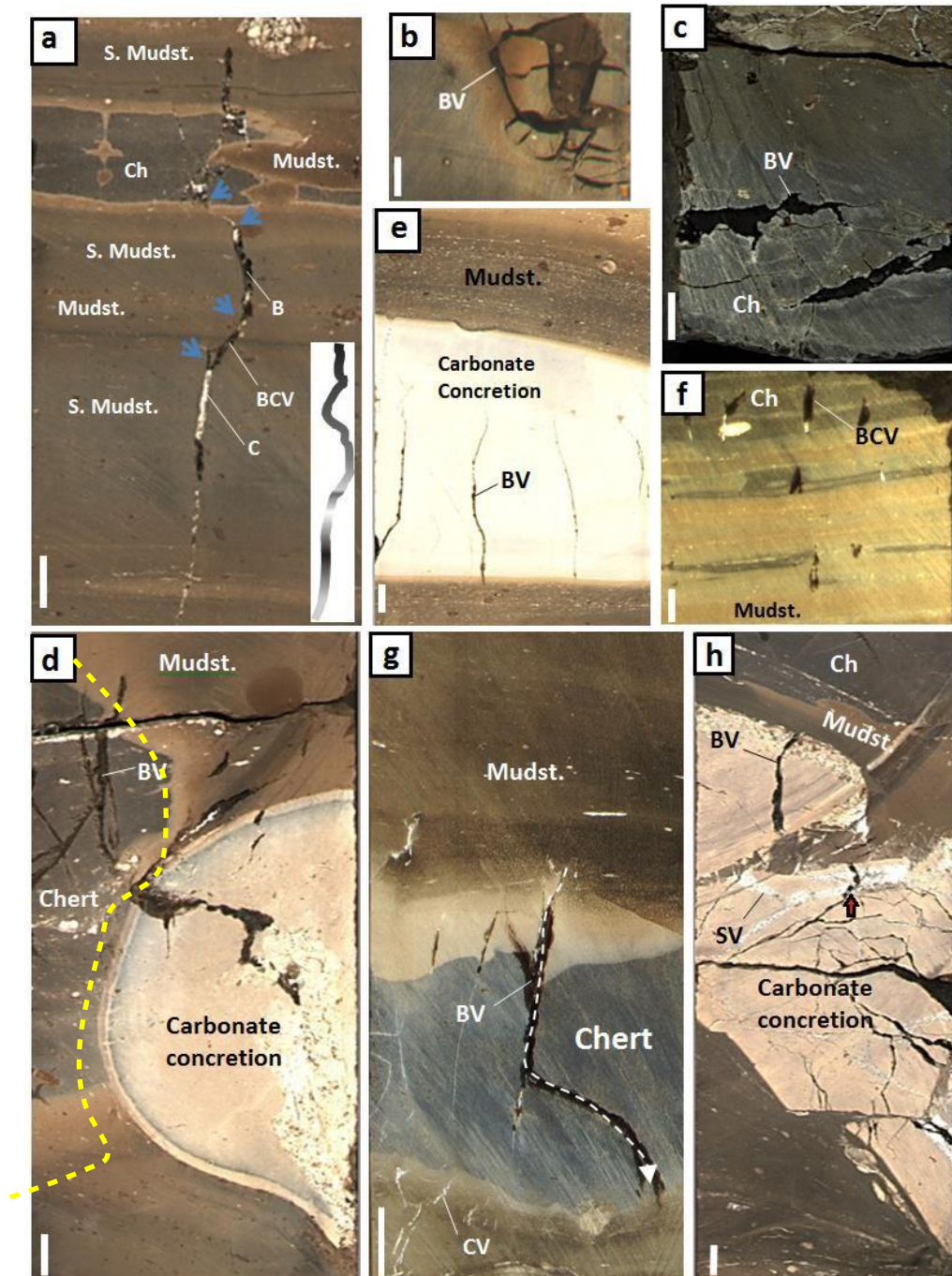


Figure 5.3 Core photographs showing examples of bitumen-bearing fractures observed in the Lower Unit of each of the fifteen studied cores. Ch, Chert, S. Mudst, Silicified Mudstone; B, Bitumen; C, Calcite; BCV, Bitumen-Bearing Calcite Vein; BV, Bitumen Vein; CV, Calcite Vein; SV, Silica Vein. a. BCV propagates across chert concretion, silicified mudstone and mudstone. Note the deflection of vein when the bed lithology changes (blue arrows and drawing). b. Mosaic brecciation by bitumen in an organic-rich mudstone. c. Horizontal BVs in Chert. d. Bitumen vein propagates through three different types of lithologies (mudstone, chert and carbonate). Note the increase in aperture size of the vein in the chert and carbonate concretions. Yellow-dotted shape shows a mudstone zone with low organic content (see text for information, Section 5.3.1.4). e. BVs confined to carbonate concretion. f. BCVs confined to chert beds. g. BV opens towards a chert concretion. h. Two fracture networks of horizontal and vertical SVs and BVs. BVs crosscut SVs (red arrow). Scale bar is 1cm.

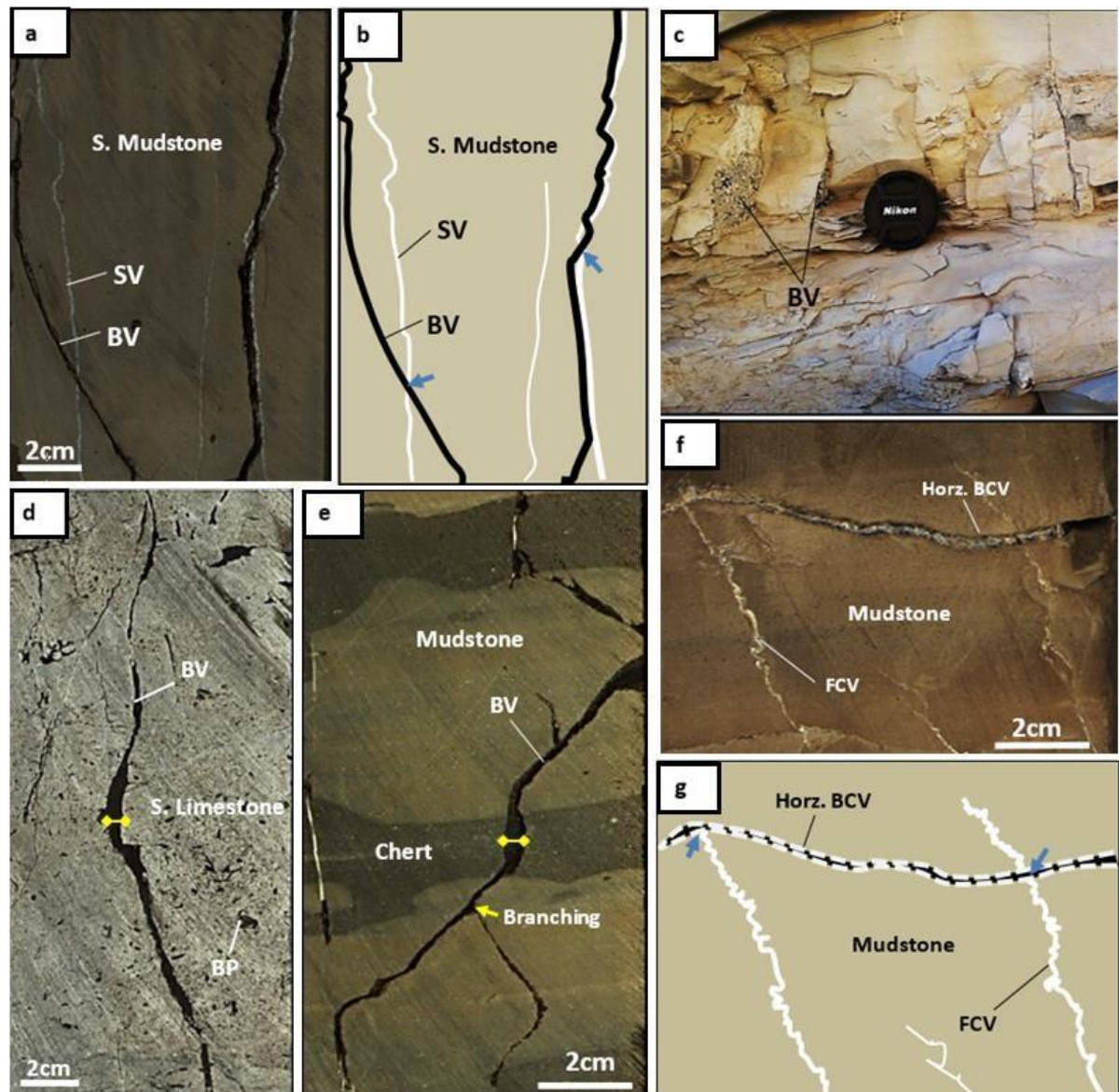


Figure 5.4 Core and field photographs showing different bitumen-bearing fractures and their crosscutting relationships with other fractures. BCV, Bitumen-bearing Calcite Vein; BV, Bitumen Vein; SV, Silica Vein; FCV, Folded Calcite Vein; S. Mudstone, Silicified mudstone. (a) SVs and BVs in organic-rich mudstone. (b) Schematic drawing of (a). Note the crosscutting relationship between silica veins and bitumen veins (blue arrows). (c) Field photograph of vertical BVs confined to the oil shale bed at the El-Lajjun quarry. (d) Bitumen in veins and pores (vugs) in silicified limestone beds in the lower unit of core C8. Yellow diamonds show the maximum aperture size of the bitumen vein. (e) Bitumen vein propagates in mudstone and chert beds. Note the branching of the bitumen vein in the mudstone bed (yellow arrow). (f) Horizontal BCV crosscut folded calcite veins in organic-rich mudstone. (g) Schematic drawing of (f). Note the crosscutting relationships between the horizontal bitumen-bearing vein and folded veins (blue arrows).

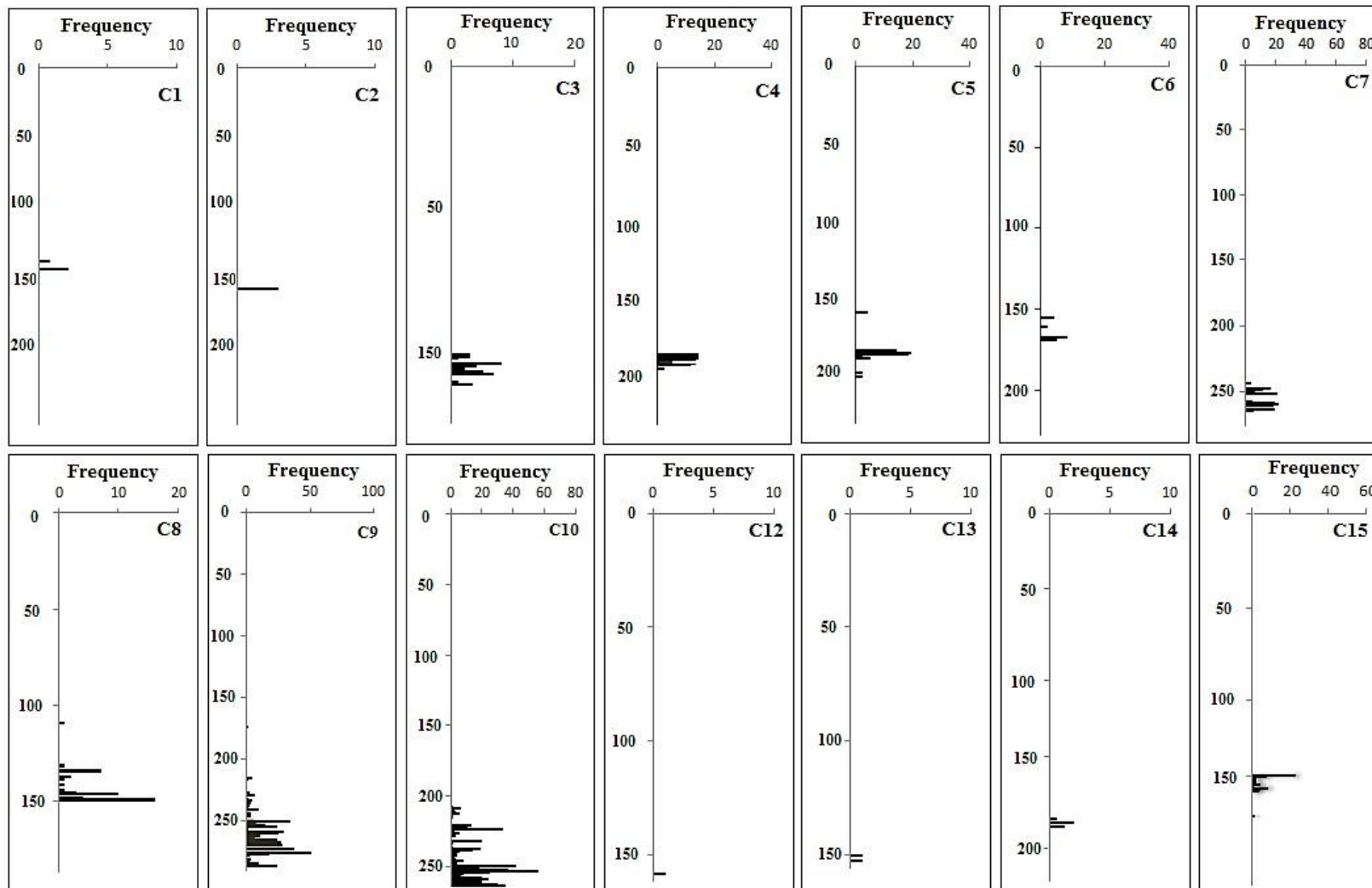


Figure 5.5 Plots of the frequency of bitumen-bearing fractures. Harrana cores (C1-C5); Azraq cores (C8-C10); Bayer cores (C12-C15). See Figure 2.8 in Chapter 2 for all core locations. Note that the fractures are mainly clustered in the Lower Unit of all cores.

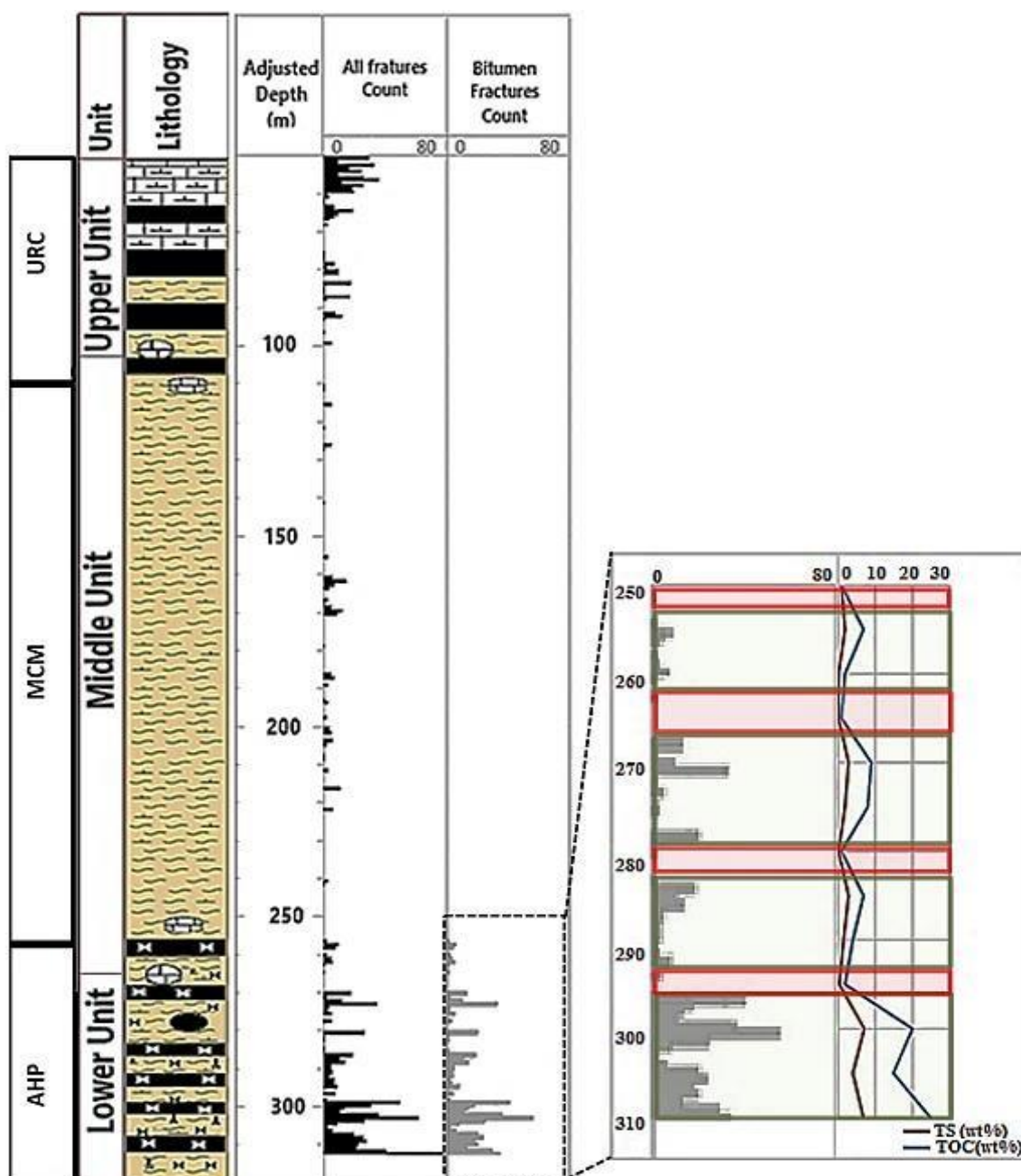


Figure 5.6 A columnar section showing the lithology of the studied cored intervals along with a representative trend (from core C10) of vertical distribution of all fractures and bitumen-bearing fractures against the three lithological units. Zoomed in area demonstrates the relationship between fracture intensity and the TOC and TS of the lower unit. Red-shaded rectangles show low fracture intensity correlate with low TOC and TS values. Green-shaded rectangles show high fracture intensity correlates with high TOC and TS values.

Type of Bitumen-bearing fracture/ vugs	Carbonate concretion	Chert Concretion/Bed	Silicified Limestone	Silicified Mudstone	Marl	Organic-rich Mudstone	Dolomite	Re-Sedimented Bed
BV	H	H	H	H	M	H	H	H
BCV	M	M	H	H	M	M	M	-
BSV	L	H	M	H	-	-	-	-
Bitumen Vugs	M	H	H	M	-	-	H	H

Table 5.1 Relative distribution of bitumen-bearing fractures and vugs (High: H, Moderate: M, Low: L) in the lower units of all the studied cores

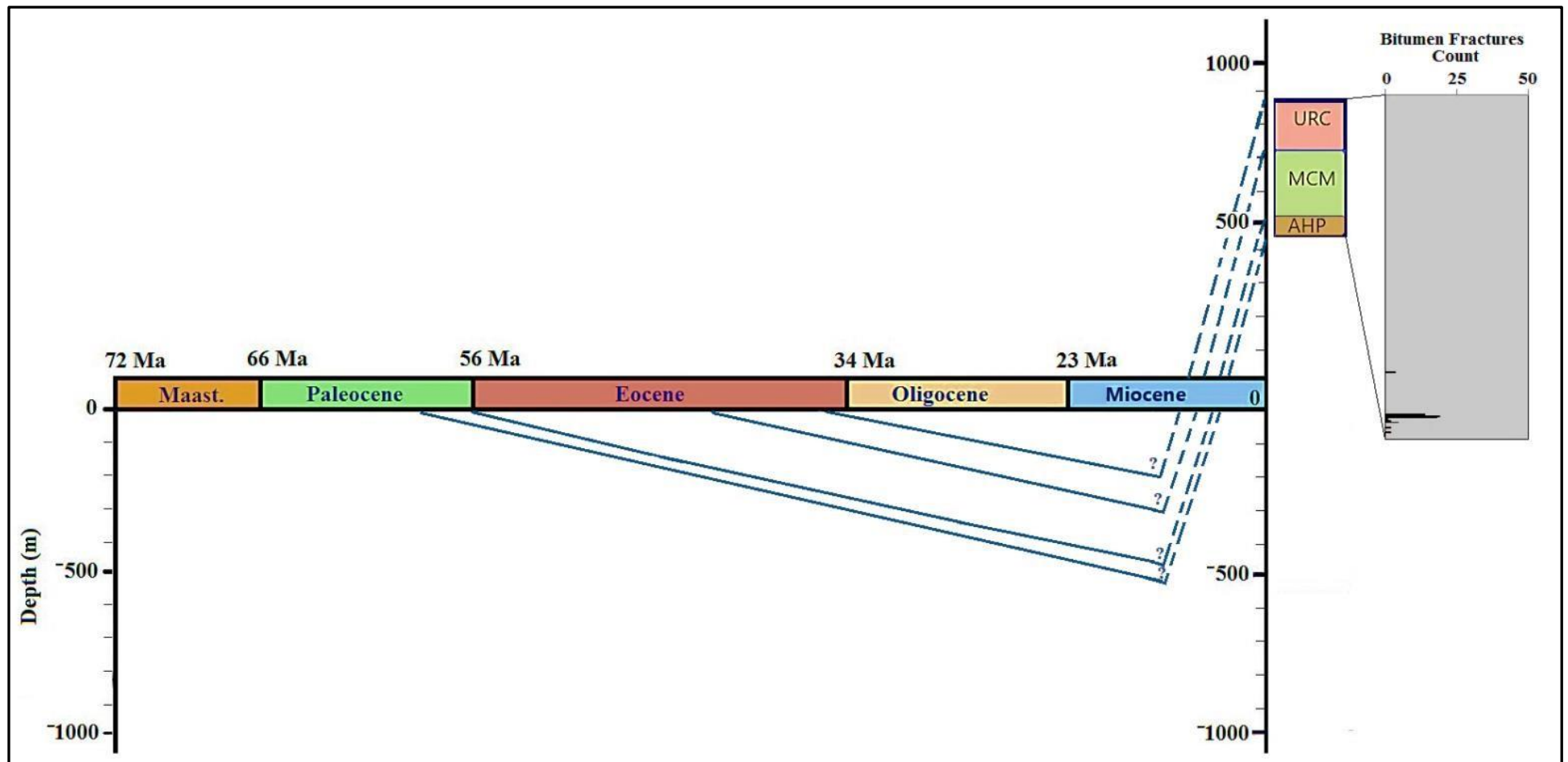


Figure 5.7 Schematic burial history curve of the JOS in the Harrana area (CW). The burial history curve was constructed based on information (formation thickness, tops, tectonic events) extracted from well and seismic data. The depth and age are relatively defined. Fracture log is from the core C5.

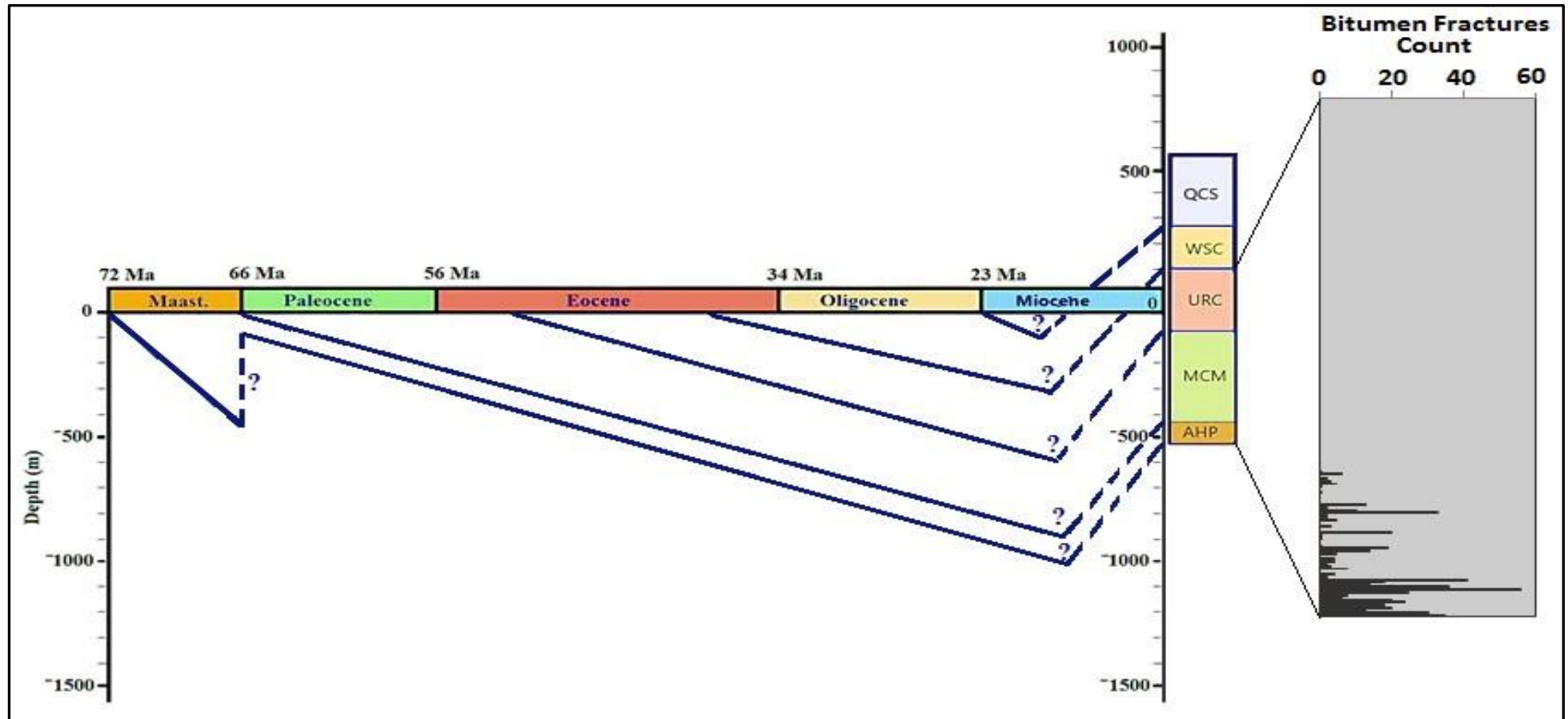


Figure 5.8 Schematic burial history curve of the JOS in the Azraq area (CE). The burial history curve was constructed based on information (formation thickness, tops, tectonic events) extracted from well and seismic data. The depth and age are relatively defined. Fracture log is from core C10.

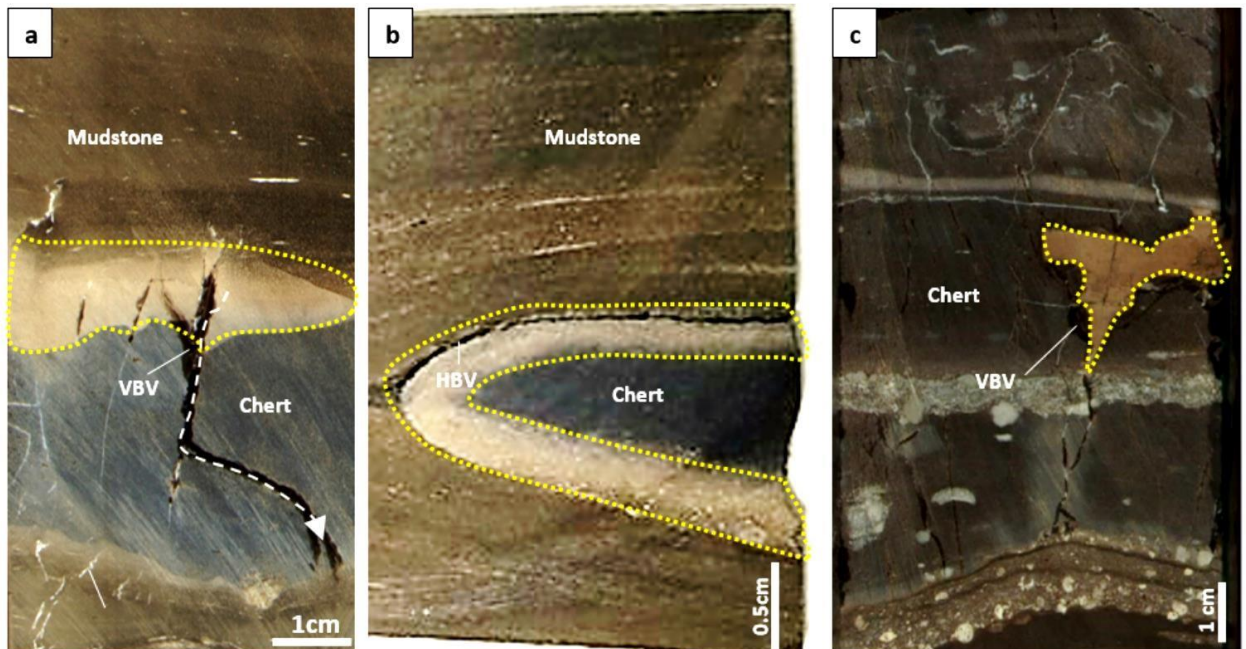


Figure 5.9 Examples of the organic-low zones (halos; yellow-dotted shapes) associated with chert concretions in the Lower Unit of the study succession. VBV, Vertical Bitumen Vein; HBV, Horizontal Bitumen Vein.

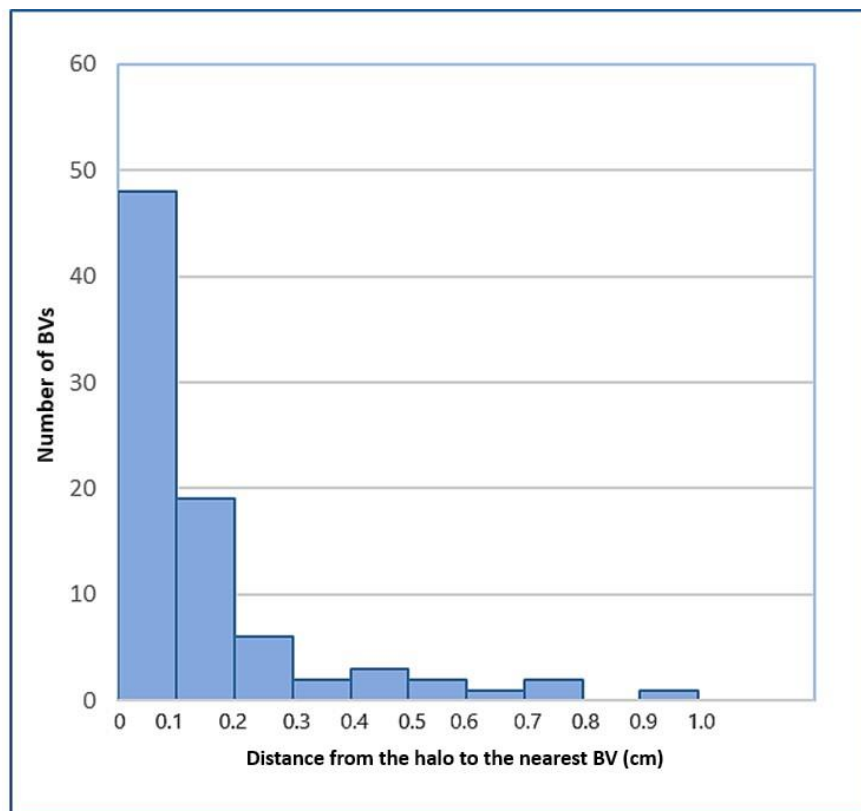


Figure 5.10 Distance from the halo to the nearest bitumen vein, measured from the Lower Unit of each of the study cores. BV, Bitumen Vein. N=84

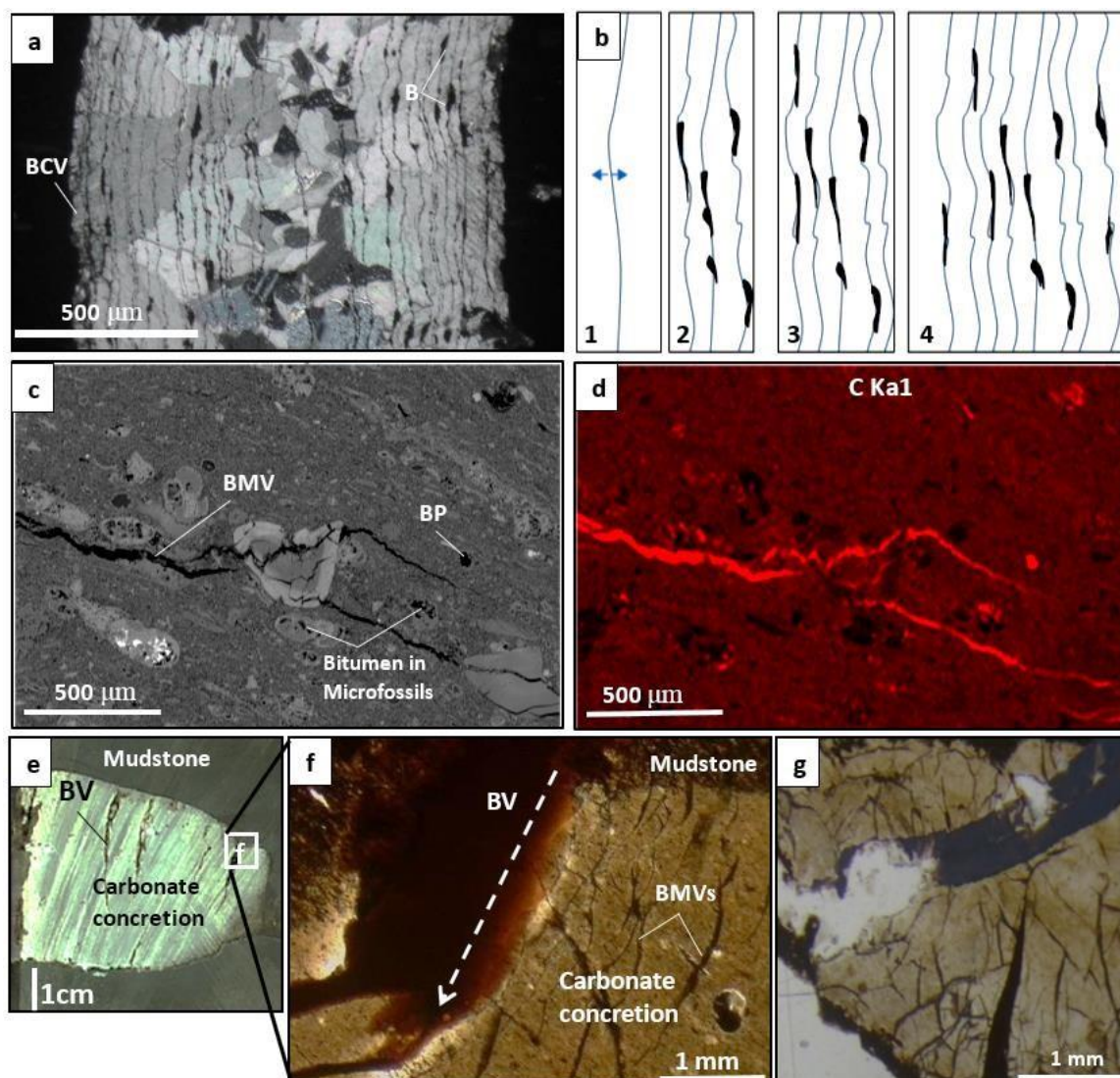


Figure 5.11 Micrographs of different bitumen-bearing fractures observed in the study cores. BCV, Bitumen-bearing Calcite Vein; B, Bitumen; BP, Bitumen-bearing pore, BMV, Bitumen Micro-Vein; BV, Bitumen Vein. a. BCV shows crack-seal texture. Note the bitumen inclusions within calcite cement. b. Schematic drawing of (a) illustrating phases of crack-seal (1-4). c. BSE image shows sub-horizontal microfractures, pores and microfossils filled with bitumen. Note Microfracturing of grains. d. EDS map of the bitumen microveins shown in (c), showing the carbon content in the bitumen microfractures and matrix in the lower unit of core. e. Core photograph showing BVs in carbonate concretions hosted by organic-rich mudstone. f. Micrograph of the inset in (e) showing the direction of bitumen migration inside both BV (white-dashed arrow) and BMV from the organic-rich mudstone bed towards the carbonate concretion. g. Mosaic brecciation of mudstone by bitumen overpressure.

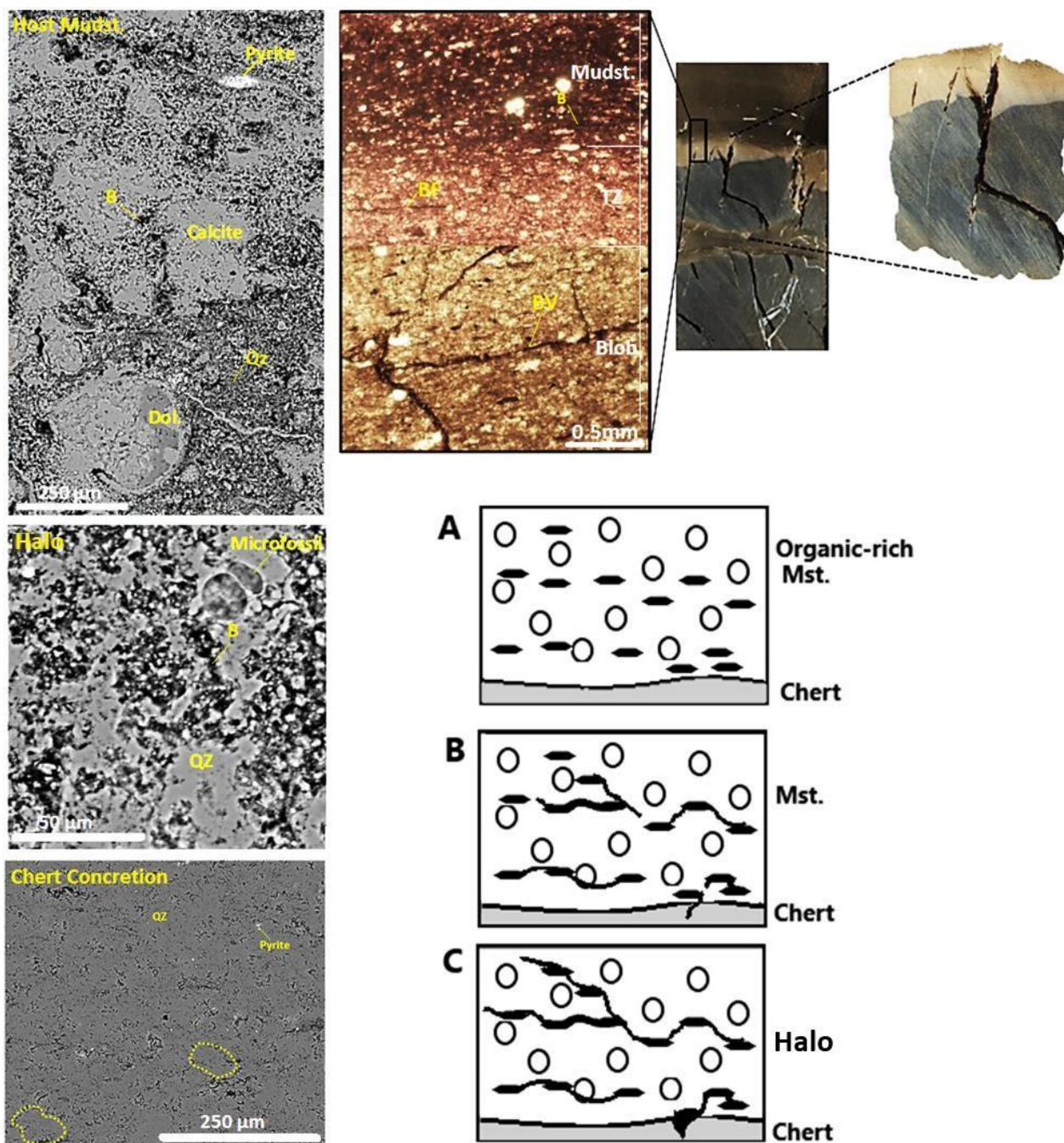


Figure 5.12 BSE images, microphotographs and a diagram illustrating the development of bitumen veins within the halos and adjacent chert concretions observed in the Lower Unit of core. B, Bitumen; Qz, Quartz; BV, Bitumen vein; BF, Bitumen Filament. Note the change in composition and bitumen content from the host mudstone towards the chert concretion. Yellow dotted shapes are fossil 'ghosts' in the chert concretion.

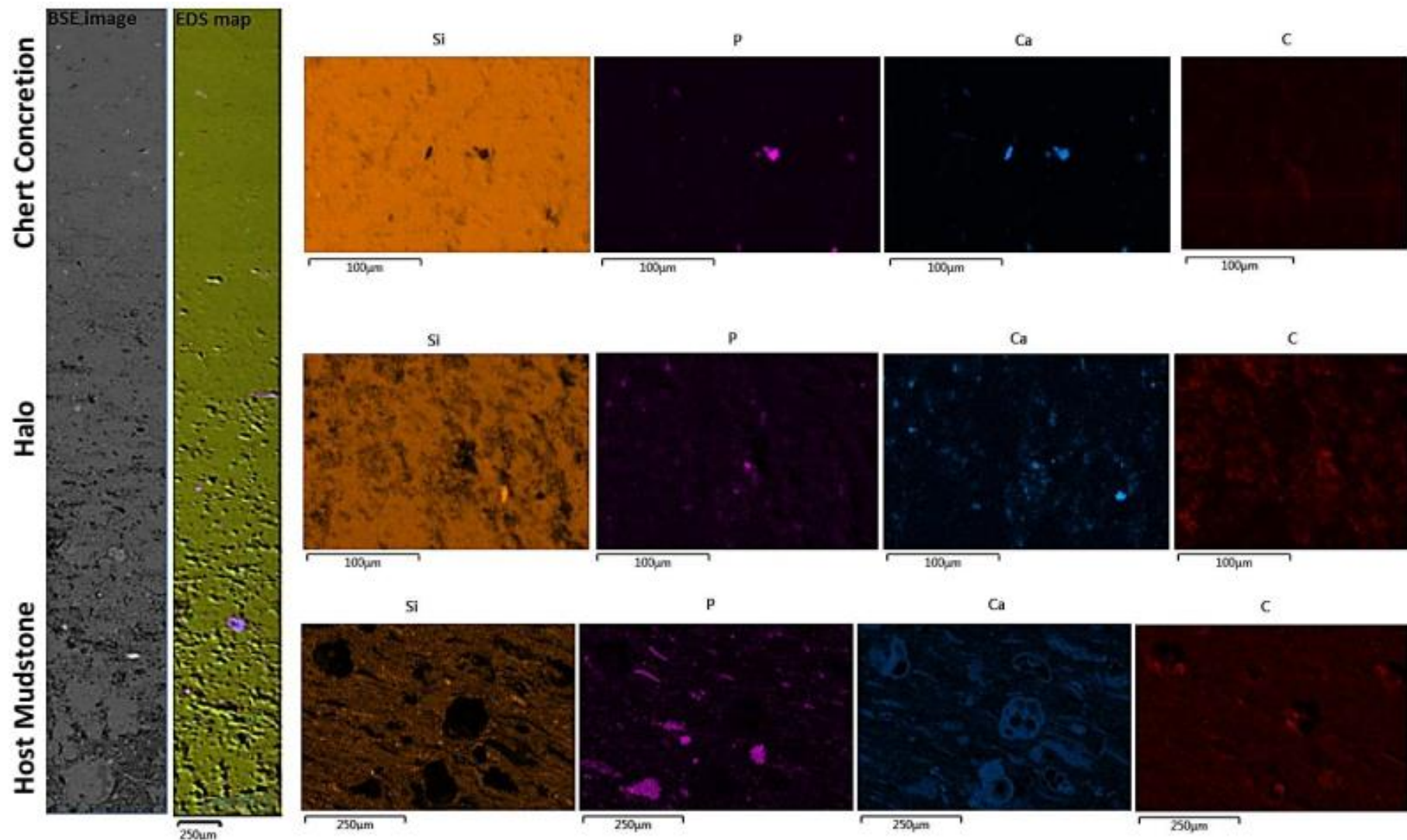


Figure 5.13 A BSE image and EDS maps showing the composition of the chert concretion, halo and host mudstone (from the centre of the concretion (top) through the halo (middle) to the mudstone (bottom)) from the Lower Unit of core. Si, Silica; P, Phosphorus; Ca, Calcium; C, carbon.

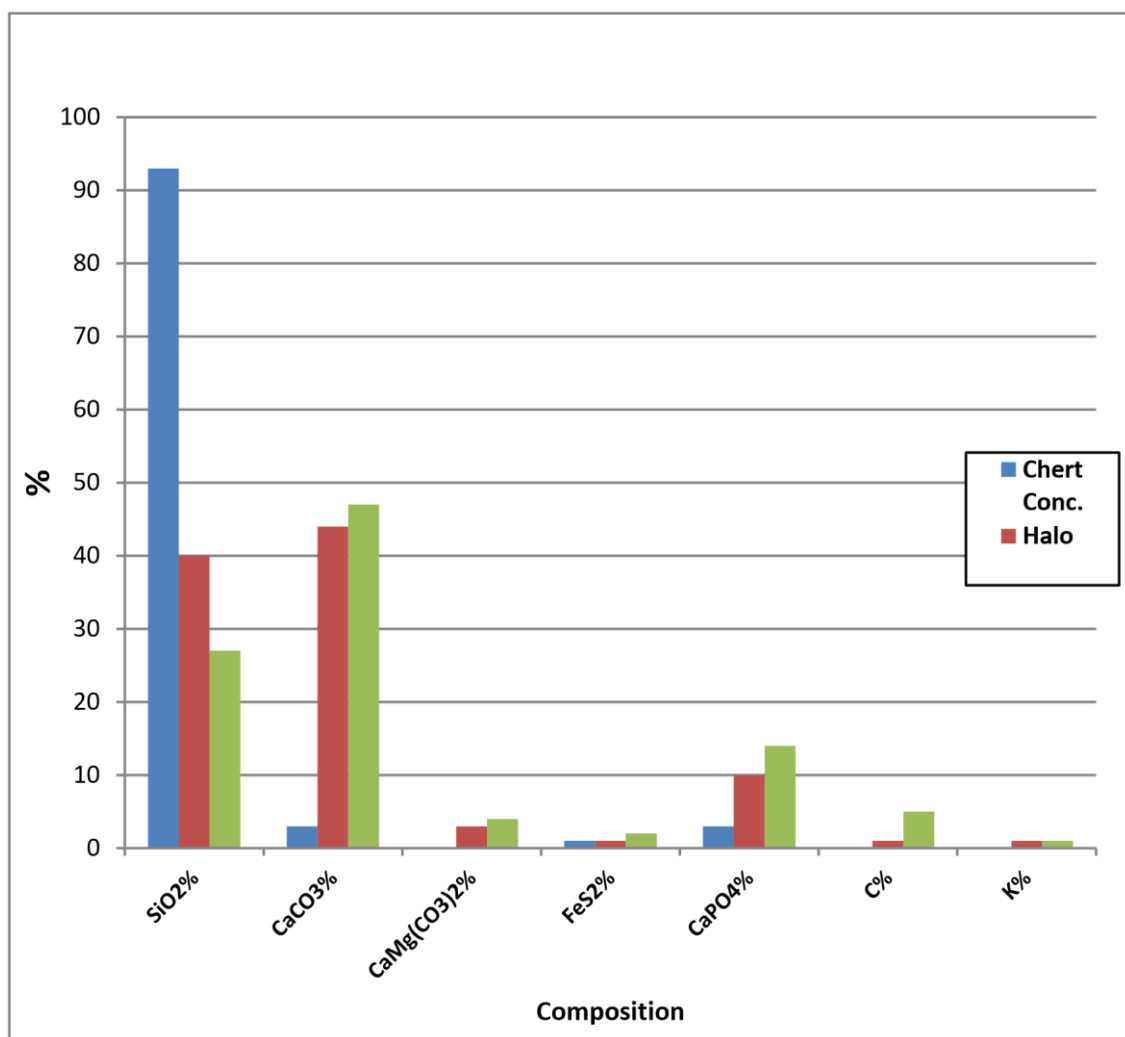


Figure 5.14 Compositions of the chert concretions (Chert Conc.), halos and host mudstones from the Lower Unit of cores, calculated from EDS maps. C, Organic Carbon; K, Potassium.

Sample	Sample Type	TOC (wt%)	TS (wt%)	S1 (mg/g rock)	S2 (mg/g rock)	T _{max} (°C)	HI (mg HC/g TOC)	OI (mg CO ₂ /g TOC)	PI
1	Core	19.5	2.8	2.7	148.2	402.0	762.0	17.4	0.02
2	Core	15.7	4.8	1.7	105.3	413.0	672.8	19.3	0.02
3	Core	16.9	3.8	2.0	115.0	420.0	680.5	15.3	0.02
4	Core	14.9	5.2	1.8	131.2	415.0	880.2	8.5	0.06
5	Core	15.9	4.5	2.5	112.4	416.0	705.7	11.4	0.02
6	Core	30.0	8.3	6.7	175.0	404.0	583.3	27.0	0.04
7	Core	14.1	2.3	0.9	104.5	416.0	743.5	9.3	0.01
8	Core	17.9	4.3	5.5	115.9	424.0	647.5	14.2	0.05
9	Core	20.2	4.3	1.9	130.0	395.0	642.9	12.9	0.04
10	Core	13.9	6.5	2.0	97.8	404.0	703.6	15.2	0.02
11	Core	14.9	3.7	1.4	100.1	405.0	670.7	12.0	0.01
12	Outcrop	21.3	3.9	4.6	132.3	416.0	621.1	13.7	0.03
13	Core	6.1	1.5	0.5	45.0	403.0	734.1	18.0	0.01
14	Core	12.4	2.8	1.8	90.3	409.0	726.6	13.0	0.02
15	Core	17.4	3.8	3.8	119.4	407.0	688.1	12.7	0.03
16	Core	11.7	4.7	2.5	90.4	389.2	771.9	14.3	0.06

Table 5.2 Bulk organic geochemical properties of the JOS matrix in the Lower Unit of the studied succession from both the cores (C1-C15, core locations are shown in Figure 2.8 in chapter 2) and outcrop (El-Lajjun Quarry)

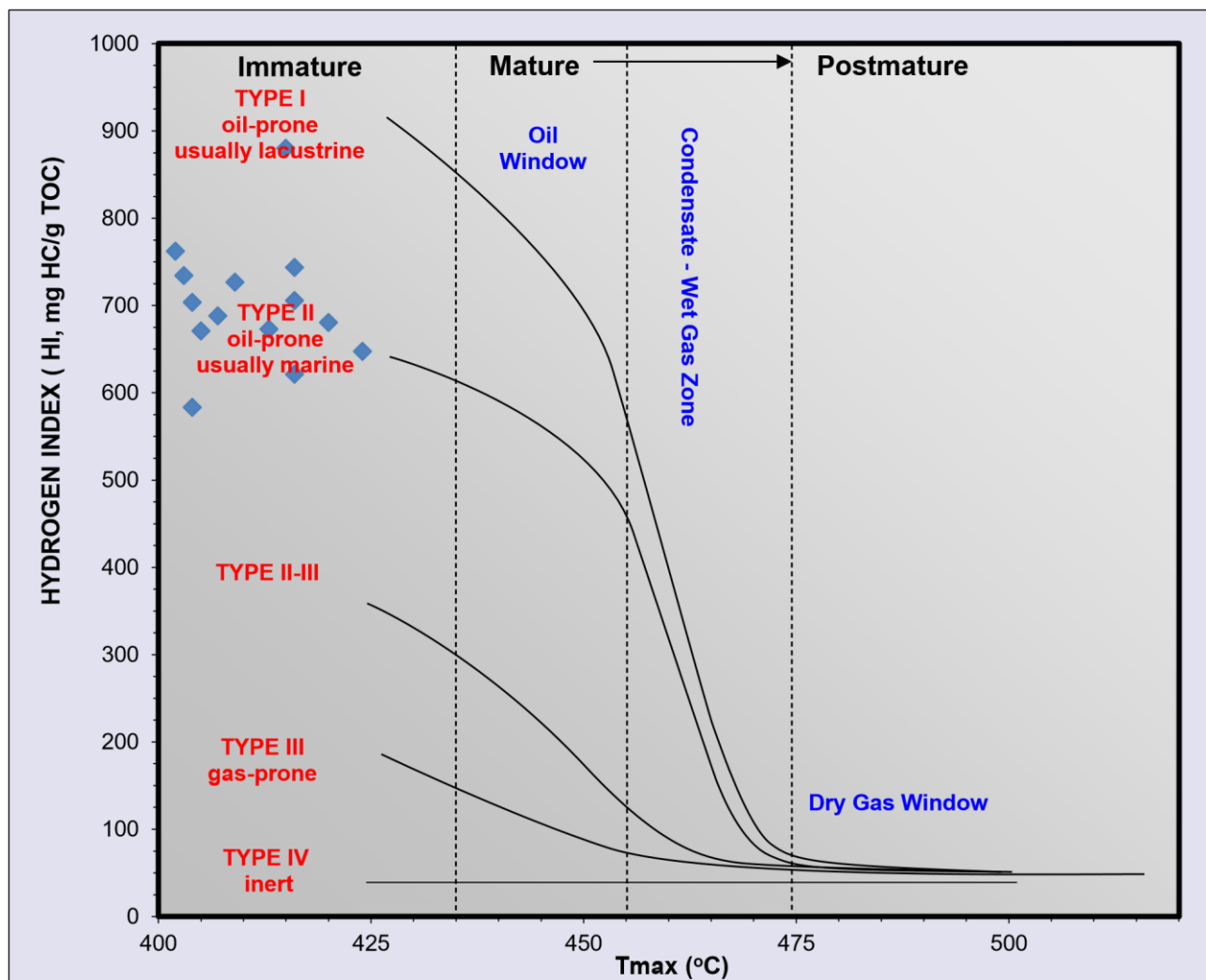


Figure 5.15 Plot of Hydrogen index (HI) versus T_{max} , showing maturity and kerogen quality of the Late Cretaceous-Paleogene oil shale samples from the Lower Unit of the study succession. These data are also shown in Table 5.2.

Sample	Hopanes Data			Steranes Data							
	Gammacerane:C30	C29:C30	22S/(22S+22R) C31 Hopanes	%C27 $\alpha\alpha\alpha$ R	%C28 $\alpha\alpha\alpha$ R	%C29 $\alpha\alpha\alpha$ R	C27:C28	C27:C29	C28:C29	$\alpha\beta\beta/(\alpha\beta\beta+\alpha\alpha\alpha)$ C28 Steranes	20S/(20S+20R) C28 Steranes
FB1	0.48	0.52	0.21	52	25	23	2.10	2.32	1.11	0.59	0.30
FB2	0.37	0.39	0.25	55	21	24	2.61	2.24	0.86	0.59	0.25
FB3	0.79	0.60	0.53	48	26	26	1.85	1.85	1.00	0.51	0.52
FB4	0.79	0.53	0.54	47	27	25	1.72	1.85	1.07	0.49	0.51
FB5	0.46	0.27	0.53	58	20	22	2.88	2.58	0.89	0.49	0.47
FB6	0.47	0.38	0.56	52	23	25	2.23	2.04	0.92	0.55	0.42
FB7	0.45	0.38	0.56	48	24	27	2.00	1.77	0.88	0.54	0.39
FB8	0.87	0.35	0.55	54	20	26	2.67	2.09	0.78	0.63	0.40
FB9	0.90	0.33	0.60	57	20	23	2.82	2.53	0.89	0.63	0.43
FB10	0.93	0.28	0.55	57	18	25	3.21	2.25	0.70	0.65	0.42
FB11	0.89	0.36	0.60	58	21	22	2.81	2.65	0.94	0.66	0.43
FB12	0.44	0.44	0.28	53	21	26	2.58	2.04	0.79	0.50	0.24
MB1	0.57	0.43	0.19	53	23	24	2.33	2.23	0.95	0.59	0.28
MB2	0.38	0.27	0.22	53	20	27	2.67	2.00	0.75	0.53	0.25
MB3	0.77	0.54	0.55	49	23	27	2.11	1.82	0.86	0.41	0.51
MB4	0.76	0.49	0.56	47	24	29	1.96	1.63	0.83	0.51	0.52
MB5	0.40	0.38	0.56	55	22	24	2.53	2.29	0.90	0.59	0.44
MB6	0.38	0.52	0.55	56	20	24	2.72	2.33	0.86	0.57	0.44
MB7	0.34	0.53	0.57	56	21	23	2.72	2.45	0.90	0.57	0.38
MB8	0.84	0.28	0.57	55	21	24	2.67	2.29	0.86	0.64	0.42
MB9	0.88	0.29	0.58	55	20	25	2.67	2.18	0.82	0.62	0.45
MB10	0.91	0.33	0.56	54	20	26	2.67	2.09	0.78	0.63	0.42
MB11	0.83	0.36	0.56	52	23	24	2.23	2.13	0.96	0.60	0.45
MB12	0.45	0.41	0.27	51	16	30	2.58	1.69	0.66	0.52	0.21

Table 5.2 Biomarker data extracted from m/z 191 and m/z 217 fragmentograms of the analysed matrix and fracture bitumen extracts. Samples from FB1-FB11 and MB1-MB11 are from the Lower Units of the cored wells located in the Harrana, Azraq and Bayer areas (core locations are shown in Figure 2.8 in Chapter 2), whereas samples FB12 and MB12 are from the studied outcrop (El-Lajjun Quarry)

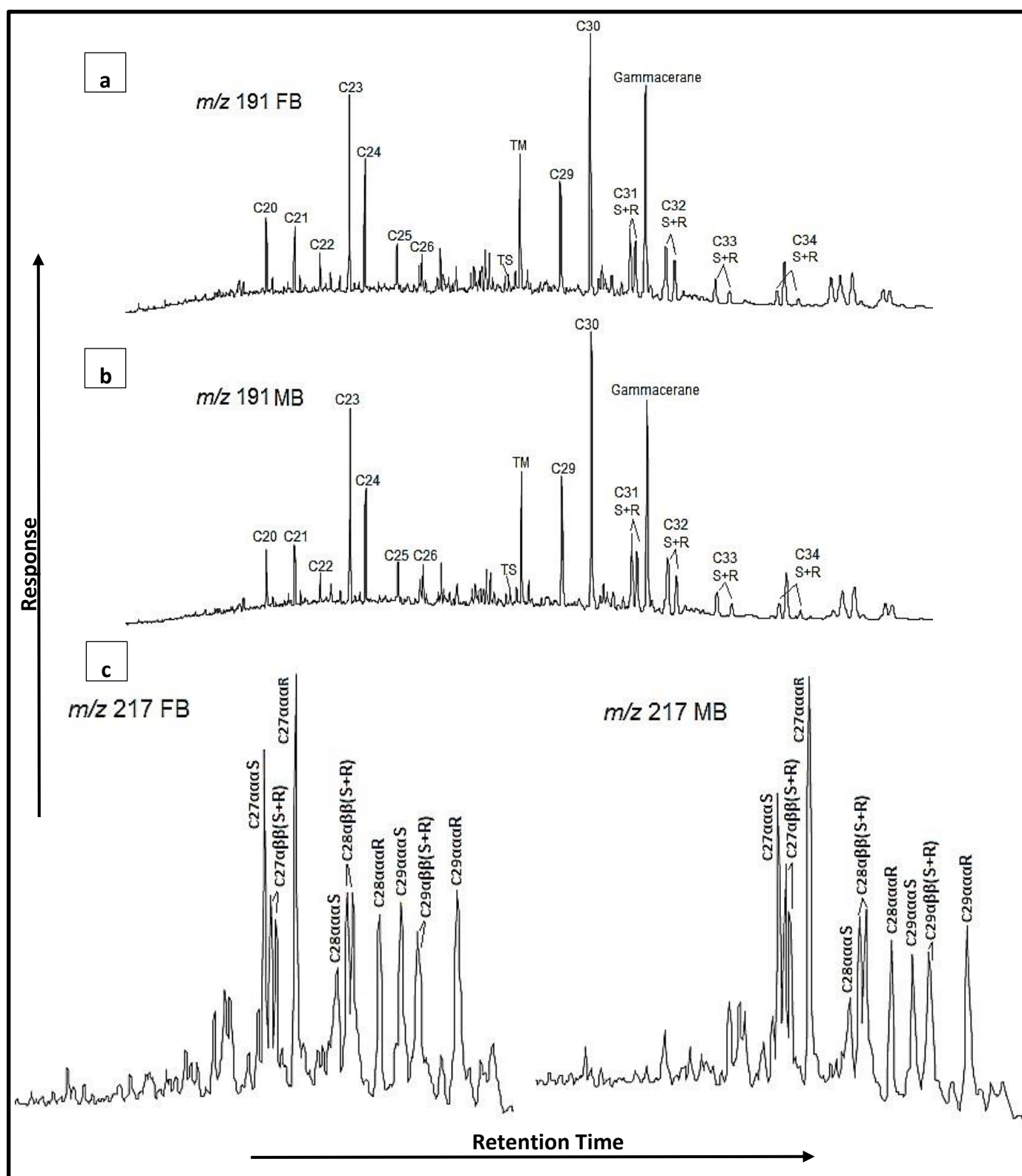


Figure 5.16 Examples of representative m/z 191 and m/z 217 fragmentograms showing the distribution of hopanes and steranes, respectively, in studied matrix and fracture bitumen samples. Note the similarity between the fracture (FB) and matrix bitumen (MB). See Appendices for all fragmentograms of all analysed samples.

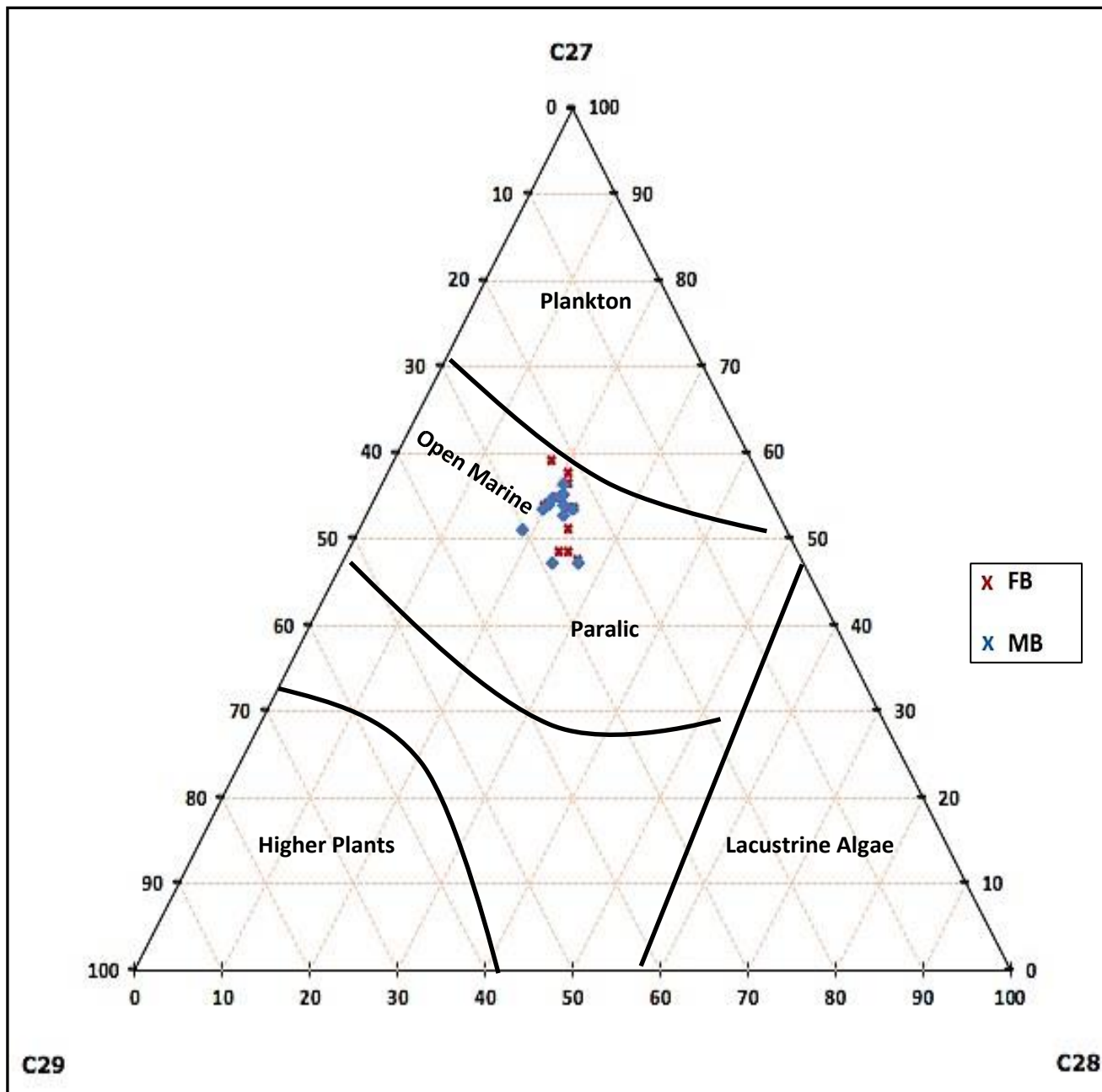


Figure 5.17 Ternary diagram showing the distribution of C27, C28 and C29 $\alpha\alpha\alpha$ -R steranes in the fracture and matrix bitumen samples. Depositional environments adapted from Huang & Meinschein 1979. FB, Fracture Bitumen; MB, Matrix Bitumen.

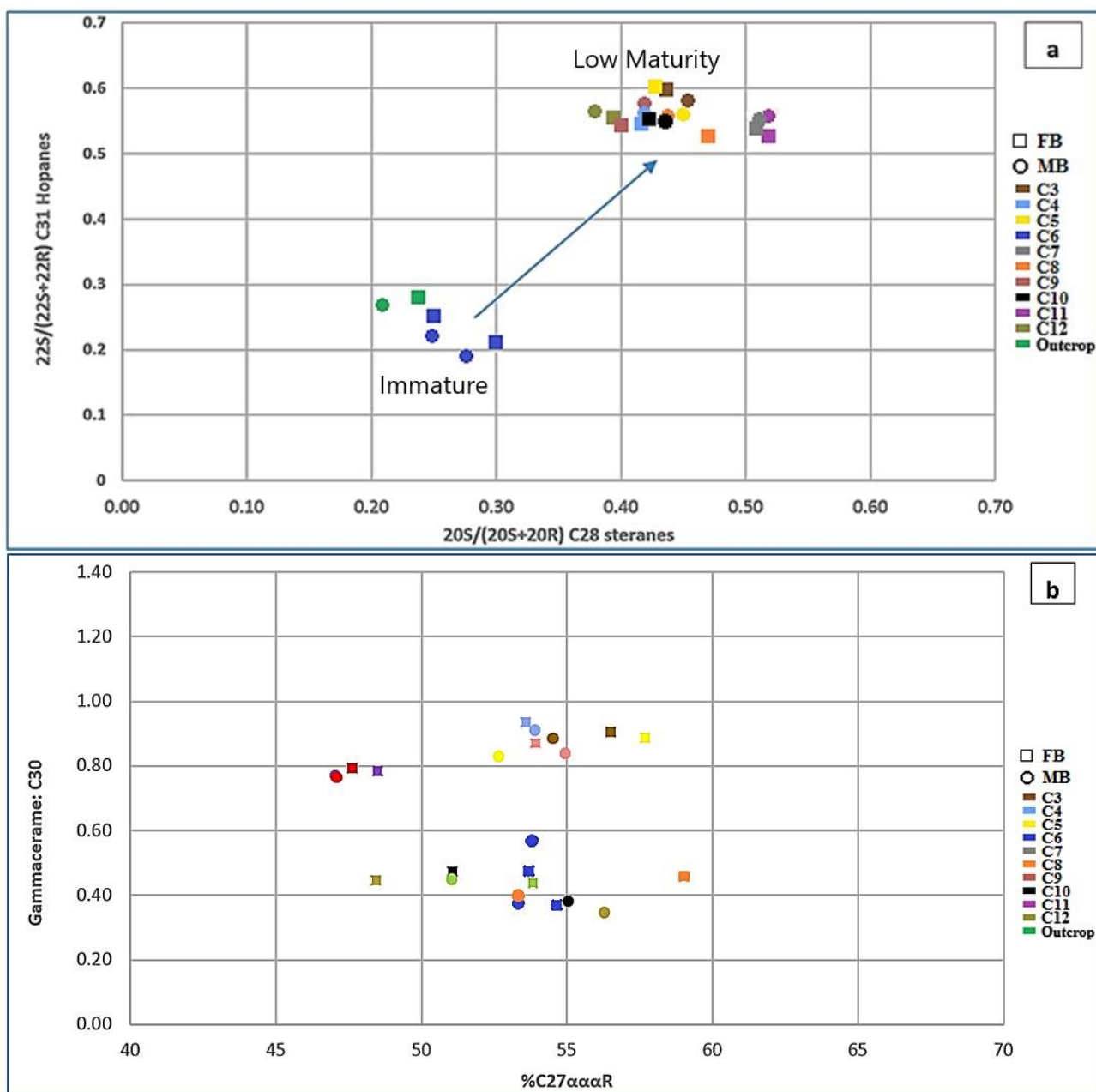


Figure 5.18 Plots of the hopane and sterane biomarkers ratios analysed from the bitumen extracts. a. $22S/(22S+22R)$ C31 Hopanes against $20S/(20S+20R)$ C28 Steranes. Note the consistency between the matrix bitumen and fracture bitumen sets of each core and the outcrop. Note the clustering of immature samples (from the core C6, in dark blue, and the outcrop, in green). b. Gammacerane: C30 against C27 $\alpha\alpha$ R. FB, Fracture Bitumen; MB, Matrix Bitumen; C3-C12, Cores from C3 to C12. The colours represent the core/outcrop numbers.

Sample Number	Halo (TOC, wt%)	Host Mudstone (TOC, wt%)
B1	4.88	11.94
B2	5.28	12.06
B3	11.88	17.89
B4	4.6	13.32
B5	8.07	11.02
B6	6.98	13.26
B7	5.5	11.1
B8	5.4	17.4
B9	6.3	17.8

Table 5.3 TOC data of the analysed halos and their host mudstone from the Lower Unit of the study succession

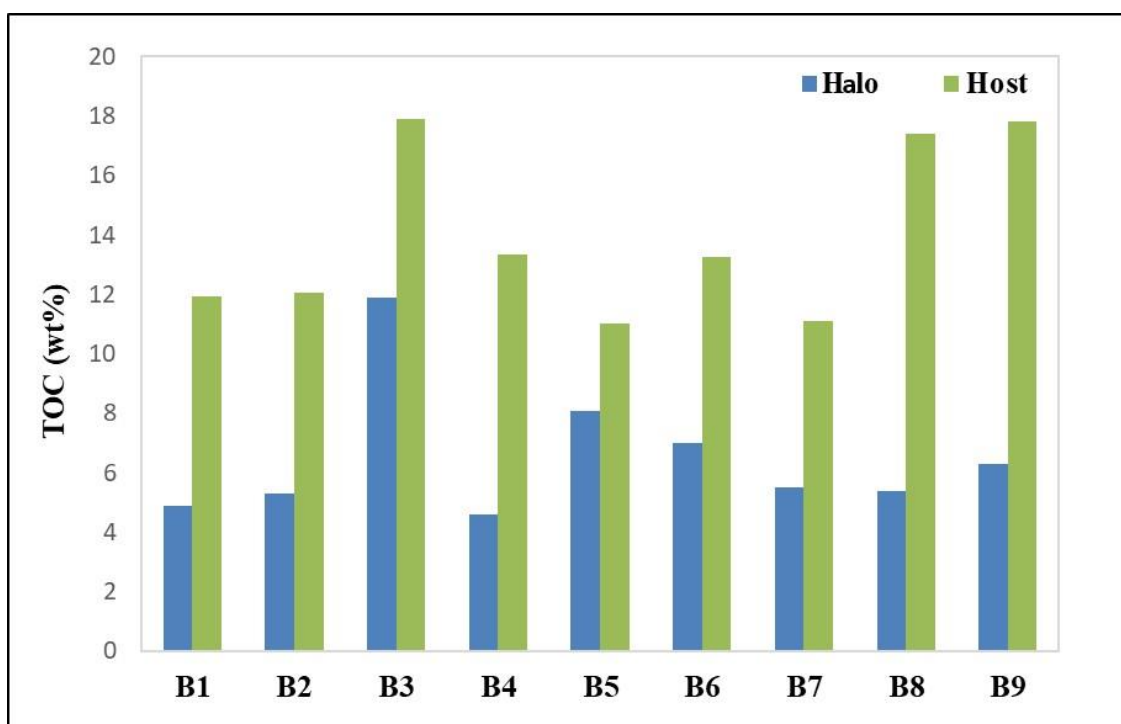


Figure 5.19 Histogram showing the TOC data of the analysed halos and their host organic-rich mudstone in the Lower Unit of cores.

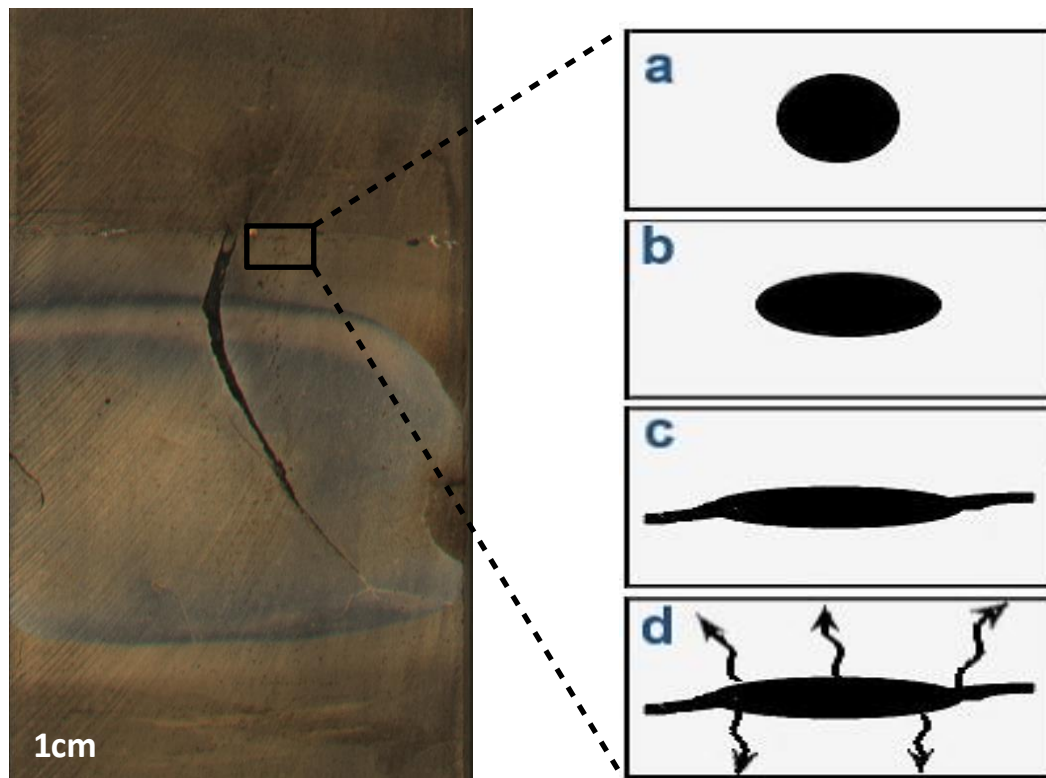


Figure 5.20 Illustration of the bitumen mobilization process (a-d). (a) Organic particle at shallow burial. (b) Mechanical compaction and water removal. (c) It becomes mechanically weak-loading and extrudes bituminized OM. Occurs preferentially in OM with early generation kinetics and high HI. (d) Saturation of the matrix followed by short-distance migration, creating fractures (the diagram modified from Pepper, 2017). OM, Organic Matter, HI, Hydrogen Index.

Chapter 6: Discussion

6.1 Development of Natural Fracture Systems in Mudrocks: their Distribution, Timing and Genesis

In all cores, BHIs and outcrops, fractured and non-fractured beds have been observed. Natural fractures have different spatial distributions through the studied section. Three main factors are interpreted here to have controlled fracture formation and distribution. First, the host-rock lithology appears to have affected fracture development in terms of mechanical properties of rocks. Rock properties have been argued to be a main control on fracture development and distribution (Gross et al., 1995; Underwood et al., 2003; Cooke et al., 2006; Laubach et al., 2009; Couples, 2013). Based on the subsurface and outcrop examination, the individual fracture systems are mostly clustered in the Lower and Upper chert-rich units and are mainly confined to their host beds. This reflects that fractures are mainly abundant in the rocks that show stiff or brittle mechanical characteristics (e.g., chert and silicified intervals). The relatively low-fracture abundance in the homogenous organic-rich marl unit (Middle) supports the observation that the rock properties have affected fracturing in the section. The occurrence of hydraulic fractures (e.g., bitumen veins; bedding-parallel calcite veins) in this unit however can be explained by the lowpermeable nature of mudstone that results in fluid overpressuring. Bed thickness is also considered as a major control on distribution, particularly on fracture spacing (Ladeira & Price, 1981; Narr & Suppe, 1991; Gross et al., 1995; Bai & Pollard, 2000; Underwood et al., 2003; Schöpfer et al., 2011). In the present study, it is evident that fractures are closely spaced in thin beds, whereas they are widely spaced in thick beds (see Figures 3.19 and 4.21 from Chapters 3 and 4, respectively).

The second important factor in fracture development is diagenesis; diagenesis (early and late; Figure 6.1) has been shown to have had a great impact on fracture development in the studied succession. This is clear from the following three points: (1) the development of folded calcite veins in the mudstone intervals that are positioned between, or close to, chert beds or concretions. The consistency in compaction ratios between folded veins and chert concretions indicates that these veins have formed at an early stage of diagenesis and were subsequently folded by sediment compaction. In addition, the timing of formation of these veins has been constrained, using combined relative and absolute dating techniques, to be very early (~45 Ma; see Figures 3.48-3.51, in Chapter 3, and Figure 6.1).

This also reflects the chronostratigraphic age of the host sediments and, in turn, supports the formation of these veins during early diagenesis. The strontium isotopic data of the calcite fracture fill (in the folded calcite veins) closely match that of the host-rock carbonate (see Table 3.5 and Figure 3.49, Chapter 3), suggesting that the strontium (and calcium) originated from diffusion from the host rocks. However, strontium ratios of folded calcite veins are close to the host rock lithology. Supportively, the data also show a strong match with the predicted range on the global seawater curve (see Figures 3.50, Chapter 3). (2) The development of bitumen veins within the low-TOC mudstone zones (halos; see Figure 5.9 in Chapter 5), that are usually present in immediate contact with chert concretions (see Figure 5.10, Chapter 5) and the presence of intermediate silica content (higher than that of the host mudstone and lower than that of the chert concretion; see Figures 5.12 and 5.14 in Chapter 5), suggest that diagenesis can result in overpressure that can subsequently create hydraulic fractures (bitumen veins). It is argued in this study that the halos have developed as a result of the formation of chert concretions, which, in turn, led to the development of BVs within these alteration rims. It is proposed that disseminated organic-matter within the original host organic-rich mudstone migrated

during concretion growth to produce the bitumen veins and concomitant reduction in TOC within the depleted zones (halos). The nucleation of BVs appears to occur in the halos with propagation into the chert concretions (see Figure 5.12, Chapter 5). A huge reduction in disseminated organic matter seems to have occurred in the halos, resulting in the formation of these halos. This is evidenced by their low TOC content compared to the surrounding mudstone (see Figure 5.19, Chapter 5). The very strong relationship between the aperture size of the BVs and the colour intensity of the halos also demonstrates hydraulic fracturing created by the bitumen overpressure (see Figure 5.20, Chapter 5). This is further supported by petrographic and biomarker analyses that indicate that the bitumen in the fractures is endogenously sourced and it represents very low-mature hydrocarbons expelled from the JOS matrix (see Figures 5.16 and 5.18a, Chapter 5). (3) An interesting example of fracturing due to late diagenetic processes can be summarised by the formation of gypsum veins (i.e., white powdery gypsum veins in the Sultani Quarry; see Figure 4.9, Chapter 4) in mudstone. Overpressuring driven by fluid expulsion during diagenetic processes can generate bedding-parallel veins. Such processes have been widely discussed in the literature for different types of fluids (e.g., Freed & Peacor, 1989; Parnell & Carey, 1995; Katahara, 2006, Parnell et al., 2017a, b). Gypsum is a typical weathering product of Cretaceous-Paleogene marine shale rocks in many semiarid regions worldwide (Hattin, 1965; Retallack & Dilcher, 1981; Retallack, 1983). The white powdery gypsum veins observed in the marl-rich unit are interpreted here to have formed as a product of a diagenetic alteration of previously formed gypsum (in the immediately adjacent beddingparallel fibrous gypsum veins). This is evident from the microcrystalline nature of gypsum within the white powdery vein (see Figure 4.25b, e, h, Chapter 4), which suggests transformation of gypsum fibres into microcrystalline gypsum crystals (Testa & Lugli, 2000). The nature of the veins indicates that the veins may have formed by direct,

precipitation of gypsum during the vein development or by fast recrystallization of gypsum fibres, from the median fibrous vein, as microcrystalline gypsum (see Figure 4.25, Chapter 4). The latter suggestion is supported by the presence of discrete gypsum fibres or fibre relics in the microcrystalline part. In both possibilities, gypsum could have been generated from local evaporitic fluids containing sulphur that is abundant in the host sulphur-rich mudstone rocks and deposited by the near-surface groundwater. The near-surface groundwater could have offered the redox state that allowed the production of sulphate that contributed to the formation of gypsum in veins.

The nature of fracture cementation is considered as an important factor that controls the fracture shape; folded veins were, always, observed to be filled only by calcite. Planar veins are filled by either calcite or by both calcite and silica. It is also observed that the filling material usually reveals the host rock lithology. Specifically, folded calcite-filled veins were observed in calcareous mudstone beds; planar silica-filled veins were detected in silicified and cherty beds; and planar calcite-filled veins are common in calcareous organic-rich marls. The general consistency between fracture minerals and their host-rock lithology suggests that the sealing of the different fracture networks has occurred in a chemically-closed system (Czerniakowski et al., 1984; Blyth et al., 2004; Hooker et al., 2017b), which supports the diagenetic control on fracture development.

The third factor that is believed in this study to have controlled or contributed to fracture development and distribution in the study succession is tectonic stresses represented by regional faulting. In most studies, for joints to form, it is suggested that weaknesses in rock can induce local tensile stress governed by external compression or extension and so create open fractures (Hoek & Bieniawski, 1965; Pollard & Aydin, 1988; Aydin, 1978). Internal fluid pressure and thermal contraction are also suggested to be the driving forces that led to the propagation of joints (e.g., Engelder, 1987; Pollard & Aydin, 1988;

Engelder, 1993). Another important factor controlling the formation of open fractures is the contrast in elastic properties of different lithologies (Bourne, 2003). In the present study, such fractures are interpreted to develop, or at least, to be controlled by tectonic stresses. This is mainly supported by the preferred NNW-SSE fracture trend of such fractures (see Figures 3.54 and 4.23 in Chapters 3 and 4, respectively), which is parallel to the trend of the regional normal faults in the area (see Figure 2.8; Chapter 2) and consistent with the present-day S_{HMAX} .

Finally, the dating techniques outlined above constrain the timing of the development of the different fracture systems. Isotope studies provide absolute ages for the fracture events (Figures 6.1). Fracture timing was also constrained by the evidence that cross-cutting relationships between the fractures and some of the diagenetic features such as concretions. Concretion growth, in general, seems to have played an important role in fracture development; growth of displacive carbonate concretions (see Figures 4.14b, Chapter 4) resulted in fracture development in adjacent chert beds that are folded (distorted) around the concretions. Folding is interpreted to have been enhanced by increased displacive mineral growth within the margins of the concretions in comparison to the concretion centres (Figure 6.2). Fracturing is also argued to be highly developed around replacive concretions, which is thought to result from differential compaction around such concretions.

To summarise, Figure 6.3 shows the sequence of events (through time) from the deposition of carbonates, organic matter and chert sols (opal CT-chert transition), through folded calcite veins, silica veins, concretion development, bitumen depleted zones (halos), and late calcite veins.

6.2 Regional Fracture Distribution

The availability of continuous vertical cores across the entire succession (Upper Cretaceous-Eocene), together with the borehole images and wireline-log data of both the cored and non-cored wells, allowed a unique opportunity for a detailed structural/stratigraphic analysis of the fracture systems. The study of the equivalent interval from outcrop analogues has also provided a comprehensive study of the succession. Regional correlations performed between the fracture data collected from the different sources, linking central west (Harrana), central east (Azraq) and southern Jordan (Bayer), have allowed the investigation of fracture intensities and distribution in the study succession.

Correlations based on core, BHIs and outcrop-derived data were carried out through regional transects (Figure 6.4). The transects were designed to compile the structural data measured from the different sources used in this study and to allow a better understanding of the variation in fracture distribution between the study areas. Moreover, these transects allow a synthesis of the regional fracture distribution with respect to the major faults in the region and place fractures in a burial/uplift history.

Lithostratigraphic and structural correlations performed between the study wells allowed a regional understanding of the fracture distribution among the study areas (Figures 6.5 and 6.6). A west-east correlation provided information on the influence of major faulting on fracture intensities and lateral distributions between the Harrana and Azraq Basins. The transect is sub-parallel to the strikes of two major faults in the area (Suwaqa and Zarqa Ma'in strike-slip faults) (Diabat, 2005, 2006). The correlation panels allowed the construction of an integrated stratigraphic-structural model that helps to understand the structural evolution of the basin (Figure 6.7).

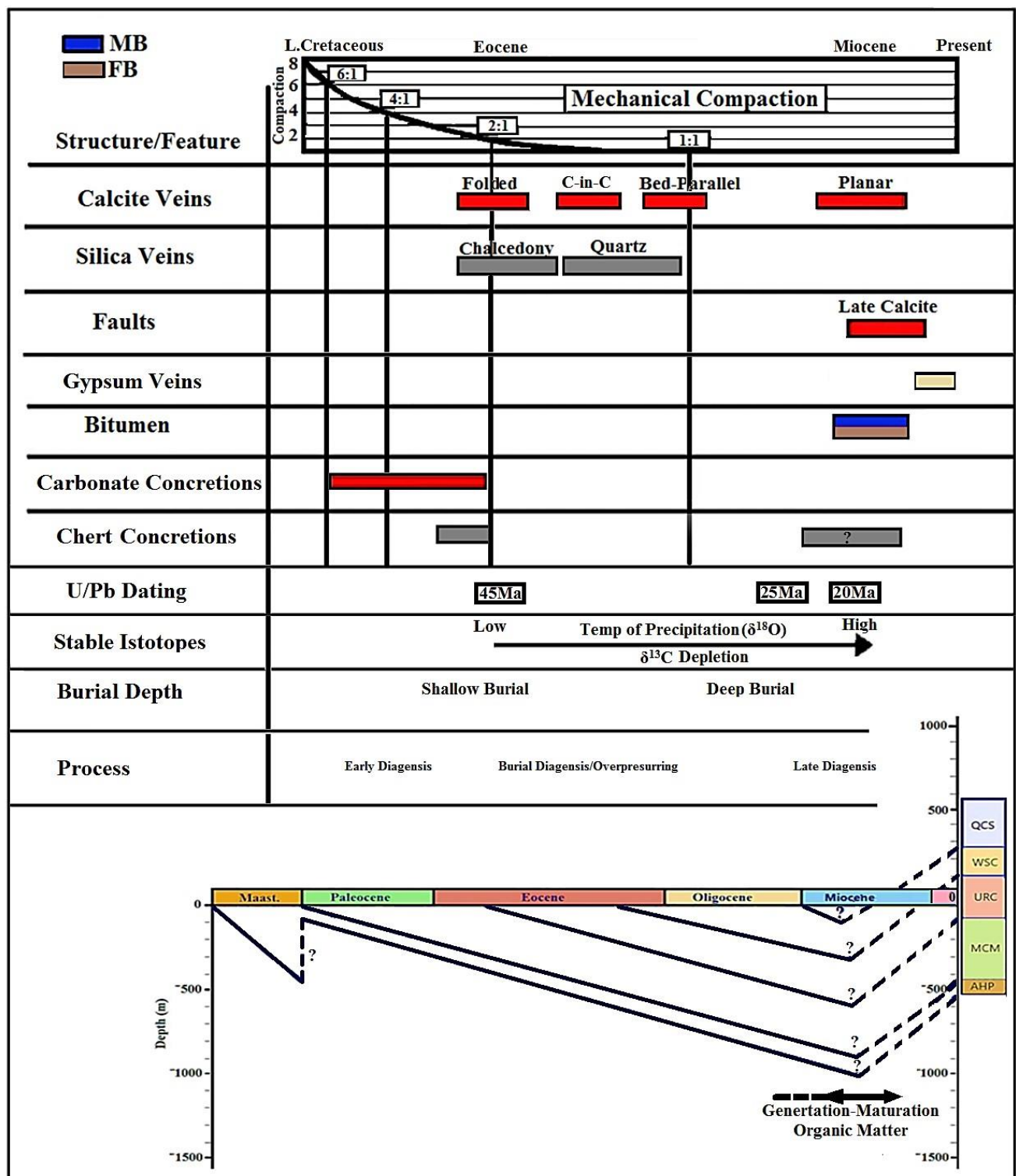


Figure 6.1 A paragenetic sequence of the different fracture cements. The timing of the different features is constrained based on both relative (crosscutting relationships, compaction ratio and stable isotopes), and absolute (U-Pb dating; for calcite structures) analyses. MB, Matrix Bitumen, FB, Fracture Bitumen; AHP, Al-Hisa Phosphorite Formation; MCM, Muwaqger Chalk-Marl Formation; URC, Umm Rijam Chert-Limestone Formation; WSC, Wadi Shallalah Chalk Formation; QCS, Qirna Sandstone Formation.



Figure 6.2 Field photographs showing the folding of chert beds around carbonate concretions. Yellow lines in the photo on the right represent silica veins that show orthogonal direction to the host chert bed that is folded around displacive carbonate concretion.

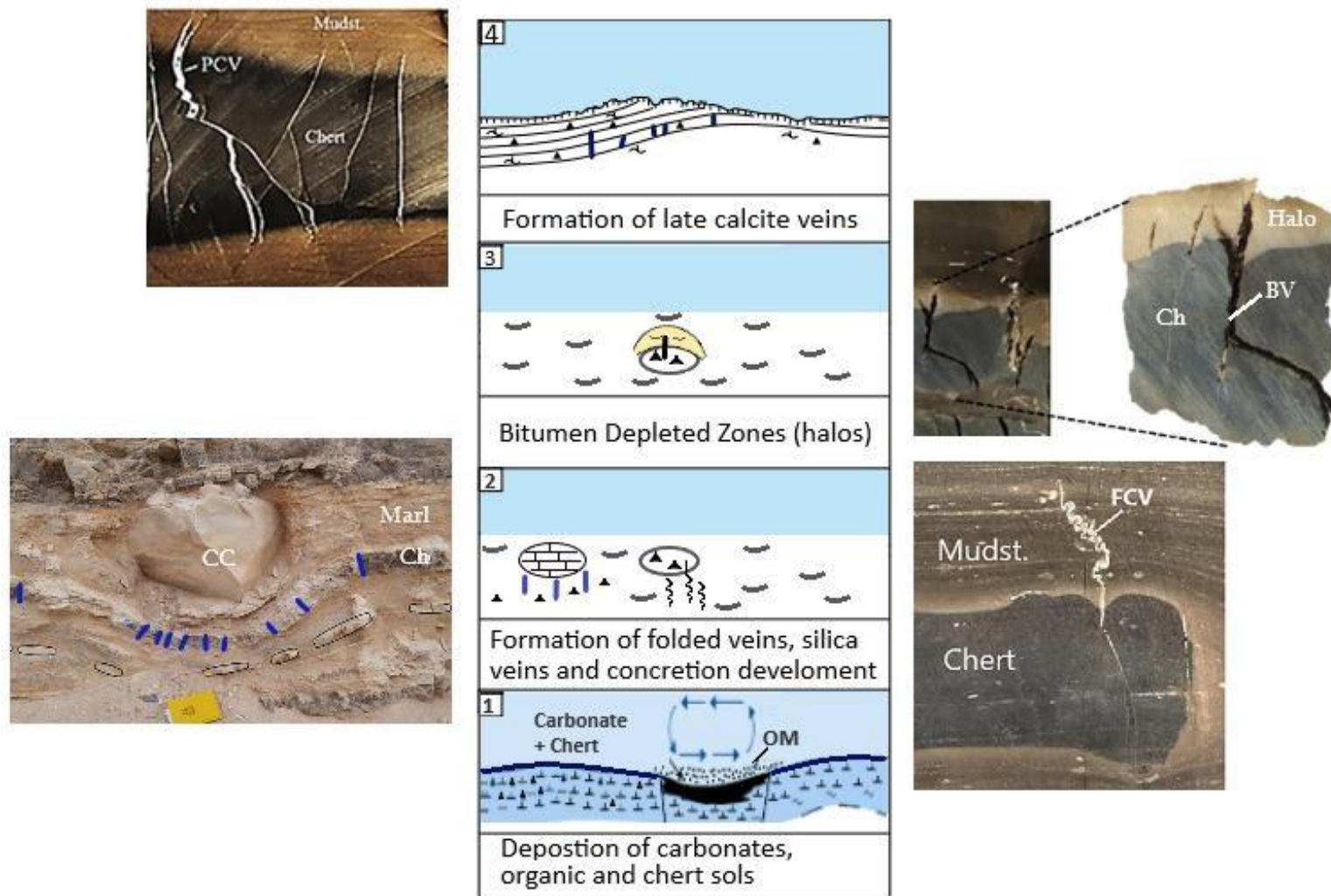


Figure 6.3 A schematic drawing with core and field photographs illustrating the sequence of events (1-4) of the study succession. PCV, Planar Calcite Vein; Ch, Chert; BV, Bitumen Vein; FCV, Folded Calcite Vein; CC, Carbonate Concretion.

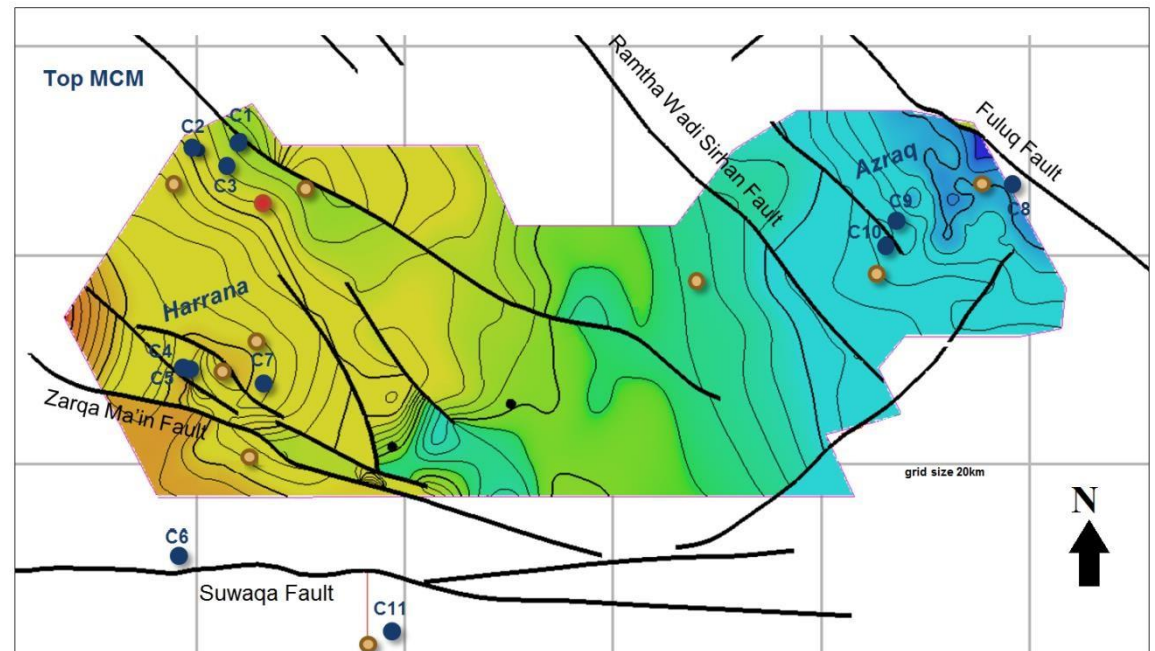
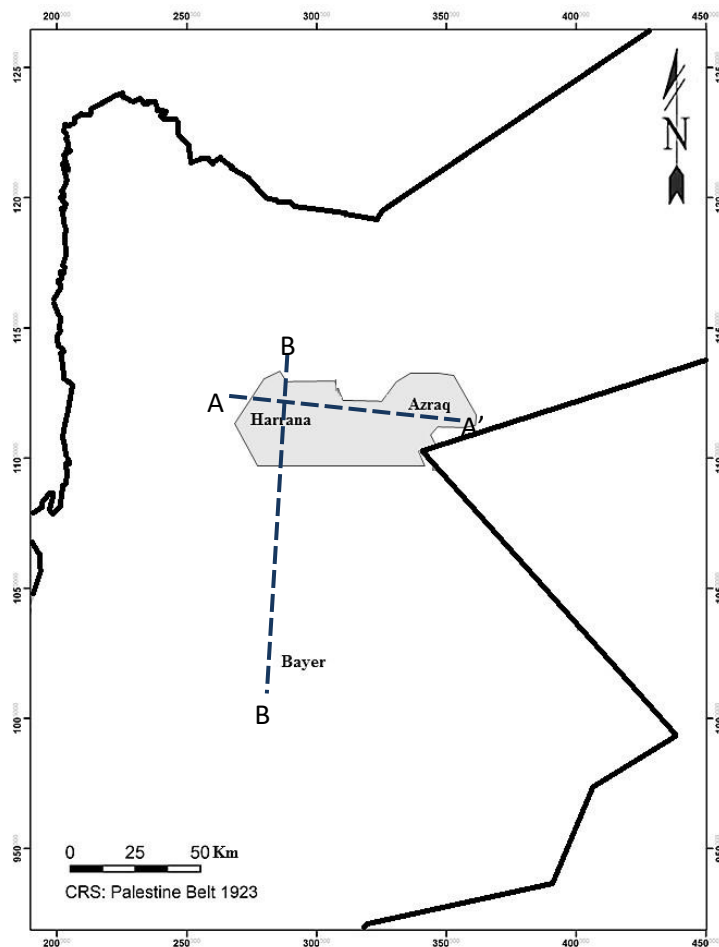


Figure 6.4 Map of Jordan showing transects designed for correlation purposes. on the right, top MCM map built by JOSCO on over 150 well tops and the main tectonic features. Coloured circles show the location of cores (blue), cutting wells (brown) and outcrop used in the W-E correlation. The Harrana and Azraq sub-basins are separated by a set of syn-sedimentary faults (generally oriented NW-SE). The MCM progressively deepens eastwards.

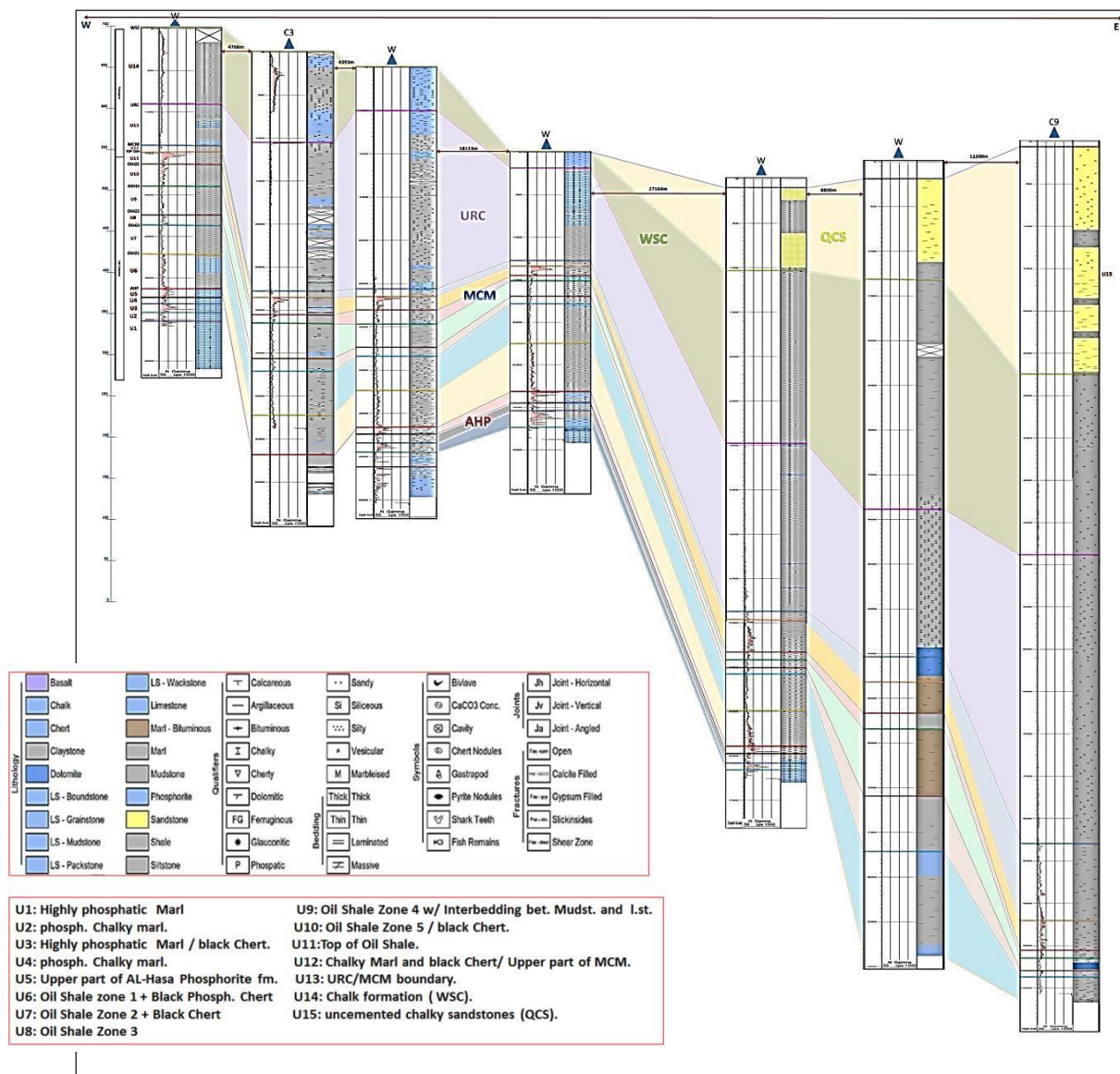


Figure 6.5 An example of W-E cross section (transect A-A' in Figure 6.4) showing the lithostratigraphic correlations conducted between the study wells. Note the change in formation thickness and depth from west to east.

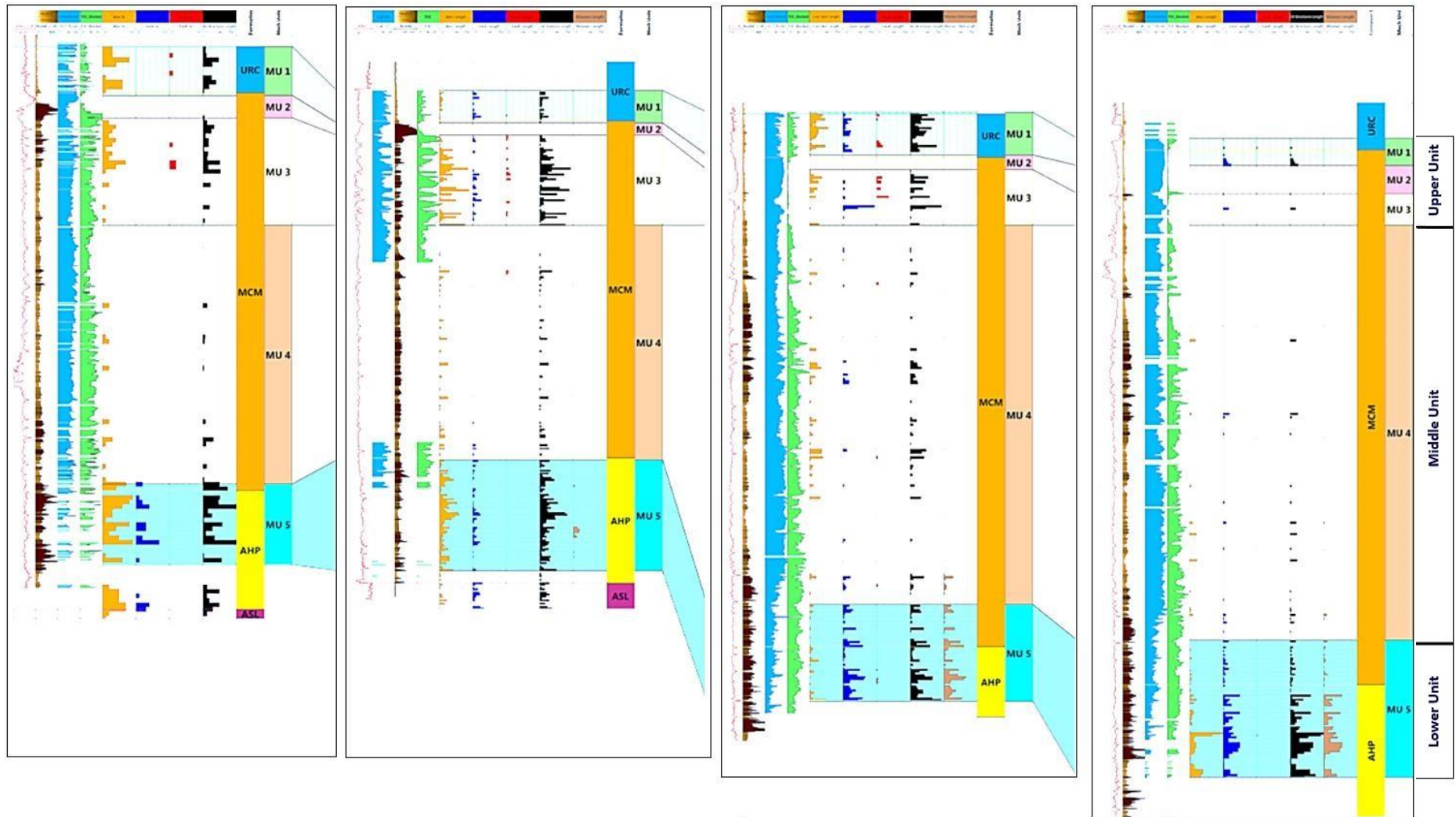


Figure 6.6 West-east regional correlation panel of fracture distribution. Cores used (from left to right) are C3, C5, C10, and C9. Core Locations are shown in Figure 6.4.

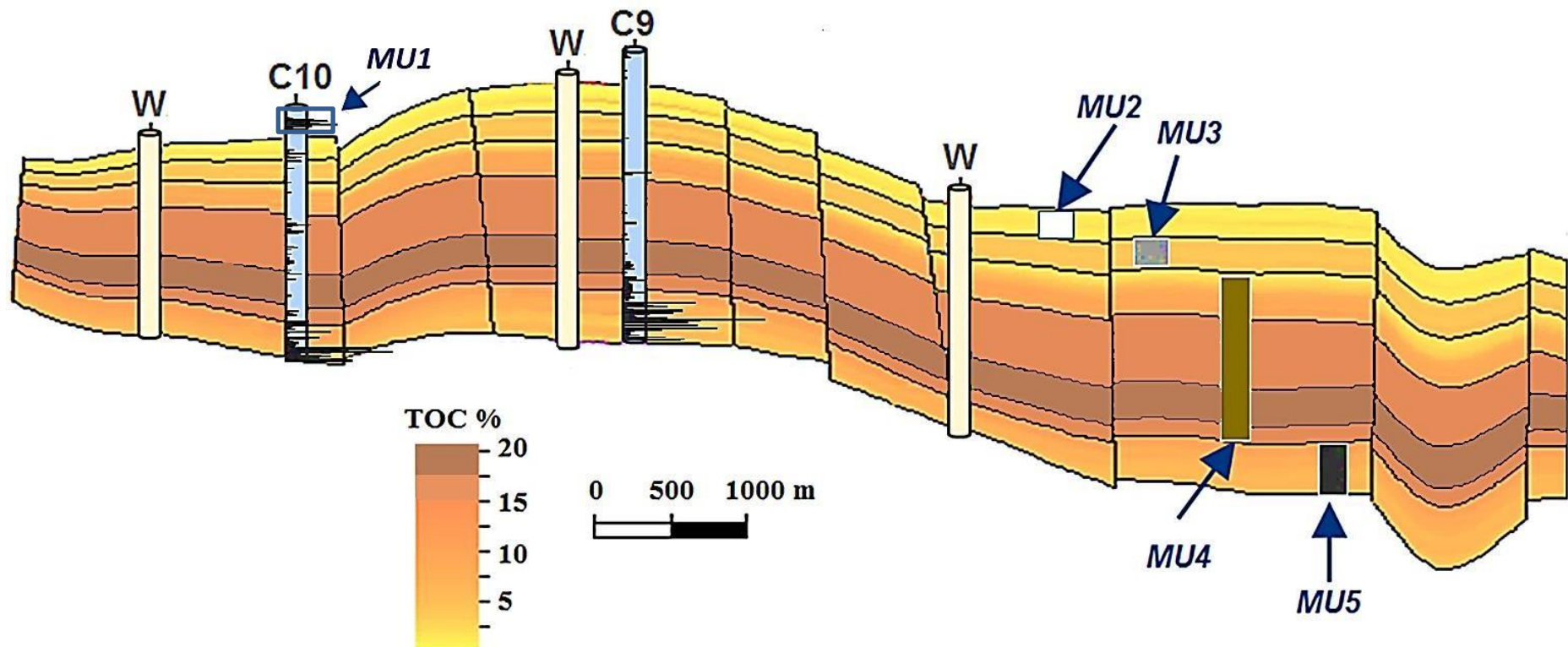


Figure 6.7 A 2D model of the Azraq Basin. Mechanical units (MU): MU2, non-fractured unit; MU3 moderately-fractured unit; MU4, Slightly-fractured unit; MU5, Highly-fractured unit. W, cuttings well. Note the heterogeneity in the vertical and lateral fracture distributions, with respect to the TOC.

Chapter 7: Conclusions and Future Work

7.1 Conclusions

This thesis research aimed to characterise the natural fracture systems, to investigate the spatial distributions of fractures and to constrain the timing of their development in unconventional shale reservoirs using integrated observational and analytical methods that vary from macroscopic characterization (field, core and borehole image logs), geochemical analyses (organic and inorganic) to microscopic (optical microscopy and SEM) investigation. The present study represents the first comprehensive study of the natural fracture systems in the Upper Cretaceous-Paleogene succession, representing the oil shale interval (MCM) and the over- and underlying chert-rich units (URC, and AHP/ASL, respectively).

The anisotropy in rock properties through the section allowed a unique opportunity to examine fracture propagation within different lithologies. Key findings of this research are as follows:

- (1) A wide variety of natural fracture subsystems are present in the study succession, comprising veins, cone-in-cone structures, joints, bitumen-filled fractures, and faults. Opening-mode fractures (veins and joints) are, however, the most commonly observed fractures.
- (2) Several different factors have controlled the distribution of fractures in the study succession. These include rock lithology, diagenesis, mechanical rock properties, and bed thickness. Based on core and outcrop studies, fractures are mostly clustered in the units that have been silicified and in the chert-rich intervals (Lower and Upper), and are mainly confined to individual beds. Low fracture abundance was observed in the homogenous organic-rich marl unit (Middle). Bed thickness is a major control on

fracture spacing and frequency; fractures are closely-spaced in thin beds, whereas wider spacing is usually present in thick beds.

- (3) Planar fractures (i.e., planar calcite veins and joints) show preferred NNW-SSE fracture orientation, a trend parallel to the regional normal faults in the area, and consistent with the present-day regional maximum horizontal stress (SH_{MAX}).
- (4) Fracture minerals are related to the host-rock lithology, generally at the bed level, and therefore, the system is considered to have been chemically closed at the time of fracture development (opening and sealing) and concretion growth.
- (5) Based on the crosscutting relationships between fractures seen in the core and at outcrop, together with petrographic investigation and geochemical analysis, the natural fracture systems in the study succession reveals multiple stages and timing of fracture development; (a) early fractures including folded calcite veins, bedding-parallel veins and silica veins and (b) late fractures such as late tectonic fractures (planar calcite veins, joints and faults), and (c) late diagenetic fractures such as the white powdery gypsum veins. Folded calcite veins formed very early in the diagenetic history, prior to any significant lithification, and were later folded by vertical compaction of the host sediments. Late calcite veins often show crack-seal texture and contain bitumen inclusions indicating that bitumen occurrence is syngenetic with the cementation of late calcite veins, and was generated relatively late in the paragenetic sequence.
- (6) $\delta^{13}C$ and $\delta^{18}O$ signatures of the early structures are consistent with the low-temperature precipitation of carbonate derived from organic material in the host sediment, while isotope results of planar calcite veins are consistent with higher-temperature precipitation and dissolution/re-precipitation of host-rock carbonates during subsequent burial phases.

- (7) The biomarker and petrographic results of fracture and matrix bitumen show that the source of bitumen within fractures and adjacent vugs are very low-mature hydrocarbons expelled from the JOS matrix, and have not migrated from another petroleum source. Organic fingerprinting of bitumen indicates a marine environment of deposition.
- (8) Different mechanisms have controlled fracture formation and development, either individually or in combination, in the succession: (a) fluid pressure (e.g., bitumen veins, gypsum veins and bed-parallel veins), (b) early diagenesis (e.g., folded calcite veins and planar silica veins) and late tectonic processes (planar veins, and joints associated with fault-damage zones), (c) late diagenetic process (e.g., white powdery gypsum veins), and (d) force of crystallization (e.g., fibrous gypsum veins).
- (9) Diagenetic processes (e.g., silica diagenesis) have contributed to the bitumen mobilization, the formation of low organic zones ('halos') and the development of bitumen veins in the Lower Unit of the succession.
- (10) The growth of concretions has also contributed to the fracture development in the succession. Displacive concretions in the Lower Unit of the studied section show good examples of fracturing as a result of, or enhanced by, the growth of concretions.

7.2 Implications for Resources Exploitation

Different properties of natural fracture systems such as distribution, orientation and fracture interaction and connectivity, have a significant impact on fluid flow and storage (Nelson, 2001; Spence et al., 2014). Determination of the different fracture properties such as type, geometry, frequency, fracture aperture, and fracture porosity, is critical for exploration, development and production of hydrocarbon reservoirs (Prioul & Jocker, 2009; Zeng et al.,

2013; Lai et al., 2017, 2018). Natural fractures can act as significant hydraulic conduits or barriers (Aguilera, 1995; Nelson, 2001).

In the present study, although the Lower and Upper caps of the studied succession are highly fractured, subsurface fractures described from both the cores and BHIs show that the majority of fractures observed in these two units are formed of completely-cemented fractures. This can be crucial for production purposes, since cemented fractures usually act as fluid barriers (Laubach, 2003), minimizing the risk of fluid loss and maintaining sufficient containment of the potential reservoir. But, they can be easily opened during operations such as hydraulic fracturing (fracking).

Open fractures (i.e., joints) are documented in many studies to form in the subsurface (e.g., Laubach, 1988), even where the least-compressive stress is compressive (Engelder, 1993). The presence of open fractures in shale reservoirs can have a great impact on reservoir recovery (Nelson, 2001). They enhance the permeability of reservoirs when they are interconnected (Leung & Zimmerman, 2012). In such cases, they can act as fluid conduits (Nelson, 1985; Laubach, 2003). In the case of the JOS, such an impact is reflected in the Middle Unit of the core from Well C10 in the Azraq Basin, where open fractures are observed interconnected in the mudstone unit (see Figure 3.7g, Chapter 3). But in both the Lower and Upper chert-rich units, open features are poorly interconnected; they are usually present either as single fractures or as unconnected, parallel arrangements, which are in both cases, confined to beds (see Figure 3.7a-b, Chapter 3). This can also help in preserving the containment required for the potential reservoir.

Bitumen-bearing fractures are observed also only in the Lower Unit, and they show higher abundance in the wells from west to east; the highest cluster of bitumen fractures is observed in cores from the Azraq Basin to the east (see Figure 5.8; Chapter 5). This suggests a positive

correlation between the burial depth and bitumen generation and migration (see Chapter 5 for more details).

7.3 Areas for Further Study

This thesis provides a general framework of the different fracture systems and concretion growth present in the JOS and the over- and underlying chert-rich intervals. Further specific studies on each fracture type, specifically, are suggested. In addition, the new observations that have been documented by this study on the displacive growth of concretions, including the formation of triangular spaces filled by chert at the edges of carbonate concretions, and the folding of chert beds and the asymmetrical strain fields around concretions, open many research questions that need to be addressed. Therefore, future study of these diagenetic features, and an understanding of their temporal relationships to fracturing is highly recommended.

In addition, there is a need to understand the relationship between bitumen fractures/pathways and high temperature (pyrometamorphic) ‘marble zones’ seen in core and in central Jordan.

There is also a scope to investigate geochemical fingerprinting of Dead Sea oil/bitumen occurrences adjacent to the DST (washed-up bitumen blocks and impregnations in the Kurnub Sandstone) in relation to the JOS bitumen to test for similar origins and/or migration/maturation histories?

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Appendices

Appendix 1: A Table lists the samples used for stable and strontium isotopic analysis

Sample ID	Core Name	Stratigraphic Unit	Sample ID	Core Name	Stratigraphic Unit
F1	C1	Upper	F22	C6	Lower
F2	C1	Middle	F23	C7	Upper
F3	C1	Middle	F24	C7	Middle
F4	C1	Lower	F25	C7	Middle
F5	C1	Lower	F26	C7	Lower
F6	C2	Middle	F27	C8	Upper
F7	C2	Lower	F28	C8	Middle
F8	C3	Lower	F29	C8	Lower
F9	C3	Lower	F30	C9	Lower
F10	C4	Lower	F31	C10	Lower
F11	C5	Lower	F32	C11	Lower
F12	C6	Upper	F33	C12	Middle
F13	C6	Upper	F34	C12	Lower
F14	C6	Upper	F35	C13	Upper
F15	C6	Middle	F36	C13	Upper
F16	C6	Middle	F37	C13	Middle
F17	C6	Middle	F38	C13	Middle
F18	C6	Middle	F39	C13	Lower
F19	C6	Middle	F40	C14	Middle
F20	C6	Lower	F41	C15	Lower
F21	C6	Lower	F42	C15	Middle

Appendix 2: *A rough summary of the procedure used for sample separation for the $^{87}\text{Sr}/^{86}\text{Sr}$ analysis.*

The column racks are such that 12 samples (11 natural samples plus a blank) can be run at once.

Resin washing is done in a batch with a dedicated teflon vial set, so that can be done at any time, including before teflon cleaning. But leaching samples and the column method require clean teflon in advance.

Teflon and resin cleaning take multiple days but do not require continuous attention; rather, 2-3 trips into the lab each day, for only several minutes at a time.

1. Teflon cleaning (~3 days):

- ✦ First step, on newly-ordered teflon only: Wyatt lab: wash in acetone. Slosh in beaker for one minute. Leave acetone to evaporate. Rinse teflon in DI water twice.
- ✦ Reflux in 1ML in each vial concentrated HNO_3 and put on hotplate overnight.
- ✦ Drain HNO_3 into waste container the next day and put Teflon into a quartz beaker.
 - ✦ Rinse Teflon with MQ water and drain.
- ✦ Fill up beaker with enough 40% HCL to submerge Teflon then put on hotplate overnight.
- ✦ Pour away acid to container the next day (save the cleaning acid!) and rinse with MQ once.
- ✦ Add 30% HNO_3 to submerge the Teflon and put on hotplate overnight.
- ✦ Next day drain the acid to a container, and rinse twice in cold MQ.
- ✦ Submerge Teflon in MQ and put on hotplate for 3 hours. Repeat 2-3 more times. ✦
Leave to dry on the bench covered in a kimwipe OR in a fume hood.

Requires (2L) jugs of cleaning acid (HNO_3 and HCL) and ~50 mL conc HNO_3

2. Weighing samples (~ 0.5 day)

3. Leaching samples, to analyze phases separately (~1 day for each leach):

Exchangeable fraction: MQ (pH ~3)

Carbonates: dilute ascectic (avoid organic)

Oxides: citric acid plus hydrous hydrazine—probably don't need

Organics: HNO_3 + H_2O_2 , heat to 85C for 2 hours

Residual: HF

Requires clean acids as noted, ~100 mL each

3. Resin cleaning (~2 days):

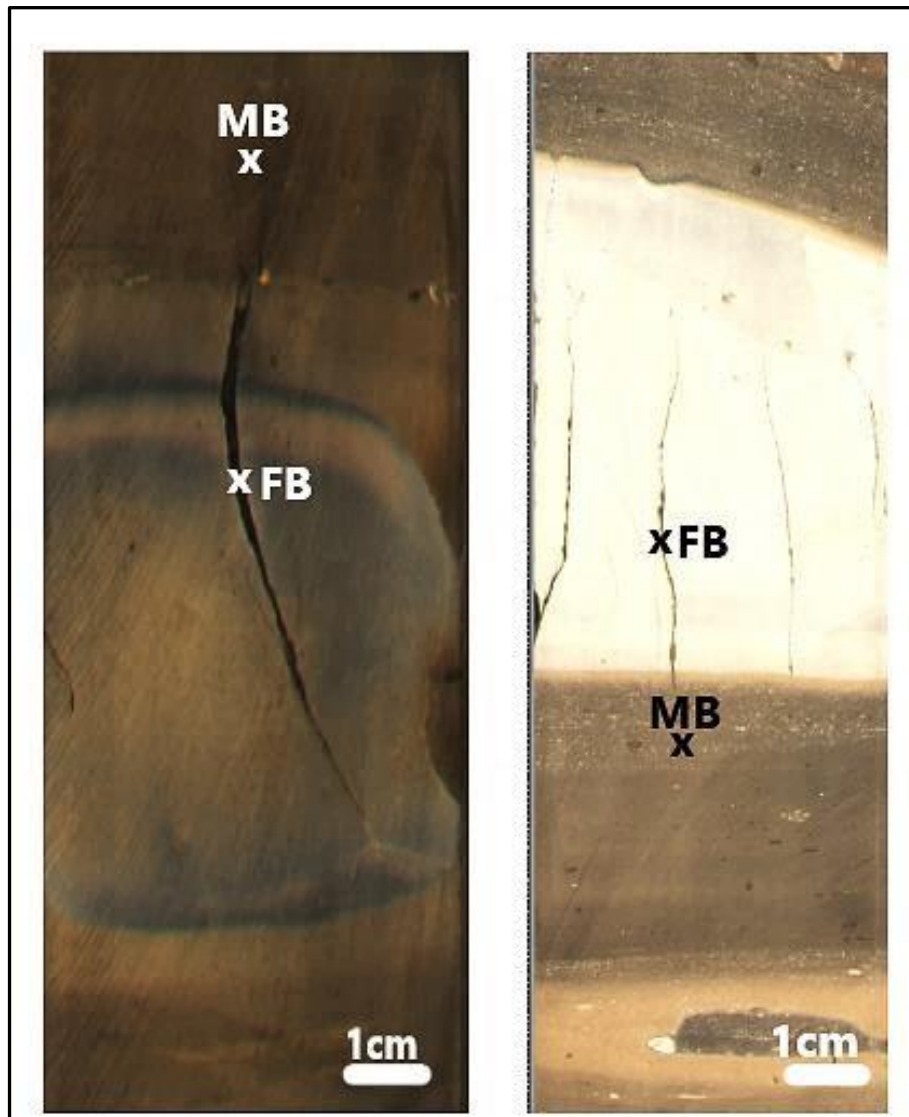
- ✦ MQ x 5, then HNO₃-MQ-HNO₃-MQ-HNO₃-MQ-HNO₃-MQ-MQ
- ✦ bold HNO₃s are 3M, non-bold are 0.1 M. Keep 3M HNO₃s to < 3 hours
- ✦ Store in “weak” HNO₃: 0.1 M

Requires ~ 100 mL clean conc. HNO₃

4. Chemical separation (column method) (~7 hours, more or less continuous):

- ✦ 150 µL resin slurry
- ✦ Wash with 200 µL MQ
- ✦ 2 mL 6M HCl
- ✦ 2X 2 mL MQ
- ✦ 1 mL 7M HNO₃
- ✦ Resin condition: 1 mL 2M HNO₃
- ✦ Sample load into two sets of 250 µL 2M HNO₃
- ✦ Wash sample out with 250 µL and 150 µL 2M HNO₃
- ✦ Elute matrix with 1 mL 7M HNO₃
- ✦ Collect Sr in 1.5 mL 0.05M HNO₃
- ✦ Dry Sr fraction overnight at 90C
- ✦ Requires ~200 mL clean conc. HNO₃, HCl
- ✦ Then you are ready to prepare your column elutions for the mass spec, which takes about a day of prep-work then a day to analyze.

Appendix 3: Sampling of Fracture Bitumen (FB) and Matrix Bitumen (MB).



Appendix 4: *m/z 191 and m/z 217 fragmentograms showing the distribution of terpanes and steranes, respectively, in studied matrix and fracture bitumen samples. Fracture Bitumen, FB; Matrix bitumen, MB.*

