Fluvial and aeolian interactions in modern and ancient dryland continental sedimentary systems: implications for reservoir heterogeneity

MOHAMMED ALI JBREEL ALMASRAHY

Submitted in accordance with the requirements for the degree of
Doctor of Philosophy

The University of Leeds
School of Earth and Environment

December 2015

Intellectual Property and Publication Statements

The candidate confirms that the work submitted is his/her own, except where work which has formed part of jointly authored publications has been included. The contribution of the candidate and the other authors to this work has been explicitly indicated below. The candidate confirms that appropriate credit has been given within the thesis where reference has been made to the work of others. See the Preface (page iv) for further information.

Acknowledgements

I would like to express my sincere gratitude to my advisor Dr. Nigel Mountney for the continuous support of my Ph.D study and related research, for his patience, motivation, and immense knowledge. His guidance helped me throughout the research programme, and in the writing of this thesis. I could not have imagined having a better advisor and mentor for my Ph.D study.

I express my profound gratitude to my sponsor Saudi Aramco Company (Exploration Organization, Reservoir Characterization Department, and Educational Partnerships Department), for giving me the opportunity to pursue my Ph.D. Special thanks are due to Dr. John Melvin, Dr. Aus Al-Tawil and Mrs Aggie Copper, for all their encouragement and support.

I am also grateful to the University of Leeds, School of Earth and Environment for all the support I received throughout my study.

My sincere thanks goes to my mother whose prayers are always with me, to my wife, my daughters and to my brothers and sisters for supporting me spiritually throughout the writing of this thesis and my life in general. Thanks are also expressed to my friend, Mahmud Alkathery.

Preface

Two chapters within this thesis (Chapters 2 and 3) have been written as papers; these have each been published as articles in the peer-reviewed journal *Aeolian Research*. Furthermore, Chapter 4 has also been written in the style of a paper. Thus, each of these three chapters is presented in this thesis as stand-alone piece of work; they are linked by a common research theme. The literature review and background sections within these chapters deal, in part, with aspects of the wider research philosophy adopted in this study programme; as such, there is a modest amount of overlap in these sections of these chapters. Such overlap is useful since it serves to remind the reader of the commonality of the overall research theme as a linked thread that runs throughout the thesis.

Publications (at time of submission)

i. **Mohammed A. Al-Masrahy, Nigel P. Mountney,** (2013). Remote sensing of spatial variability in aeolian dune and interdune morphology in the Rub' Al-Khali, Saudi Arabia. Aeolian Research, v. 11, 155-170.

ii. **Mohammed A. Al-Masrahy, Nigel P. Mountney,** (2015). A classification scheme for Fluvial-aeolian system interaction in desert-margin settings. Aeolian Research, v. 17, 67-88.

Abstract

Quantitative stratigraphic prediction of the three-dimensional form of sedimentary architectures and associated heterogeneities arising from fluvial and aeolian interactions and preserved as accumulated stratigraphic successions is notoriously difficult, meaning that prediction of 3D stratigraphic architectures in subsurface fluvial and aeolian reservoirs is challenging.

This study comprises four discrete but related research components: (1) analysis of aeolian dune-field geomorphology through a remotely sensed analysis of four parts of the Al Rub' Al-Khali Desert, Saudi Arabia; (2) analysis of types of aeolian-fluvial system interaction in modern dune-field margins through study of the morphological expression and areal distribution of 130 examples of fluvial-aeolian interaction mapped by high-resolution satellite imagery from 60 deserts around the world; (3) analysis of the preserved stratigraphic expression of an ancient mixed aeolian and fluvial succession via analysis of the upper part of the Wilmslow Sandstone and the lower part of the overlying Helsby Sandstone formations, Sherwood Sandstone Group, UK; (4) development of a series of predictive, semi-quantitative facies models with which to account for the geological complexity and origin of mixed aeolian-fluvial successions.

Principal finding are as follows: (1) observations from the Rub' Al-Khali Desert have enabled the spatial rate of change of morphology of aeolian sub-environments to be characterized and described through a series of empirical relationships; (2) aeolian-fluvial interaction case-study examples have been classified to propose a framework of ten distinct types of system interactions; (3) outcrop analysis of an ancient preserved succession reveals mechanisms for the accumulation and preservation of aeolian and fluvial successions, and demonstrates the role of water table on the development and preservation of a water-table influenced aeolian system; (4) results from this study have enabled the development of facies models that serve as the basis for gaining an improved understanding of controls governing the detailed sedimentary architecture of preserved aeolian-fluvial successions.

Table of Contents

List of Tables

List of Figures

Chapter One
Introduction

This chapter provides a general introduction to the research topic, discusses the rationale that underpins the research, states the aim and objectives, summarises the data types and methods used for the research, and describes the thesis structure. Each subsequent chapter is outlined in turn in order to summarise the thesis.

1.1 Introduction

Throughout the 1980s and 1990s, studies of the geomorphology of desert sand dunes were dominated by field studies of wind flow and sand flow over individual dunes (e.g. Livingstone and Warren, 1996). More recently, attention has shifted to some extent to modelling approaches, and progress has been made in developing models that relate the geomorphology of dune fields and their temporal and spatial development to likely preserved sedimentary architecture (e.g. Mountney, 2006a and b, 2012; Rubin and Carter, 2006). It is clear, however, that such models require substantiation using quantitative field observations if they are to be constrained and used effectively for the development of improved predictions regarding the relationship between dune-field morphology and gross-scale preserved stratigraphic architecture.

Many sand seas exhibit clear patterns of combinations and transitions of dune types, dune size and spacing, crest orientation and sediment thickness (Breed et al., 1979; Edgell, 1989; Lancaster 1999; Ewing et al. 2006; Al-Masrahy and Mountney, 2013). Since the late 1970s, satellite images have been used to observe and illustrate spatially and temporally changing patterns of aeolian sedimentation (dune and interdune type and style of interaction), including bedform distributions in major sand seas, such as the Rub' Al-Khali, Saudi Arabia and surrounding region (e.g. Breed et al., 1979). Large compound and complex bedforms (dunes and draa) of the Rub' Al-Khali and other major sandy deserts are imaged in detail on the latest generation of public-release satellite imagery. The near worldwide coverage and ease of availability of such data sources, which offer high-resolution imagery, provides a valuable opportunity for the development of new and innovative research methods.

Interactions between aeolian and fluvial systems take place within the marginal areas of many aeolian dune-fields, and in the central parts of others. The location of zones where aeolian-fluvial system interactions take place may vary in time and space. In particular, aeolian-fluvial interactions can affect the extent, shape and boundaries of an individual dune field (Al-Masrahy and Mountney, 2015). In dryland settings, both fluvial and aeolian processes may serve to erode, transport and deposit sources of sediment (Visser et al., 2004); sediment eroded and transported by one type of process (water or wind) may become material available for subsequent transport and eventual deposition by the other type (Belnap, et al., 2011). Perennial or ephemeral rivers can act to intercept, trap or erode aeolian sediment, such that fluvial channels may disrupt or even halt the downwind movement of aeolian sediment (Bullard and McTainsh, 2003; Thomas et al. 1997). Similarly, aeolian processes may act to winnow the fine sand and silt fractions from ephemeral fluvial channel or sheet-like deposits during dry episodes (e.g., Al-Farraj and Harvey, 2000, Prospero and Lamb, 2003; Painter et al., 2010); clay, silt, sand, and even very small pebbles that form the deposits of deposits channel bases, sandbars or fan surfaces in fluvial and alluvial systems are susceptible to aeolian erosion processes. Thus, fluvial deposits can form an important source of sediment for later aeolian dune construction (e.g., Blair and McPherson, 2009; Draut, 2012).

Interactions between subaqueous and aeolian processes in aeolian dune fields may occur as a consequence of flooding by streams or rivers, by tidal or storm surge flooding in coastal dune fields, or by flooding in response to groundwater-table rise (Fryberger, 1990). Such interactions generate distinctive associations of facies (Mountney et al., 2006) and architectural-element arrangements (e.g., Jordan and Mountney, 2010; Blakey et al., 1994). Although these types of interaction are documented from subsurface successions (e.g., Meadows and Beach, 1993), the nature of the stratigraphic relationships arising from such interactions are difficult to determine from subsurface datasets alone. Constraining the spatial and temporal scales at which aeolian and non-aeolian systems interact is important to improve our understanding of the evolution of landscape geomorphology arising from different types of system interaction. From an applied perspective, such studies are required to increase our ability to generate more accurate subsurface geological models to describe heterogeneity in subsurface reservoirs.

Fluvial-aeolian interactions result from both short-term interplay between contiguous aeolian and fluvial systems, and much longer-term expansion and contraction of entire aeolian dune fields in response to climatic, tectonic or eustatic effects (Herries, 1993). At an intermediate spatial scale, active fluvial channels in aeolian dune fields may cut through aeolian sand dunes or sand sheets to form extensive networks (Langford, 1989). Water from these channels may collect and form interdune ponds; where the groundwater table is elevated, interdune areas in entire sand seas may become flooded, and, as a result, the aeolian landforms may be deflated (Trewin, 1993; Mountney, 2006b; Belnap, 2011). Over intermediate time scales, during arid episodes, aeolian sediment may accumulate over abandoned fluvial channel forms. Rare flood events can act to flush aeolian sand from choked fluvial channels (Belnap, 2011). This usually takes place when rivers flow only intermittently and the intervening dry period is sufficiently long for aeolian deposits to accumulate (Bullard and Livingstone 2002; McIntosh, 1983).

Where fluvial channels pass into aeolian dune fields, flows commonly experience a reduction in competence meaning that sand- and gravel-grade sediment is commonly deposited at the outer dune-field (erg) margin. By contrast, finer fractions (clay and silt) are commonly carried predominantly as suspended load further inside the aeolian dune fields via flood flows that exploit open interdune corridors; such sediments commonly accumulate as clay-prone pans or continental sabkhas where dissolved salts are precipitated as flood waters evaporate (Langford, 1989; Glennie, 2005; Edgill, 2006).

The preserved sedimentary signature of aeolian and fluvial interactions has been widely recognised in the ancient rock record (e.g. Andrews, 1981; Loope, 1985; Langford and Chan, 1988, 1989; Trewin, 1993; Heries, 1993; Meadows and Beach, 1993; Jones and Blakey, 1997; Haig et al., 1997; Howell and Mountney, 1997; Mountney et al. 1998; Sweet, 1999; Stanistreet and Stollhofen, 2002; Bullard and McTainsh, 2003; Mountney and Jagger, 2004; Scherer and Lvina, 2005; Veiga and Spalletti, 2007; Simpson et al., 2008; Rodriguez-Lopez et al., 2010; Jordan and Mountney, 2010; Spalletti et al., 2010; Bongiolo and Scherer, 2010; Cain and Mountney, 2011; Mountney, 2012; East et al., 2015). However, generalised facies models have yet to be developed which are able to serve as predictive tools with which to account for lateral changes between aeolian and fluvial facies and elements in aeolian dune-field margins settings where a complex set of interacting processes are known to take place.

From both contemporary and palaeoenvironmental studies, it is recognised that aeolian systems are commonly dependent upon fluvial systems for their sediment supply (e.g., Clarke and Rendell, 1998; Kocurek and Lancaster, 1999). Bullard and Livingstone (2002) suggest that the three main factors controlling the transfer of sediment between fluvial and aeolian systems are moisture availability, the nature of the sediment supply and the magnitude and frequency characteristics of both fluvial and aeolian events in dryland environments. The nature of the sediment supply is an important aspect of the relationship between aeolian and fluvial processes (Bullard and Livingstone, 2002). The magnitude of sand supply is affected by fluvial transport processes, which sort and deliver sand to sites where it is susceptible to subsequent aeolian reworking. Such interaction may determine rates of aeolian erosion or accumulation in downwind areas (Laity, 1995; Gillette and Chen, 2001). For example, sand delivered to the Sinai Desert (Egypt) comprises sediment transported from eastern Africa by the River Nile (Tsoar, 1978). The large quantities of sediment required to construct major sand seas must be delivered over long time periods. This means that other variables operating over long periods of time, such as tectonic activity and eustasy, will also affect the sediment supply (Winspear and Pay, 1995). Basin subsidence plays an important role in generating the accumulation space for long-term aeolian accumulation and preservation, as well as promoting a suitable wind regime for aeolian deposition by encouraging airflow deceleration into developing basinal areas (Wilson, 1971; Fryberger and Ahlbrandt, 1979).

The rate and nature of sediment supply to aeolian systems from fluvial systems is not only a function of sediment production and sorting but is also strongly dependent upon the nature of the channel via which sediment is transported. Flow regimes where low or zero flow is the norm, and channels which are wide, shallow or braided are more likely to be sites of active and regular aeolian deflation. Muhs and Holliday (1995) developed a process-response model to summarise the relationship between climate change and fluvial and aeolian activity in the Great Plains, USA, which associated the formation of wide, braided channels with periods of increased aeolian activity. Lancaster (1997) also found that periods of aeolian construction in the Coachella Valley, south-central California, were determined by stream channel dynamics. At this site, increased sediment supply is associated with periods of channel entrenchment. Sediment is supplied to the dune field via alluvial fans. During periods of increased rainfall and storm intensity, channels on the fans become entrenched and sediments entrained during this process are transported to the distal areas of the alluvial fans, close to the dune field.

At the northern limit of the Namib Sand Sea, large aeolian dunes advance northwards into the Kuiseb Valley near Gobabeb at a rate of approximately 2 m per year (Ward, 1983). However, annual flood events that pass down the Kuiseb River are sufficient to rework and remove any aeolian sand that has moved into the channel. Thus, annual flooding of the ephemeral Kuiseb River prevents dunes from blocking or crossing the valley (Goudie, 1972) and the river serves to delimit and abruptly define the northern margin of the

Namib Sand Sea.

Satellite imagery of dunes and interdunes in desert dune fields has provided the basis of an approach for qualitative and quantitative studies of patterns of arrangement of large-scale aeolian bedforms and adjoining interdunes in large and widely distributed sand seas. Collection and analysis of data relating to primary landform morphology has enabled an improved understanding of modern desert sedimentary systems and the spatial arrangement of various sub-environments within these systems. In particular, the morphological changes and distributions of aeolian bedforms and interdunes across dune-field systems provides important information with which to improve our understanding of the likely arrangement of architectural elements in ancient aeolian preserved successions, several of which form important reservoirs for natural resources such as hydrocarbons or water.

Observations from modern dune-field margins have enabled the spatial rate of change of morphology of aeolian sub-environments to be characterised and described through empirical relationships. Results enable the proposition and development of a range of dynamic facies models for aeolian systems that can be used as predictive tools for subsurface reservoir and aquifer characterisation. A combination of morphological and architectural data from a range of modern dune fields and their ancient counterparts preserved as successions in the geologic record can be used to constrain forward stratigraphic models for the prediction of aeolian reservoir and aquifer heterogeneity. Such heterogeneity is likely to vary in three-dimensions within a reservoir volume.

In ancient aeolian settings, non-aeolian deposits that are known to have accumulated in dune-field margin settings can provide important information about the environmental conditions that prevailed during or between episodes aeolian system construction (Jones and Blakey, 1997). Such evidence is documented from many ancient and recent erg deposits (e.g. Mckee and Moiola, 1975; Ahlbrandt and Fryberger, 1981; Kocurek, 1981; Fryberger et al., 1988; Mountney and Thompson, 2002; Mountney and Jagger, 2004; Rodriguez-Lopez et al., 2010; Cain and Mountney, 2011).

Non-aeolian fluvial or alluvial-fan systems also provide important evidence to demonstrate the relative timing of episodes of deposition and deflation within and adjacent to ancient erg systems (Horne, 1975; Andrew, 1981; Ahlbrandt and Fryberger, 1981; Loope, 1985; Langford and Chan, 1989; Porter, 1987; Anderson and Anderson, 1990).

1.2 Project rationale

Quantitative stratigraphic prediction of the three-dimensional form of sedimentary architecture and associated heterogeneities arising from fluvial and aeolian interaction and preserved as an accumulated stratigraphic succession is notoriously difficult: (i) the preserved products of system interactions observed in one-dimensional cores and well-log data typically do not yield information regarding the likely lateral extent of sand bodies; (ii) stratigraphic heterogeneities typically occur on a scale below seismic resolution and cannot be imaged using such techniques. A database recording the temporal and spatial scales over which aeolian and fluvial events operate and interact in a range of present-day and ancient desert-margin settings has been collated as part of this study using high-resolution satellite imagery, aerial photography and field observation. Together, these data have been used to develop a series of dynamic facies models to predict the arrangement of architectural elements that define gross-scale system architecture. Case-study examples from both modern and ancient systems have enabled the construction of a series of depositional models to account for the diversity of styles of fluvial and aeolian system interactions.

From an applied perspective, aeolian dune and interdune successions form important reservoirs for hydrocarbons, including the Permian Rotliegened Group of North Sea (Howell and Mountney, 1997), the Triassic Ormskirk Sandstone Formation of the East Irish Sea (Herries and Cowan, 1997; Meadows, 2006), the Jurassic Norphlet Sandstone of the Gulf of Mexico (Kugler and Mink, 1999) and the Permian Unayzah Formation of Saudi Arabia (Melvin et al., 2010; Al-Masrahy et al., 2012). Dune facies and elements are typically the most productive lithofacies in aeolian reservoir systems, whereas interdune facies and elements tend to have lower porosities and permeabilities and may act as baffles or barriers to flow. In dune-field margin settings, where fluvial system activities take place, the arrangement of preserved architectural elements can form a complex mosaic of fluvial and aeolian types, and predicting lateral change in reservoir quality in the preserved examples of such successions is challenging. Therefore, the ability to predict the geometry and degree of interconnectivity of these basic element types and changes from central to marginal palaeo-dune-field environments is essential in assessing likely reservoir quality and the distribution of potential baffles and barriers to fluid flow (Weber, 1987; Chandler et al., 1989; Herries, 1993; Stanistreet and Stollhofen, 2002; Taggart et al., 2010; Mountney, 2012).

Significant lithological heterogeneities arise in aeolian successions from the juxtaposition of dune elements with generally favourable reservoir properties against interdune elements that may act as baffles to flow. The size, geometry and type of inter-relationship between aeolian dune and interdune elements tends to change markedly from dune-field centre to dune-field margin settings (Figure 1.1). Prediction of the arrangement of such elements in subsurface successions is therefore important in developing aeolian reservoir models. However, such predictions are difficult because the preserved thickness, continuity and internal facies composition of both dune and interdune elements typically vary spatially both locally and regionally due to the action of a complex set of both autogenic and allogenic controls. Important controls on spatial architectural variability include the morphology and migratory behaviour of the original bedforms and their intervening interdunes at the time of accumulation, as well as climatic, tectonic, water table and sediment supply controls.

Development of an improved understanding of the sedimentological processes that give rise to preserved products of aeolian and mixed aeolian-fluvial successions is essential for creating more realistic reservoir models that better describe subsurface architectures. This study seeks to investigate aeolian depositional systems from both modern and ancient settings. An initial part of this research study focuses on gaining an improved understanding of the distribution of aeolian dunes, interdunes and closely

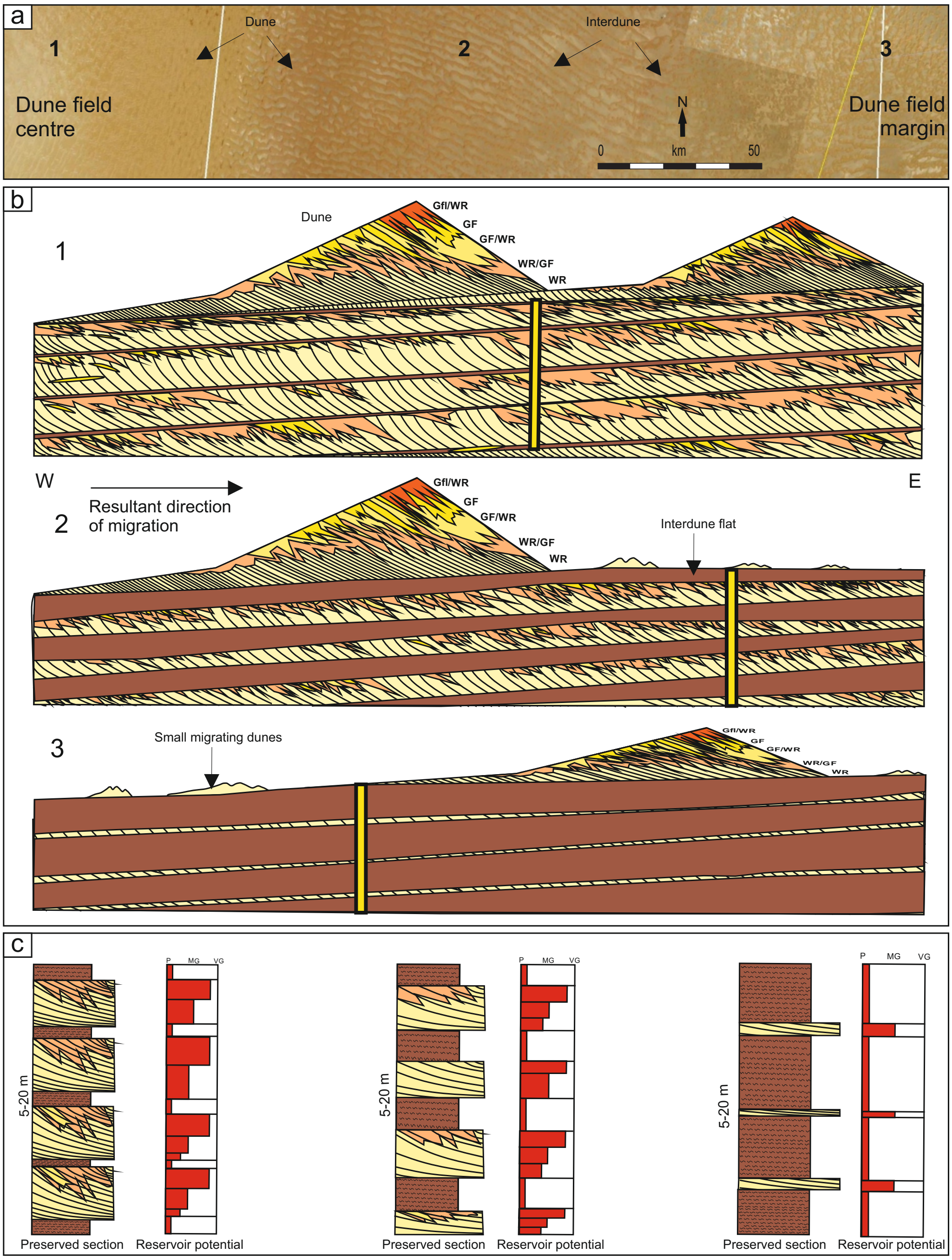


Figure 1.1: Spatial variation of aeolian dune and interdune morphology across the zone transition from a dunefield centre to its margin. a) showing the part of the Rub' A Al-Khali Sand Sea, numbers showing the location of the cross sections in (b); note the change in dune and interdune morphology from the centre to the margin of the dune filed . b) Predicted cross section illustrating the change in lithofacies across the dune field, the lines showing the location of the logs in (c),GF: grainflow, Gfl: grainfall, WR: wind ripple. c) Preserved section illustrating the changes in lithofacies and reservoir potential from the central area where aeolian dunes dominate to the marginal area where interdune flats dominate.

associated fluvial landforms in a range of modern dryland desert systems. This is accomplished through analysis of satellite and aerial photograph imagery. Such data have enabled study of the relationships between the different types of dunes and fluvial landforms. A second part of this research study has focused on the analysis of data from ancient preserved aeolian and fluvial successions that have been studied at outcrop in order to build a series of facies models that can be used to improve understanding of controls on stratigraphic heterogeneity in aeolian and mixed aeolian-fluvial reservoirs.

1.3 Aims and objectives

The aim of this study is to gain an improved understanding of controls governing the detailed sedimentary architecture of preserved aeolian and fluvial successions in dryland settings. This is accomplished through the integration of a novel set of case-studies of modem and ancient aeolian and fluvial systems in order to propose a set of general models. This overarching aim is fulfilled through the implementation of four closely related yet independent studies, each of which has its own set of research objectives.

1.3.1 Remote sensing of spatial variability in aeolian dune and interdune morphology in the Rub' Al-Khali, Saudi Arabia

Work-block 1 seeks to quantify the form of geomorphic relationships between aeolian dune and interdune sub-environments in the central and marginal parts of four modern dune fields of the Rub' Al-Khali desert, Saudi Arabia. The aim is to document how and explain why dune- and draa-scale aeolian bedforms and their adjoining interdunes systematically change form from central to marginal dune-field areas in terms of their morphology, geometry (scale), orientation and style of bedform linkage (i.e. the extent to which interconnected and amalgamated aeolian bedform complexes are developed).

Specific objectives of this research are as follows: (i) to assess the geomorphic complexity and variety of dune types present in the Rub' Al-

Khali desert; (ii) to demonstrate and quantify types of spatial variation in dune and interdune type and geometry for a series of major dune fields; (iii) to consider how a series of external factors that collectively define the sediment state of the system act to dictate spatial changes in dune and interdune morphology and geometry.

1.3.2 A classification scheme for fluvial-aeolian system interaction in desert-margin settings

Work-block 2 seeks to propose a generalised framework with which to account for the diverse styles of interaction known to exist between coevally active aeolian and fluvial depositional systems, and to discuss the significance of these interactions for the geomorphological and sedimentological evolution of mixed aeolian-fluvial systems.

Specific objectives of this research are as follows: (i) to illustrate the principal types of aeolian-fluvial interaction documented from the world's major dryland systems and to propose a framework for their classification; (ii) to demonstrate how the orientation of fluvial systems relative to the trend of aeolian bedforms present at the leading edge of dune fields controls the nature of aeolian-fluvial system interaction; (iii) to consider the role of open versus closed interdune corridors in controlling the style and distance of incursion of fluvial systems into aeolian dune fields; (iv) to consider how different types of aeolian-fluvial interaction give rise to complex geomorphic arrangements of landforms and to consider the implications of such arrangements for the palaeoenvironmental reconstruction of ancient preserved counterparts.

1.3.3 Outcrop architecture of ancient preserved aeolian and fluvial successions: Triassic Wilmslow Sandstone and Helsby Sandstone formations, Sherwood Sandstone Group, Cheshire Basin, UK

Work-block 3 seeks to demonstrate the preserved expression of aeolian-fluvial system interactions via documentation of a case-study example from an ancient outcropping succession.

The research aim is to document the preserved record of aeolian and fluvial successions, to further develop our understanding of processes that operate in aeolian and fluvial systems, and to propose a novel facies model for the mechanism of preservation of aeolian and fluvial deposits that accumulated in arid and semi-arid depositional settings.

Specific objectives of this research are as follows: (i) to describe and interpret the sedimentary facies of both ancient dryland fluvial deposits and wet aeolian deposits that are present in an outcropping ancient succession; (ii) to develop high-resolution, three-dimensional facies models for the studied successions; (iii) to develop a discussion that investigates the aeolian system construction, accumulation and preservation; (iv) to investigate the relationship between aeolian dune and interdune morphology; (v) to investigate the relationship between preserved aeolian set thicknesses, grainflow thicknesses and original aeolian dune bedform size.

This research is significant because the temporal and spatial scales over which processes related to aeolian-fluvial interactions occur are highly varied and complex. Understanding the different interaction styles between the two systems is important in the development of generic facies models to explain the rock record in terms of geomorphic landscape evolution and formative palaeoenvironment, and for predicting heterogeneity in preserved subsurface reservoir successions.

1.3.4 Modelling system interactions in aeolian dune-field margin successions

Work-block 4 seeks to summarise the spatial and temporal and spatial scales over which aeolian and fluvial sediment accumulation events operate and interact through the development of a series of generalised facies models that account for the range of stratigraphic architectures documented from a range of modern desert-margin settings and equivalent ancient preserved successions.

Specific objectives of this research are as follows: (i) to collate the data acquired from modern systems using high-resolution satellite imagery as part of work-blocks 1 and 2 into a database that can be used to constrain graphical facies models; (ii) to employ the data acquired from ancient aeolian and fluvial systems present in an outcropping succession, along with data collated from wider literature case studies as examples of aeolian and fluvial system interaction, to better understand the variable distribution of facies generated by aeolian and fluvial systems in aeolian dune-field margin settings; (iii) to show how the database of case-study examples can be employed to develop a series of quantitative facies models with which to account for dynamic spatial and temporal aspects of aeolian-fluvial system behaviour; (iv) to show how a database approach can serve as a tool for system classification and quantification.

From an applied standpoint, quantitative depositional models arising from this database-driven approach serve to minimise uncertainties relating to stratigraphic heterogeneity in subsurface reservoir settings and aid inter-well correlation and prediction.

1.4 Data and methods

The four principal work blocks that comprise this research each utilise a substantial set of data.

1.4.1 The study of the Rub' Al-Khali desert system

The Rub' Al-Khali of south-eastern Saudi Arabia is covered by the latest generation of public-release satellite imagery, which reveals a varied range of dune types, the morphology of which changes systematically from central dune-field areas to marginal areas where aeolian interdunes, sand sheets, and ephemeral fluvial systems dominate. This study has entailed work in four distinct geographic areas of the Al Rub' Al-Khali, which collectively cover an area of 73,200 km^2. These areas were selected for study according to the following specific criteria: (i) chosen locations document spatial changes in the morphology of dunes and interdunes from the central part of a dune field to its outer margin; (ii) public-release satellite imagery used for examination of the dune forms is available for these areas at a resolution that is sufficiently high to enable detailed quantitative measurements to be made regarding various morphological attributes of dunes and interdunes.

Morphological and geometrical attributes relating to 555 dunes and 1415 interdunes from the 4 selected study areas have been characterised through the collection of 10,100 measurements made through the examination of satellite imagery provided by Google Earth Pro software and datasets. A series of quantitative approaches have been employed to characterise the complexity present in a range of dune-field settings where large, morphologically complex and compound bedforms gradually give way to smaller and simpler bedform types at dune-field margins. Parameters collected describe the following: dune bedform height and elevation of interdune flats; the along-crest length of a dune segment; bedform spacing; dune wavelength; maximum and minimum dune wavelength; amplitude of along-crest sinuosity; bedform long-axis orientation; the distance from the dune-field centre of the studied dune fields to their outer margins. Additionally, attributes recorded for interdunes are as follows: interdune length; interdune width; interdune long-axis orientation; distance from the centre of the studied dune fields to their outer margins; style of connectivity to neighbouring interdunes.

1.4.2 The study of the Interaction between aeolian and fluvial systems

The morphological expression and areal distribution of flood deposits present between aeolian dunes located in the outer margins of a series of desert dune fields from around the world have been mapped using high-resolution satellite imagery. This study has analysed the morphological expression and areal distribution of 130 examples of fluvial-aeolian interaction that have been mapped using high-resolution satellite imagery from 60 desert dune fields around the world. Case study examples have been classified to propose a framework of ten distinct types of system interaction. Studied desert systems include the Namib Desert and Skeleton Coast (Namibia), Taklamakan Desert (northwest China), Rigestan Desert (southwestern Afghanistan), Sahara Desert (North Africa), Algodones (southeastern California), White Sands (New Mexico), Rub' Al-Khali and An Nafud sand seas (Saudi Arabia), and Wahiba Sands (Oman), Great Sandy, Great Victoria, and Simpson Deserts (Australia). Google Earth Pro software provides public-release imagery that covers remote desert regions and provides a high-resolution images that can be exported as seamless tiles that are each up to 4,800 pixels wide. Such images provide an opportunity to analyse styles of aeolian-fluvial interaction that operate at a range of spatial scales.

1.4.3 Study of the ancient aeolian and fluvial systems

This study examines an outcropping mixed aeolian and fluvial succession from the Triassic Sherwood Sandstone Group of the Cheshire Basin, northwest England. The studied outcrop section extends laterally for 230m and is up to 13 m high. The outcrop provides an extensive section that exposes strata of both aeolian and fluvial origin, representing the upper part of the Wilmslow Sandstone Formation and the lower part of the overlying Helsby Sandstone Formation, respectively. Detailed analyses of lithofacies of aeolian and fluvial origin, facies associations and architectural elements within the studied outcrop succession have been undertaken. Fourteen lithofacies are defined based on their lithology, sedimentary texture and the range and type of sedimentary structures present within the studied section. Field sketches and photomosaics have been used to generate composite architectural panels from which the distribution and relationship of various aeolian and fluvial architectural elements present in the outcrop have been determined. Two-hundred palaeocurrent readings comprising cross-bedding foreset dip magnitude and azimuth data were collected from both aeolian and fluvial deposits. Architectural elements of aeolian and fluvial origin have been defined and utilised to generate a series of palaeo-depositional system and palaeoenvironment models. Further, a suite of data have been collected from preserved aeolian dune bedsets, including measurements of duneset thickness and thickness of individual grainflow deposits.

1.4.4 Modelling the aeolian and fluvial system interactions

Data used to construct the series of predictive, semi-quantitative facies models with which to account for the geological complexity and origin of mixed aeolian-fluvial successions form a database populated using the results from work blocks 1, 2 and 3, and supplemented through literature-derived case studies (54 ancient case studies). These data are used to assign ranges of values to describe the expected distribution of facies generated by different types of aeolian and fluvial system interaction at aeolian dune-field margins and the geometry and facies composition of preserved elements in ancient successions. Ten distinctive semi-quantitative geological models that describe types of aeolian-fluvial system interaction are presented.

1.5 Thesis structure

This thesis represents a discussion based around the following: (i) two papers that investigate modern aeolian and fluvial system interactions, both published in *Aeolian Research*, an internationally recognised academic journal, (ii) a chapter that investigates ancient aeolian and fluvial system interactions; (iii) a chapter that introduces an approach to modelling aeolian and fluvial system interactions through the development of a series of three-dimensional graphical, semi-quantitative facies models; (iv) a concluding chapter that seeks to summarise the generic findings that have arisen from the research. By virtue of the way that the research has been undertaken, parts of the various chapters contain modest overlap where key background information is reiterated.

1.5.1 Chapter One: Introduction

This chapter sets out the aims and objectives of the thesis and describes its structure. The key research questions that are used to meet the specific work objectives are introduced, and the data and methods used to address these research questions are explained; each chapter is outlined in turn in order to summarise the thesis content.

1.5.2 Chapter Two: Remote sensing of spatial variability in aeolian dune and interdune morphology in the Rub' Al-Khali, Saudi Arabia

This chapter investigates aspects of spatial variability of dune and interdune morphology in desert dune fields. Analysis of geomorphic relationships between dune and interdune sub-environments within a series of modern dunes fields of the Rub' Al-Khali has been undertaken to document how the morphology, geometry, internal facies arrangement and relationship of the various depositional architectural elements produced by these geomorphic features vary over space from dune-field-centre to dune-field-margin settings. Analysis of this active modern dune-field system shows a characteristic reduction in aeolian dune size and degree of connectivity and a corresponding increase in interdune size and degree of connectivity towards outer dune-field margins. This study has entailed work in four distinct geographic areas of the Rub' Al-Khali Sand Sea, Saudi Arabia and surrounding area, which collectively cover an area of 73,200 km^2. Morphological and geometrical attributes relating to 555 dunes and 1415 interdunes from the 4 selected study areas. The collection of data relating to primary landform morphology has enabled an improved understanding of the sediment system state of the modern Rub' Al-Khali desert sedimentary system to be gained.

1.5.3 Chapter Three: A classification scheme for fluvial-aeolian system interaction in desert-margin settings

This chapter provides a classification scheme for fluvial-aeolian system interactions in desert-margin settings. This is achieved through the examination of 60 desert regions around the world. This study has analysed the morphological expression and areal distribution of 130 examples of fluvial-aeolian interaction that have been mapped using high-resolution satellite imagery. Case-study examples have been classified to propose a framework of ten distinct types of system interaction. The proposed generalised framework is used to account for the diverse types of interaction known to exist between coevally active aeolian-fluvial depositional systems. The developed framework serves as a tool with which to discuss the significance of system interactions within the context of the geomorphological and sedimentological evolution of mixed aeolian and fluvial systems. This study presents a novel classification scheme for the description of types of interaction between fluvial systems and adjoining aeolian dune systems and their marginal areas into which flood events episodically extend. Ten distinct styles of interaction are recorded and illustrated by a set of case-study examples from around the world.

1.5.4 Chapter Four: Outcrop architecture of ancient preserved aeolian and fluvial successions: Triassic Wilmslow Sandstone and Helsby Sandstone formations, Sherwood Sandstone Group, Cheshire Basin, UK

This chapter describes and interprets the various lithofacies present in sections of the upper part of the Wilmslow Sandstone Formation and the lower part of the overlying Helsby Sandstone Formation of the Triassic Sherwood Sandstone Group, Cheshire Basin, UK. Specifically, this study examines outcrops of the Runcorn Expressway road-cut, a laterally continuous outcrop that exposes the boundary between these two formations. The study reveals that accumulation of the aeolian system represented by the Triassic Wilmslow Sandstone Formation was controlled by water table, and accumulation via both climbing and non-climbing mechanisms took place. The fluvial succession in the overlaying Helsby Sandstone Formation represents a preserved dryland ephemeral braided fluvial succession.

1.5.5 Chapter Five: Modelling system interactions in aeolian dune-field margin successions

This chapter employs the ten types of aeolian-fluvial system interactions described from modern desert systems (chapters Two and Three), together with the ancient outcrop case-study example (Chapter Four) and a suite of literature-derived data to generate a series of ten semi-quantitative geological facies models with which to account for the nature and origin of stratigraphic complexity present in aeolian dune-field margin successions that arise from the interplay of both autogenic and allogenic controls.

The geological models developed as an outcome of this study can be used to predict the arrangement of architectural elements that define gross-scale system architecture in a variety of mixed aeolian-fluvial system types. Results demonstrate the significance of aeolian dune type and orientation relative to fluvial-system type and orientation in determining the style of fluvial incursion into dune fields.

1.5.6 Chapter Six: Conclusions and future work

This final chapter summarises the general findings of the research study and postulates possible future related research that could be undertaken to further advance current understanding of the processes that control types of interaction between aeolian and fluvial depositional systems and their resultant accumulated sedimentary products.

Chapter Two
Remote sensing of spatial variability in aeolian dune and interdune morphology in the Rub' Al-Khali, Saudi Arabia

---

This chapter investigates aspects of spatial variability of dune and interdune morphology in desert dune fields. This study has entailed work in four distinct geographic areas of the Rub' Al-Khali Sand Sea, Saudi Arabia and surrounding area, which collectively cover an area of 73,200 km^2. Morphological and geometrical attributes relating to 555 dunes and 1415 interdunes from the 4 selected study areas have been characterised through collection of 10100 measurements made via the examination of satellite imagery provided by Google Earth Pro software and datasets. The collection of data relating to primary landform morphology has enabled an improved understanding of the sediment system state of the modern Rub' Al-Khali desert sedimentary system to be gained.

2.1 Abstract

The Rub' Al-Khali aeolian sand sea of south eastern Saudi Arabia – also known as the Empty Quarter – covers an area of 660,000 km² and is one of the largest sandy deserts in the world. The region is covered by the latest generation of public-release satellite imagery, which reveal spatially diverse dune patterns characterised by a varied range of dune types, the morphology, scale and orientation of which change systematically from central to marginal dune-field areas where non-aeolian sub-environments become dominant within the overall desert setting. Analysis of geomorphic relationships between dune and interdune sub-environments within 4 regions of the Rub' Al-Khali reveals predictable spatial changes in dune and interdune morphology, scale and orientation from the centre to the outer margins of dune fields. A quantitative approach is used to characterise the complexity present where large, morphologically complex and compound

bedforms gradually give way to smaller and simpler bedform types at dune-field margins. Parameters describing bedform height, spacing, parent morphological type, bedform orientation, lee-slope expression, and wavelength and amplitude of along-crest sinuosity are recorded in a relational database, along with parameters describing interdune size (long- and short-axis dimensions), orientation, and style of connectivity. The spatial rate of change of morphology of aeolian sub-environments is described through a series of empirical relationships. Spatial changes in dune and interdune morphology have enabled the development of a model with which to propose an improved understanding of the sediment system state of the modern Rub' Al-Khali desert sedimentary system, whereby the generation of an aeolian sediment supply, its availability for aeolian transport and the sand transporting capacity of the wind are each reduced in dune-field margin areas.

2.2 Introduction

Significant advances in our understanding of the spatial arrangement of aeolian dune patterns have been made possible through the increasing availability of high-resolution satellite imagery in recent years (e.g. Blumberg, 2006; Hugenholtz and Barchyn, 2010). Aeolian dune-field patterns are a product of self-organizing systems (Kocurek and Ewing, 2005; Wilkins and Ford, 2007; Ewing and Kocurek, 2010a) in which the development of simple or complex distributions of genetically related groups of aeolian bedforms and their adjoining interdunes is characterised by systematic and predictable changes in dune type, size, morphology, orientation and spacing from dune-field centre to dune-field margin settings (Werner and Kocurek, 1997, Kocurek and Ewing, 2005; Ewing et al. 2006; Bullard et al. 2011).

Several previous studies have documented spatial variation in bedform type and associated spatial changes in aeolian lithofacies distributions in desert dune fields (e.g. Breed and Grow, 1979; Sweet et al., 1988; Kocurek and Lancaster, 1999; Atallah and Saqqa, 2004; Baas, 2007; Bullard et al. 2011). However, relatively few studies have attempted to quantitatively document the form of spatial variability of dune and interdune morphology from the centres of aeolian dune-field systems to their margins (Kocurek and Ewing, 2005; Wilkins and Ford, 2007; Ewing and Kocurek, 2010a, b; Kocurek et al. 2010; Hugenholtz and Barchyn, 2010).

This study utilizes the latest generation of public-release satellite imagery to quantify the form of geomorphic relationships between dune and interdune sub-environments in both the central and marginal parts of four modern dunes fields of the Rub' Al-Khali (Empty Quarter) of Saudi Arabia. The overall aim of this work is to document how and explain why dune- and draa-scale aeolian bedforms and their adjoining interdunes systematically change form from central to marginal dune-field areas in terms of their morphology, geometry (scale), orientation and style of bedform linkage (i.e. the extent to which interconnected and amalgamated aeolian bedform complexes are developed). Specific objectives of this research are as follows: (i) to assess the geomorphic complexity and variety of dune types present in the Rub' Al-Khali desert; (ii) to demonstrate and quantify styles of spatial variation in dune and interdune type and geometry for a series of major dune fields; (iii) to consider how a series of external factors that collectively define the sediment state of the system act to dictate spatial changes in dune and interdune morphology and geometry.

This research is important because understanding the morphology and architectural distribution of the deposits of aeolian dune and interdune sub-environments serves to constrain the development of models with which to explain the principal controls on desert dune distributions. Further, the establishment of spatial trends in dune morphology and geometry aids the reconstruction of ancient aeolian palaeoenvironments and guides the prediction of sedimentary architecture in subsurface stratal successions. Understanding the morphological complexity present in a range of modern aeolian desert systems is a primary control on preserved stratigraphic complexity and is the first step in developing a series of generic models with which improve our understanding of the mechanisms by which complex sedimentary architectures arise in ancient preserved aeolian successions. Thus, there exists a need to document the morphology of modern desert systems to understand how spatial morphological changes in dune type and size might impact preserved stratigraphic architecture (Mountney, 2012).

2.3 Background

Modern aeolian dune-field systems are composed of complex arrangements of geomorphic elements, including dunes, draa and interdunes, which occur on a range of scales and are characterised by a variety of morphologies and geometries (Warren and Knott, 1983; Kocurek and Havholm, 1993; Lancaster, 1994; Rubin and Carter 2006; Ewing and Kocurek, 2010a). In many dune-field systems, the form of geomorphic elements and their relationship with adjacent elements varies systematically and predictably as a function of position within the overall aeolian system, especially in downwind directions and from the centre to outer margins of dune fields (Breed et al. 1979; Lancaster, 1983, 1994). Indeed, groups of genetically related aeolian dunes and intervening interdunes represent some best examples of patterned landscapes in nature (Kocurek and Ewing, 2005).

Few aeolian desert sand dunes exist in isolation. Most cluster, with many examples forming large dune fields in which systematic patterns of groups of genetically related dunes can be recognised, in some cases repeating with spatial regularity or with one or more defining attribute of the dune-form changing progressively in a given direction from, say, the centre of a dune-field to its margin. Groups of dunes collectively form larger geomorphic elements typically referred to as sand seas, dune fields (Livingstone and Warren, 1996) or ergs (Wilson, 1973). Although Cooke et al. (1993) define the lower size limit for a sand sea at 30 000 km^2, this being an inflexion point on the distribution curve of sand-sea size given by Wilson (1973), in modern usage (post-1995), no lower size limit is formally applied by way of definition.

Dune fields are not necessarily continuously covered with active aeolian sand dunes and most additionally include other morphological bodies of aeolian-derived or aeolian-related sediment deposits, including interdunes, sand sheets (which lack distinctly recognisable larger bedforms), areas of soil cover, lacustrine systems (e.g. playa lakes), and fluvial systems (typically ephemeral), some developed between active aeolian dunes (Lancaster, 1989). Thus, dunes in sand seas, including those in the Rub' Al-Khali, are commonly separated from each other by geomorphic elements whose well-defined shapes are, in part, dictated by the shapes of adjoining dune bedforms of different types (e.g. McKee and Bigarella, 1979).

The construction of aeolian dune-field systems and the spatial variation in the form of their internal components (e.g. dunes and interdunes) from central to marginal areas is governed by numerous controlling parameters that dictate sediment state (Berg, 1986; Kocurek, 1998, 1999; Kocurek and Lancaster, 1999). At a regional scale, the sediment state of aeolian dune fields is defined by separate components of sediment supply, sediment availability and transport capacity of the wind (Kocurek and Lancaster, 1999), and together these factors govern where and when aeolian system construction via the growth of dunes occurs.

2.4 Study area

The Rub' Al-Khali of south-eastern Saudi Arabia – also known as the Empty Quarter – is one of the largest continuous sand deserts in the world and comprises a series of dune fields, some spatially discrete and some merging into neighbouring fields, within which self-organised patterns of aeolian bedforms and adjoining interdunes are developed (Bishop, 2010). The name for the Arabian desert – Rub' Al- Khali or the Empty Quarter – was introduced by the Swiss geographer Burckhardt (1829) in his book "Travels in Arabia" and used later by Doughty (1888). Early research by Thesiger (1949), Beydoun (1966), Holm (1960, 1968), Glennie (1970) and Breed et al. (1979) each documented the presence of different bedform types and noted general spatial variations in dune types between different parts of the overall desert system.

In total, the Rub' Al-Khali covers approximately 660,000 km^2, rising to 776,000 km^2 of continuous active sand cover if adjoining sand seas (e.g. Jafura, Dahna and Nefud in Saudi Arabia) are additionally included (Breed et al. 1979; Edgell, 1989, 2006). Indeed, the wider desert region, which additionally incorporates the Wahiba Sand Sea of the Sultanate of Oman (Laity, 2009), covers an area of 795,000 km^2. Within the main the Rub' Al-Khali, active aeolian dunes and interdunes cover an area of 522,340 km^2 (Edgell, 2006), extending from United Arab Emirates and Oman in the east, to south-western Saudi Arabia and northern Yemen (Figure 2.1; Wilson, 1973; Glennie, 2005; Edgell, 2006). The largely unconsolidated sand dune deposits of the Rub' Al-Khali are characterised by large bedforms (dunes and draa), individual examples of which range from 50 to 300 m in height (Brown et al. 1963; Abd El Rahman, 1986; Edgell, 2006), and the majority of which are each separated by broad interdune flats, some up to 5 km in width in dune-field margin settings (Figure 2.2). The majority of the Quaternary sediments of the Rub' Al- Khali are composed chiefly of aeolian-reworked Pliocene alluvial sediments (McClure, 1978), though a secondary sand component is likely to have been additionally sourced from local modern alluvial (wadi) sediments (Holm, 1960; Brown, 1960).

The Rub' Al-Khali basin is a combined physiographical and tectonic feature (Bagnold, 1951; Powers et al., 1966; McClure, 1976, 1978; Edgell, 1989, and Clark, 1989) that forms a structural depression characterised as an embayment with a structural axis trending from northeast-to-southwest, bordered to the northwest and west by the Arabian Shield, and to the south and southeast by the Hadramawt-Dhofar Arch or Plateau (Figure 2.1b). The northern end of Rub' Al-Khali basin opens into the Arabian Gulf through the United Arab Emirates (Edgell, 2006). The desert is additionally constrained by the arc of the Oman Mountains to the northeast and by the Qatar Arch to the northwest (Figure 2.1b). The area occupied by active sand seas extends from the United Arab Emirates and Oman in the east, to south-western Saudi Arabia and the area directly north of Yemen.

2.5 Quaternary evolution of the Rub' Al-Khali

The Rub' Al-Khali formed in response to cyclic episodes of aridity driven by climatic fluctuations throughout much of the Quaternary period (Edgell, 1989, 2006; Glennie, 1998; Goudie, et al. 2000). The application of optical dating methods to sand samples from both linear and crescentic dunes has

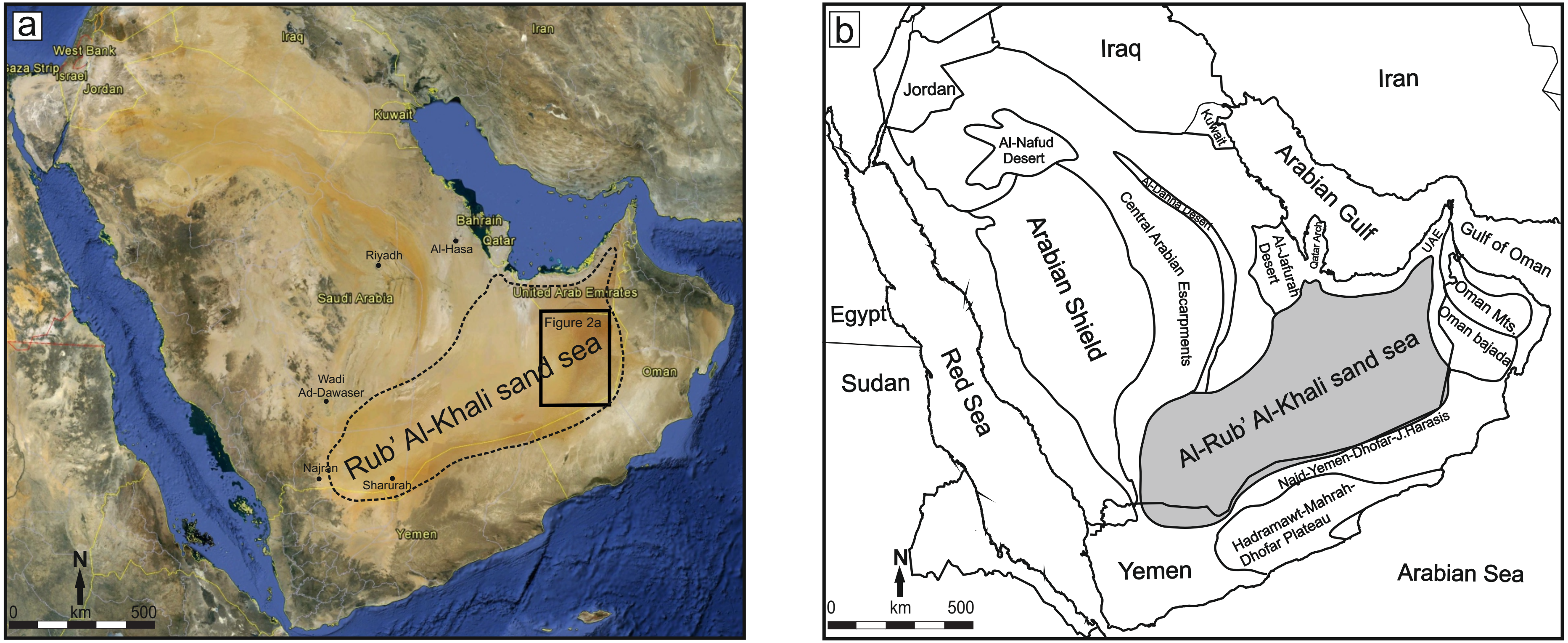


Figure 2.1: (a) Map of the Arabian Peninsula showing the location of the Rub' Al-Khali sand sea, dune-fields within which are the focus of this study.. Image from Google Earth Pro. (b) Map of Arabian Peninsula outlining geomorphological referred to in the text. Modified after Edgell (2006). Location of area of detailed study shown in figure 2.2a is indicated.

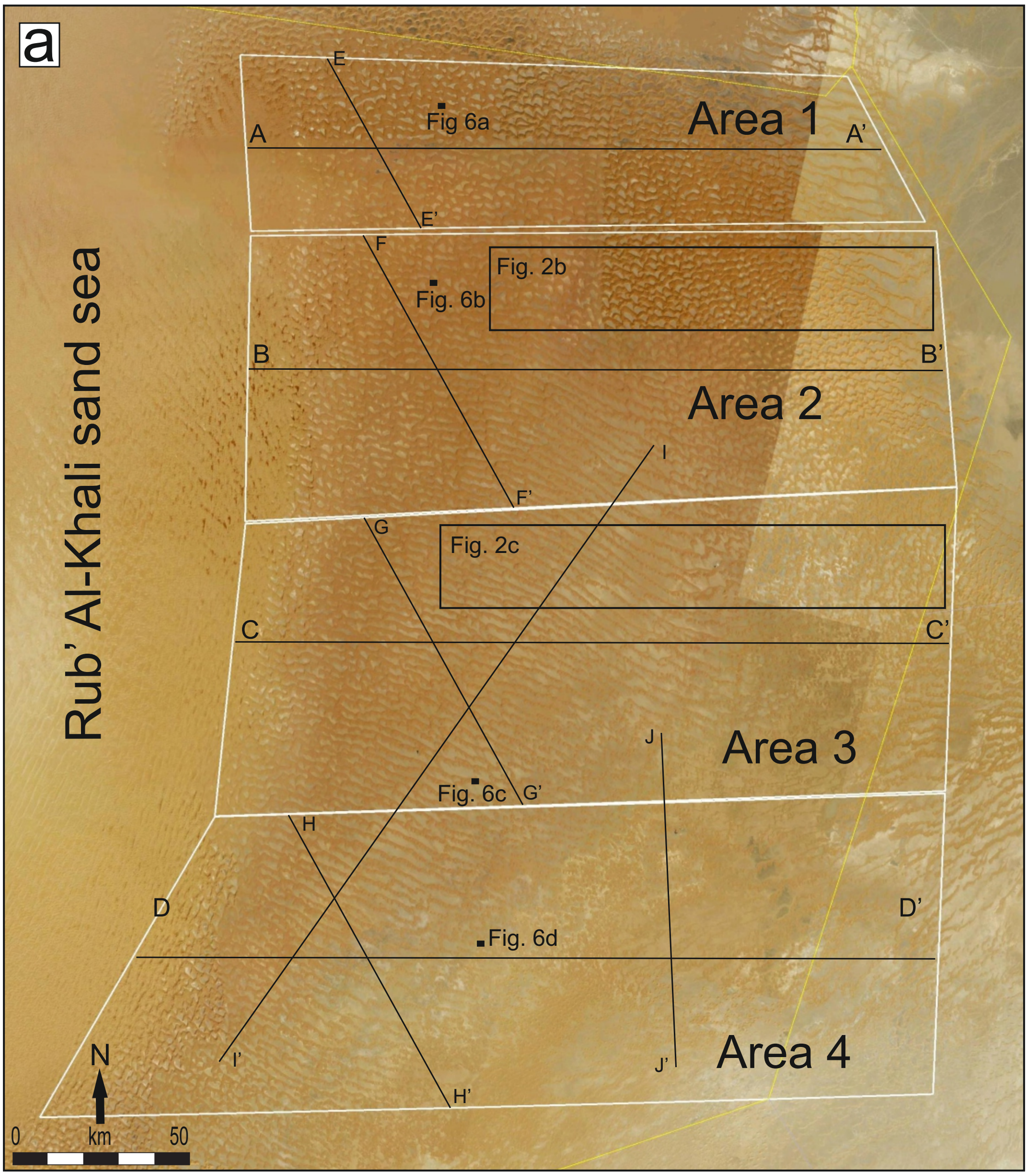


Figure 2.2: (a) Map of part of south-eastern Saudi Arabia showing the location of the areas named 1, 2, 3, and 4 for this study. Location of regions shown in figures 2.2b and 2.2c are indicated. Letters A-D refer to the location of the images shown in figure 5. Lines A-A' to J-J' are transcets across different parts of the study areas and are shown in figures 6 and 7.

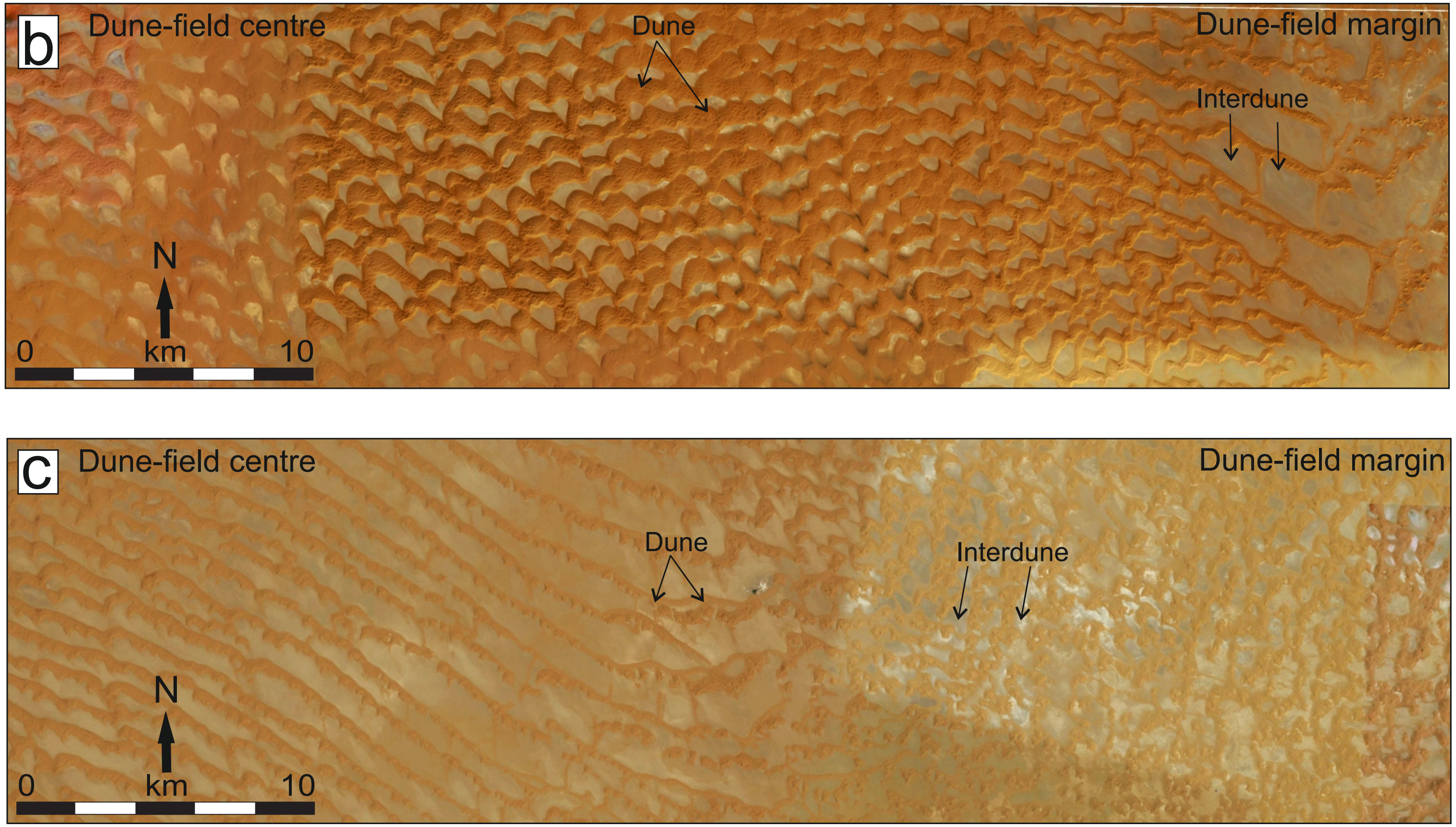


Figure 2.2 (Cont.): (b) Satellite images from the study area depicting the typical geomorphology of the dune fields, and the variation in dune morphology and distribution from the central part of the dune fields toward their margins. Note the reduction in dune size in a direction toward the dune-field margins, and the concomitant increase in the extent and connectivity of interdunes and playa areas. Images from Google Earth Pro.

demonstrated that the majority of the dunes of the Rub' Al-Khali formed during cold, arid intervals associated with high latitude Quaternary glacial cycles and concurrent sea-level lowering (Glennie, 1998; Glennie and Singhvi, 2002; Preusser et al., 2002; Lancaster and Tchakerian, 2003; Edgell, 2006; Bishop, 2010). Most dune fields of the Rub’ Al-Khali were more extensive than their present-day distribution during earlier parts of the Quaternary (Pye and Tsoar, 2009), though were less extensive than at present during the middle Holocene (Sarnthein, 1978). Indeed, the evolution of the Rub’ Al-Khali is known to have been influenced by climatic changes from arid, to semi-arid, to sub-humid throughout much of the Quaternary, with changes in mean annual precipitation and wind direction (Anton, 1984), and changes in sea level (Lancaster, 1998) each exerting a control on landscape evolution.

Dunes of the Rub’ Al-Khali (and indeed other deserts of the Arabian Peninsula) underwent several phases of spatial and temporal evolution in response to Quaternary climate and environmental change (Alsharhan et al., 1998, Anton, 1984, 1985; Hotzl et al., 1978; Barth, 2003). Firstly, an increase in aridity, especially during the late Pleistocene, encouraged the construction of dune fields whereby the availability of sediment for aeolian transport and regional variation in wind directions governed dune morphology such that complex dunes and draa developed. Secondly, increased climatic humidity during the early Holocene enabled lakes to develop and vegetation cover to partly stabilize the substrate, thereby restricting the availability of sediment for aeolian transport, limiting further dune construction, encouraging partial deflation in areas where the sand carrying capacity of the wind was undersaturated with respect to its potential, and encouraging the growth of non-aeolian regions. Some of the larger compound bedforms present in the Rub’ Al- Khali (e.g. those in the Liwa area of the United Arab Emirates) are known to have been constructed in response to the effects of repeated Quaternary climatic changes whereby multiple generations of superimposed dunes are delineated by calcrete horizons that record repeated episodes of bedform construction, partial deflation and stabilization indicative of alternations between relatively arid and relatively more humid climatic conditions (El-Sayed, 1999, 2000).

Throughout much of the Quaternary, spatial and temporal changes in dune morphology and geometry, and their spatial relationships are known to have been controlled significantly by changes in position, direction and intensity of Shamal and Indian Ocean Monsoon wind regimes (Preusser, 2009). The Shamal wind is a hot and dry wind with substantial sand-transporting capacity (Barth, 2001; Edgell, 2006) that forms as part of a large-scale flow of air toward a low-pressure centre that develops over Pakistan each year (Edgell, 2006). Although active in both summer and winter, it is most intense in June and July when it blows almost continuously at velocities that commonly reach and exceed 50 km h^{-1}. These winds are the principal agent for aeolian transport of sand and dune formation in much of the Arabian Peninsula (Membery, 1983).

2.6 Data and methods

This study has entailed work in four distinct geographic areas of the Al Rub' Al-Khali, herein called Areas 1, 2, 3 and 4 (Figure 2.2a), which collectively cover an area of 73,200 km². These areas were selected for study according to the following specific criteria: (i) chosen locations document spatial changes in the morphology of dunes and interdunes from the central part of a dune-field to its outer margin; (ii) public-release satellite imagery used for examination of the dune forms is available for these areas at a resolution that is sufficiently high to enable detailed quantitative measurements to be made regarding various morphological attributes of dunes and interdunes.

Morphological and geometrical attributes relating to 555 dunes and 1415 interdunes from the 4 selected study areas were collected through examination of satellite imagery provided by Google Earth Pro software and datasets, a business- and scientific-oriented mapping service (Figure 2.3). Satellite imagery from the studied areas has a spatial resolution of resolution 15 m per pixel, derived from 15 to 30 m-resolution multispectral Landsat data that have been pan-sharpened with panchromatic Landsat image processing software. Individual high-resolution images are each 4800 x 2442 pixels and images recording adjacent areas have been seamlessly tiled to render larger visualizations of each study area. Elevation data are derived

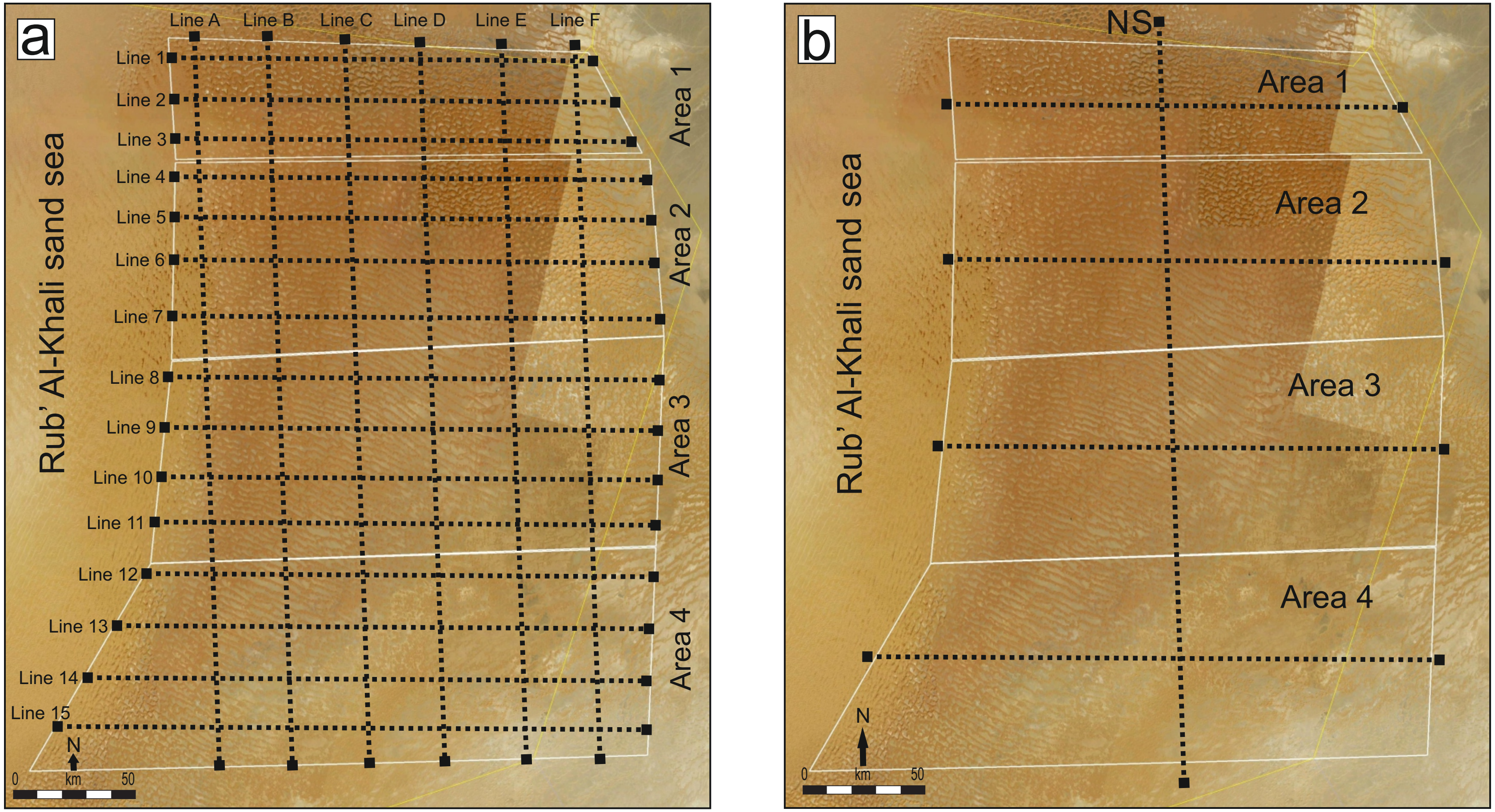


Figure 2.3: Maps of the study areas depicting the locations from which quantitative data regarding dune and interdune morphology and geometry were collected. (a) Location of transects from which data relating to interdune morphology were collected. (b) Location of transects from which data relating to dune morphology were collected; transects relate to areas 1-4 and also to a north-south line (NS). See supplementary data entries 1 and 2 for detailed information regarding sites of data collection.

from Shuttle Radar Topography Mission (SRTM) data, which has an absolute vertical accuracy of 16 m and a relative vertical accuracy of 10 m (Falorni et al., 2005).

Collected dune and interdune data have been recorded in a relational database and this has been used to discern trends between measured parameters. Spatial variation in both dune and interdune size and shape in directions both close to parallel and close to perpendicular to the overall direction of net sand transport has been recorded, with the resultant net direction of sediment transport having been identified from the analysis of dune bedform type and slipface orientation and through reference to Resultant Drift Direction calculations made by Fryberger and Dean (1979).

Attributes recorded for dunes are as follows and as depicted in Figure 2.4: bedform height is based on relief change from the regional level of the desert surface indicated by the elevation of interdune flats in the outer dune-field margin area to the crests of the bedforms; the along-crest length of a dune segment is a measure of bedform continuity whereby dune segments are terminated by major re-entrants or scours; bedform spacing is the crest-to-crest (or toe-to-toe) distance between adjacent bedforms in an orientation perpendicular to the trend of elongate bedform crestlines; dune wavelength records the extent of a bedform in an orientation perpendicular to the trend of the bedform crestline and this may vary from a maximum dune wavelength to a minimum dune wavelength within one dune segment as a function of bedform sinuosity; the wavelength and the amplitude of along-crest sinuosity observed in plan form together define crestline sinuosity (Rubin, 1987); bedform long-axis orientation describes the trend of dune crestlines; the distance from the dune-field centre is a relative measure of distance from the centre of the studied dune fields to their outer margins. Attributes recorded for interdunes are as follows and as depicted in Figure 2.4: interdune length is a measure of the distance that a single interdune corridor extends in an orientation parallel to the trend of the crestlines of the dunes that bound the interdune; interdune width is a measure of the width of an interdune flat developed between two dunes in an orientation perpendicular to interdune length; interdune long-axis orientation describes

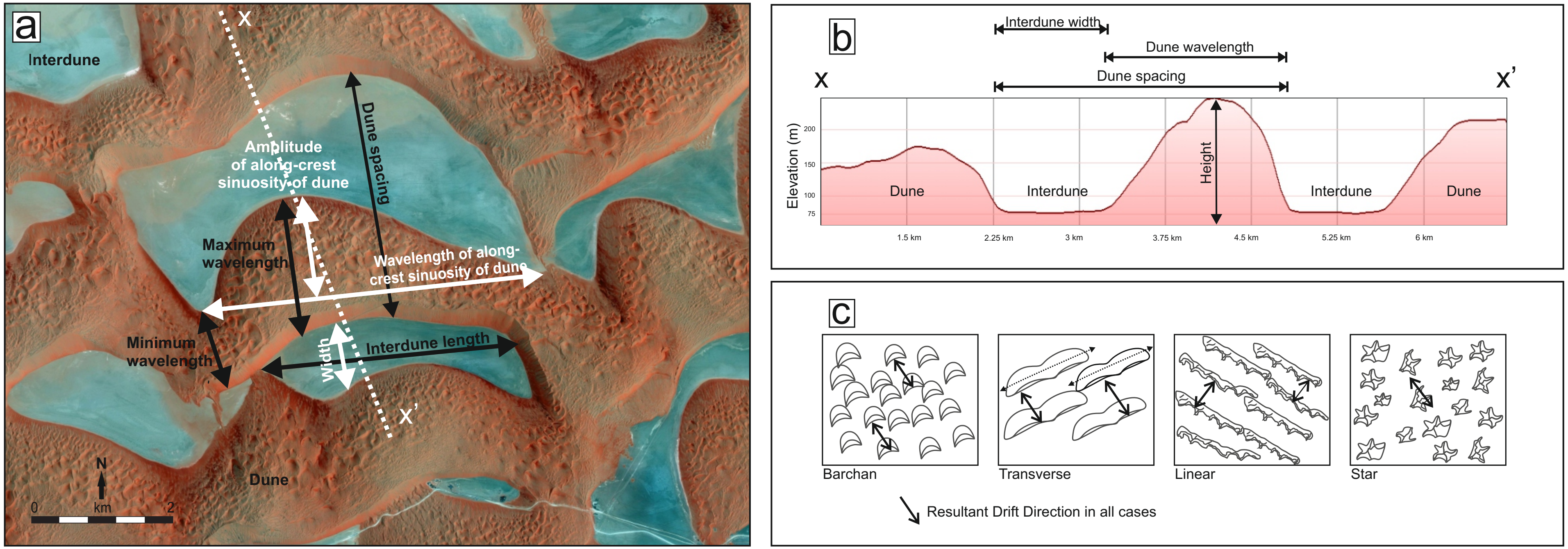


Figure 2.4: Example of dunes and interdunes of the Rub' Al-Khali, including definitions of the terminology used in this study to quantitatively describe their morphology and geometry. (a) Satellite imagery showing plan-form morphology. (b) Cross-sections derived from DEM data. (c) Schematic illustration of how bedform spacing has been determined for different morphological bedform types, dashed line on transverse dunes indicate the bedform long-axis orientation. Measurements made using Google Earth Pro software; data are recorded in a relational database, which has been queried to determine common trends and determine styles of spatial change in geometry and morphology across the study region.

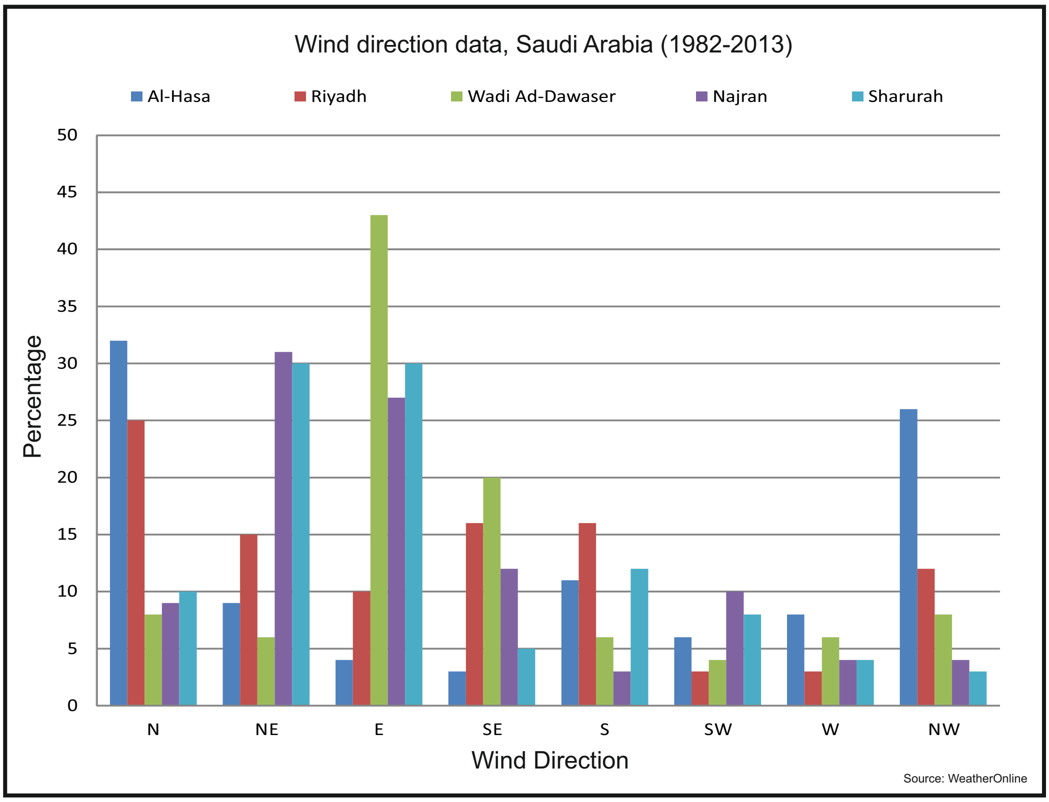
the trend of an interdune; the distance from the dune-field centre is a relative measure of distance from the centre of the studied dune fields to their outer margins.

The terminology applied in this study is derived from that introduced by Rubin (1987) and that used in a slightly modified form more latterly by a variety of authors including Mountney and Thompson (2002), Rubin and Carter (2006) and Mountney (2006a, 2012). It is important to note that some authors view the term “bedform wavelength” to be synonymous with “bedform spacing” (cf. Breed and Grow, 1979) but for the purposes of this study, and also as used by the aforementioned authors, the two terms differ in that bedform wavelength defines the distance from the leeward toe to the rearward (stoss) toe of the bedform in an orientation parallel to the direction of downwind migration. By contrast, bedform spacing defines the crest-to-crest (or toe-to-toe) distance from one bedform to an adjacent bedform, again in an orientation parallel to the direction of downwind migration. Thus, bedform spacing additionally incorporates the size of any adjoining interdune flat present between two bedforms (Figure 2.4).

Further to the wind directional data documented by Fryberger and Dean (1979), additional data relating to wind direction were obtained for the period 1982 to 2013 from 5 stations located in the central and eastern parts of the Rub’ Al-Khali and these indicate a prevailing Resultant Drift Direction to the southwest (Figure 2.5).

2.7 Dune morphology in the Rub’ Al-Khali

Satellite imagery for the Rub’ Al-Khali reveals a varied range of dune types, the morphology of which changes systematically from central dune-field areas to marginal areas where ephemeral fluvial systems become dominant (Figure 2.2b). The spatial variability of the dunes and interdunes present in this desert region is significant and was the principal reason for the choice of this region for this study. Key attributes describing the nature of the dunes and interdunes present in the studied areas are summarised in Tables 2.1 and 2.2.



	Al-Hasa	Riyadh	Wadi Aldawaser	Najran	Sharurah
N	32	25	8	9	10
NE	9	15	6	31	30
E	4	10	43	27	30
SE	3	16	20	12	5
S	11	16	6	3	12
SW	6	3	4	10	8
W	8	3	6	4	4
NW	26	12	8	4	3

Figure 2.5: A summary of dominant wind directions for the central and eastern part of the Rub' Al-Khali, based on data from 1982 to 2013 collected from 5 stations. Values in table are in percent. See Figure 1a for location of stations. Source: WeatherOnline.

Although aeolian dune forms in the Rub' Al-Khali are many and varied, three principal forms are defined at a fundamental level: transverse, linear and star forms, depending on whether the orientation of dune crest-lines is close to perpendicular or close to parallel to the dominant wind direction, or whether the forms exhibit a multi-facetted, pyramid-like morphology, respectively (Figure 2.6; Hunter et al., 1983; Kumar and Mahmoud, 2011). These fundamental bedform types are broadly distributed into three regions, as identified by Breed et al. (1979) through their analysis of Landsat imagery and Skylab photographs, and as summarized by Glennie (2005): crescentic transverse dunes dominate in the northeast part of the study region, linear dunes dominate throughout the western half, and star dunes dominate along the eastern and southern margins (Figure 2.2b; cf. Glennie, 1970). Additionally, zones occupied by non-classified complex dune forms and sand sheets and streaks are also recognised (Breed et al., 1979). Edgell (1989) used Landsat 7 near-infrared imagery, large-format camera images and field observations to divide the three primary bedform classes into 17 specific bedform types, though for this study the basic three-fold classification is used for the sake of simplicity.

Compound crescentic transverse dunes – some termed 'giant crescentic massifs' and 'compound crescentic dunes' (Breed and Grow, 1979; Breed et al., 1979) – in the northern and eastern Rub' Al-Khali, including in the Uruq al Mutaridah sub-basin (Figure 2.2b, c; Glennie, 2005), have a mean horn-to-horn width of 2.8 km, a mean length of 2.1 km (cf. Breed et al., 1979), a mean bedform long-axis orientation of 88.54 degrees, and exhibit a vertical relief of up to 160 m above the surrounding desert floor that is characterised by interdune flats and up to 230 m above sea-level datum (Figure 2.7). A 'hooked' variety of compound crescentic dune is described from the north-central and northeastern Rub' Al-Khali (Breed and Grow, 1979; Holm, 1960). Compound linear dunes in the western and south-western Rub' Al-Khali are aligned from northwest to southeast, parallel to Shamal winds (Glennie, 2005), and have different types of smaller dunes including star and barchan forms superimposed on them (Breed and Grow, 1979). These linear forms have a mean dune spacing of 3.9 km, are in places in excess of 130 m high above the surrounding desert floor that is characterised by interdune flats

Table 2.1: Summary of data relating to geometry of 555 dunes in the studied part of the Rub' Al-Khali sand sea.

Dune measured parameters	**Statistical analysis**				
	Minimum	Maximum	Mean	Standard Deviation	Coefficient of Variance
Amplitude of along-crest sinuosity (km)	0.03	2.39	0.69	0.52	75.23
Wavelength of along-crest sinuosity (km)	0.11	4.99	2.02	1.17	57.89
Minimum wavelength (km)	0.01	2.04	0.52	0.40	78.18
Maximum wavelength (km)	0.05	3.34	1.22	0.72	58.89
Spacing (km)	0.63	8.41	3.05	1.21	39.78
Height (m)	67	218	144.42	31.49	21.80
Orientation (degrees)	13	182	88.54	33.47	37.81

Table 2.2: Summary of data relating to geometry of 1415 interdunes in the studied part of the Rub' Al-Khali sand sea.

Interdune measured parameters	**Interdune Statistical analysis**				
	Minimum	Maximum	Mean (km)	Standard Deviation	Coefficient of Variance
Length (km)	0.20	53.15	6.97	8.19	117.54
Width (km)	0.06	6.22	1.49	0.88	58.95
Orientation (degrees)	5	221	106.39	35.63	33.49
Elevation (m)	60	107	76.17	9.96	13.08

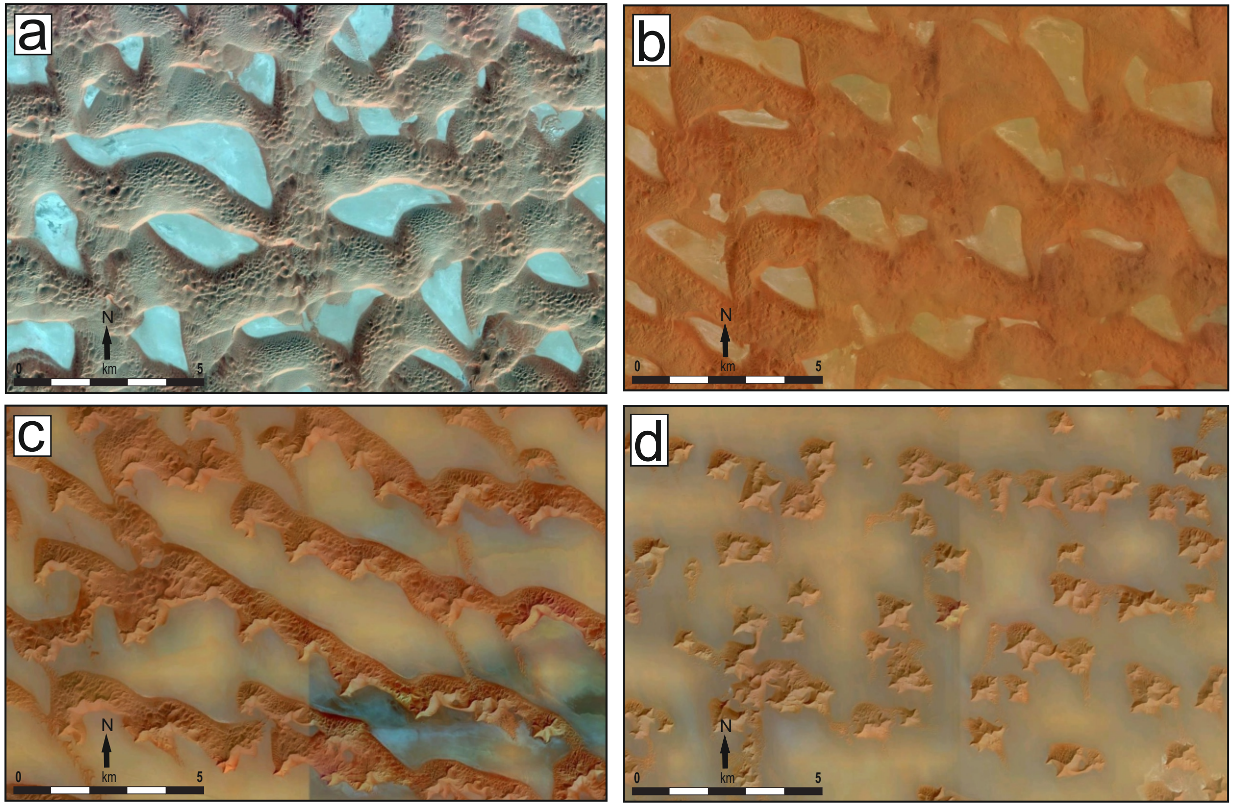


Figure 2.6: Satellite images from different locations across the Rub' Al-Khali desert depicting typical variations in dune and interdune morphology. Note the contrast in dune form and size between each image. All images depicted at the same scale. See figure 2a for locations. (a) Image from the northern part of Study Area 1, north-eastern Rub' Al-Khali, showing rows of laterally linked mega-barchan dunes with intervening interdunes (salt flats). (b) Image from the northern part of Study Area 2, north Rub' Al-Khali, showing a region dominated by complex giant barchan dunes with superimposed crescentic dune forms and parallel interdune corridors. (c) Image from the southern part of Study Area 3, southeast Rub' Al-Khali, showing an example of compound linear dune ridges, each separated by wide and parallel interdune corridors. (d) Image from the central part of Study Area 4, southeast Rub' Al-Khali, depicting an area characterised by pyramidal dunes (star bedforms) with distinctive star-like plan-form shapes and surrounded by extensive interdune areas.

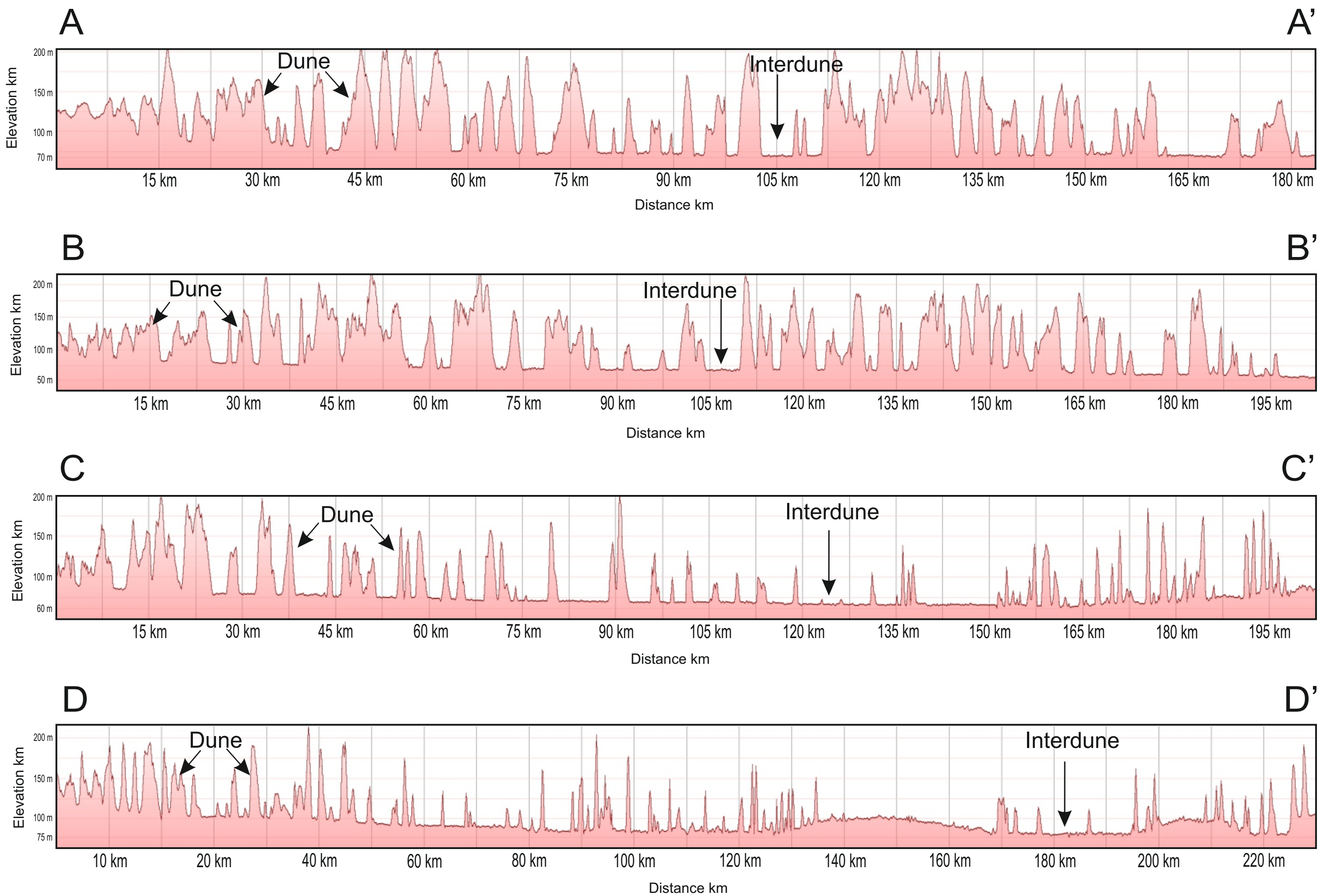


Figure 2.7: Cross sections from study areas 1-4 with elevation data reflecting variation in dune and interdune morphology and spacing from the centre of the dune field to its margin. The location of the cross sections is shown in Figure 2.2a. Digital Elevation data from Google Earth Pro are accurate to +/- 10 m.

and up to 200 m above sea-level datum; they form linked chains of bedforms, some of which extend uninterrupted for distances up to 50 km (Figure 2.6c, and Figure 2.8, transect I-I'; Bunker,1953; Holm, 1960; Breed et al., 1979). Neighbouring linear bedforms are separated by interdune flats (see below) that themselves each have a mean width of up to 1.4 km. Bedforms at the eastern and south-eastern margin of Rub' Al-Khali are mainly star dunes with mean widths of 2.1 km and heights up to 120 m above the surrounding desert floor that is characterised by interdune flats (Figure 2.6d and Figure 2.8, transect J-J'); these types merge with barachanoid dunes forming complex bedforms with slipfaces generally oriented toward the southeast (McKee and Breed, 1976).

The majority of dunes in the studied sand seas are oriented separated from each other by extensive interdune-flat areas whose shapes are at least partly dictated by the morphology of adjoining dunes of different types (cf. McKee, 1979). Interdunes in the Rub' Al-Khali vary in shape and in size (Figure 2.6), with the size and continuity typically increasing toward the margins of the dune fields (Figure 2.7), where the supply of sand and its availability for aeolian bedform construction is less, especially in areas where the water table lies close to the accumulation surface, such that the draw-up of moisture from the shallow subsurface via capillary action leaves the surface damp, thereby encouraging adhesion of sand (cf. Kocurek and Fielder, 1982; Olsen, et al. 1989). Open interdune corridors vary in length from 0.5 km in the central parts of dune fields to in excess of 50 km at dune-field margins; widths vary from 0.2 to 6 km.

2.8 Morphological and geometrical relationships between dunes

Dune bedform wavelength is a simple measure of bedform size herein defined as the extent of a dune bedform in an orientation perpendicular to its crestline. In this study both maximum and minimum dune wavelength are recorded for individual dune segments (Figure 2.4) as a measure of size. The difference between maximum and minimum wavelength for a single dune segment is also a measure of along-crest crest variability, with similar

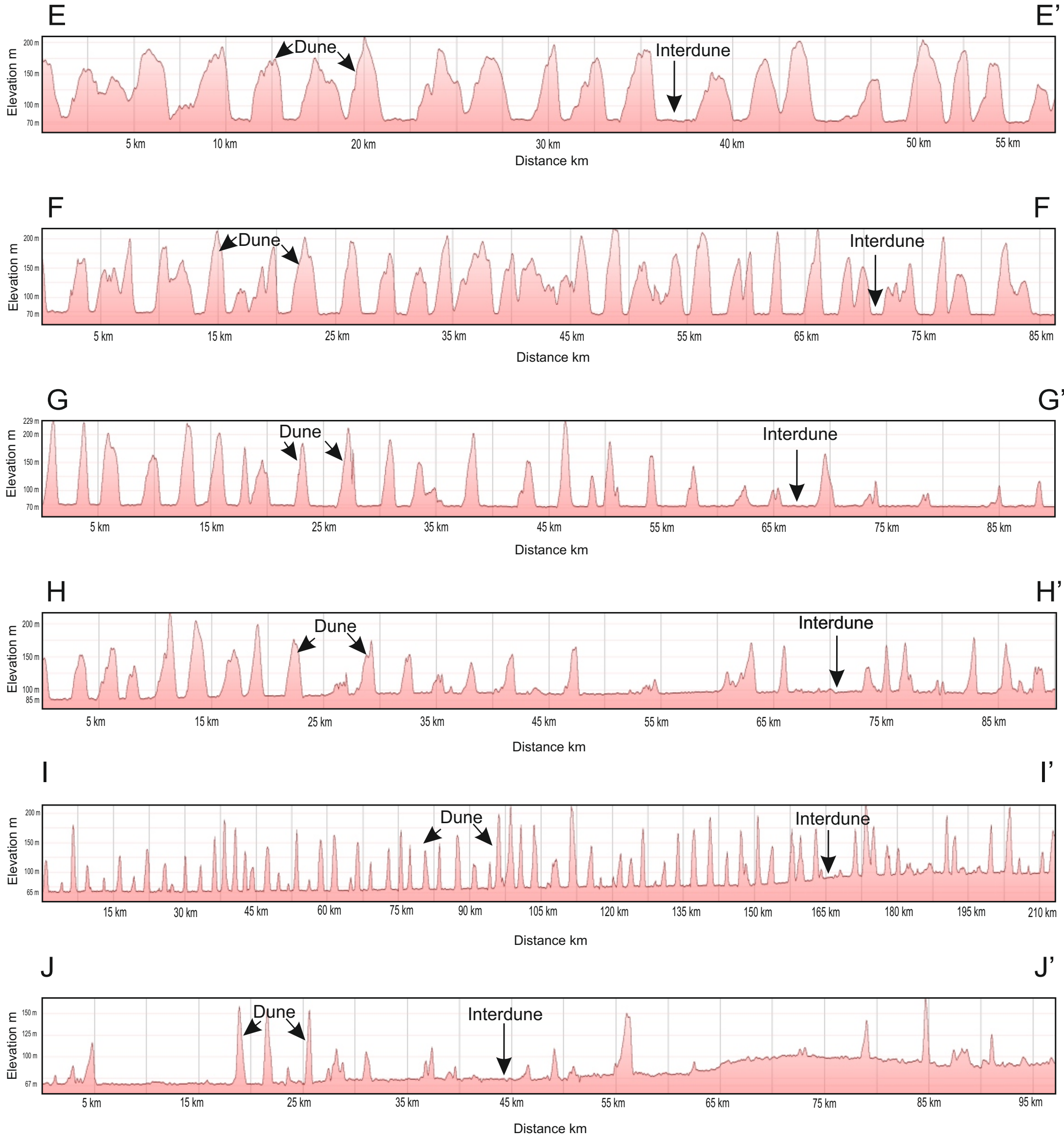


Figure 2.8: Cross sections from different positions in the study areas 1-4 with elevation data reflecting variation in dune and interdune morphology and spacing. The location of the cross sections is shown in Figure 2.2a. Sections E-E', F-F', G-G' and H-H' are aligned close to parallel to the regional resultant drift direction (toward the southeast); Section I-I' is aligned close to perpendicular to the regional resultant drift direction across a series of linear bedforms. Section J-J' reveals spatial changes in the morphology and geometry of star bedforms in the southern part of the study area. Digital Elevation data from Google Earth Pro are accurate to 10 m. Note that the dune wavelengths and interdune widths depicted in the cross sections are apparent since the orientation of the bedforms is oblique to the regional resultant drift direction in most cases.

values representing relatively straight-crested bedforms and values with greater differences reflecting increasing bedform crestline sinuosity (cf. Rubin, 1987; Rubin and Carter, 2006). A strong positive correlation exists between maximum and minimum dune wavelength (Figure 2.9a). Dunes from all areas exhibit a decrease in mean dune wavelength with increasing distance away from the dune-field centre and towards the outer margin (Figure 2.9b). Combined results from all 4 study areas demonstrate that over a distance of 300 km average dune wavelength decreases from 1.5 km at the dune-field centre to ~0.1 km at the dune-field margin.

Dune spacing is the distance between successive dunes in a train (measured, for example, between successive bedform crests), and includes both the wavelength of a dune bedform plus the width of the adjoining interdune (Figure 2.4). In contrast to dune wavelength, dune spacing exhibits little or no change with increasing distance from a dune-field centre toward its margin (Figure 2.7, and Figure 2.9c), though recorded values of dune spacing demonstrate considerable spread and vary mainly between 2 and 4 km. Given the lack of discernible change in mean dune spacing from the central parts of the dune fields to their margins, yet the systematic decrease in mean dune wavelength (i.e. bedform size), progressively smaller bedforms in dune-field margin settings are compensated for by progressively wider interdunes.

Wavelength and amplitude of along-crest plan-form sinuosities are together a simple measure of bedform crestline sinuosity (Rubin, 1987). For each of the 4 study areas, over a distance of 300 km, mean amplitude of along-crest sinuosity decreases from 2.5 km at the centre of dune-field to less than 0.5 km at the outer margin where the dunes become smaller (Figure 9d). Mean wavelength of along-crest sinuosity decreases from 5 km at the dune-field centre to ~0.2 km at the dune-field margin (Figure 2.9e). A positive relationship between dune along-crest amplitude and wavelength demonstrates little change of the form of along-crest sinuosity across the dune-field (Figure 2.9f).

Bedform height (defined as the difference in relief between the crest of a bedform and the general level of the desert surface where interdune flats are

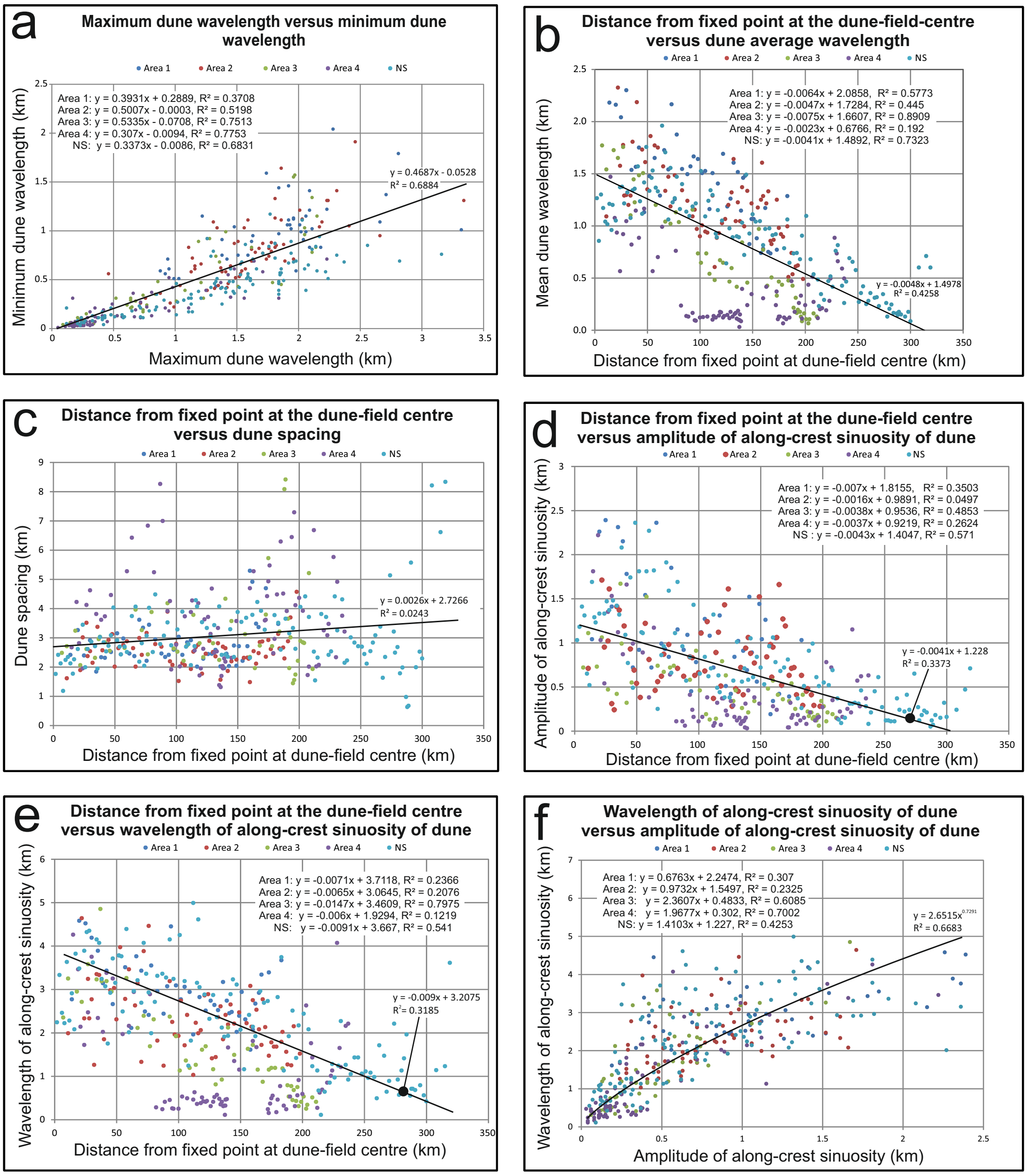


Figure 2.9: Examples of data demonstrating relationships present in aspects of dune bedform morphology in the Rub' Al-Khali dune field, showing the relationship between different parameters measured in the study area. Dune heights record the relief change between bedform crests and the generalised level of the desert floor as defined by the level of interdune flats in marginal areas of the dune fields. Best-fit lines shown on graph are for all data; separate best-fit equations are additionally shown for individual data sets from each study area. See text for further explanation.

present) varies as a function of both bedform type and location within the dune-field in terms of proximity to the outer dune-field margin (cf. Lancaster, 1988). Data collected from this study reveal bedforms with heights that range from <5 m at the dune-field outer margin to >155 m in the dune-field centre. The relationship between bedform height and spacing is complex (Figure 2.10a) and reveals no predictable trend because dune spacing is largely independent of bedform size. By contrast, a positive correlation exists between bedform height and wavelength (Figure 2.10b), though the spread in these data likely reflect contrasts between dunes of fundamentally different morphological types. Although bedform height generally decreases from the dune-field centre to its outer margin (Figure 2.10c), the trend is not straight-forward because bedforms tend to undergo changes in morphology along such transects.

2.9 Morphological and geometrical relationships between interdunes

The relationship between interdune long-axis length and width exhibits a positive correlation (Figure 11a), with the rate of increase of length relative to that of width being largely independent of the study area. Interdunes have a tendency to become longer and wider with increasing distance from a fixed point in the dune-field centre towards the eastern dune-field margin (Figure 2.11b, c), though considerable variability exists. For the data from Area 3, as interdune lengths become very large (> 40 km) their widths stabilize at 1.5 to 3.5 km in the dune-field margin areas (Figure 2.7, Section C-C'; Figure 2.11c). In these marginal areas, interdunes become the dominant landform type and they effectively partition the dune-field with the dune bedforms being subordinate and in some cases spatially isolated landforms. Considering the entire dataset, interdune long-axis orientation systematically varies (mimicking the trend of bedform crestlines); interdune long-axes rotate systematically counter-clockwise from ~130-310 degrees in the central dune-field areas to 070-250 degrees at the dune-field margin, 300 km away (Figure 2.11d). Two dominant orientation trends are evident: interdunes in Areas 1, 2 and 3 are mostly oriented between 0 and 200 degrees, with a

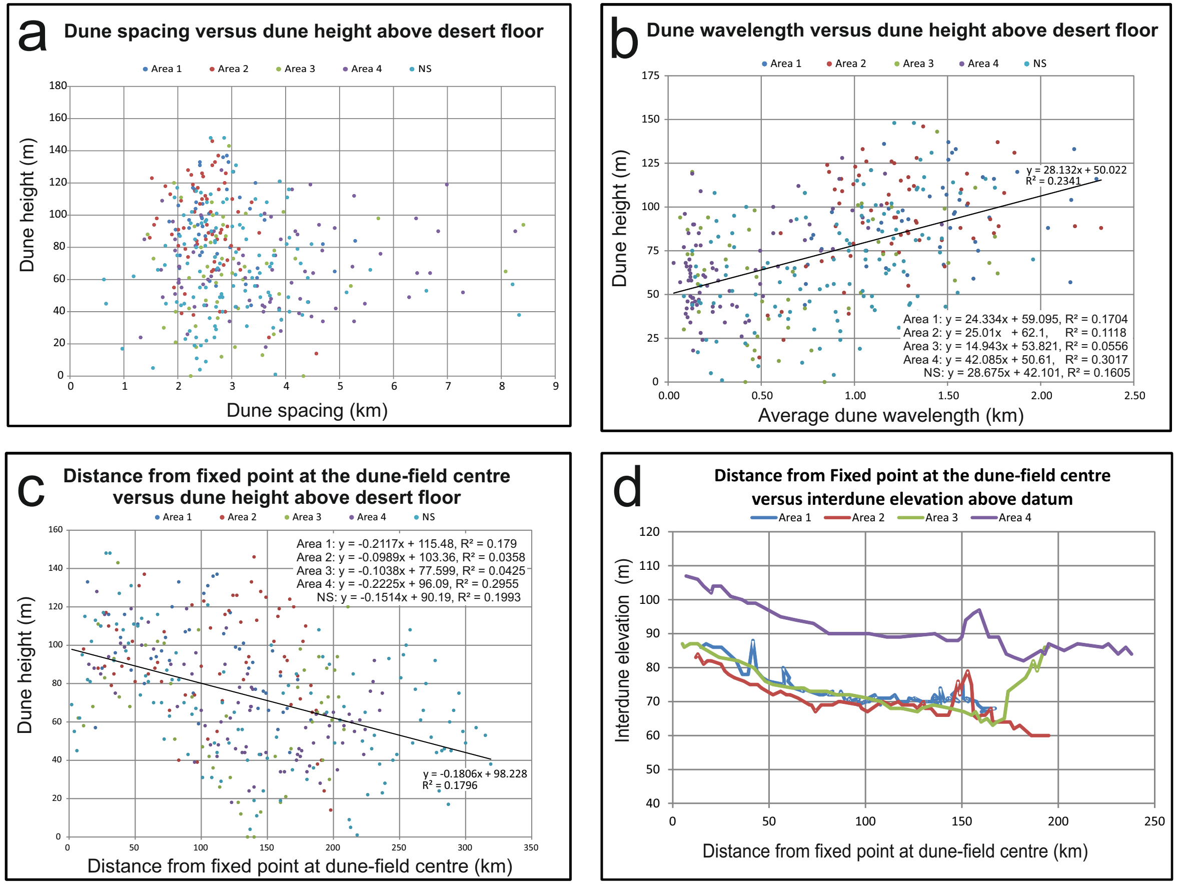


Figure 2.10: Examples of data demonstrating relationships present in aspects of dune bedform height and interdune elevation in the Rub' Al-Khali dune field, showing the relationship between different parameters measured in the study area. Best-fit lines shown on graph are for all data; separate best-fit equations are additionally shown for individual data sets from each study area. Dune heights have been calculated based on relief above the regional level of the desert floor in the dune-field margin areas; note that this level is higher in Area 4 where the desert system is constructed on a slightly elevated basement, as shown in graph d. See text for further explanation.

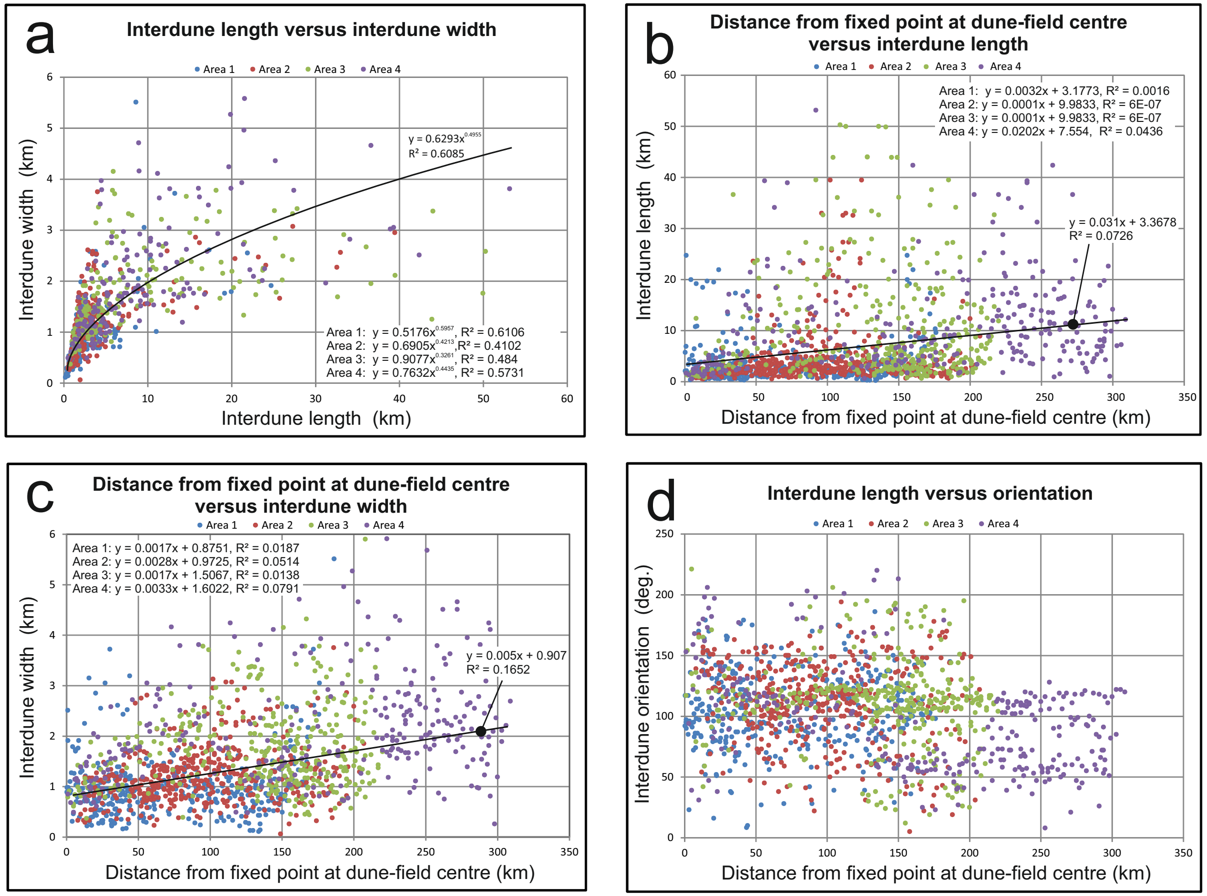


Figure 2.11: Examples of data demonstrating relationships present in aspects of interdunes of the Rub' Al-Khali dune field. The scatter plots demonstrate several relationships between measured interdune parameters. Best-fit lines shown on graph are for all data; separate best-fit equations are additionally shown for individual data sets from each study area. See text for further explanation.

cluster of interdunes trending between 050 and 130 degrees; in Area 4 interdunes have long-axis orientations preferentially trending between 200 and 300 degrees (Figure 2.11d). The spatially isolated interdune depressions present between bedforms in the central dune-field region are elevated up to 25 m above the regional level (Figure 2.7, and Figure 2.10d) and this demonstrates that bedforms in these central regions are climbing over one another to generate an accumulation. The elevation of interdunes in Area 4 is ~20 m higher than that in Areas 1-3 (Figure 2.7, and Figure 2.10d) because bedforms and interdunes in this most southerly study area are constructed on a topographically elevated basement.

2.10 Discussion

Use of satellite imagery for the analysis of changes in morphology, geometry, orientation and related attributes of dunes and interdunes present along a series of transects from central to marginal positions within 4 areas of the Rub’ Al-Khali represents a quantitative approach to the characterisation of the changing spatial distribution of aeolian geomorphic elements and sub-environments in desert settings. The observed spatial variation in patterns of dune bedform and interdune arrangement arise as a consequence of several controls that operate to determine aeolian sediment system state in the studied dune fields. Although results from the analysis of this dataset do not necessarily enable quantitative determination of the nature of these controls, they do allow for the statement of a series of generalised discussion points.

Relationships between dune wavelength, interdune length and width, and position in the dune-field highlight how interdune morphology interacts with the spatial distribution of dunes. The increase in interdune size and connectivity in dune-field margin areas and the corresponding decrease in dune size results from an overall reduction in either (i) the rate of generation of a sand supply for aeolian bedform construction in dune-field margin settings, (ii) the availability of that supply for aeolian bedform construction, or (iii) a downwind reduction in the sediment transport capacity of the wind (Kocurek and Lancaster, 1999). In dune-field margin settings, where areas of interdune flats are dominant, supply-limited and availability-limited aeolian systems are common and are controlled by factors such as the presence of a damp substrate due to an elevated, near-surface water table (e.g. Hotta et al., 1984) and/or the action of surface-stabilizing agents such as vegetation and precipitated crusts (e.g. Glennie, 2005; Edgell, 2006; Kumar and Mahmoud, 2011). Downstream reduction in the transport capacity of the wind at dune-field margin may also limit potential for dune construction (e.g. Ash and Wasson, 1983; Anderson and Haff, 1991). The presence of a near-surface water table in the outer margins of the studied dune fields such that its capillary fringe acts as a wicking effect to maintain a damp surface effectively renders these parts of the study area wet aeolian systems (*sensu* Kocurek and Havholm, 1993) such that, although a source of sand-grade sediment that is potentially suitable for aeolian transport and dune construction may be present within interdune flats, this supply is not available for transport. Thus, water-table level plays a fundamental role in limiting dune construction and protecting interdune flats from deflation. Indeed, a progressive rise in the relative level of the water table in the outer dune-field margin areas may potentially enable accumulation of packages of interdune strata between the accumulations of migrating dunes (cf. Mountney and Russell, 2006, 2009).

The positive correlation between interdune length and width, and distance from a fixed point in the dune-field centre, which summarizes the form of generalised changes in dune and interdune morphology across the dune field, arises as a result of a change in the sediment state components. The relations reveal significant geomorphological changes in the interdune areas and adjoining dunes across the dune field, and these represent a systematic spatial reduction in sediment supply and availability for aeolian transport, most likely because sediment is stored in upwind, central parts of the dune-field rendering the wind undersaturated with respect to its potential sand transporting capacity in downwind dune-field margin areas, and therefore potentially capable of sediment deflation.

Dunes in the Rub’ Al-Khali desert exist in great variety of morphologic types that change systematically across the dune field. Variation in dune form is the primary control on the morphology of adjacent interdunes, especially in dune-field centre regions where the shape and extent of each interdune form is governed and defined by the geometry and spacing of surrounding dune forms.

Spatial variation in the arrangement of dune patterns in the Rub' Al-Khali takes the form of gradational transitions from complex to simple bedform types, and a decrease in dune size and an associated increase in interdune size from the centre to the outer-margin areas of dune fields. Such changes reflect the interaction between sediment supply and transport capacity. The availability of sand for dune construction has long been recognised as a primary control on dune morphology (e.g. Wilson, 1972), whereby simple barchan dunes tend to evolve in systems where sand supply is limited and therefore are the main bedform type in marginal areas of many dune fields including the eastern part of the Rub' Al-Khali studied here. By contrast, the central and northern part of the studied dune-field is dominated by interlinked barchanoid dune types, whose presence records a greater sand supply.

Interdune orientation varies systematically as a function of geographic location, which in turn reflects the distribution and form of surrounding dunes, development of which is governed by prevailing wind type, directionality and intensity (Resultant Drift Potential and Resultant Drift Direction) and seasonality (Fryberger and Dean, 1979). Study Area 1, which is located in the northeast of the Rub' Al-Khali, is dominated by large linked barchanoid dune ridges; Study Area 4, which is located in the southeast of the Rub' Al-Khali, is characterised by a systematic change from connected to isolated star dunes and associated changes in interdune morphology whereby the observed dune-form variability arises partly because of low length-to-width ratios of the dune wavelength. The direction and the rate of aeolian sand transport are strongly governed by the wind regime, velocity and direction (cf. Pye and Tsoar, 2009). Data depicting dune cross-sectional morphology (Figure 2.7) demonstrate increasing variability in the range of bedform height and spacing in several marginal dune-field areas (e.g.

transects C-C’ and D-D’), which likely results from a sediment state that is not in equilibrium for these parts of the studied system.

The Rub’ Al-Khali region is influenced by winds with a high drift potential (the energy of surface winds in term of their capability to induce sand transport), chiefly because of the action of trade winds in mid-latitude depressions (Fryberger and Ahlbrandt, 1979). Directional variability of effective winds – south-southwest in winter and northwest in spring and summer (the so-called Shamal wind) – influences both the sand transporting potential of the wind (and therefore the bedform migration rate) and the Resultant Drift Direction (itself a control on dune migration direction) in the Rub’ Al-Khali. This explains the high system activity and the pronounced variety of bedform patterns, sizes and orientations. Unidirectional winds are responsible for the construction of large crescentic (barchan) dunes, which attain heights in excess 130 m in some areas in northern part of the dune field. Seasonally varying Shamal winds form linear dune ridges; more complex multi-directional winds form the star dune complexes that dominate in the southern part of the dune field.

The extracted elevation data describing surface topography, which were acquired as a series of transects recording changes in dune spacing, height and morphology from different locations and in different orientations across the study areas (Figure 2.7, and Figure 2.8), show clear examples of dune and interdune variability in the Rub’ Al-Khali sand sea. The central part of the dune-field contains the largest and most connected dune forms, many of which exceed 130 m in height (up to 160 m high), and this reflects bedform construction enabled by a large sand supply. Transects in Figure 2.7 each show spatially isolated interdune depressions within the central dune-field regions that are elevated up to 25 m above the regional level and this demonstrates that bedforms in these central regions are climbing over one another to generate an accumulation whereby the general interdune level is elevated above the regional level of the desert floor observed in more marginal dune-field areas (Figure 2.10d; *sensu* Kocurek, 1999).

In transects oriented in an upwind-to-downwind direction located in more central parts of the dune-field (e.g. transects E-E’ and F-F’, Figure 2.2a and

Figure 2.8) no discernible downwind change in mean bedform height, wavelength or spacing is evident. By contrast, in transects oriented in an upwind-to-downwind direction but located in the zone of transition between the central and marginal parts of the dune-field (e.g. transects G-G' and H-H', Figure 2.2a, and Figure 2.8), a general reduction in dune height and wavelength (Figure 2.9b and Figure 2.10c), and an associated increase in interdune width (Figure 2.11c) in a downwind direction are evident and such changes are indicative of a spatial reduction in the availability of sand for bedform construction in downwind dune-field margin regions.

2.11 Conclusions

The latest generation of high-resolution, public-release satellite imagery and SRTM digital elevation data has provided the basis for a quantitative analysis of patterns of arrangement of large-scale aeolian bedforms and adjoining interdunes in a series of large sand seas present as desert dune fields in the Rub' Al-Khali of south-eastern Saudi Arabia. Image analysis documents a varied range of dune types, the morphology of which changes systematically from central dune-field areas to marginal areas where aeolian interdunes, sand sheets, and ephemeral fluvial systems dominate. Analysis of geomorphic relationships between dune and interdune sub-environments within 4 modern dune fields documents how dune and interdune morphology, geometry and orientation varies over space from dune-field-centre to dune-field margin settings. Results demonstrate a characteristic reduction in aeolian dune size and degree of connectivity and a corresponding increase in interdune size and degree of connectivity towards outer dune-field margins. The collection of data relating to primary landform morphology has enabled an improved understanding of the sediment system state of the modern Rub' Al-Khali desert sedimentary system. Observed trends arise as a function of spatial changes in the sediment state of the system whereby sediment supply, the availability of that supply for transport and the sediment transporting capacity of the wind each combine to dictate the geomorphology of dune and interdune forms, which vary from thick accumulations of sands in the form of coalesced compound and complex barchanoid bedforms in dune-field centre settings, to spatially discrete star dunes and small, spatially isolated barchan dunes separated by extensive water-table-controlled interdune flats in dune-field margin settings. Observations from this modern dune-field system have enabled the spatial rate of change of morphology of aeolian sub-environments to be characterised and described through a series of empirical relationships.

Results of this study have implications for developing an improved understanding of the likely controls on the detailed sedimentary architecture of preserved aeolian successions by enabling the proposition and development of a range of dynamic facies models for aeolian systems. This has wider applied implications and significance: for example, the morphological changes in the distribution of aeolian bedforms and interdunes across dune-field systems provides important information with which to improve our understanding of the likely arrangement of architectural elements in ancient aeolian preserved successions, several of which form important reservoirs for hydrocarbons. This work is therefore an important step in the development of improved models for the characterisation of stratigraphic complexity and heterogeneity in aeolian reservoirs.

Chapter Three
A classification scheme for fluvial-aeolian system interaction in desert-margin settings

This chapter provides a classification scheme for fluvial-aeolian system interaction in desert-margin setting. This is achieved through the examination of 60 desert regions around the world. This study has analysed the morphological expression and areal distribution of 130 examples of fluvial-aeolian interaction that have been mapped using high-resolution satellite imagery. Case-study examples have been classified to propose a framework of ten distinct types of system interaction. The proposed generalised framework is used to account for the diverse types of interaction known to exist between coevally active aeolian-fluvial depositional systems. The developed framework serves as a tool with which to discuss the significance of system interactions within the context of the geomorphological and sedimentological evolution of mixed aeolian fluvial systems.

3.1 Abstract

This study examines 130 case examples from 60 desert regions to propose a generalised framework to account for the diverse types of interaction known to exist between active aeolian and fluvial depositional systems at modern dune-field margins. Results demonstrate the significance of aeolian and fluvial system type, orientation of aeolian versus fluvial landforms, distribution of open versus closed interdune corridors, and fluvial flow processes in controlling the distance and type of penetration of fluvial systems into aeolian dune-fields.

Ten distinct types of fluvial-aeolian interaction are recognised: fluvial incursions aligned parallel to trend of linear chains of aeolian dune forms; fluvial incursions oriented perpendicular trend of aeolian dunes; bifurcation of fluvial flow between isolated aeolian dune forms; through-going fluvial

channel networks that cross entire aeolian dune-fields; flooding of dune-fields due to regionally elevated water table levels associated with fluvial floods; fluvial incursions emanating from a single point source into dune-fields; incursions emanating from multiple sheet sources; cessation of the encroachment of entire aeolian dune-fields by fluvial systems; termination of fluvial channel networks in aeolian dune-fields; long-lived versus short-lived modes of fluvial incursion.

Quantitative relationships describing spatial rates of change of desert-margin landforms are presented. The physical boundaries between geomorphic systems are dynamic: assemblages of surface landforms may change gradationally or abruptly over short spatial and temporal scales. Generalised models for the classification of types of interaction have application to the interpretation of ancient preserved successions, especially those known only from the subsurface.

3.2 Introduction

Desert dune-fields are not necessarily covered with aeolian bedforms; most are also characterised by other morphological bodies of aeolian-derived or aeolian-related sediment deposits, including interdunes, sand sheets, soils, lacustrine systems, and perennial, intermittent or ephemeral fluvial systems. These geomorphic forms are commonly developed between active aeolian dunes, else they define the limits of dune-fields, with sharp or gradational boundaries. Figure 3.1 depicts common depositional processes that operate at dune-field margins, many of which control the mechanisms by which successions accumulate to form bodies of preserved strata. Significant diversity in the arrangement and type of interaction of competing depositional sedimentary systems is recognised in modern desert dune-fields and their marginal areas, and these give rise to complex yet predictable geomorphological patterns that commonly vary over space and time (e.g. Lancaster, 1989; Cooke et al., 1993; Bullard and Livingstone, 2002; Al-Masrahy and Mountney, 2013). The record of these interactions is also recognised in the ancient sedimentary record (e.g. Langford and Chan, 1989; Kocurek, 1991; Spalletti and Veiga, 2007), where spatial and temporal changes in the type of interaction between aeolian dune and associated

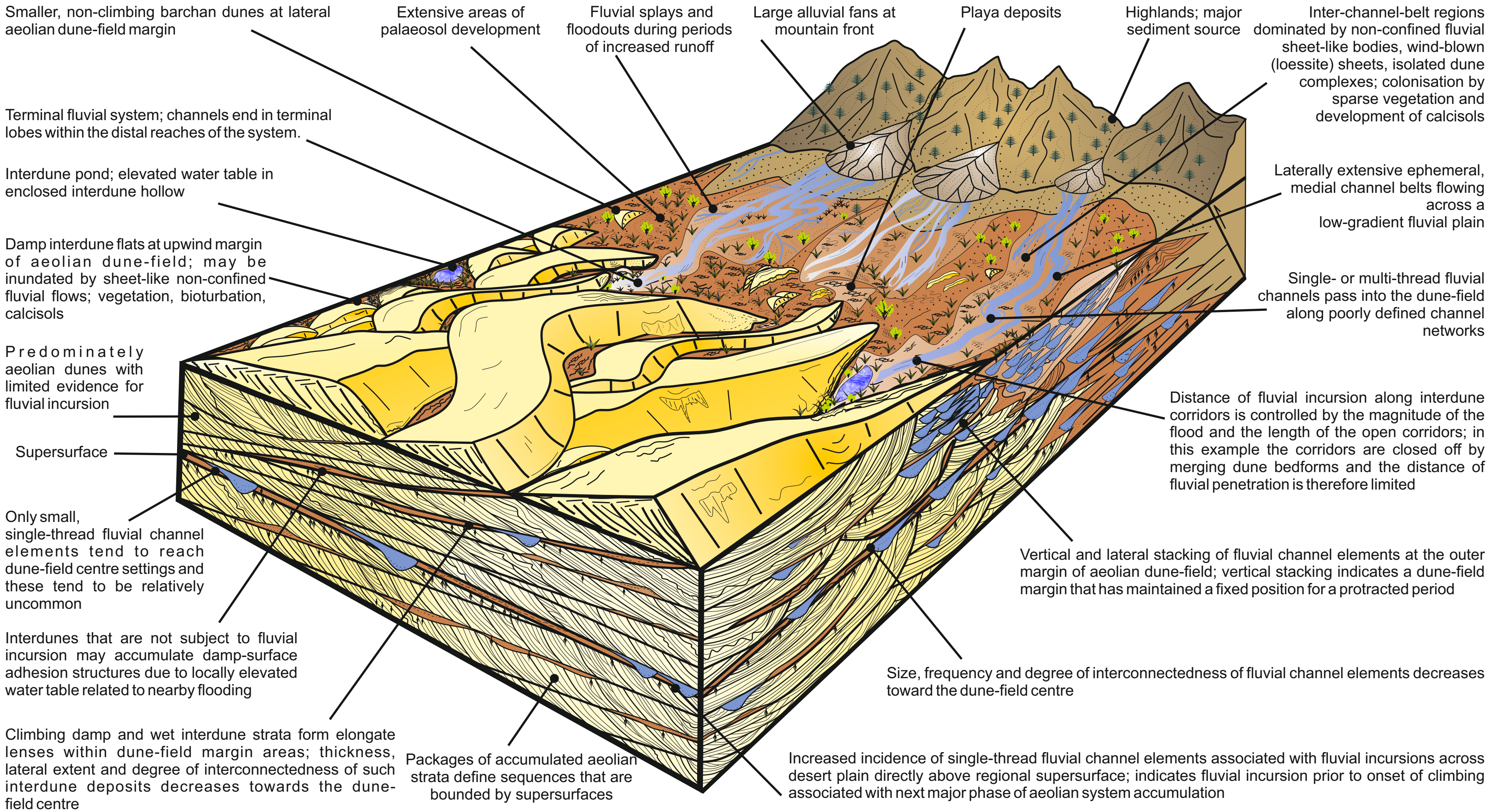


Figure 3.1: Schematic model illustrating common depositional processes that operate at dune-field margins, and resultant stratigraphic relationships. No particular scale implied.

desert sub-environments are known to have resulted in the preservation of complex arrangements of sedimentary deposits and stratigraphic architectures (Mountney, 2006a, 2012).

Permanent, intermittent and ephemeral fluvial systems occur in many dryland regions (Powell, 2009), including in parts of Australia, India, Saudi Arabia, and the Southwestern United States (e.g., Schenk and Fryberger, 1988; Tooth, 2000a, b; Glennie, 1987, 2005; Nanson et al., 2002), and many such systems exhibit complex and long-lived interactions with aeolian dunes. Some fluvial systems serve to generate significant supplies of sediment that are subsequently available for aeolian-dune construction, as in the Kelso dune-field, Mojave desert of California (Sharp, 1966; Kocurek and Lancaster, 1999). Similarly, alluvial-fan systems that form laterally extensive bajada may contribute significant sources of sediment for aeolian landform construction, as is the case for the Mojave River, southeastern California (Blair and McPherson, 2009; Belnap et al., 2011), and the alluvial-fan systems that border parts of the Rub' Al-Khali sand sea, Saudi Arabia (Figure 3.2). Other fluvial systems limit the spatial extent of dune-fields and serve to remove significant volumes of sediment transported into river beds via aeolian processes from desert sedimentary systems (e.g. The Kuiseb River, Namibia, Goudie, 1972; Ward, 1983).

The role of fluvial systems in aeolian-dominated deserts is significant: they are important landscape-forming and developing agents in many dryland systems (Wainwright and Bracken, 2011). Although many studies have documented types of interaction between aeolian and fluvial systems in both modern systems (e.g. Langford, 1989; Trewin, 1993; Stanistreet and Stollhofen, 2002; Bullard and McTainsh, 2003) and their ancient preserved successions recognised in the geological record (e.g. Langford and Chan, 1988; 1989; Herries, 1993; Chakraborty and Chaudhuri, 1993; Mountney and Jagger, 2004; Jordan and Mountney, 2010; Spalletti et al., 2010), relatively few geomorphological studies have explicitly focused on types of interaction between contemporaneously active aeolian and fluvial systems (e.g. Frostick and Reid, 1987; Cooke et al., 1993; Tooth, 2000a,b; Bull and Kirkby, 2002; Parsons and Abrahams, 2009; Reid and Frostick, 2011; Liu and Coulthard, 2015). Analysis of types of aeolian-fluvial system interaction

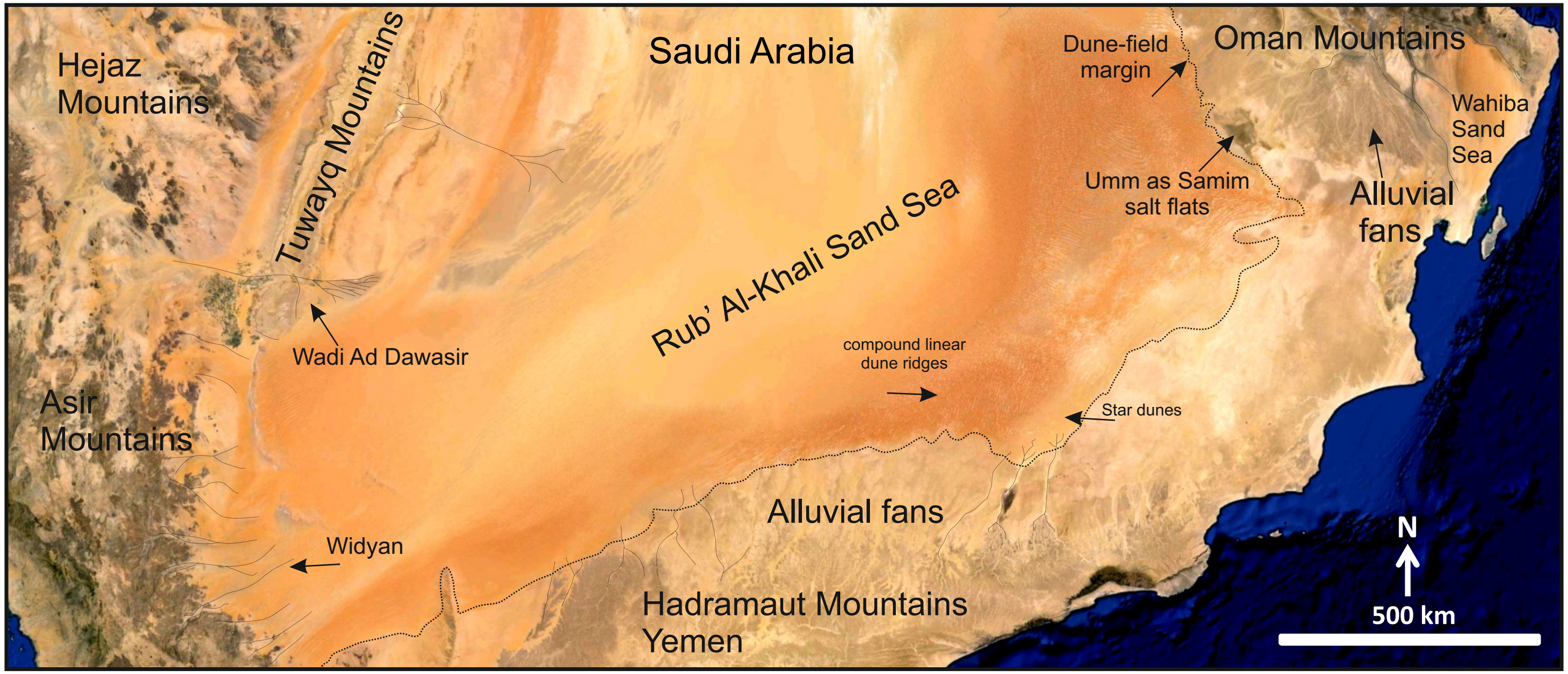


Figure 3.2: Google Earth image from southern Arabian Peninsula showing the location of the Rub' Al-Khali sand sea and surrounding mountains. Note the presence of alluvial systems with catchments in the mountainous regions that surround the dune fields and the fluvial drainage networks that enter the dune fields.

has implications for gaining an improved understanding of the effects of climate change. Furthermore, such analysis aids in the reconstruction of ancient palaeoenvironments (cf. Trewin, 1993; Herries, 1993; Yang et al., 2002; Al Farraj and Harvey, 2004; Simpson et al., 2008; Jordan and Mountney, 2010).

The increasing availability and global coverage of high-resolution satellite and aerial-photograph imagery through resources such as Google Earth (Butler, 2006; Yu and Gong, 2012; Fisher et al., 2012) has enabled the study of geomorphological relationships in detail for remote dryland settings (e.g. Tooth, 2006; Bullard et al., 2011; Al-Masrahy and Mountney, 2013). Significantly, the global coverage of such data means that comprehensive analyses can now be undertaken. This study utilises the latest generation of remotely sensed imagery to investigate the nature of aeolian and fluvial system interactions in a representative set of desert systems.

The aim of this study is to propose a generalised framework with which to account for the diverse types of interaction known to exist between coeval aeolian and fluvial depositional systems, and to discuss the significance of these interactions for the geomorphological and sedimentological evolution of mixed aeolian-fluvial systems. Specific objectives of this work are: (i) to illustrate the principal types of aeolian-fluvial interactions documented from the world's major dryland systems; (ii) to propose a framework for their classification; (iii) to demonstrate how the orientation of fluvial systems relative to the trend of aeolian bedforms present at the leading edge of dune-fields controls the nature of aeolian-fluvial system interaction; (iv) to document how open and closed interdune corridors act to control the type and extent of incursion of fluvial systems into aeolian dune-fields; (v) to consider how different types of aeolian-fluvial interaction give rise to complex geomorphic arrangements of landforms; and (vi) to consider the implications of such arrangements for the palaeoenvironmental reconstruction of ancient preserved counterparts (Figure 3.1).

This research is significant because it presents a robust framework to account for all the commonly identified types of aeolian-fluvial interaction in desert systems, which can be used as a tool to predict the likely spatial extent over which such interactions occur in both modern systems and their ancient counterparts preserved in the rock record.

3.3 Methodology

The morphological expression and areal distribution of 130 examples of fluvial-aeolian interaction have been mapped using high-resolution satellite imagery of 60 desert dune-fields around the world (Figure 3.3). Case study examples have been classified to propose a framework of ten distinct types of system interaction. Studied desert systems include the Namib Desert and Skeleton Coast (Namibia), Taklamakan Desert (northwest China), Rigestan Desert (southwestern Afghanistan), Sahara Desert (North Africa), Algodones (southeastern California), White Sands (New Mexico), Rub' Al-Khali and An Nafud sand seas (Saudi Arabia), and Wahiba Sands (Oman), Great Sany, Great Victoria, and Simpson deserts (Australia).

The Google Earth Pro software tool provides global coverage of remotely sensed imagery, including for desert regions that are generally not readily accessible by land. The satellite imagery used is from multiple sources and is of variable age; study sites have been selected in part on the availability of high-quality imagery with spatial resolution of resolution 15 m per pixel, derived from 15 to 30 m-resolution multispectral Landsat data that have been pan-sharpened with panchromatic Landsat image processing software. The software and its associated datasets have been used to generate a high-resolution images in the form of tiles, each up to 4800 x 2442 pixels, that have been near-seamlessly stitched together to yield detailed composite mosaic images that are well suited to detailed analysis of desert landforms.

3.4 Types of fluvial-aeolian interaction in aeolian dune-fields

The following discussion presents a novel classification scheme for types of interaction between fluvial systems that are present both within and at the margins of aeolian dune-field systems. Ten distinct types of interaction are recorded and illustrated by 130 case-study examples from 60 deserts around the world.

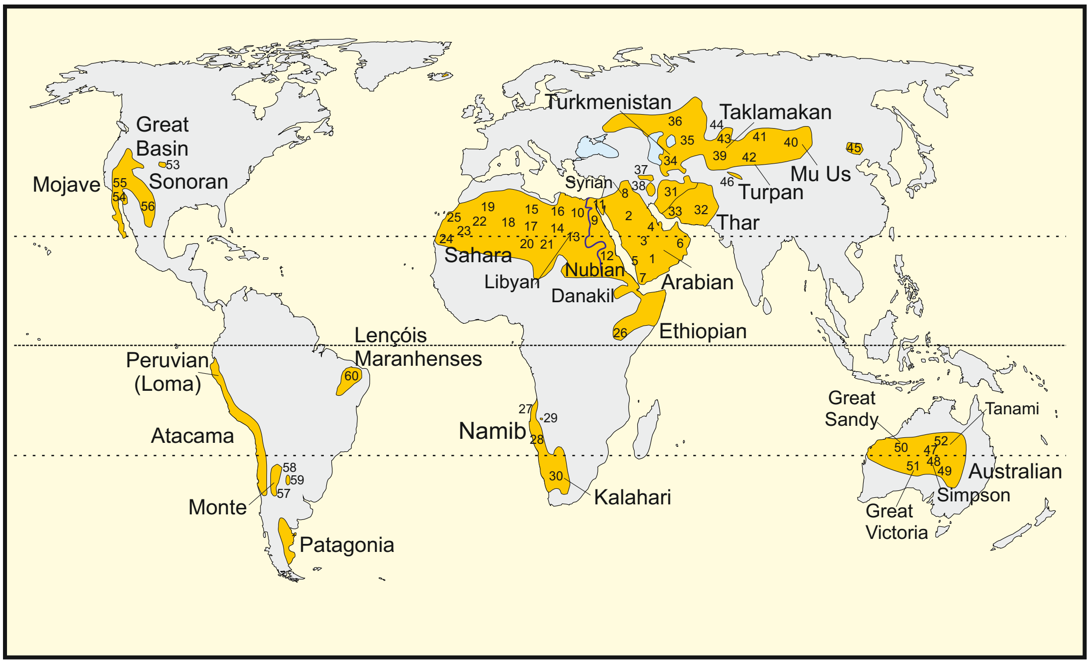


Figure 3.3: Geographic locations of the sixty studied desert systems: 1 – Rub' Al-Khali Desert, 2 – An Nafud Desert, 3 – Ad Dahna Desert, 4 – Al Jafurah Desert, and 5 – Tihama Dune Fields Saudi Arabia; 6 – Wahiba Sands, Oman; 7 – Coastal Dune Field southern Yemen; 8 – Syrian Desert, Syria; 9 – Eastern Desert, 10 – Western Desert, and 11 – Sinai Desert, Egypt; 12 – Nubian Desert, northern Sudan; 13 – Libyan Desert, eastern Sahara Desert; 14 – Idhan Murzuq Desert, Sahara Desert, Libya; 15 – Grand Erg Occidental Desert, 16 – Grand Erg Oriental Desert, 17 – Tassili-N-Ajjer Desert, 18 – Erg Iguidi Desert, and 19 – Hamada Du Draa Desert, Sahara Desert, Algeria; 20 – Tassili-Oua-Ahaggar Desert, Sahara Desert, Niger; 21 – Tenere Desert, Southern Sahara Desert, Chad; 22 – El Djouf Desert, 23 – Akchar Desert, and 24 – Trarza Desert, Sahara Desert, Mauritania; 25 – Western Sahara 26 – Chalbi Desert, Kenya; 27 – Namib Desert, 28 – Skeleton Coast, and 29 – Giribes Plain, Namibia; 30 – Kalahari Desert, South Africa; 31 – Rigestan Desert, Afghanistan; 32 – Thar Desert, 33 – Kharan Desert, Baluchistan, Pakistan; 34 – Garagum Desert, Turkmenistan; 35 – Qizilqum Desert, Uzbekistan; 36 – Betpaqdala Desert, Southern Kazakhstan; 37 – Kavir Desert , and 38 – Lut Desert, Iran; 39 – Taklamakan Desert, 40 – Mu Us Desert, 41 – Gobi Desert, 42 – Turpan Desert, 43 – Gurbantünggüt Desert, 44 – Junggar Basin, and 45 – Horqin Desert, Inner Mongolia, China; 46 – Dune Fields northern Tibetan Plateau, China; 47 – Simpson Desert, 48 – Tirari Desert, 49 – Strzelecki Desert, 50 – Great Sandy Desert, 51 – Great Victoria Desert, and 52 – Tanami Desert, Australia; 53 – White Sand Desert, New Mexico, 54 – Algodones Dune Field southeastern California, and 55 – Mojave Desert, California, United States; 56 – Sonoran Desert, Northeastern Mexico; 57 – Marayes Dune Field, and 58 – Vallecito Dune Field, Monte Desert, Argentina 59 – Salinas Grandes Desert, Argentina; 60 – Lençóis Maranhenses, or Brazilian Sahara, Brazil.

3.4.1 Fluvial incursions oriented parallel to trend of aeolian dune forms

In cases where the configuration of aeolian dunes is such that they form elongate ridges with crestlines aligned close to parallel to the direction of fluvial flow and where neighbouring dune ridges are separated by interdune flats, fluvial systems are typically able to penetrate along the interdune corridors and into the aeolian dune-field, in some cases for many tens of kilometres. One example of this type of interaction is the northern margin of the Simpson Desert, Australia (Nanson et al., 1995), where fluvial systems flow along open interdune corridors with an average width of 450 m, between linear dunes (Figure 3.4a). A second example is the Kharan Desert, Pakistan, where fluvial systems flow along open interdune corridors with an average width of 1250 m between barchanoid and transverse dune ridges (Figure 3.4b). These and other representative examples are listed in Table 3.1.

Where interdune corridors between dunes are open, they serve to guide flood waters and provide the required paths for water to advance significant distances into aeolian dune-fields. Where interdune corridors narrow but nevertheless remain open, they may promote a localised increase in stream power as floods of a given discharge are forced through a narrow constriction, which may result in localised erosion, either laterally from the toes of adjoining aeolian dunes or via scour on the bed of the interdune corridor. Where erosion of aeolian deposits occurs, the nature of the sediment load being carried by flood waters will change, and this will influence the sedimentary character of resultant flood deposits. Where interdune corridors become closed, for example where two neighbouring dune ridges meet, flood waters will pond, giving rise to standing water bodies that gradually desiccate in the aftermath of the flood event; Sossusvlei in the Namib Desert is one such example. Where aeolian sand is blown over the course of river channels during dry episodes, the fluvial course may be progressively diverted with each successive flood event (Figure 3.5a) or terminated (Figure 3.5b).

This type of interaction results in the deposition of ribbon-like fluvial deposits in cases where the aeolian dunes that funnel the flood waters into specific

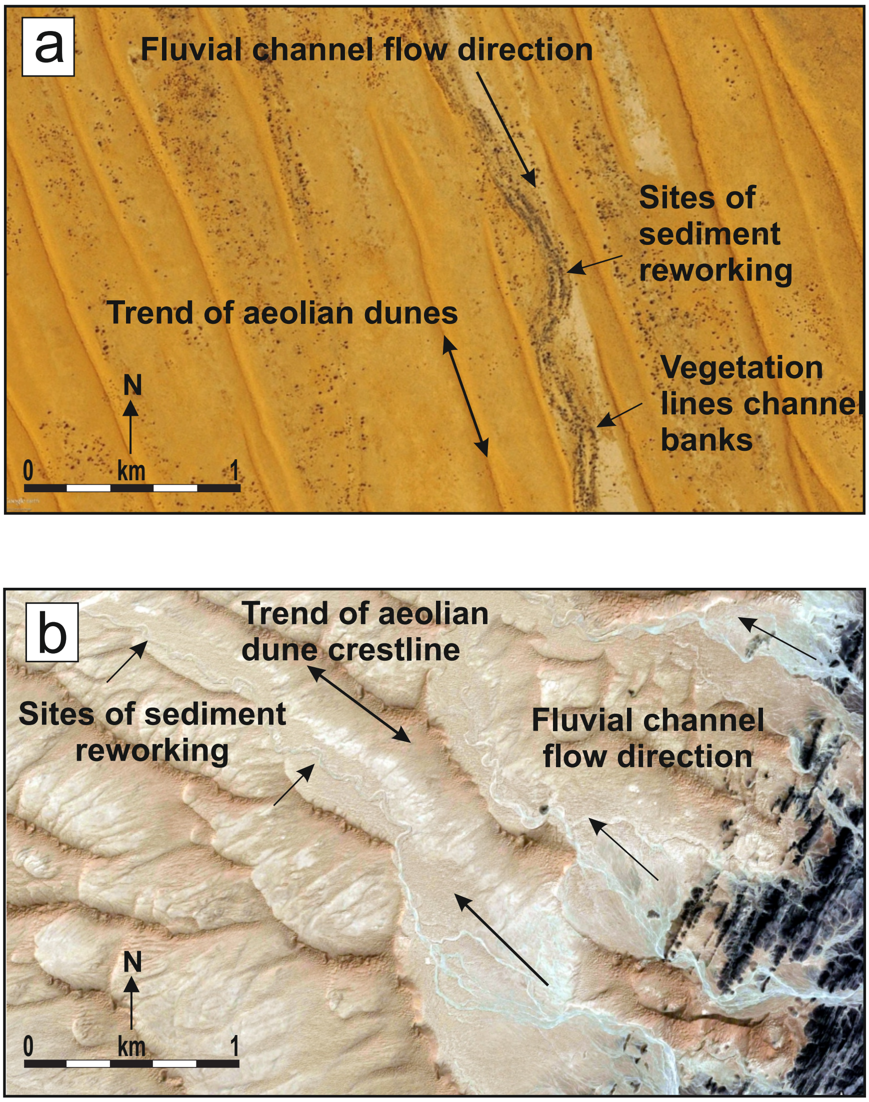


Figure 3.4: Examples of fluvial incursions oriented parallel to trend of the crestlines of aeolian dune forms. (a) Northern Simpson Desert, Australia (24 23 07 S 135 28 24 E); (b) Kharan Desert, Baluchistan Province, Pakistan (28 16 54 N 65 29 20 E). See text for discussion. Image source: Google Earth Pro.

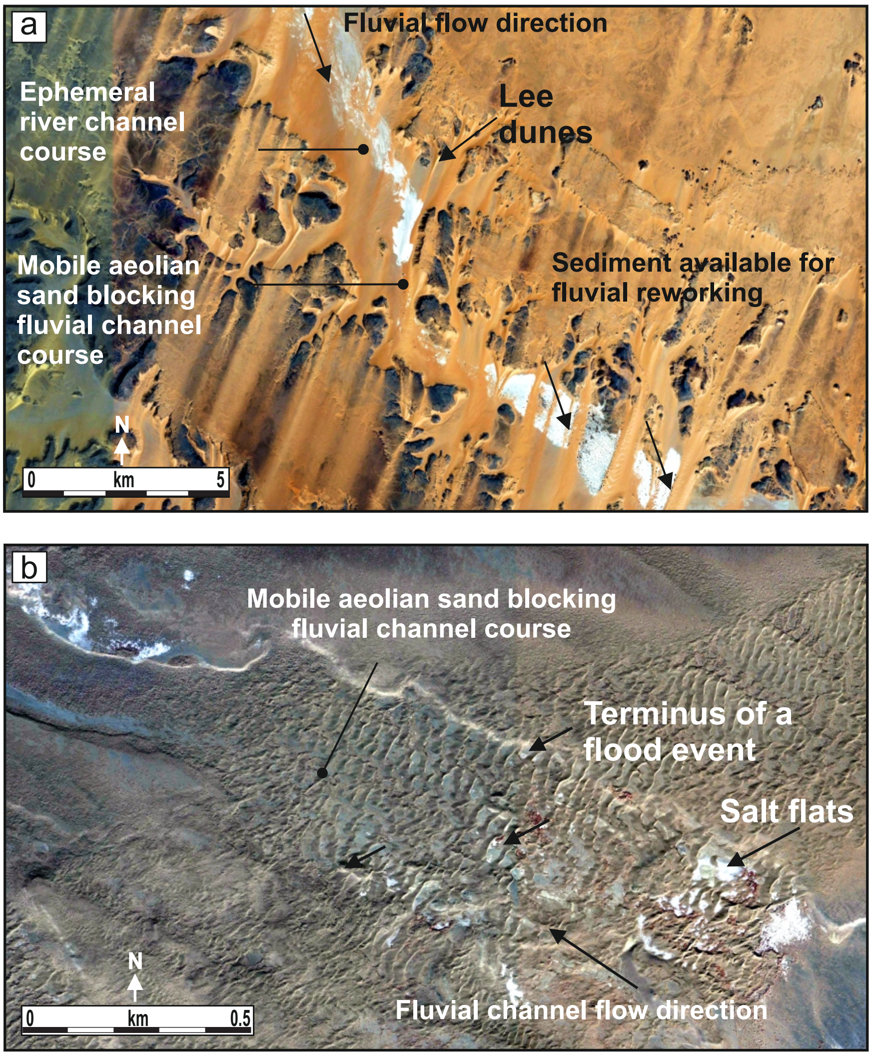


Figure 3.5: Examples of mobile dunes occupying fluvial channel courses. (a) Sahara Desert, Northern Chad (19 59 03 N 19 31 19 E); (b) Gurbantünggüt Desert, northwestern China (44 24 03 N 91 05 17 E). See text for discussion. (Image source: Google Earth Pro).

Table 3.1: Scheme for the classification of types of aeolian-fluvial system interaction, with 130 notable case-study examples documented from 60 modern desert systems. Column labelled “Fig. 3” provides a cross-reference to the desert locations shown in Figure 3. Abbreviations for aeolian bedform types: S – star; Cs – complex star; Br – barchan; Bi – barchanoid ridges; SB – superimposed barchanoid ridges; T – transverse; L – linear; P – parabolic; Cb – compound barchan; R – reverse; D – dome; SS – Sand sheets.

Interaction type	Case Study No.	Example Desert	Desert No. (see Fig. 3)	Case Study Location	Dune spacing at outer dune-field margin (km)	Dune spacing at inner dune-field margin (km)	Interdune width at outer dune-field margin (km)	Interdune width at inner dune-field margin (km)	Mean fluvial channel width (m)	Fluvial channel extent within dune-field (km)	Dominant aeolian bedform type
1: Fluvial incursions oriented parallel to trend of aeolian dune forms	1	Southern El Djouf Desert, Mauritania	22	18 04 17 N 11 11 09 W	2.26	1.51	2.21	1.40	244	114	L/T
	2	Western Idhan Murzuq Desert, Libya	14	24 34 52 N 11 43 59 E	3.38	1.64	3.31	1.35	21	13	Bi
	3	Southwestern Rub’ Al-Khali Desert	1	16 46 26 N 45 25 39 E	4.32	1.42	4.10	1.52	68	58	L
	4			17 01 50 N 45 16 23 E	4.65	1.17	4.20	0.35	80	34	L
	5	Grand Erg Oriental Desert, Algeria	16	29 00 56 N 04 36 20 E	2.40	1.30	2.10	1.62	31	15	L
	6	Northern Simpson Desert , Australia	47	24 23 07 S 135 28 24 E	0.51	0.19	0.45	0.45	67	29	L
	7			24 03 46 S 135 55 26 E	1.91	0.18	1.82	0.42	62	33	L
	8	Kharan Desert, Baluchistan, Pakistan	33	28 16 54 N 65 29 20 E	1.40	0.45	1.25	1.01	57	27	T/Bi
2: Fluvial incursions oriented perpendicular to the trend of aeolian dune forms	9	Mu Us Desert, China	40	40 22 29 N 109 18 00 E	1.6	0.49	1.40	0.18	50	07	Bi
	10	Wahiba Sand Sea, Oman	6	22 25 19 N 58 49 11 E	1.90	1.60	1.60	1.07	327	120	L
	11	Eastern Rub’ Al-Khali Desert, Saudi Arabia	1	19 10 06 N 44 24 58 E	0.18	0.13	0.15	0.10	36	06	Bi/T
	12	Namib Desert, Namibia	27	23 40 59 S 15 14 16 E	2.21	1.60	1.90	1.45	205	147	L
	13	Southern Simpson Desert, Australia	47	27 13 18 S 137 56 43 E	0.81	0.17	0.75	0.45	124	89	L
	14	Northern Simpson Desert, Australia	47	24 15 14 S 135 35 09 E	0.83	0.18	0.72	0.42	148	16	L
	15	Strzelecki Desert, Australia	49	28 25 11 S 138 56 35 E	1.35	0.78	1.12	0.40	133	198	L
	16	Kharan Desert, Baluchistan, Pakistan	33	27 46 16 N 63 48 19 E	2.31	0.24	2.11	0.04	165	86	Bi/L
	17	Tassili-N-Ajjer Desert, Sahara, Algeria	17	26 32 28 N 07 53 49 E	3.39	2.80	2.90	1.21	460	53	CS/L
	18	West Salinas Grandes Desert, Argentina	59	31 45 04 S 67 04 05 W	0.15	0.13	0.10	0.07	260	20	L
3: Bifurcation of fluvial flow between isolated aeolian dune forms	19	Rub’ Al-Khali Desert, Oman	1	18 31 24 N 53 22 06 E	1.30	1.27	0.90	1.67	18	32	S
	20			18 27 00 N 53 12 06 E	1.40	1.10	1.00	1.19	82	20	S
	21			18 35 23 N 53 25 35 E	1.85	1.38	1.50	1.47	73	39	S/Cs
	22	Rub’ Al-Khali Desert, Northeastern Yemen	1	18 37 12 N 51 24 40 E	3.03	1.57	2.60	2.43	120	7.61	S/CS/D
	23	Tenere Desert, Southern Sahara Desert, Chad	21	13 40 22 N 16 16 34 E	2.70	1.87	2.10	2.20	188	135	T/Br

Table 3.1: Cont.

	24	Taklamakan Desert, China	39	38 22 42 N 81 53 46 E	1.46	1.58	1.20	1.67	74	161	Cb/SB
	25	Horqin Desert, Inner Mongolia, China	45	43 12 45 N 118 48 25E	0.69	0.48	0.53	0.22	134	06	T
	26	Dune Field southern Tibetan Plateau, China	46	29 55 57 N 83 31 48 E	0.66	0.34	0.73	0.14	63	60	Br
4: Through-going fluvial channel networks that cross entire aeolian dune fields	27	Mu Us Desert, China	40	40 15 15 N 109 46 35E	0.16	0.15	0.11	0.04	74	17	Bi
	28	Eastern Grate Victoria Desert, Australia	51	28 57 46 S 135 56 51 E	0.98	0.62	0.85	0.35	138	98	L
	29	Tirari Desert, Australia	48	27 49 13 S 137 37 34 E	1.17	0.48	1.12	0.26	182	162	L
	30	Southern Libyan Desert, Sudan	13	15 42 49 N 26 27 06 E	0.49	0.31	0.35	0.08	116	97	Bi/SS
	31	Nile River, eastern Sahara Desert/Sudan	12-13	18 55 06 N 30 33 47 E	0.49	0.31	0.41	1.23	620	800	Br
	32	Tihama Dune Fields, Saudi Arabia	5	19 26 36 N 41 06 29 E	0.35	0.11	0.31	0.24	142	34	Br/SS
	33	Sinai Desert, Egypt	11	30 56 43 N 33 57 26 E	0.49	0.32	0.42	0.17	57	56	Cb/L
	34	Southern El Djouf Desert, Mali	22	16 57 07 N 01 52 06 W	2.04	0.96	1.90	1.80	561	500	L
	35	Holtan River, Taklamakan Desert, China	39	39 15 31 N 80 52 22 E	2.28	0.12	2.10	0.10	157	396	Bi/SB/Cb
	36	Garagum Desert, Turkmenistan	34	40 17 42 N 61 50 23 E	2.45	0.16	2.20	0.18	475	275	SB/Br
	37	Kalahari Desert, South Africa	30	25 06 26 S 20 20 37 E	0.75	0.26	0.68	0.21	90	602	L
	38	Horqin Desert, Inner Mongolia, China	45	43 07 51 N 119 17 45E	0.28	0.22	0.14	0.07	48	42	T/Br
	39	Grand Erg Occidental Desert, Algeria	15	29 07 22 N 01 01 50 W	6.90	0.11	5.09	0.03	115	289	Cs/T/SB
5: Fluvial flooding of aeolian dune fields associated with elevated water-table level	40	Gobi Desert, northern China	41	39 46 11 N 102 09 00 E	4.23	2.99	1.13	2.59	NA	NA	SB
	41			38 40 42 N 104 54 15 E	2.70	3.22	1.90	0.86	NA	NA	T/Br
	42	Taklamakan Desert, China	39	40 54 43 N 85 30 27 E	1.42	1.24	0.95	0.51	NA	NA	Cb
	43	Al Jafurah Desert, Eastern Saudi Arabia	4	25 47 17 N 49 48 28 E	0.62	0.51	0.38	0.15	NA	NA	T/Br
	44	North-eastern Rub Al-Khali, Saudi Arabia	1	24 26 14 N 51 09 37 E	0.83	0.91	0.67	0.45	NA	NA	Br/T
	45	Western Desert, Egypt	10	29 08 13 N 25 26 33 E	2.64	0.69	1.30	0.51	NA	NA	L/SB
	46	Northern Grand Erg Oriental Desert ,Tunisia	16	33 37 35 N 07 56 32 E	2.30	0.38	0.97	0.15	NA	NA	L
	47	Libyan Desert, Northeastern Chad	13	18 56 52 N 20 51 36 E	NA	0.57	NA	0.21	NA	NA	Br/Bi
	48	Tenere Desert, Southern Sahara Desert, Chad	21	14 35 38 N 14 42 29 E	3.85	2.65	2.40	0.63	NA	NA	T
	49	Betpaqdala Desert, Southern Kazakhstan	36	43 32 33 N 72 18 11 E	1.65	2.80	0.97	1.65	NA	NA	Br/Bi
	50	Thar Desert, Pakistan	32	26 23 01 N 69 45 01 E	4.11	0.67	2.10	0.22	NA	NA	SB/Bi
	51	Lençóis Maranhenses Desert, Brazil	60	02 34 31 S 42 57 03 W	0.43	0.54	0.27	0.27	NA	NA	Br/Bi/T
	52	Great Victoria Desert, Australia	51	28 39 16 S 128 20 58 E	1.31	1.42	0.65	0.63	NA	NA	L
	53	Dune Field northern Tibetan Plateau, China	46	37 04 20 N 90 33 05 E	1.14	0.70	0.53	0.22	NA	NA	Cs/Bi
	54	Mu Su Desert, China	40	39 14 43 N 108 50 36 E	0.52	0.34	0.37	0.18	NA	NA	T/Br/Bi
	55	Horqin Desert, Inner Mongolia, China	45	42 57 47 N 119 33 38E	0.71	0.61	0.42	0.23	NA	NA	T/Br/Bi
6: Fluvial incursions into aeolian dune fields associated	56	Giribes Plain, Namibia	29	19 01 34 S 13 21 34 E	NA	NA	NA	NA	32	2.91	SS
	57	Southern Kavir Desert ,Iran	37	33 36 37 N 53 45 55 E	2.14	1.26	1.42	0.25	26	17	L/Bi
	58	Simpson Desert, Australia	47	24 08 55 S 135 13 56 E	1.74	1.68	1.33	1.24	15	05	L/SS
	59	Sonoran Desert, Northwestern Mexico	56	32 03 08 N 113 37 37 W	0.38	0.56	0.28	0.09	13	04	L

Table 3.1: Cont.

with a single point source	60			34 12 03 N 115 16 50 W	0.31	0.38	0.12	0.04	04	06	D/SS
	61	White Sand Desert, USA	53	32 51 54 N 106 12 11W	0.21	0.16	0.10	0.06	07	05	T/Br/P
	62	Grand Erg Occidental Desert, Algeria	15	32 26 55 N 00 10 53 E	3.35	3.16	2.20	0.67	262	135	T/L/Br
	63			32 30 19 N 00 08 39 W	0.19	0.80	0.12	0.04	63	03	L
	64	Libyan Desert, central Sahara Desert, Libya	13	23 55 43 N 19 46 42 E	3.57	3.07	2.73	0.49	105	09	L
	65	Tenere Desert, central Sahara Desert, Chad	21	19 23 40 N 16 37 02 E	1.00	0.88	0.57	0.36	187	93	T/Bi/SB
	66	Akchar Desert, Mauritania	23	20 42 53 N 11 59 50 W	0.32	0.12	0.27	0.03	127	10	Br/Bi/L
	67	Erg Iguidi Desert, Algeria	18	27 31 26 N 03 45 48 W	2.30	3.60	1.20	2.30	112	22	S/L
7: Fluvial incursions into aeolian dune fields associated with a multiple sheet source	68	Sonoran Desert, Northwestern Mexico	56	31 28 13 N 112 55 36 W	0.56	0.26	0.46	0.19	228	08	L/Bi
	69			31 45 29 N 113 08 19 W	0.37	0.17	0.27	0.07	397	07	L/Bi
	70	Algodones Dune Field, south California, USA	54	33 06 00 N 115 14 44 W	1.48	0.43	0.75	0.15	244	02	T/Bi
	71	Tassili-N-Ajjer Desert, Algeria	16	26 43 05 N 06 54 04 E	6.08	2.96	3.94	0.96	240	02	Cs/SB/L
	72	Hamada Du Draa Desert, Algeria	19	29 54 39 N 03 08 59 W	0.71	1.14	0.48	0.34	385	1.6	Cs
	73	Akchar Desert, Mauritania	23	21 26 46 N 11 42 37 W	0.53	0.32	0.35	0.07	387	1.74	Bi/SB/L
	74	Southern Kavir Desert ,Iran	37	33 32 33 N 53 56 43 E	1.74	1.25	1.09	0.41	580	06	Br/Bi
	75	Lut Desert, Iran	38	30 03 51 N 59 37 57 E	0.43	2.41	0.12	0.51	1200	07	T/S/L
	76	Kharan Desert, Baluchistan, Pakistan	33	28 47 10 N 64 23 01 E	0.31	0.11	0.09	0.05	289	2.4	T/Bi/SB
	77	Betpaqdala Desert, Southern Kazakhstan	36	44 17 57 N 68 43 37 E	0.82	1.12	0.53	0.49	269	540	Bi/T
	78	Gobi Desert, northern China	41	41 36 31 N 101 58 43 E	0.83	1.44	0.38	0.24	197	0.86	S/Bi
	79	Mojave Desert, California	55	34 56 27 N 115 39 10 W	0.35	0.18	0.17	0.05	377	0.91	Bi/SB
8: Cessation of encroachment of aeolian dune fields by fluvial systems	80	Qizilqum Desert, Uzbekistan	35	44 12 28 N 66 08 20 E	1.30	0.27	0.64	0.09	326	589	T/Bi
	81	Kuiseb River, Namib Desert	27	23 30 21 S 14 59 00 E	2.28	2.25	0.97	0.77	307	150	L/Bi
	82	Swakop River, Namib Desert	27	22 41 14 S 14 32 36 E	0.12	0.21	0.13	0.08	185	04	L/Bi
	83	Kunene River, Namib Desert	27	17 15 29 S 11 49 17 E	0.42	0.49	0.17	0.03	180	63	Bi/SB/Br
	84	Hoarusib River, Skeleton Coast, Namibia	28	19 01 15 S 12 39 07 E	0.61	0.41	0.37	0.17	274	26	Bi/SB/Cb
	85	North Namib Desert, Angola	27	16 17 40 S 12 16 23 E	0.17	0.28	0.04	0.18	119	08	Bi/SB
	86			15 46 50 S 11 59 01 E	0.18	0.27	0.04	0.06	462	86	Bi
	87	Yellow River, Mu Us Desert , China	40	40 04 26 N 106 44 06 E	0.28	0.27	0.19	0.04	687	147	Bi/SB/T
	88			40 06 38 N 110 40 57 E	0.33	0.11	0.19	0.03	53	10	Bi/T
	89	Irtysh River, Junggar Basin, Northwestern China	44	47 57 22 N 85 42 40 E	0.22	0.40	0.33	0.08	342	100	Bi/T
	90	Tuolahai River, Northern Tibetan Plateau, China	46	36 42 06 N 94 30 03 E	0.33	0.17	0.41	0.09	232	24	Br/T
	91	Vallecito Dune Field, Monte Desert, Argentina	58	31 52 15 S 67 49 43 W	1.98	2.34	1.04	0.35	116	50	L/Bi/SB
	92	Marayes Dune Field ,Monte Desert, Argentina	57	31 22 32 S 67 29 52 W	1.07	1.47	0.41	0.17	258	27	L/Bi
	93	Helmand River, Rigestan Desert, Afghanistan	31	31 22 34 N 65 53 27 E	0.22	0.18	0.09	0.04	218	176	Bi/SB
	94	Euphrates River, Northern Syrian Desert, Syria	8	34 50 25 N 40 24 35 E	NA	NA	NA	NA	391	65	SS
	95	Chalbi Desert, Kenya	26	02 51 35 N 37 45 13 E	0.26	0.13	0.16	0.02	75	16	L/SS

Table 3.1: Cont.

	96	Nile River, Western Desert, Egypt	10	28 12 00 N 30 31 26 E	0.59	0.23	0.31	0.04	643	364	Bi/SB/SS
	97	Northern Hamada Du Draa Desert, Morocco	19	31 33 00 N 04 31 21 W	0.07	0.04	0.05	0.02	55	13	T/Br
9: Termination of fluvial channel networks in aeolian dune fields	98	Coastal Dune Field southern Yemen	7	14 17 22 N 47 54 39 E	3.42	1.34	1.83	0.15	189	11	Bi/SB
	99	An Nafud Desert, Saudi Arabia	2	24 22 58 N 46 14 14 E	2.07	1.08	0.72	0.29	53	04	Bi/SB/D
	100	Tassili-Oua-Ahaggar Desert, Sahara , Niger	20	20 06 00N 08 37 51 E	1.73	0.43	0.74	0.05	40	03	S/R/Bi
	101	Tenere Desert, Sahara, Niger	21	19 20 29 N 16 34 23 E	0.58	0.27	0.38	0.12	165	101	T/Bi
	102	Ad Dahna Desert, Saudi Arabia	3	25 17 40 N 47 24 12 E	5.42	7.38	2.46	2.69	68	03	Bi/SB
	103			25 20 58 N 47 17 31 E	5.63	8.56	3.34	2.28	67	01	Bi/SB
	104	Taklamakan Desert, China	39	37 41 22 N 82 41 28 E	3.53	3.07	1.24	1.35	199	95	L/Br
	105	Turpan Desert, China	42	42 31 06 N 90 21 53 E	1.12	1.99	0.65	0.37	95	04	Cs/T/R
	106	White Sand Desert, USA	53	32 57 57 N 106 14 03 W	0.59	0.10	0.59	0.03	46	12	T/Br/P
	107	Sonoran Desert, Northwestern Mexico	56	31 50 15 N 113 11 39 W	0.42	0.24	0.33	0.18	78	06	Bi/L/SS
	108	Gurbantünggüt Desert, Northwestern China	43	44 26 50 N 89 20 33 E	0.36	0.24	0.18	0.06	102	53	L/Bi
	109	Lut Desert, Iran	38	29 37 35 N 58 50 09 E	2.60	2.05	1.78	0.21	108	04	Bi/L
	110	Great Sandy Desert, Australia	50	22 09 41 S 122 55 00 E	1.50	0.33	1.02	0.26	188	68	L
	111	Simpson Desert, Australia	47	24 10 29 S 135 15 53 E	0.63	0.35	0.55	0.25	115	11	L
	112	Vallecito Dune Field, Monte Desert, Argentina	58	31 49 59 S 67 53 02 W	1.23	1.47	0.63	0.19	41	08	L/Bi/SB
	113	Tsondabvlei, Namib Desert, Namibia	27	23 55 37 S 15 22 36 E	2.67	2.08	1.89	0.98	40	60	L/S/Bi
	114	Namib Desert, Angola	27	16 22 05 S 12 09 36 E	0.15	0.22	0.09	0.03	135	1.5	SB/T
	115	Trarza Desert, Mauritania	24	19 33 58 N 13 19 54 W	7.22	3.45	3.69	1.32	61	05	Bi/SB/Br
	116	Skeleton Coast, Namibia	28	19 57 15 S 13 12 24 E	0.88	0.27	0.64	0.12	99	1.4	SB/Bi/L
	117			20 01 46 S 13 16 17 E	0.38	0.26	0.29	0.06	145	03	SB/Bi/L
	118	Western Sahara	25	27 09 45 N 13 15 10 W	0.84	0.14	0.73	0.08	40	58	Bi/Br/Cb
10: Examples of short-term versus long-term fluvial-aeolian interaction	119	Southeastern Libyan Desert, Sudan	13	15 39 11 N 26 25 44 E	0.48	0.42	0.28	0.20	84	09	T/SS
	120	Western Libyan Desert, North Chad	13	19 59 03 N 19 31 19 E	0.14	0.35	0.05	0.11	374	30	Br/Cb/L
	121	Hamada Du Draa Desert, Algeria	19	28 58 03 N 04 02 14 W	6.56	4.81	3.84	3.86	217	23	Bi/S/L
	122			28 52 38 N 04 02 13 W	5.35	2.46	2.89	1.12	410	NA	Bi/S/L
	123	Eastern Sahara Desert, Egypt	9	23 09 39 N 30 42 44 E	0.57	0.28	0.34	0.04	NA	NA	Bi/Cb/D
	124	Great Sandy Desert, Australia	50	22 38 00 S 123 18 36 E	1.23	0.76	0.97	0.40	NA	NA	L
	125			22 18 10 S 128 56 12 E	2.87	0.26	5.56	0.18	NA	NA	L
	126	Tanami Desert, Australia	52	19 23 02 S 131 35 10 E	2.04	1.05	1.67	0.75	NA	NA	L
	127	Gurbantünggüt Desert, Northwestern China	43	44 24 03 N 91 05 17 E	0.22	0.23	0.13	0.09	NA	NA	T/Bi
	128	Betpaqdala Desert, , Southern Kazakhstan	36	43 34 11 N 72 12 56 E	8.71	4.01	5.28	1.21	NA	NA	Bi
	129	Taklamakan Desert, China	39	37 55 41 N 81 28 49 E	1.48	2.18	0.98	0.61	NA	NA	Cb/SB
	130			37 56 35 N 81 32 18 E	1.48	2.18	0.98	0.61	NA	NA	Cb/SB

interdune corridors are fixed in position. Alternatively, in cases where the dunes and their intervening interdunes gradually migrate laterally between successive flood events, fluvial deposits arising from successive floods may expand laterally to form more sheet-like depositional elements (cf. Langford and Chan, 1988). In both cases, the opportunity for aeolian reworking of flood deposits is significant, and winnowing of sand and finer fractions by the wind is likely, resulting in the generation of armoured lag deposits (Krapf et al., 2005; Simpson et al., 2008). Thus, fluvial incursion along interdune corridors can generate a local supply of sediment suitable for later aeolian construction. Conversely, the deposition of mud drapes through suspension settling in ponded flood waters may limit the availability of underlying sand substrates for later aeolian transport (Cain and Mountney, 2009, 2011).

3.4.2 Fluvial incursions oriented perpendicular to the trend of aeolian dune forms

In cases where the configuration of aeolian dunes is such that they form elongate ridges with crestlines aligned close to perpendicular to the direction of fluvial flow, aeolian topography will exert a significant control on fluvial flood pathways, and the nature of the flooding event. In cases where such a configuration is present at the outer margin of an aeolian dune-field, flood events may be prevented from passing into the dune-field and may instead become ponded or be diverted in orientations parallel to the trend of the dunes at the outer dune-field margin (Figure 3.6). Where flood waters pond, the water level may rise to a point where saddles (cols) between neighbouring dune crests are breached, thereby allowing fluvial incursion into the inner part of a dune-field. Fluvial breaching at specific sites will rapidly lead to erosion and incision as flow is forced through a narrow gap between dunes. Three examples where this process is documented are the interaction between sand dunes of the Mu Us Desert and the Sala Us River, Inner Mongolia, China (Li et al., 2012), ephemeral rivers of the Skeleton Coast, northwestern Namibia, including the Hoanib, Uniab, and Hunkab rivers (Stanistreet and Stollhofen, 2002), and the Todd River, northwestern Simpson Desert, Australia (Hollands et al., 2012). The interaction of Wadi Batha Oman with aeolian dunes of the Wahiba Sand Sea (Warren, 1988; Figure 3.6a) records a 120 km-long fluvial system that flows eastwards along

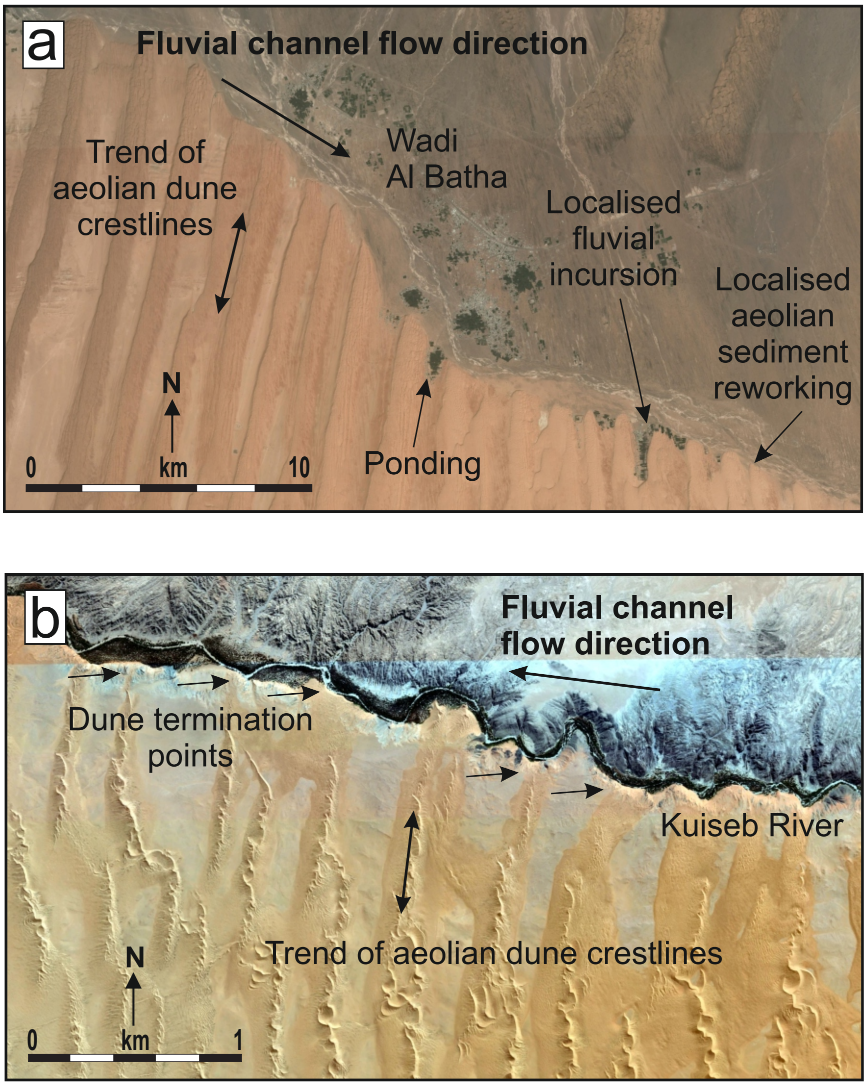


Figure 3.6: Examples of fluvial incursions oriented perpendicular to trend of the crestlines of aeolian dune forms. (a) Wahiba Sand Sea, Oman (22 25 19 N 58 49 11 E); (b) Namib Desert, Namibia (23 40 59 S 15 14 16 E). See text for discussion. (Image source: Google Earth Pro).

the northern margin of a dune-field composed of north-south trending linear dunes with an average dune spacing of 1900 m. Fluvial incursion into the dune-field is restricted to the outermost 1 to 2 km of open interdune corridors where localised ponding of floodwater occurs. The northern and eastern boundaries of the dune-field are delineated by the Wadi Batha, which maintains a course close to perpendicular to the tip-out points of the large linear dunes. At the northern margin of the Namib Desert, Namibia (Figure 3.6b), the northward advance of large linear dunes of the Namib sand Sea is curtailed by the Kuiseb River, which intermittently flows westwards: aeolian sand blown into the river channel during dry episodes is periodically flushed up to 147 km downstream during major seasonal flood events. These and other representative examples are listed in Table 3.1.

This type of interaction is typically expressed as a sharp boundary between adjoining fluvial and aeolian environments. Where fluvial flood waters repeatedly pond against the leading edge of an aeolian dune-field, fine-grained, mudstone layers will progressively accumulate (e.g., Wadi Al Ayn and Wadi Al Batha, Oman: Glennie, 2005). In cases where flood waters are saline and where ponded water evaporates or infiltrates only slowly, salts such as calcium carbonate, gypsum, halite or potash may be precipitated (Valyashko, 1972). For example, the salt flats of Umm as Samim, close to the eastern border of the Rub’ Al-Khali Sand Sea, Oman, occur in a low-lying area between the alluvial fans to the north, the aeolian dunes of the Rub' Al Khali to the west and south (Figure 3.2; Goodall et al. 2000). If the outer edge of the aeolian dune-field gradually expands over time via dune migration, aeolian deposits may become juxtaposed over flood deposits. Conversely, if the outer edge of the aeolian dune-field gradually retreats (contracts), aeolian deposits may become overlain by flood deposits.

3.4.3 Bifurcation of fluvial flow between isolated aeolian dune forms

In cases where fluvial flood waters pass into the outer parts of aeolian dune-fields that are characterised by isolated bedforms or small clusters of bedforms of variable size, orientation and spacing, the physical organisation of the dunes (or dune clusters) may encourage flood waters to bifurcate around the topographic obstacles on both sides. This process is common in the southeastern part of the Rub' Al-Khali Desert, Oman (Figure 3.2), which is dominated by fields of simple and compound star dunes that are bordered by the mountains of Oman from which flood events emanate. The distance of penetration of these fluvial systems is 20 to 40 km (Figure 3.7a), and this is governed by the flow frequency and magnitude, surface topography, substrate type (which governs infiltration rate and capacity) and aeolian bedform morphology. In some examples, such as the Keriya River in the Taklamakan Desert, China, intricate threading of fluvial channels between migrating but spatially isolated aeolian dunes is widespread (Figure 3.7b): in this example aeolian bedforms or clusters of bedforms that comprise small dune-fields are fixed in position by well-established fluvial courses. Similar types of interaction are also common in non-desert aeolian settings, including on Skeiðarársandur, southern Iceland (Mountney and Russell, 2009). These and other representative examples are listed in Table 3.1.

The presence of flowing water in such settings may affect sand dunes either directly through erosion or indirectly by generating a local supply of sediment suitable for later aeolian construction. In cases where episodic flooding results in a water table level that remains permanently close to the aeolian accumulation surface, such that the dune-field margin may be classed as a wet aeolian system (*sensu* Kocurek and Havholm, 1993), the long-term preservation potential of migrating but spatially isolated aeolian bedforms may be enhanced (cf. Mountney and Russell, 2009).

3.4.4 Through-going fluvial channel networks that cross entire aeolian dune-fields

In cases where fluvial systems pass through entire aeolian dune-fields, the presence of a fluvial course may act to effectively partition the dune-field by disrupting or limiting aeolian sediment transport pathways (Figure 3.8a; cf. Ward, 1987; Krapf et al., 2003). Such fluvial channel networks (or non-channelised fluvial pathways) may be either permanent (e.g. Nile River, Sudan), intermittent (e.g. Saoura River, Algeria) or ephemeral (e.g. Uniab River, Skeleton Coast, Namibia and Wadi Juweiza, United Arab Emirates). Such fluvial systems may operate as an agent of aeolian erosion; seasonally active fluvial courses may be filled with aeolian-derived sediment during dry episodes, and this sediment will be flushed downstream out of the dune-field

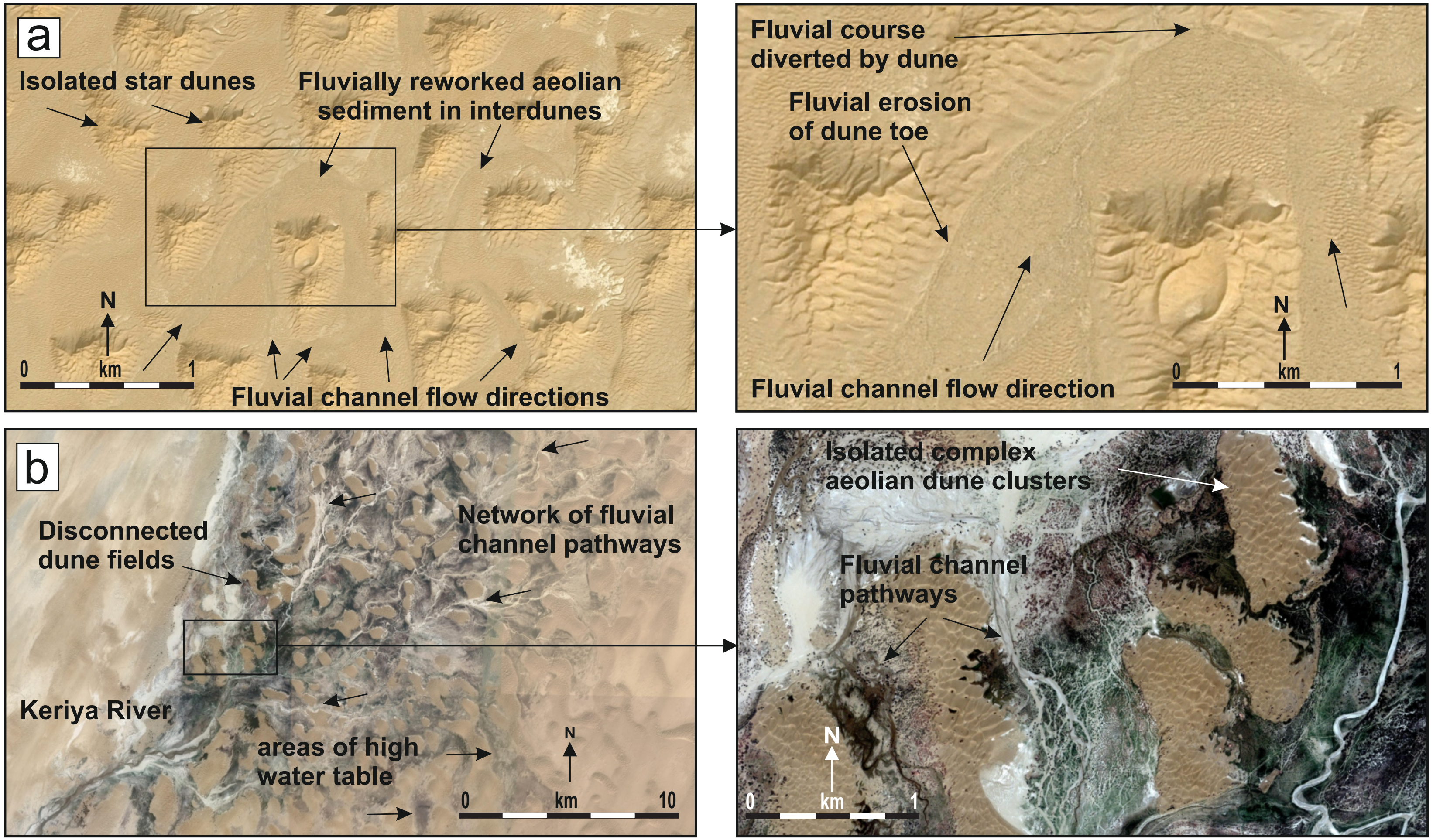


Figure 3.7: (a) Example of ephemeral fluvial channel network between star draa, southeastern Rub' Al-Khali Desert, Oman (18 31 24 N 53 22 06 E). (b) Example of intricate threading of fluvial channels between migrating aeolian dunes and small disconnected dune fields in the Taklamakan Desert, China (38 22 42 N 81 53 46 E). (Image source: Google Earth Pro).

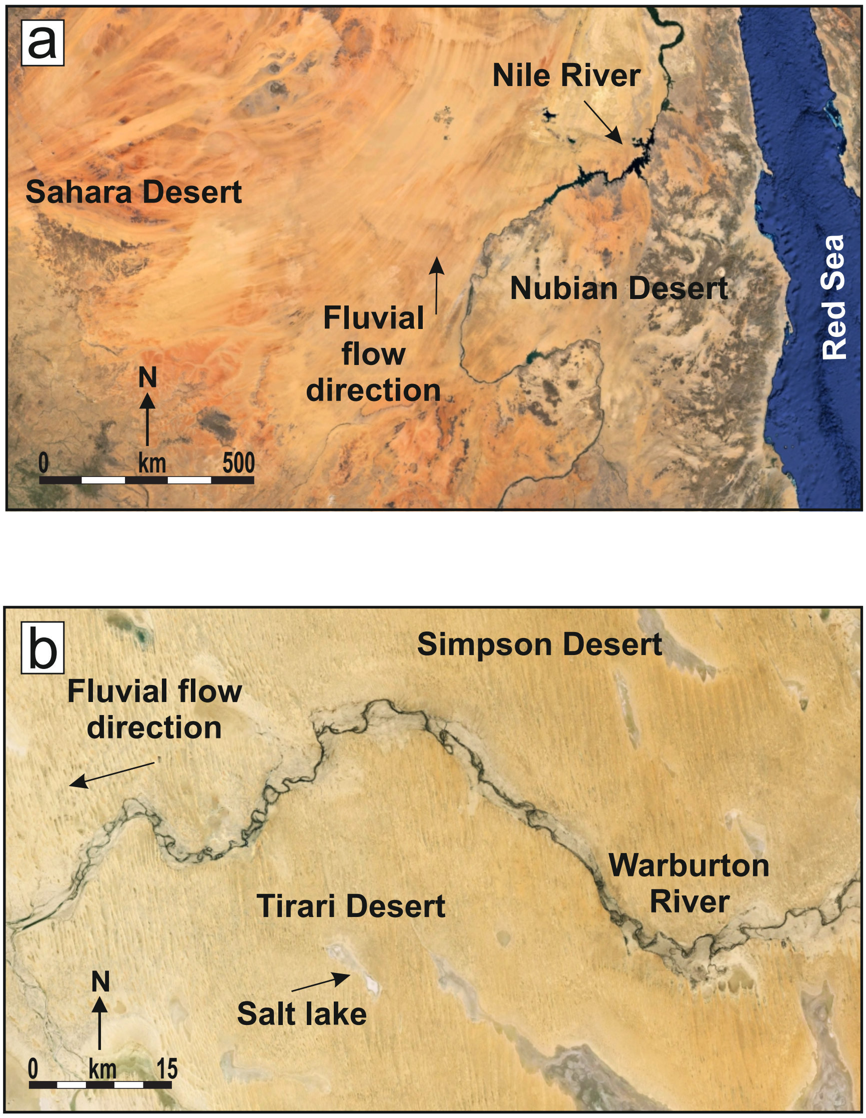


Figure 3.8: Examples of through-going fluvial channel networks that cross entire aeolian dune fields. (a) Eastern Sahara Desert (18 55 06 N 30 33 47 E); (b) Tirari Desert, Australia (27 49 13 S 137 37 34 E). (Image source: Google Earth Pro).

during each flood event. In some cases, this acts to transport sediment suitable for aeolian construction to parts of the dune-field further downstream. In cases where fluvial flooding along the fluvial flow pathway is frequent and regular, repeated flushing of sediment may severely limit the availability of sediment for aeolian construction to the part of the dune-field lying downwind of the river course (Figure 3.4). Alternatively, through-going fluvial systems may act to generate a localised supply of sediment for further aeolian construction, especially if they undergo a downstream reduction in flow competency. Where aeolian dunes are prevented from migrating across fluvial courses, the aeolian bedform character (size, morphological type, sediment composition) will be markedly different on the downwind side of fluvial course. The world's largest example is the 2000 km-long course of the Nile River through the eastern Sahara Desert (Figure 3.8a), which separates dune-fields of the Nubian Desert from those in the main Saharan sand seas. A second example is Warburton River, which separates the Simpson Desert from the Tirari Desert, Australia: average channel width is 182 m (Figure 3.8b). These and other representative examples are listed in Table 3.1.

The sedimentary record of these types of interactions is predictable. Aeolian sand transported into river courses will provide a source detritus that will typically be composed of well-sorted, fine sand suitable for fluvial transportation; fluvial deposits lying downstream from the dune-field will reflect this character. By contrast, aeolian deposits in areas downwind from the fluvial course may have a sediment composition that reflects the fluvial source.

3.4.5 Fluvial flooding of aeolian dune-fields associated with elevated water table level

In aeolian dune-fields where floods of relatively high magnitude and frequency occur, or where charge to subsurface aquifers is high due to either direct or indirect precipitation, interdune areas may be inundated by water not only during flood events. The local water table may remain permanently at or close to the accumulation surface such that low-lying interdune flats remain wet or damp between successive flood events (Nash, 2011). Thus, aeolian dunes may be surrounded for protracted episodes by wet (i.e. flooded) or damp interdunes (Figure 3.9). Such wet aeolian systems

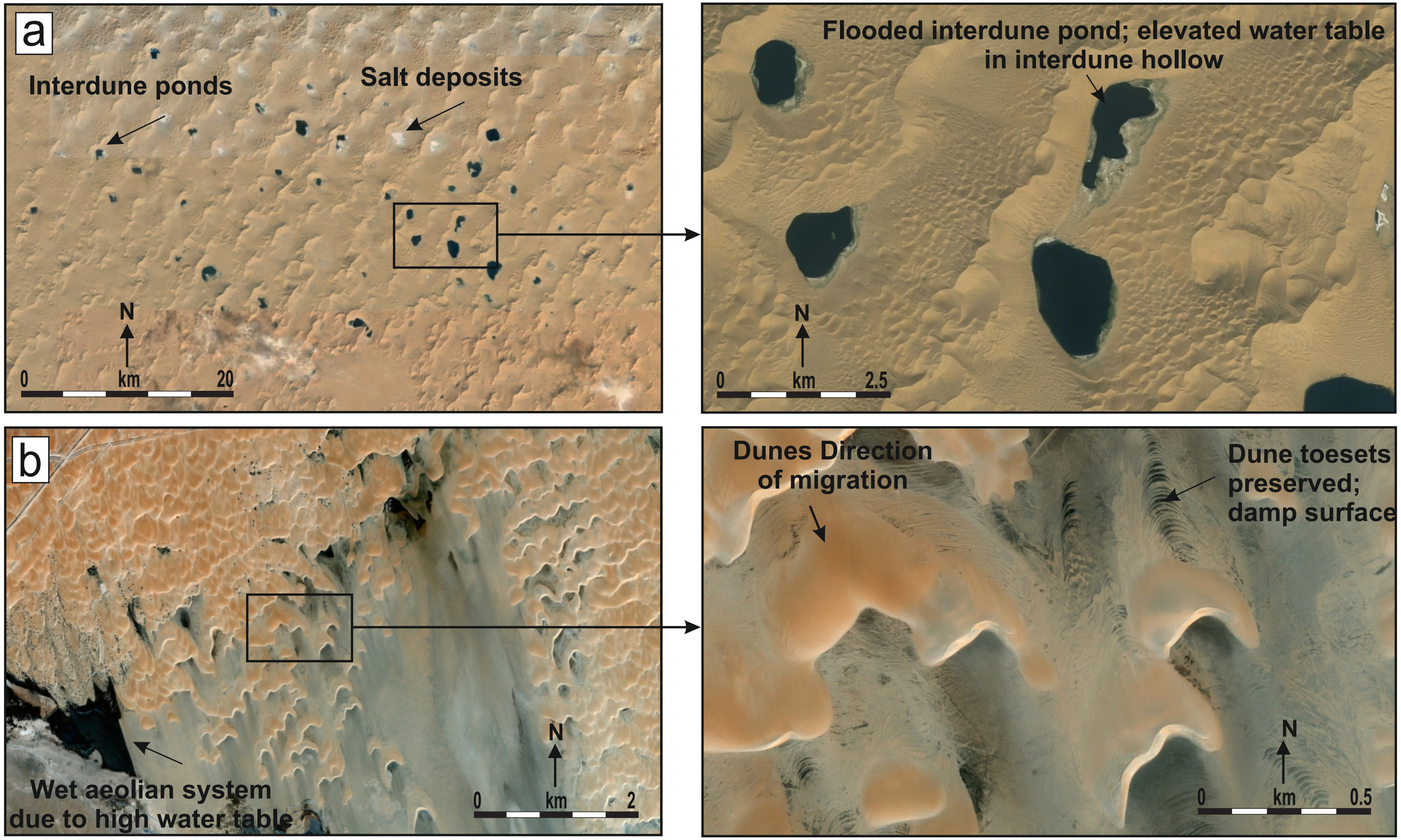


Figure 3.9: Examples of fluvial flooding of aeolian dune fields associated with elevated water-table level. (a) Gobi Desert, northern China (39 46 11 N 102 09 00 E); average interdune width is 1.13 km. (b) Al Jafurah Desert, eastern Saudi Arabia (25 47 17 N 49 48 28 E); progressive migration of barchan dunes across a damp, water table-controlled surface. Note how lee-slope strata of the lowermost flanks for the migrating barchans have been left as a record of the passage of the dunes. (Image source: Google Earth Pro).

(*sensu* Kocurek and Havholm, 1993) undergo aeolian construction and accumulation in a manner that differs from dry aeolian systems. Aeolian sediment transport across wet and damp sediment surfaces is severely restricted (Hotta et al., 1984; Good and Bryant, 1985; Crabaugh and Kocurek,1993; McKenna and Scott, 1998; Mountney and Russell, 2009), which limits the volume of sediment available for aeolian dune construction. Airflows in wet aeolian systems are therefore commonly under-saturated with respect to their potential sand transport capacity, rendering dry sand on existing aeolian dunes susceptible to erosion as the wind attempts to entrain more sediment. If direct precipitation in the dune-field acts to render dune surfaces damp for protracted periods, the effects of aeolian deflation may be limited. Rates of aeolian dune migration may be low or zero where flooded interdunes prevent bedform advancement. Fluctuations between relatively higher and lower water table levels can allow interdunes to change from a dry, to damp, to wet state on a seasonal basis and associated aeolian activity will reflect these changes. For example, the Lençóis Maranhenses dune-field, Brazil, is characterised by the presence of chains of barchanoid and transverse dunes separated by interdune lakes and lagoons that flood during the wet season (Parteli et al., 2006; Luna et al., 2012). Sauda Nethil Sabkha, Qatar (Ashour, 2013) and Chott Rharsa playa lake basin (Blum et al., 1998) are other similar examples. Other examples of wet aeolian systems in which interdune depressions are flooded in response to a high water table level include parts of the Gobi Desert of northern China (Figure 3.9a) and part of the Al Jafurah Desert, eastern Saudi Arabia (Figure 3.9b). In this latter example, a progressive rise in relative water table is enabling preservation of the toesets of aeolian dunes that pass over the damp surface. These and other representative examples are listed in Table 3.1.

Damp and wet interdune deposits typical of this type of interaction include adhesion structures (adhesion ripples, adhesion warts and adhesion plane beds), aqueous-ripple structures, wavy laminations, contorted structures and brecciated laminae (Kocurek, 1981; Kocurek and Fielder, 1982). Elevated water table levels promote aeolian accumulation and long-term preservation, especially in systems where aeolian dune-fields are constructed in subsiding sedimentary basins: slow but progressive basin subsidence will gradually cause the aeolian dune deposits to sink beneath a static but relatively high water table via a so-called relative water table rise (*sensu* Kocurek and Havholm, 1993), as is the case for the Skeiðarársandur dune-fields in southern Iceland (Mountney and Russell, 2009) and part of the Al Jafurah Desert, eastern Saudi Arabia (Figure 3.9b). An elevated water table also limits the effects of aeolian deflation (Fryberger et al., 1988).

3.4.6 Fluvial incursions into aeolian dune-fields associated with a single point source

The arrangement of landforms at the margins of desert sedimentary basins can act as a fundamental control on the nature of fluvial-aeolian interaction (Mountney, 2005). In many desert settings fluvial systems emanate from basin-bounding highland areas to pass as single-thread systems into the receiving desert basin in which aeolian dune-fields are developed, as is the case for wadis at the southern edge of the Rub' Al-Khali (Glennie, 1970). Thus, fluvial systems commonly intersect aeolian dune-fields at specific points along their margins. One common scenario is where an aeolian dune-field lies in front of a valley where a mountain stream emerges from its catchment. The confinement of the stream within a valley system, the short distance from the catchment to the aeolian dune-field, and the generally high gradient of the fluvial profile each act to reduce the opportunity for fluvial avulsion, thereby confining the river to a single point for a protracted period. Thus, the site of fluvial incursion of such single-thread fluvial systems into an aeolian dune-field remains fixed. Where such fluvial systems intersect the leading outer edge of an aeolian dune-field, their ability to penetrate the dune system will be dictated by factors such as the magnitude and frequency of the flood events, together with the orientation and continuity of dune ridges present at the dune-field margins. The areal extent over which dune-field flooding associated with single-thread fluvial channels operates tends to be limited, as is the case in examples from the White Sand Desert, New Mexico (Figure 3.10a). In cases where several single-thread channels enter into an aeolian dune-field, the lateral spacing of such fluvial courses dictates the types of fluvial-aeolian interaction, as is the case in the Grand Erg Occidental Desert, North Sahara Desert, Algeria (Figure 3.10b). These and other representative examples are listed in Table 3.1.

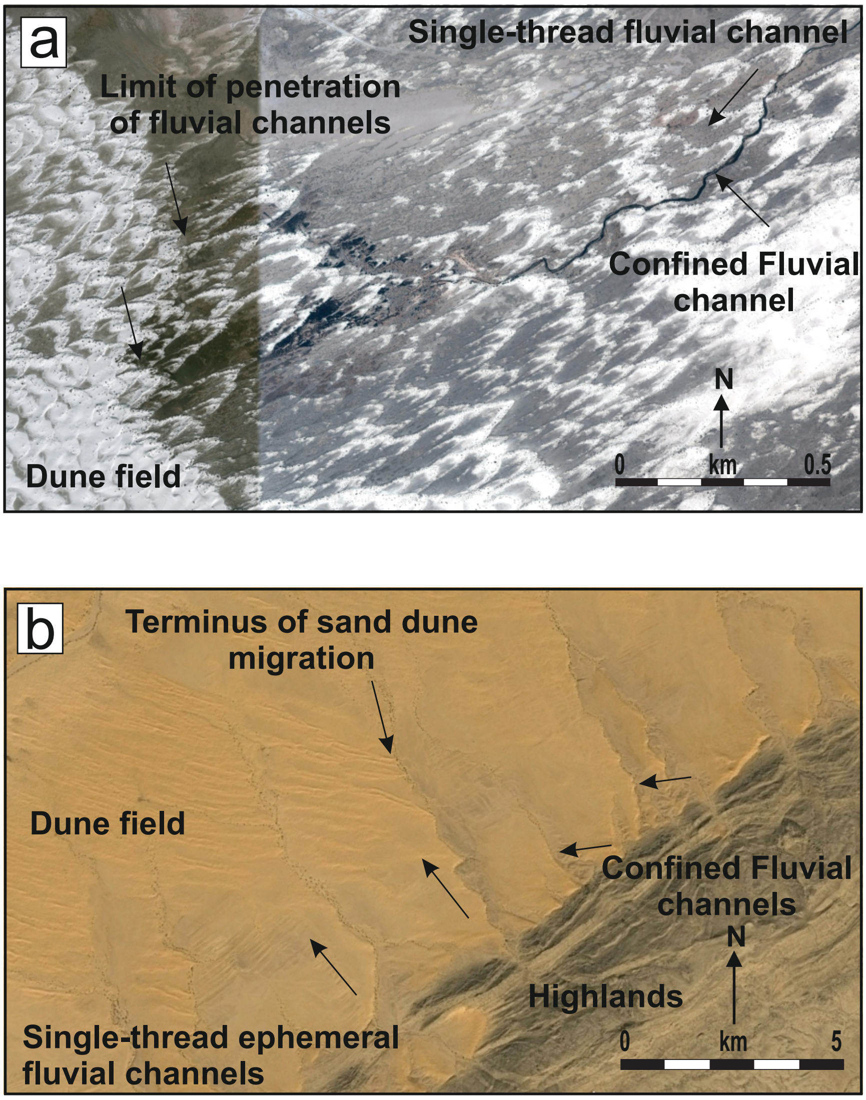


Figure 3.10: Examples of fluvial incursions into aeolian dune fields associated with a single point source. (a) White Sands, New Mexico, USA (32 51 54 N 106 12 11 W); (b) Grand Erg Occidental Desert, northern Sahara Desert, Algeria (32 30 19 N 00 08 39 W). The maximum extent of fluvial channel penetration into the dune field is 5 km. (Image source: Google Earth Pro).

The sedimentary expression of single-thread fluvial channels will be limited to the zone of penetration of the fluvial system into an aeolian dune-field, and this will tend to be present over a limited area in cases where the fluvial systems are fixed in position for protracted episodes. Consequently, the preserved sedimentary record may reveal limited lateral variations.

3.4.7 Fluvial incursions into aeolian dune-fields associated with a multiple sheet source

Alluvial fans commonly form extensive bajada where multiple catchments are present in close proximity along mountain fronts in arid settings (e.g., Padul Depression bajada, Spain, Calvache et al., 1997; bajada of northern Oman, Rodgers and Gunatilaka, 2002; Death Valley, Nevada, USA, Harvey, 2011). Similarly, distributive fluvial systems form networks of channels where they pass out onto low relief desert plains (cf. Hartley et al., 2010; Weissmann et al., 2011). Fluvial networks in such systems are commonly arranged into broad areas occupied by poorly-defined channels and are in some cases subject to non-confined flow over low-gradient surfaces (Hampton and Horton, 2007). Where such systems meet aeolian dune-field margins, they typically do so as sheet-like sources that may be active across distances of many tens of kilometres. Examples include part of the Sonoran Desert, northwestern Mexico (Figure 3.11a), and part of the Gobi Desert, northern China (Figure 3.11b). Aeolian-fluvial system interactions of this type occur over wide areas and multiple fluvial incursions may occur at many places along the dune-field margin. Non-confined sheet-like flood flows are typical, especially in the immediate aftermath of rainstorms. High-magnitude rainfall events, catchment area and relief, the low infiltration capacity of the substrate, the short run-off length from catchment to receiving basin the lack of appreciable relief on the basin plain, and the general absence of dense vegetation cover that might otherwise act to subdue run off, are all factors that contribute to sheet-like floods over large areas (Blair and McPherson, 1994; Blair, 1999; Arzani, 2005; Goudie, 2013). Such non-confined flows typically pass into dune-fields penecontemporaneously along multiple open interdune corridors with access gained from multiple points along the dune-field margin. Representative examples are listed in Table 3.1.

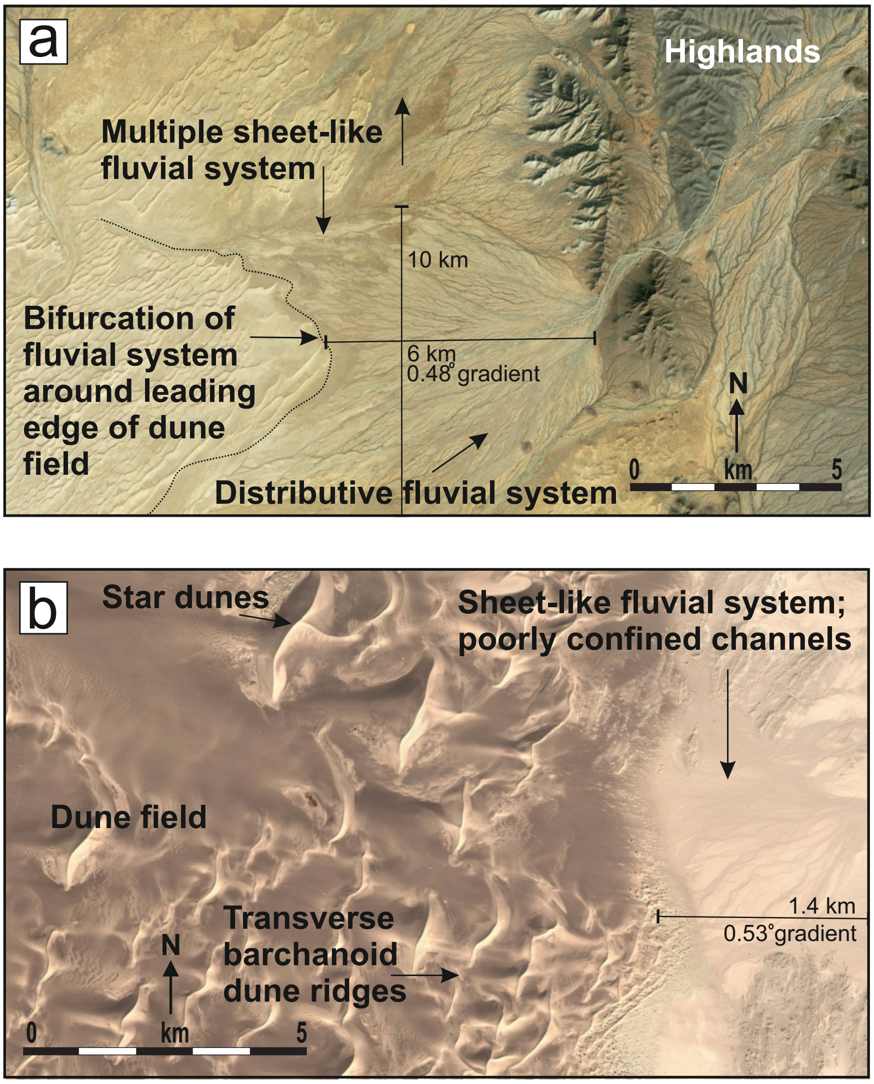


Figure 3.11: Examples of fluvial incursions into aeolian dune fields associated with a sheet source. (a) Sonoran Desert, northwestern Mexico (31 28 13 N 112 55 36 W); (b) Gobi Desert, north China (41 36 31 N 101 58 43E). Note the area of fluvial encroachment into the aeolian system. (Image source: Google Earth Pro).

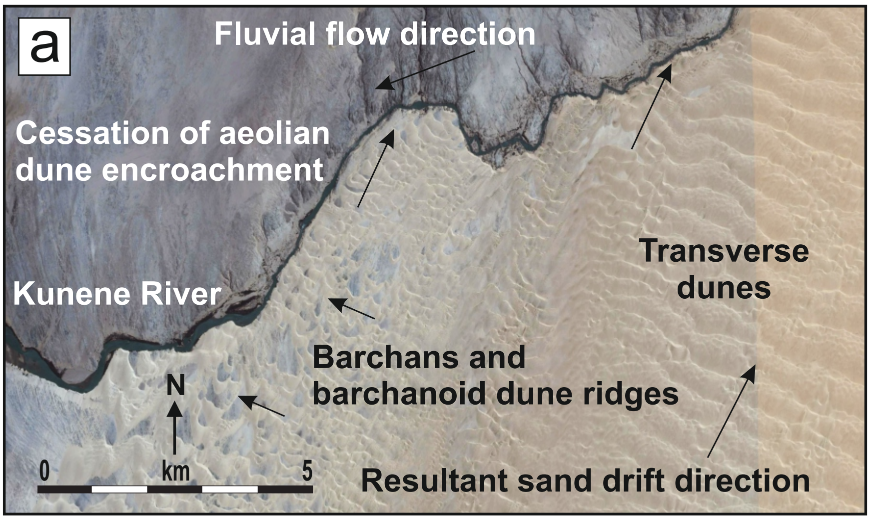
This type of aeolian-fluvial system interaction results in the widespread distribution of fluvial-derived sediment within dune-fields. Flooding over a wide spatial area means that the energy of the flow at any one location will be reduced. As such, the capacity of such flood events to erode aeolian bedforms will tend to be limited, except where non-confined flows locally coalesce into channels, for example where they are funnelled into narrow interdune corridors. Such flood deposits may serve to generate a localised supply of sediment for later aeolian dune construction.

3.4.8 Cessation of encroachment of aeolian dune-fields by fluvial systems

The downwind margins of several very large aeolian dune-fields are defined as spatially abrupt boundaries due to the presence of ephemeral or perennial fluvial systems that are effective in limiting the downwind encroachment of the dune-field. One large-scale example is the eastern boundary of the Sahara Desert, which terminates at the Nile River (Figure 3.8a). Even relatively small ephemeral fluvial systems may be effective in halting dune-field encroachment, as is the case for the Kuiseb River at the northern (downwind) margin of the Namib Sand Sea (Figure 3.6b). Other examples include the northern limit of the Skeleton Coast Dune-field, Namibia, which terminates at the Kunene River (Figure 3.12a), and the Mu Us Desert, northern China, which terminates at the Yellow River (Figure 3.12b). Flash floods passing down channel networks are commonly of sufficient magnitude to flush aeolian sand downstream, in some cases to a long-term sediment sink – the Atlantic Ocean in the case of the Kuiseb River that defines the northern margin of the Namib Sand Sea and the Kunene River that defines the limit of the Skeleton Coast Dune-field (both Namibia). These and other representative examples are listed in Table 3.1.

3.4.9 Termination of fluvial channel networks in aeolian dune-fields

Where fluvial systems terminate within the inner parts of aeolian dune-fields they do so in a variety of ways (e.g., Al Farraj and Harvey, 2004). A common type of fluvial termination is associated with a transformation from channelised to non-channelised flow, which tends to reduce flow competence, thereby expediting flow termination. Such conditions are



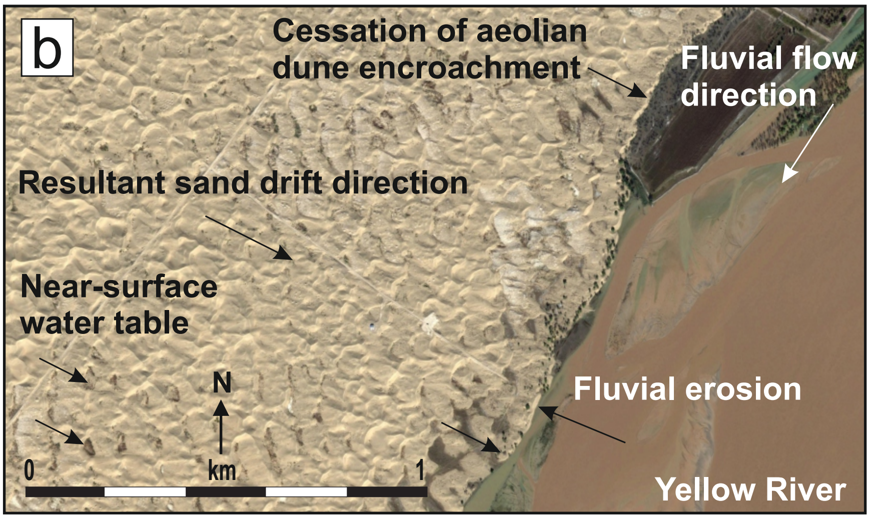


Figure 3.12: Examples of the cessation of encroachment of aeolian dune fields by fluvial systems. (a) Namib Desert, Namibia (17 15 29 S 11 49 17 E); (b) Mu Us Desert, northern China (40 04 26 N 106 44 06 E). Note the direction of the resultant aeolian sand drift direction. (Image source: Google Earth Pro).

common in ephemeral systems and may occur in any part of the aeolian dune-field depending on the energy of the flow. At the point of fluvial termination, suspended sediment comprising clay and fine silt sediment fractions are deposited (Reid and Frostick, 1987; Reid, 2002) to form mud layers in interdunes and playas. During dry seasons, aeolian sediment may to migrate over fluvial channels, thereby blocking the fluvial channel course and reducing the opportunity for future flood events to breach into the central parts of aeolian dune-fields during subsequent wet seasons (e.g., Mountney, 2006b). Examples include the Skeleton Coast Dune-field, Namibia (Figure 3.13a), the Simpson Desert, Australia (Figure 3.13b), and the Trarza Desert, Mauritania (Figure 3.13c). These and other representative examples are listed in Table 3.1.

3.4.10 Examples of short-term versus long-term fluvial-aeolian interaction

In modern dryland systems, there exist many examples of short-term aeolian-fluvial interaction (see Lancaster, 1995) whereby fluvial channels that are subject to ephemeral or intermittent flow that have been blocked by encroaching aeolian dunes or sand sheet deposits. Damming of fluvial courses typically occurs during the dry seasons or during drought episodes that are sufficiently long-lived to allow aeolian deposits to accumulate in fluvial channels (e.g., Figure 3.5; Glennie, 1970). One such example is where aeolian dunes have partially migrated across a playa lake basin at the terminus of an ephemeral river in part of the eastern Sahara Desert, Egypt (Figure 3.14a). Another example is in the Hamada Du Draa Desert, Algeria (Figure 3.14b). Episodic floods commonly act to flush out the system. Such fluvial flood deposits typically have a sedimentary character similar to that of the surrounding aeolian deposits, though grains are usually more tightly packed, producing lower primary porosities and permeabilities sandstones.

Over longer time scales, the impact of climate variation on depositional environments tends to be pronounced and significant, since it influences sediment yield, aeolian transport capacity of the wind, and the availability of sediment for aeolian transport. Together these factors govern the aeolian sediment state of the system (e.g., McKee et al., 1967; Herries, 1993; Kocuerk, 1999; Kocurek and Lancaster, 1999; Robinson et al., 2007). Short-

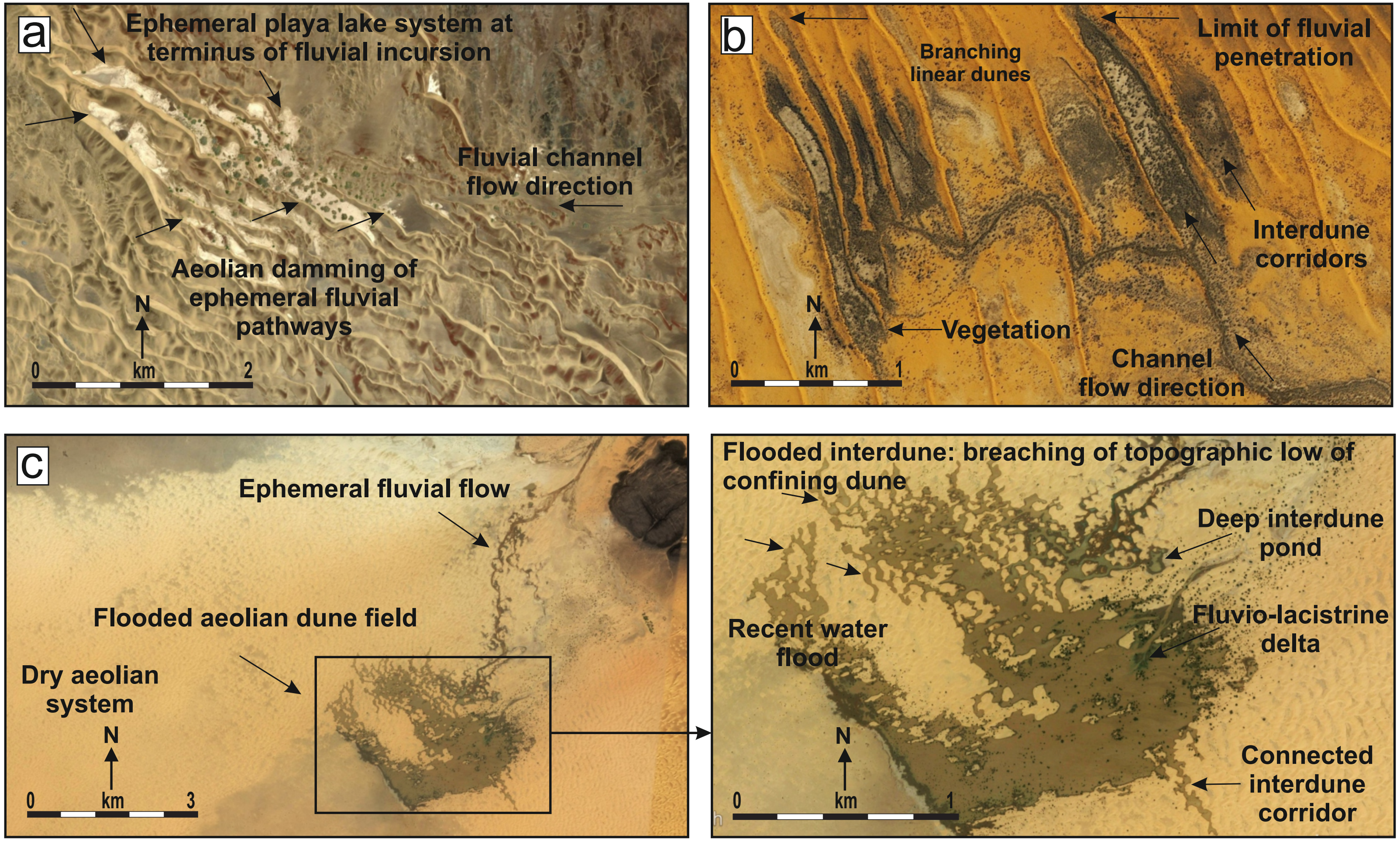


Figure 3.13: Examples of termination of fluvial channel networks in aeolian dune fields. (a) Skeleton Coast, Namibia (20 01 46 S 13 16 17 E); (b) Simpson Desert, Australia (24 10 29 S 135 15 53 E); (c) Trarza Desert, Mauritania (19 33 58 N 13 19 54 W) showing the recent flooding. (Image source: Google Earth Pro).

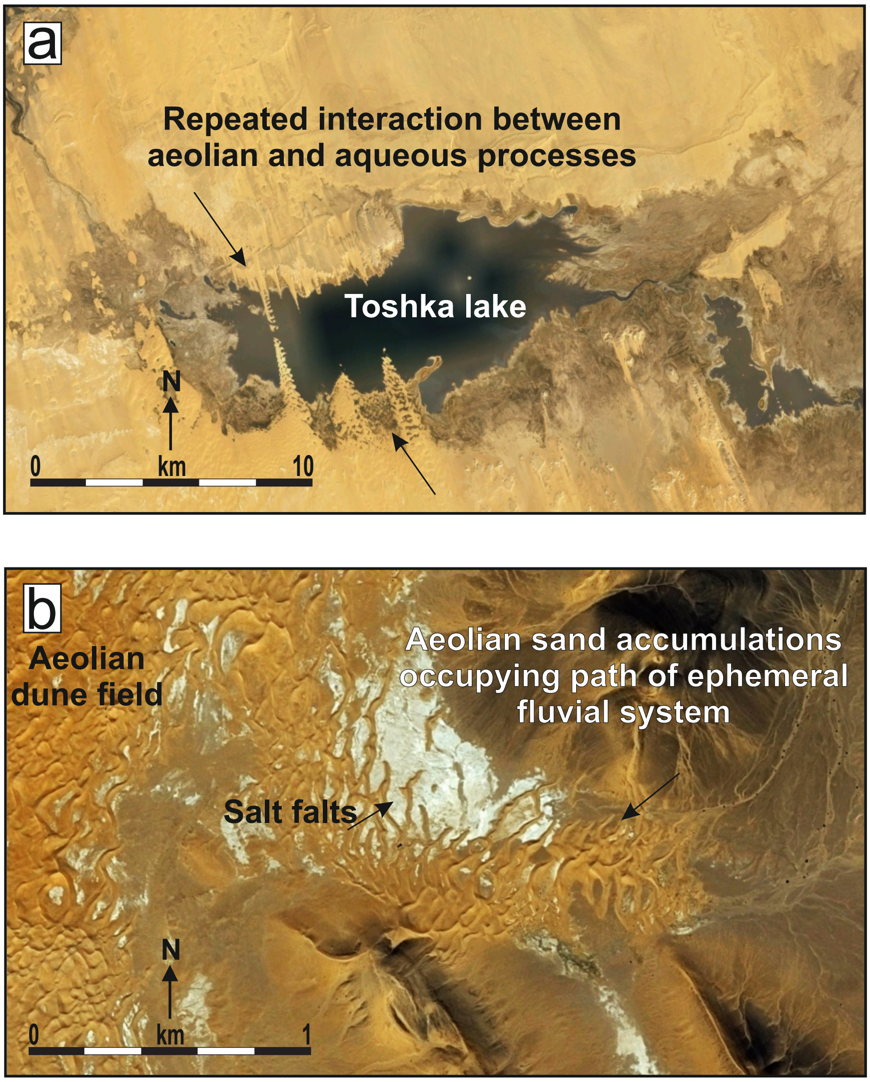


Figure 3.14: Examples of long-term versus short-term fluvial-aeolian interaction. In modern dryland systems many examples demonstrate how fluvial channels subject to ephemeral flow have been blocked by encroaching aeolian sediment. This usually occurs during the dry season or during drought episodes that are sufficiently long-lived to allow aeolian deposits to accumulate in fluvial channels. Episodic floods act to flush out the system and promote the development of vegetation at later stage. (a) Eastern Sahara Desert, Egypt (23 09 39 N 30 42 44 E); (b) Hamada Du Draa Desert, Algeria (28 58 03 N 4 02 14 W). (Image source: Google Earth Pro).

term or long-term shifts in the positions and form of the boundaries between aeolian and fluvial systems are controlled by the competition between fluvial flood events and sites of aeolian dune construction, which are subject to the external (allogenic) control of climate change (cf. Porter, 1986). During relatively more arid episodes, for example, accumulated sedimentary successions tend to be characterised by dry aeolian deposits such as dunes and sand sheets (Kocurek and Nielson, 1986; Basilici et al., 2009). During relatively more humid episodes, fluvial process tend to dominate, generating more heterogeneous successions (e.g., Stanistreet and Stollhofen, 2002). Representative examples are listed in Table 3.1.

3.5 Discussion

3.5.1 Geomorphic and sedimentary impact of fluvial-aeolian system interactions

Where externally sourced fluvial systems cannot reach the interior parts of dry aeolian systems because of the great density of aeolian dunes present and the closed nature of associated interdune depressions, the opportunity for aeolian sediment reworking via fluvial processes is limited. Minor fluvial streams may, however, develop in such settings in response to localised surface run-off associated with rainfall events that occur within the dune-field itself. Streams associated with intra dune-field flooding are highly ephemeral; reworking of aeolian sediment by such flows will be limited in extent and resultant deposits will be composed solely of fluvially reworked aeolian sand (Svendsen et al., 2003; Stromback et al., 2005).

Where externally sourced fluvial systems are able to penetrate into the interior of aeolian dune systems (Figures 3.15, and Figure 3.16), the principal morphological controls on the distance and type of fluvial incursion are as follows: (i) morphological dune type, which defines the length and continuity of individual dune segments; (ii) the orientation of dunes relative to the direction of fluvial flooding; (iii) the form of interdune corridors that are present between dune segments, which are defined in terms of their width and length, and spatial changes in these parameters that dictate whether such features are classed as open or closed morphological elements (Table

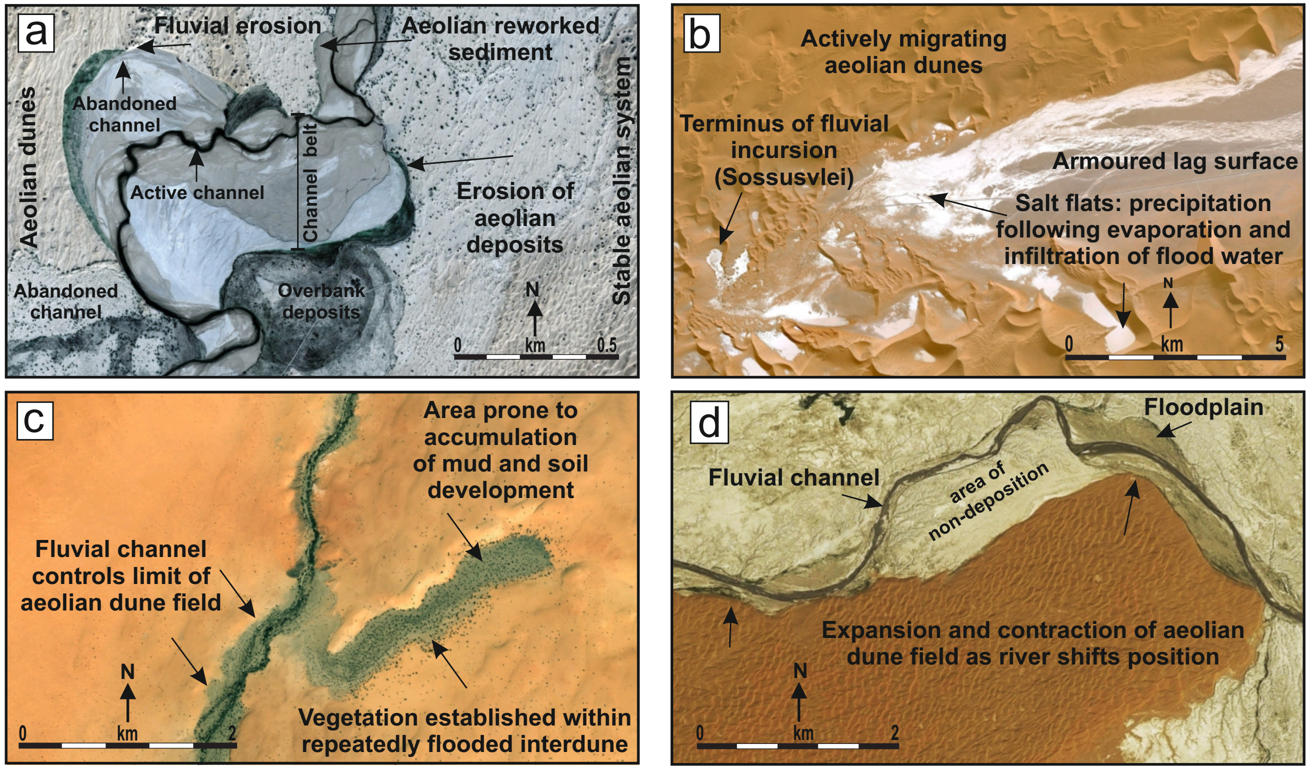


Figure 3.15: Examples of aeolian system expansion and contraction. (a) Taklamakan Desert, China. (37 46 00 N 81 27 30 E). (b) Namib Desert, Namibia (24 43 41 S 15 20 40 E); depicts various types of fluvial-aeolian system interaction and their geomorphic and sedimentary impact. Note the fluvial terminations within the dune fields, where large-scale dune bedforms have acted to pond flood waters and limit the extent of fluvial incursion. Playa deposits result in the generation of a significant surface crust of calcrete or gypcrete (white colour on the image) where flood waters have repeatedly ponded. (c) Southeastern Sahara, Sudan (15 39 11 N 26 25 44 E); shows vegetation development within a repeatedly flooded interdune and on the lower flanks of adjacent aeolian dunes; the presence of vegetation may act to partially stabilize the aeolian system. (d) Rigestan Desert, Afghanistan (31 22 26 N 65 53 19 E); demonstrates the role of fluvial flooding in controlling aeolian dune-field expansion and contraction. (Image source: Google Earth Pro).

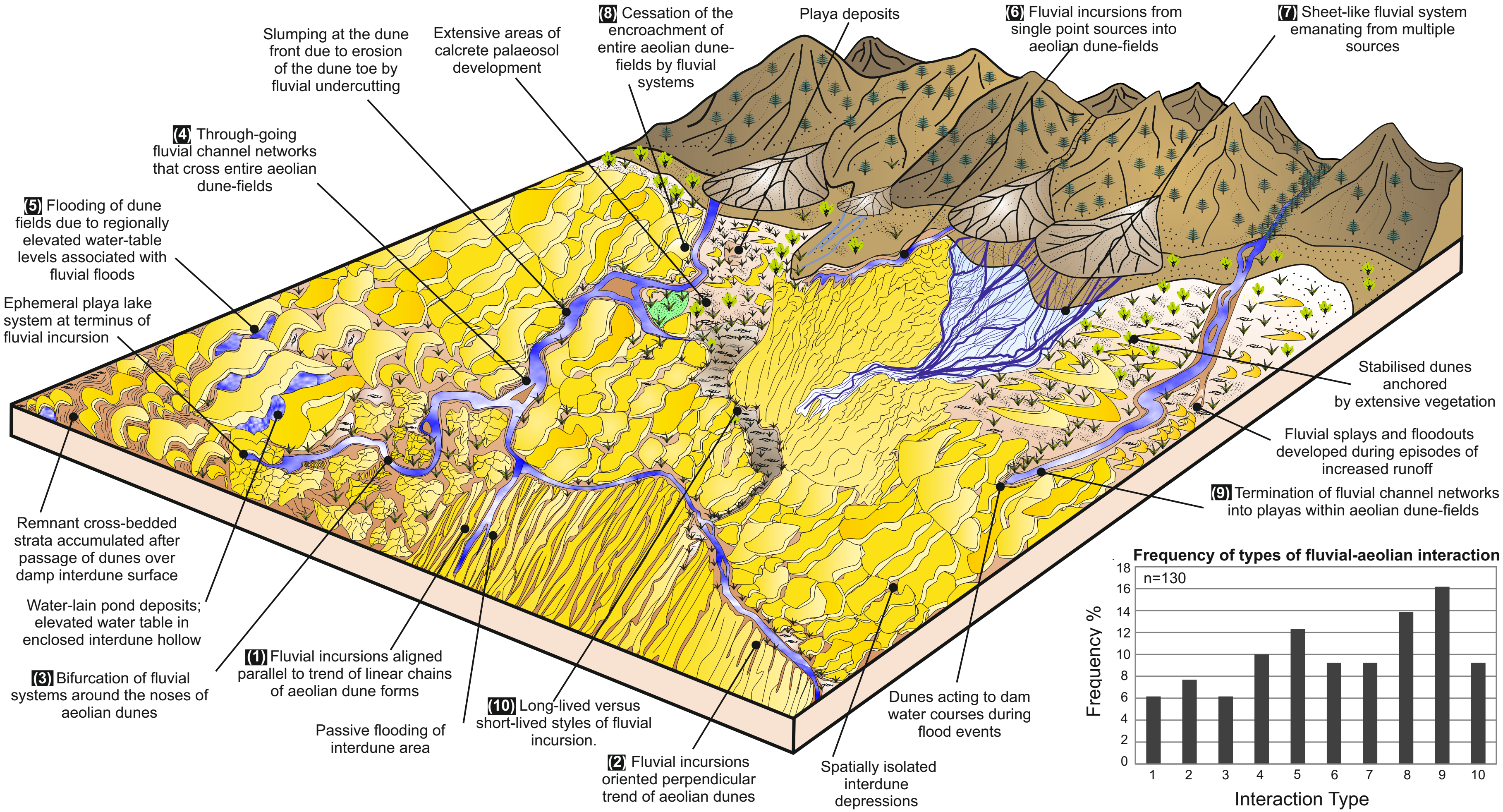


Figure 3.16: Schematic model summarising the classification of types of aeolian-fluvial system interaction. Numbers in black boxes relate to the ten types of fluvial-aeolian system interaction discussed in the text and listed in Table 1. The frequency of types of interaction for the 130 case studies listed in Table 1 is indicated.

1); (iv) the type and rate of aeolian dune and interdune migration relative to the frequency of fluvial flood events.

Accumulation and preservation of the sedimentary record of aeolian-fluvial interactions requires an appropriate mechanism to enable accumulation of both aeolian and fluvial deposits. One such mechanism is the gradual and progressive subsidence of the system within an evolving sedimentary basin (Blakey, 1988; Mountney et al., 1999). The nature of preserved types of interaction will be dictated in part by both the *spatial* arrangement of interdune corridors along which fluvial systems penetrate into aeolian dune-fields and the *temporal* change in the morphology of these interdune corridors (Mountney, 2012). Additionally, the nature of preserved types of interaction will also be dictated by the frequency and intensity of the flood events. The spatial extent of fluvial incursions may vary over time between successive floods as aeolian dunes and their intervening interdunes migrate, or as the intensity of successive flood events wax or wane in response to external controls such as climate change.

3.5.2 The role of fluvial flooding in controlling aeolian dune-field expansion and contraction

Although climatic aridity is a dominant factor that controls the distribution and extent of many sandy deserts, aeolian dune-fields are present not just in arid and semi-arid settings but also in a range of humid, non-climatic desert settings where sediment supply, sediment availability for transport, and the potential sediment transport capacity of the wind are sufficient to enable aeolian bedform construction. Climate exerts a fundamental control on the relative dominance of fluvial versus aeolian processes and plays a primary role in governing how aeolian dune-field margins expand or contract over time (e.g., Herries,1993; Clarke and Rendell, 1998; Yang and Li Ding, 2013).

Increases in either the frequency or magnitude of fluvial flood events in dune-field margin areas in response to climate change will impact continued aeolian dune-field construction in a number of ways. Increased fluvial discharge and stream power will promote erosion of older aeolian deposits. Fluvial reworking of aeolian sediment, its transport downstream and its ultimate re-deposition in areas where floods terminate will influence the supply and availability of sediment of a calibre suitable for later aeolian construction (Figure 3.15). Increased fluvial flood activity will limit the potential for aeolian dune migration (e.g., Pickup et al., 2002; Bullard and McTainsh, 2003). The availability of water provides conditions suitable for vegetation colonisation, thereby promoting stabilisation of interdune flats and limiting the capability of the wind to erode such substrates (e.g., Levin et al., 2009). Similarly, the deposition of mud drapes via settling from suspension over wide areas in the aftermath of repeated flood events will also limit the availability of underlying sediment for aeolian transport. Frequent floods will act to charge the ground water table beneath the aeolian dune-field, thereby raising the water table, possibly to the level whereby formerly dry interdunes become damp or wet (Figure 3.13, Figure 3.15, and Figure 3.16). An elevated water table tends to limit the availability of sediment for aeolian transport. However, it also increases the preservation potential of the aeolian bedforms that gradually subside beneath it (e.g., Mountney and Russell, 2009).

3.5.3 Controls on the form and spatial extent of fluvial incursion into aeolian dune-fields

The distance that fluvial systems are able to penetrate into dune-fields is partly dependent on bedform morphological type and spacing, which itself controls interdune width and shape (Figure 3.16). Further, the orientation of open interdune corridors relative to the angle of incidence of fluvial floods also plays a significant role, as does the rate of lateral migration of the dunes and their adjacent interdunes. The distance of penetration of fluvial incursion into the margins of aeolian dune-fields is greatest for regularly-spaced trains of relatively straight-crested aeolian dunes for which bedforms are separated by broad interdune flats and where fluvial systems impact the dune-field margin at an angle whereby flood waters associated with high-magnitude events can pass relatively unhindered along open interdune corridors.

Open interdune corridors play an important role where they occur adjacent to the path of fluvial systems passing into aeolian dune-fields (e.g., Hoanib River in Skeleton Coast, northwestern Namibia; Stanistreet and Stollhofen, 2002): they act as a catchment for excess water during flood events, thereby acting to buffer flood discharge (Figure 3.15b,c). In cases where interdune corridors terminate in closed depressions, they typically host ponded flood waters, the suspended-load deposits of which commonly form mudstone or salt layers that are relatively resistant to erosion due to their cohesive nature (Figure 3.15b; Loope et al., 1995; Bloomfield et al., 2006 McKie et al., 2010; Höyng et al., 2014). This has an important impact on sediment preservation potential. From an applied perspective, understanding the distribution of such layers in ancient preserved successions is important because they act as stratigraphic heterogeneities that restrict flow in water aquifers and hydrocarbon reservoirs, thereby compartmentalising subsurface bodies (e.g., Fryberger, 1990a; Mountney 2006a).

3.5.4 Controls on the accumulation and preservation of mixed aeolian and fluvial deposits

In modern desert dune-field settings, the relative dominance of aeolian versus fluvial activity is highly variable over a range of spatial and temporal scales, and this gives rise to complex arrangements of aeolian and fluvial morphological landforms and their deposits. In systems subject to infrequent or low-magnitude flood events, aeolian processes tend to dominate; conversely in systems subject to high-frequency, high- magnitude floods, fluvial processes dominate.

The frequency and persistence of fluvial flooding controls the period of occupancy of interdune corridors by active fluvial systems; in cases where aeolian dunes continue to migrate whilst flooding is on-going, the preserved architectural elements of fluvially-flooded interdunes tend to expand laterally as successive flood deposits develop in-front of advancing aeolian dunes. In non-climbing (i.e., non-accumulating) aeolian systems, such behaviour favours the development of sheet-like bypass supersurfaces (e.g. flood surfaces of Langford and Chan, 1988); in aeolian systems that climb at low angles (i.e., where a modest component of vertical accumulation is coincident with on-going aeolian dune and interdune migration), thin intercalations of vertically stacked, sheet-like fluvial and aeolian elements tend to accumulate (Mountney, 2012). The scale and connectivity of fluvial flood deposits tends to diminish with increasing distance toward the aeolian dune-field centre (Figures 3.1, and Figure 3.16), though exceptions occur where aeolian dunes act as natural dams, thereby encouraging floodwaters to pond creating temporarily lakes over large areas within more central parts of dune-fields. This type of interaction tends to be characterised by the accumulation of clay and silt deposits, and potentially of salt if the water salinity is high. The accumulation of such fine-grained or crystalline deposits is important from an applied perspective because elements composed of such material have the potential to form laterally extensive and continuous low-permeability baffles or barriers to flow in subsurface water aquifers and hydrocarbon reservoirs (e.g., Fryberger, 1990a; Bloomfield et al., 2006; Bongiolo and Scherer, 2010; McKie et al., 2010; Höyng et al., 2014; Romain and Mountney, 2014).

3.6 Conclusions

Fluvial and aeolian processes in desert-margin settings rarely operate independently: they are usually dynamically linked and exhibit a range of sedimentary interactions between fluvial and aeolian systems that are important and widespread in modern deserts. The diverse range of system interactions gives rise to considerable complexity in terms of geomorphology, sedimentology and preserved stratigraphy. Ten distinct types of fluvial-aeolian interaction are recognised (Figure 3.16, Table 3.1): fluvial incursions aligned parallel to trend of linear chains of aeolian dune forms; fluvial incursions oriented perpendicular to trend of aeolian dunes; bifurcation of fluvial systems around the noses of aeolian dunes; through-going fluvial channel networks that cross entire aeolian dune-fields; flooding of dune-fields due to regionally elevated water table levels associated with fluvial floods; fluvial incursions emanating from a single point source into dune-fields; incursions emanating from multiple sheet sources; cessation of the encroachment of entire aeolian dune-fields by fluvial systems; termination of fluvial channel networks in aeolian dune-fields; and long-lived versus short-lived types of fluvial incursion. These interaction types form the basis for a classification scheme that can be applied to desert dune-field systems generally.

The varied range of temporal and spatial scales over which aeolian-fluvial processes interact means that simple generalised models for the classification of types of interaction must be applied with caution when interpreting ancient preserved successions, especially those known only from the subsurface. By understanding the nature and surface expression of various types of aeolian and fluvial interaction, and by considering their resultant sedimentological expression, predictions can be made about how the preserved deposits of such interactions might be recognised in the ancient stratigraphic record and assessment can be made of the spatial scale over which such interactions are likely to occur.

Chapter 4
Outcrop architecture of ancient preserved aeolian and fluvial successions: Triassic Wilmslow Sandstone and Helsby Sandstone formations, Sherwood Sandstone Group, Cheshire Basin, UK

This chapter describes and interprets the various lithofacies present in sections of the upper part of the Wilmslow Sandstone Formation and the lower part of the overlying Helsby Sandstone Formation of the Triassic Sherwood Sandstone Group, Cheshire Basin, UK. Specifically, this study examines outcrops of the Runcorn Expressway road-cut, a laterally continuous outcrop that exposes the boundary between these two formations. The research aim is to document the preserved record of aeolian and fluvial successions, to further develop our understanding of processes that operate in aeolian and fluvial systems, and to propose a novel facies model for the mechanism of preservation of aeolian and fluvial deposits that accumulated in arid and semi-arid depositional settings.

4.1 Abstract

The Runcorn Expressway road-cut of northern Cheshire, England, provides an extensive section that exposes strata of both aeolian and fluvial origin. Aeolian and fluvial lithofacies, facies associations and architectural elements within this studied outcrop succession have been characterised in detail. Interdunes occur between dunes in most aeolian dune-fields, but the relationship of the preserved deposits of these low-lying and low-relief sub-environments to the deposits of adjacent dunes have not been adequately studied in the preserved sedimentary record, particularly in systems where the water table played a major role in governing sediment accumulation. The research objectives of this study are as follows: to document the preserved record of a wet aeolian system and an associated fluvial succession; to develop further our understanding of processes that operate in aeolian and

fluvial systems in arid and semi-arid depositional settings; to develop high-resolution, three-dimensional facies models with which to account for the type and mechanisms of preservation of fluvial and aeolian deposits in a manner whereby the resultant models have predictive potential; to investigate the relationship between aeolian dune and interdune morphology by relating primary depositional facies and associations of such lithofacies to specific processes of sediment transport and deposition; to investigate the relationship between preserved aeolian set thicknesses, grainflow thicknesses and original aeolian dune bedform size, to discuss the type of fluvial system responsible for generating the studied fluvial deposits.

This study reveals that accumulation of the aeolian system represented by the Triassic Wilmslow Sandstone Formation was controlled by water table: the formation is considered to largely be a so-called “wet aeolian system”. This study also indicates the importance of the nature of the strata that comprise interdune elements present between aeolian dune elements in determining the relationship of interdunes to adjacent dunes forms at the tie of accumulation, particularly in systems where water table played a significant role. The succession represented by the Wilmslow Sandstone Formation accumulated via both climbing and non-climbing mechanisms. The investigated fluvial facies and elements present in the overlying Helsby Sandstone Formation reflect the development of a dryland fluvial system in an ephemeral braided river setting.

4.2 Introduction

Studies of both modern aeolian systems and their ancient preserved successions have revealed a number of important conceptual advances in recent years regarding the mechanisms by which sedimentological and stratigraphic complexity is manifest in the accumulations of desert dune and interdune systems (e.g., Brookfield and Ahlbrandt, 1983; Loope, 1985; Fryberger, 1990a; Kocurek and Havholm, 1993; Lancaster, 1995; Howell and Mountney, 2001; Mountney and Jagger, 2004; Rubin and Carter 2006; Ewing, 2010, Al-Masrahy et al., 2012; Al-Masrahy and Mountney, 2013; Rodriguez-Lopez et al., 2014; White et al., 2015, Lancaster, et al., 2015).

Several referenced studies document the great diversity of fluvial styles within dryland settings (e.g, Mabbutt, 1977; Graf, 1988; Cook et al. 1993; Thornes, 1994a, 1994b; Miall, 1996, McCarthy and Ellery, 1998; Tooth and Nanson, 2000a, 2000b; 2004; Powell, 2009; Bourquin et al., 2009; Thomas, 2011). Dryland rivers typically have intermittent flows or ephemeral flows, many of which fail to reach the ocean, instead terminating within dryland settings such as low-relief alluvial plains, in playa basins or among aeolian dune-fields (Tooth and Nanson, 2011).

The construction, accumulation and preservation of aeolian desert systems does not require extreme aridity. Although aeolian system activity tends to be more extensive in arid and hyper-arid environmental settings (e.g., Glennie, 1970; Breed et al, 1979; Mountney and Howell, 2000; Scherer, 2001; Mountney, 2006a), desert aeolian systems may also be constructed and accumulate deposits in areas where the groundwater table is elevated such that the accumulation surface (*sensu* Kocurek and Havholm, 1993) may be damp or even wet (i.e. flooded). These are so-called wet aeolian systems (Kocurek, 1981a; Fryberger, 1990a, b, c; Kocurek and Havholm, 1993; Crabaugh and Kocurek, 1993; Mountney and Thompson, 2002). Such conditions are common at the margins of aeolian dune-fields and their preserved successions, where aeolian processes occur alongside synchronous fluvial processes, give rise to a range of types of aeolian-fluvial interaction (see Chapter Three, Section 3.4: types of aeolian-fluvial interactions, Figure 3.16). Most dryland regions support river systems (e.g., deserts of southeast Arabia, Glennie, 2005; the Rub' Al-Khali sand sea, Al-Masrahy and Mountney, 2013; (Chapter Two, Figure 2.1 and Chapter Three, Figure 3.2); Namib Desert, Stanistreet and Stollhofen, 2002; Al-Masrahy and Mountney, 2015; (Chapter Three, Figure 3.15b); Skeleton Coast, Krapf et al., 2003; Al-Masrahy and Mountney, 2015; (Chapter Three, Figure 3.13a). Dryland rivers play an important role in landscape-forming processes (Reid and Frostik, 2011). In some settings fluvial activities control aeolian processes and landforms through eroding parts of aeolian dunes, breaching dune barriers or providing sediment for aeolian systems (e.g., Nanson et al., 1995; Wainwright and Bracken, 2011; Tooth and Nanson, 2011). In other settings, aeolian activities influence fluvial channel pathways, for example by acting to pond fluvial flow (e.g., Lancaster and Teller, 1988, Al-Masrahy and Mountney, 2015).

Ephemeral and intermittent fluvial systems occur in geographic regions that are influenced by arid to semiarid climatic conditions and are common on both present-day active alluvial systems and successions preserved in the ancient rock record (Picard and High, 1973; Rust, 1981; Jones et al., 2005; Jones and Frostick,2008; McKie et al., 2010; McKie, 2011a; Banham and Mountney, 2013, 2014).

Desert margin settings are sensitive to climatic change (cf. Swezey et al., 1999). For example, during episodes of increased precipitation, ephemeral and intermittent fluvial systems tend to become more active and may penetrate further into aeolian dune-field margins. Fluvial rivers commonly serve as both sources and sinks of aeolian sediment (cf. Draut, 2012). Dryland rivers influence aeolian systems if they develop contemporaneously; they can play an important role in defining the type of aeolian system. Fluvial systems may act to charge the subsurface water table that lies beneath aeolian dune-fields. Where the groundwater table rises to a level where it influences the accumulation surface, it will dictate how aeolian sediments accumulate and become preserved, providing an opportunity for wet aeolian systems to develop (cf. Kocurek and Havholm, 1993; Blakey et al., 1996).

The presence of the competing (coeval) aeolian and fluvial systems in the same geographic vicinity will give rise to intercalated depositional settings (cf. Mountney and Jagger, 2004; Bourquin et al., 2009; Yan et al., 2015). Understanding the distribution of lateral and vertical arrangements of the ancient river system architectural elements in the preserved rock record is fundamental to the development of facies models with which to advance understanding of river behaviours and the factors controlling the gross-scale of architecture of fluvial systems (Bridge and Tye, 2000; Gibling, 2006; Colombera et al., 2012); the understanding of the river behaviours will guide the reconstruction of palaeo-drainage basins (e.g. Nichols and Hirst 1998; Jones, 2004).

The exposures selected for analysis in this study provide a valuable example of the preserved remnants of an ancient water-table influenced aeolian system – a so-called wet aeolian system (sensu Kocurek and Havholm, 1993). Interdune units preserved between aeolian dune units record the impact of the water table on aeolian system construction and development and sediment accumulation and preservation (Kocurek et al, 1992; Carr-Crabaugh and Kocurek, 1998). This locality also provides access to an outcropping succession that is interpreted to represent an example of a preserved dryland fluvial system.

The principal aim of this study is to document the preserved record of aeolian and fluvial successions, specifically, to gain an improved understanding of the mechanisms of accumulation and preservation of wet aeolian systems present at the marginal parts of aeolian dune-field (erg) systems where they interact with fluvial environments, and to further develop our understanding of processes that operate in aeolian and fluvial systems in arid and semi-arid depositional settings. Fulfilment of this aim will help gain an improved understanding of the palaeoenvironmental factors that act as primary controls on sedimentation in such settings. Specific research objectives are as follows: 1) to describe and interpret the sedimentary facies of both ancient dryland fluvial system and wet aeolian deposits that developed in aeolian erg margins present in an outcropping ancient succession; 2) to develop high-resolution, three-dimensional facies models with which to account for the style and mechanism of preservation of fluvial and aeolian deposits in a manner whereby the resultant models have predictive potential (cf. Howell and Mountney, 2001; Al-Masrahy and Mountney, 2015); 3) to develop a discussion that investigates mechanisms of wet aeolian system construction, accumulation and preservation based on analyses of facies relationships preserved in outcrop (cf. McKee and Moiola, 1975; Loope, 1985; Loope and Simpson, 1992; 1993; Kocurek and Crabaugh,1993; 4) to investigate the relationship between aeolian dune and interdune morphology which have not been so far adequately studied, by relating primary depositional facies and associations of such lithofacies to specific processes of sediment transport and deposition (cf. Kocurek and Dott, 1981; Kerr and Dott, 1988; Romain and Mountney, 2014); 5) to investigate the relationship between preserved aeolian set thicknesses, grainflow thicknesses and original aeolian dune bedform size; 6) to discuss the type of fluvial system responsible for generating the studied fluvial deposits.

4.3 Background

Many previous studies have investigated both modern aeolian depositional systems and their ancient preserved successions, and several have documented the range of sedimentary processes that are known to operate in water-table-influenced aeolian depositional settings (e.g., McKee, 1966; Thompson, 1970, a and b; Wilson, 1971; McKee and Moiola, 1975. Kocurek, 1981a; Fryberger et al, 1983; Simpson and Loope, 1985; Hummel and Kocurek, 1984; Lancaster and Teller, 1988; Glennie, 1990; Fryberger, 1990a, b and c; Fryberger et al. 1990; Øxnevad, 1991; Kocurek, et al, 1992; Loope and Simpson, 1992; Kocurek and Havholm, 1993; Kocurek and Crabaugh,1993; Meadows and Beach, 1993; Herries and Cowan, 1997; Mountney and Thompson, 2002; Granja et al., 2008; Mountney and Russell, 2009; Luna et al., 2012; Al-Masrahy and Mountney, 2015). Although these studies document the impact of water table on the accumulation and preservation of aeolian bedforms, understanding the interaction between aeolian bedforms and adjacent damp or wet interdunes requires additional detailed consideration. The origin and significance of resultant preserved facies relationships arising from damp/wet interdune and aeolian dune interaction in such systems requires further investigation. Better understanding of such settings will improve the interpretation of palaeoclimatic conditions in such setting and, more specifically, will enable the more precise interpretation of subsurface aeolian reservoir geometries. Prediction of reservoir facies and architectural-element variability in three-dimensions is a fundamental requirement for quantitative reservoir characterisation (e.g. Sweet, 1996; Liu et al., 2002; Fischer et al., 2007). Therefore, more accurate modelling of the lateral and vertical arrangements of damp and/or wet interdune architectural elements within aeolian reservoirs will aid the development of models describing the internal facies characterisation and mapping of lateral continuity of the producing zones (e.g. aeolian dune elements) and baffling zones (e.g. interdune elements) within reservoir bodies.

Accumulation and preservation of the deposits of migrating aeolian dune bedforms in wet aeolian systems requires a particular set of conditions: 1) a net rise per unit time in the water table, either via relative rise, where the water table remains static but the accumulating sediment gradually subsides through it, or via absolute rise, where the change of the water table is in response to climatic change, that may be gradual (i.e. progressive) or punctuated (i.e. episodic) (Kocurek and Havholm, 1993; Carr-Crabaugh and Kocurek, 1998; Mountney, 2012; Bristow and Mountney, 2013; Rodriguez-Lopez et al., 2014); 2) placement of accumulated aeolian sediment deposits beneath the baseline of erosion (Kocurek and Havholm, 1993). In some cases, accumulation and preservation of an aeolian system may arise through the inundation of the aeolian system during marine transgression, for example as is the case in aeolian systems adjacent to coastal environments, such as parts of the coastal fringe of the Namib Sand Sea. The proximity of aeolian systems to marine systems leads to inter-relationships among competing processes and depositional products (Eschner and Kocurek, 1986; Kocurek et al., 1992). In such settings, marine waters will directly influence accumulation in interdune hollows between dunes by charging water table level (Hunter, 1981; Hummel and Kocurek, 1984; Kocurek et al., 2001), thereby enhancing the preservation potential of the aeolian deposits.

The ratio between the rate of relative water-table rise and the rate of aeolian bedform migration is an important factor that governs the expansion or contraction of aeolian dune bedforms and adjacent interdunes areas over time; it also controls the angle of climb at which accumulating aeolian systems (dunes and adjoining intedunes) aggrade (cf. Rubin and Hunter, 1982; Kocurek and Havholm, 1993; Crabaugh and Kocurek, 1993; Kocurek , 1999; Mountney and Thompson, 2002, Mountney, 2012). In wet aeolian systems, for example, with a relative rise in the water-table level, both dune and interdune deposits accumulate, the rate of vertical accumulation can potentially match rate of the water-table rise. Where water-table rise acts to reduce the local availability of loose, dry sand suitable for aeolian dune construction, damp and wet interdunes will tend to expand at the expense of adjacent dunes (Kocurek and Havholm, 1993; Kocurek, 1996). If sediment supply increased with the rise of water table, interdune flats will have the tendency to maintain its thickness over large distance in climbing aeolian erg setting (Kocurek and Havholm, 1993; Mountney and Thompson, 2002).

Recognising evidence to determine the interaction between aeolian dune and interdune development and relative changes in relative water-table level in preserved examples of wet aeolian systems is not straightforward; this research seeks to address this issue.

4.4 Geological setting

The Cheshire Basin forms part of a major north-south trending rift system present throughout much of England and the surrounding region. Rifting was initiated in early Permian times (Glennie, 1995; Chadwick, 1997), and enabled the development of a series of rift basins in response to faulting and uplift at the end of the Late Carboniferous Variscan orogeny. This episode of extension was associated with an early phase of Atlantic opening (Griffiths et al., 2003; Ziegler and Dèzes, 2006).

The Cheshire basin itself forms a half-graben bounded to the east by the Wem-Red Rock fault system (Chadwick and Evans, 1995, Chadwick, 1997), whereas at its western margin the Permian and Triassic infill succession of the basin thins by depositional onlap onto Pre-Permian basement (Colter and Barr, 1975; Chadwick, 1997).

The Permo-Triassic fill of northern Cheshire Basin comprises a sequence of major units. The lowermost part of the fill of the Cheshire Basin (Kinnerton Sandstone Formation) comprise Rotliegend Group equivalent deposits that are likely of Permian age (Figure 4.1; Warrington et al., 1980; Griffiths et al., 2002). This is overlain by the Sherwood Sandstone Group (Permo-Triassic, Zechstein-Ladinian), which is in turn overlain by the Mercia mudstone Group (Triassic) (Warrington et al., 1980; Warrington and Ivimey-Cook, 1992). The Sherwood Sandstone Group comprises predominantly arenaceous units (the Chester Pebble Beds, Wilmslow Sandstone and Helsby Sandstone Formations), which are overlain by predominantly argillaceous deposits of the Mercia Mudstone Group. Elsewhere in the Cheshire region, though not in

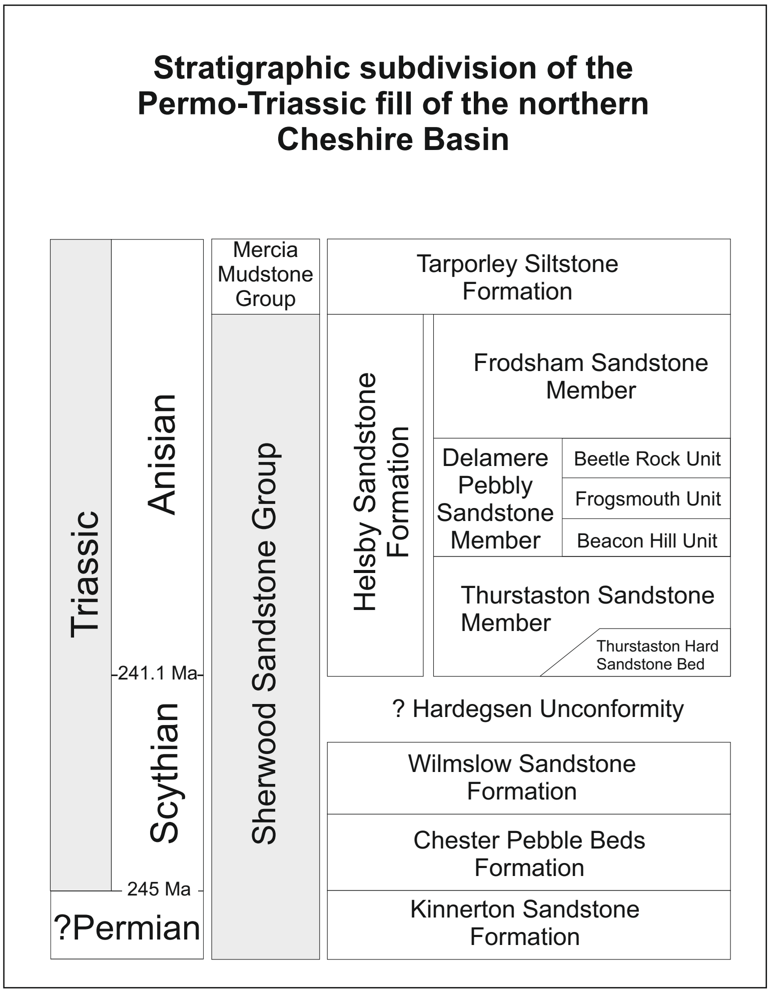


Figure 4.1: Stratigraphic subdivision of the Permo-Triassic fill of the northern Cheshire Basin. Mapped units within the Delamere Member of the Runcorn District are assigned names on an informal basis (Beacon Hill Unit, Frogsmouth Unit and Beetle Rock Unit). Based on discussion in Mountney and Thompson (2002). Ages from Harland et al. (1990).

the northern Cheshire Basin, deposits of the Merica Mudstone Group are overlain by deposits of the Penarth Group.

Regionally, the Sherwood Sandstone Group comprises much of the fill of the series of genetically related and partially interlinked rift basins that run from the Wessex Basin in the south of England, northwards through the Worcester Graben, the Knowle Basin, the Stafford and Needwood basins, and into the Cheshire Basin; the rift system continues further northwards into the Lancashire and West Cumbria basins, and north-northwest-wards via the Deemster sub-basin into the East Irish Sea Basin (Chadwick and Evans, 1995). These basins had varying maximum burial depths, ranging from less than one kilometre to more than three kilometres deep (Burley, 1984; Griffiths et al. 2002).

During permo-Triassic times northwest Europe occupied a latitudinal position similar of the Sahara deserts today (Glennie, 1983). Within these basins the Sherwood Sandstone Group accumulated a series of mixed fluvial and aeolian successions. Generally, deposits of the group comprise the following broad facies associations: 1) fluvial channel conglomerate and sandstone deposits and overbank sand-, silt- and mudstone deposits; 2) aeolian sandstone and interdune sandstone, siltstone and mudstone deposits.

Part of the Permo-Triassic Sherwood Sandstone Group is well exposed in the Runcorn Expressway road-cut, an outcrop that is 13 m high and that extends laterally for 230 m along a slipway of the A557. This outcrop reveals well-exposed examples of both an aeolian and a fluvial succession within the upper part of the Wilmslow Sandstone Formation (of aeolian dune and interdune origin) and the lower part of the overlying Helsby Sandstone Formation (of fluvial origin), respectively (Figure 4.1, and Figure 4.2).

Several detailed previous studies have sought to document the preserved stratigraphic architecture of the Sherwood Sandstone Group in the Cheshire Basin generally, and in the Wilmslow and Helsby formations, in particular. Such studies have attempted to establish lateral and vertical trends in lithofacies and architectural-element distribution in these rock units (e.g., Strahan, 1882; Thompson, 1969; 1970a; and 1970b; Øxnevad, 1991; Benton et al., 1994; Mountney and Thompson 2002). However, no detailed

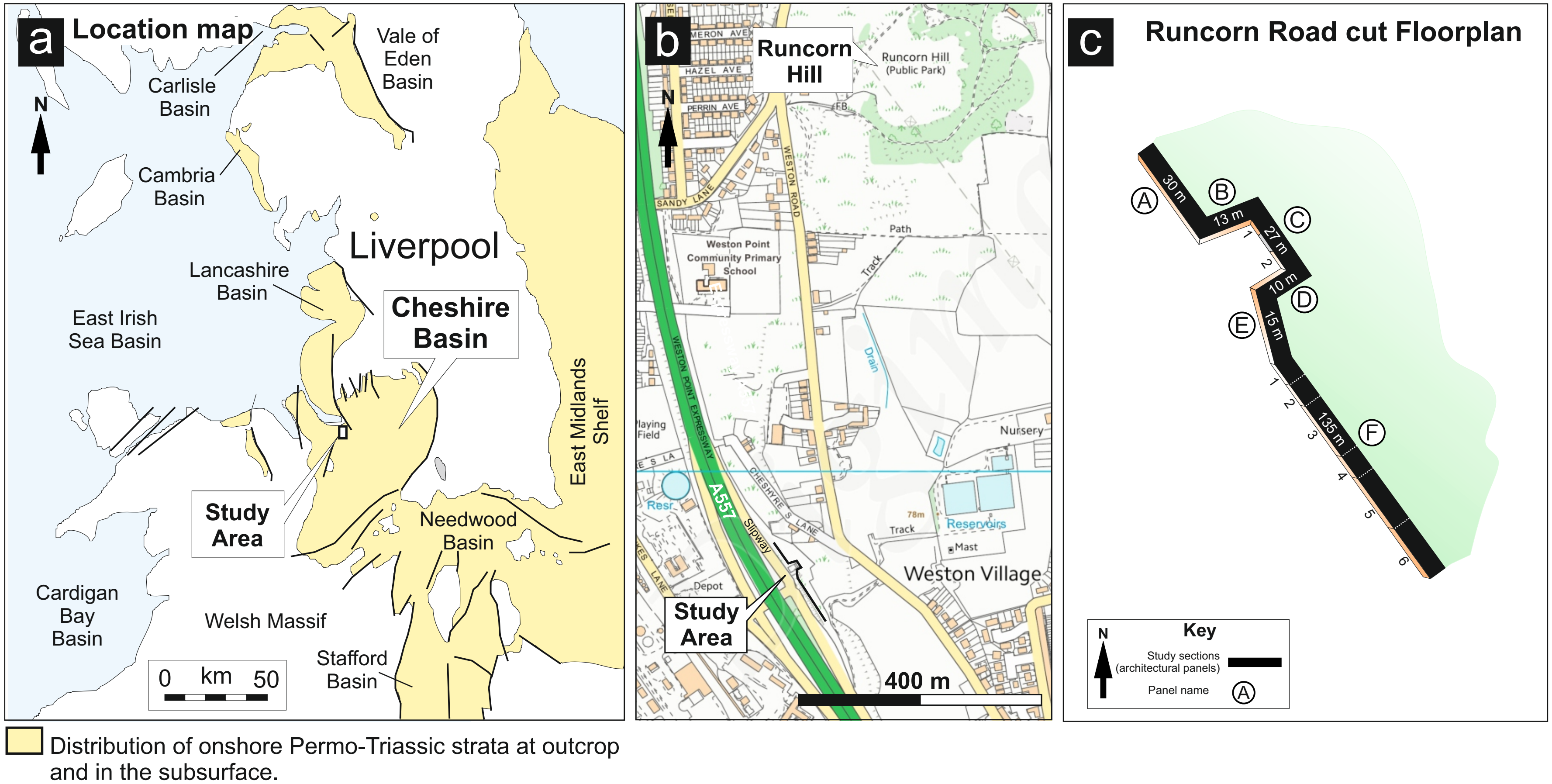


Figure 4.2: (a) Onshore outcrop of Permo-Triassic rocks in central and northern England, Wales and southern Scotland, showing the location of the study area. (b) Location of the study area (c) Plan of Runcorn Expressway road cutting outcrop with individual study panels labelled.

account has been published previously on the stratigraphic architecture of the Runcorn Expressway road-cut locality.

The Wilmslow Sandstone Formation in the Cheshire Basin is interpreted as being of mixed fluvial and aeolian in origin, and records deposition under the influence of generally arid climatic conditions (Thompson, 1970a). The Helsby Sandstone Formation forms the uppermost unit of the Sherwood Sandstone Group in the Cheshire Basin; it occurs above what is marked locally in northern Cheshire by the Hardegsen unconformity (intra-Triassic unconformity, late Scythian age), that likely resulted from a syn-extensional regional uplift accompanying lithospheric thinning during mid-Triassic times (Warrington, 1970; Evans et al., 1993). The contact between the underlying Wilmslow Sandstone Formation and the overlying Helsby Sandstone Formation in the study area is represented by this Hardegsen Unconformity (Warrington, 1970; Evans et al., 1993).

In the northern part of the Cheshire Basin, in the area around Helsby, Frodsham, Runcorn and north-westwards to Wirral, the Helsby Sandstone Formation can be divided into three members based on lithofacies: the Thurstaston Member (dominantly aeolian, though locally fluvial in places), the Delamere Member (dominantly fluvial) and the Frodsham Member (dominantly aeolian) (Thompson, 1969, 1970b; Warrington et al., 1980; Mountney and Thompson, 2002).

This study documents the stratigraphic architecture of the Runcorn Expressway road-cut, which yields an extensive section that exposes strata of aeolian origin within the upper part of the Wilmslow Sandstone Formation and fluvial origin within lower part of the overlying Helsby Sandstone Formation–equivalent to a unit known on Wirral as the Thurstaston Hard Sandstone Bed of the Thurstaston Member (Thompson, 1970b, Howard et al., 2007). Stratigraphically, the studied Expressway road-cut section lies directly beneath the Delamere Member of Helsby Sandstone Formation, which is exposed in several quarries on Runcorn Hill (Figure 4.2b), and which was the focus of the study by Mountney and Thompson (2002).

4.5 Data and methods

Analysis of aeolian and fluvial lithofacies, facies associations and architectural elements within the studied outcrop succession have been defined; the attributes of the aeolian and fluvial deposits (predominantly sandstones) have diverse characteristics, which have been the subject of detailed facies analysis.

The field-based lithofacies analysis has been used to describe the detailed characteristics of fourteen distinct lithofacies recognised within the deposits of Wilmslow Sandstone and Helsby Sandstone formations exposed in an outcrop located adjacent to the A557 Runcorn Expressway road, northwest England (Figure 4.1). The fourteen lithofacies are defined based on their lithology, sedimentary texture and the range and type of sedimentary structures present within. Six separate log sections measured from the study area characterise a total of 42 metres of sedimentary succession. Recorded information includes lithology, sediment texture and sedimentary structures. Collectively, these graphic logs provide a generalised complete vertical profile for the studied outcrop; the general log section covers a vertical section of 13 metres (the full thickness of the outcrop) (Figure 4.3). Six, two-dimensional architectural panels depict the sedimentary architectural relationships present in the entire of the outcrop, which extends laterally for 230 metres. Panels have been generated from detailed field sketches and photomosaics to determine the distribution and relationship of various aeolian and fluvial architectural elements present in the outcrop. The panels depict the stratigraphic architecture of the upper part of the Wilmslow Sandstone Formation, which primarily comprises aeolian deposits, and the lower part of the overlying Helsby Sandstone Formation, which primarily comprises fluvial deposits in the studied section. Panels were arranged to form a composite correlation panel that have been manipulated digitally to generate a pseudo-three-dimensional view of stratigraphic architecture. This has enabled the tracing of the individual sets in three-dimensions (Figure 4.4). Two-hundred readings of palaeocurrent data (comprising cross-bedding foreset dip magnitude and azimuth data) were collected from both the aeolian and fluvial deposits. Summaries of these data are presented as rose diagrams for each section (Figure 4.3). Two aeolian and five fluvial

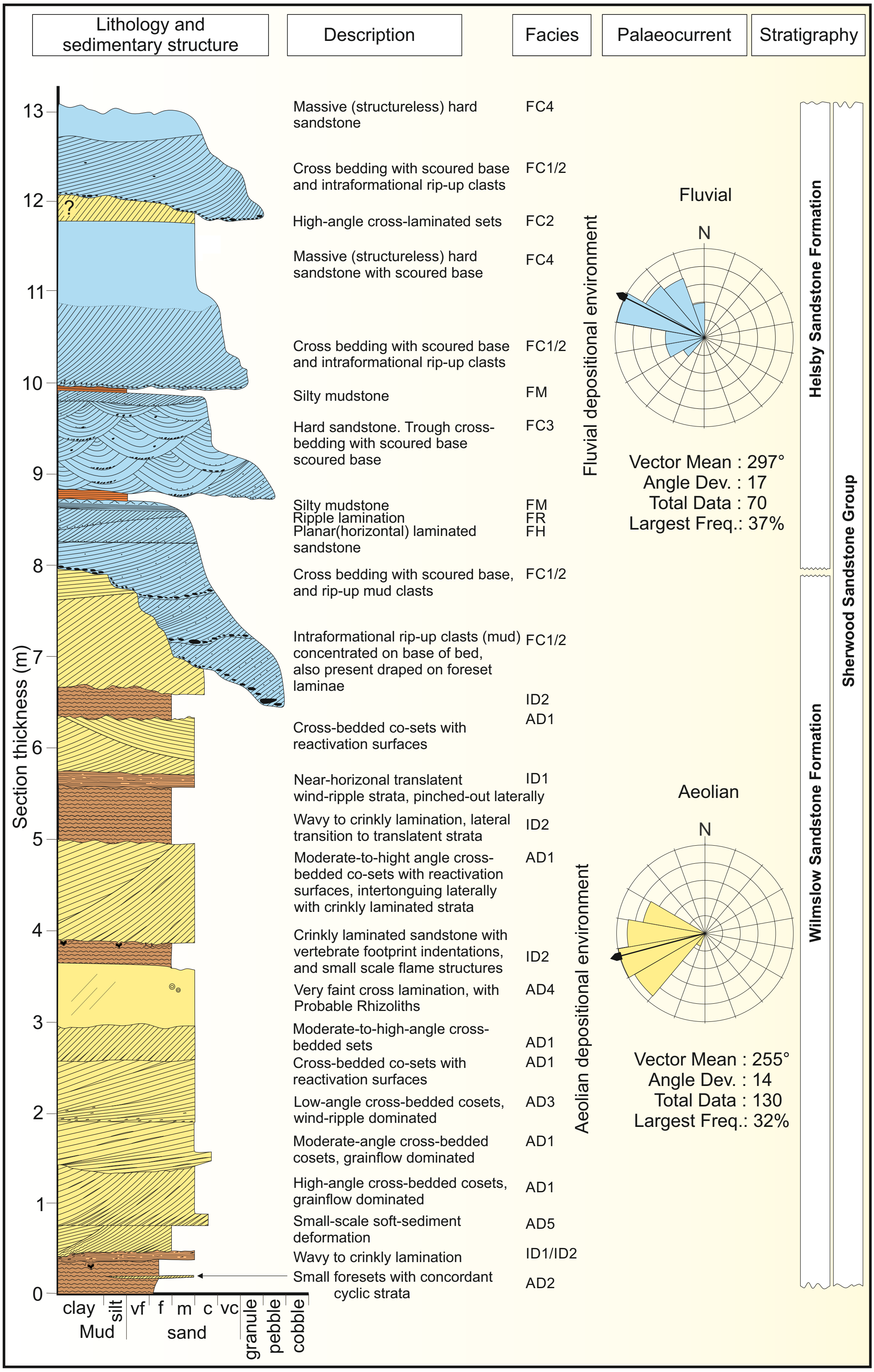


Figure 4.3: Summary log section for the study interval at Runcorn Expressway road cut.

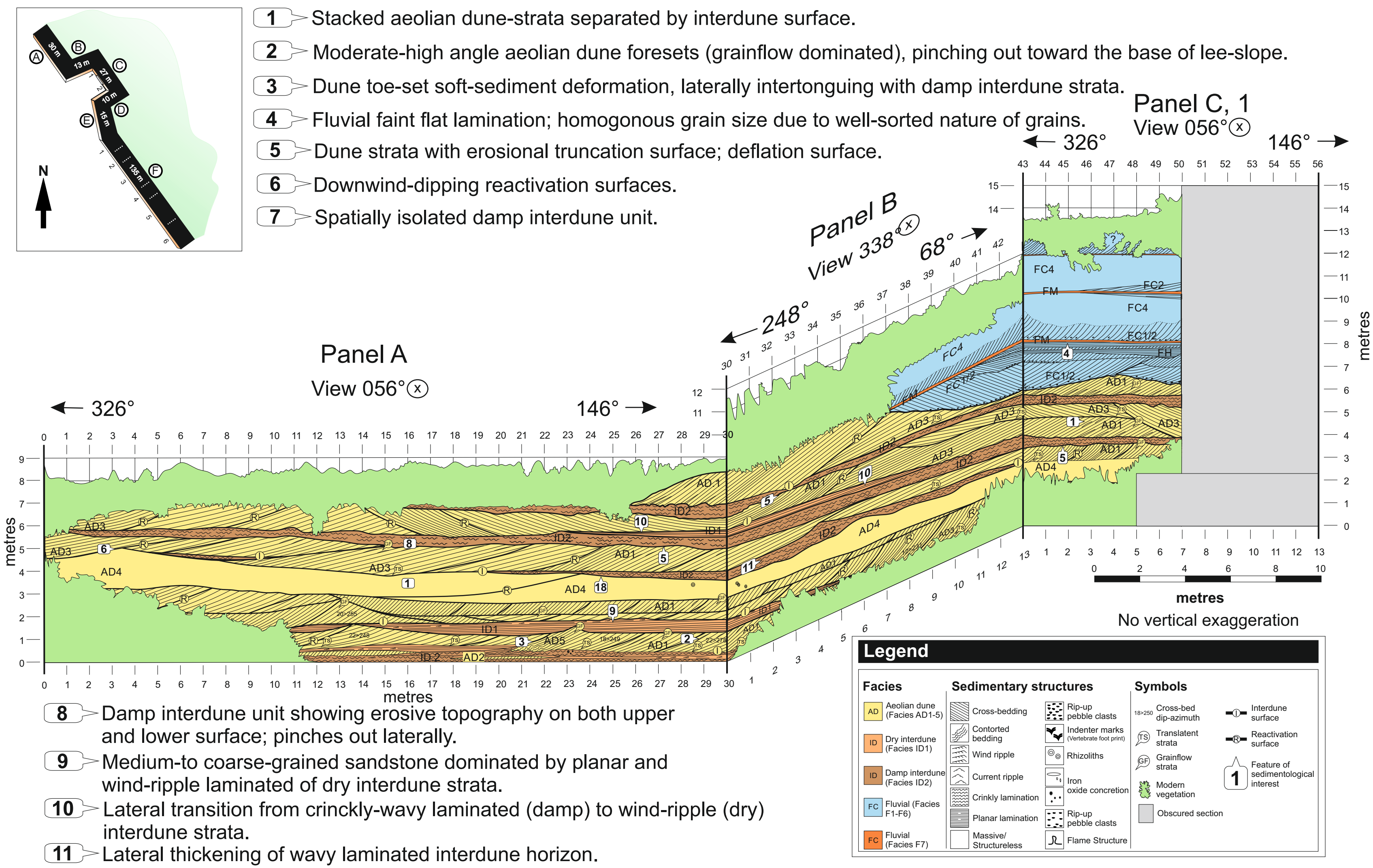
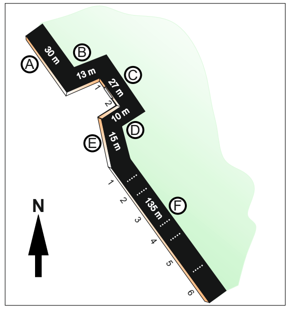


Figure 4.4: Composite architectural panels depicting the stratigraphic architecture of the Wilmslow Sandstone and Helsby Sandstone Formations as observed in Runcorn Expressway road cutting outcrop.



12 Low-angle inclined aeolian dune foresets; wind-ripple dominated (translatent wind-ripple strata common).

13 Near horizontal, bi-modal grain size (pinstripe lamination) forming clear grain size segregation unit; pinches out laterally and merges with interdune facies.

14 Stacked aeolian dune foreset deposits that lack clear stratification due to grain size homogeneity.

15 Down-folded lamination in the uppermost layers of the damp interdune represent cross-sectional view of vertebrate footprint structure.

16 Fluvial channel base; erosional bounding surface.

17 Intraformational rip-up clasts composed of red (laminated silty mudstone); highly variable size and shape. Basal lag deposits associated with channel-fill element.

18 Lense-shaped interdune element.

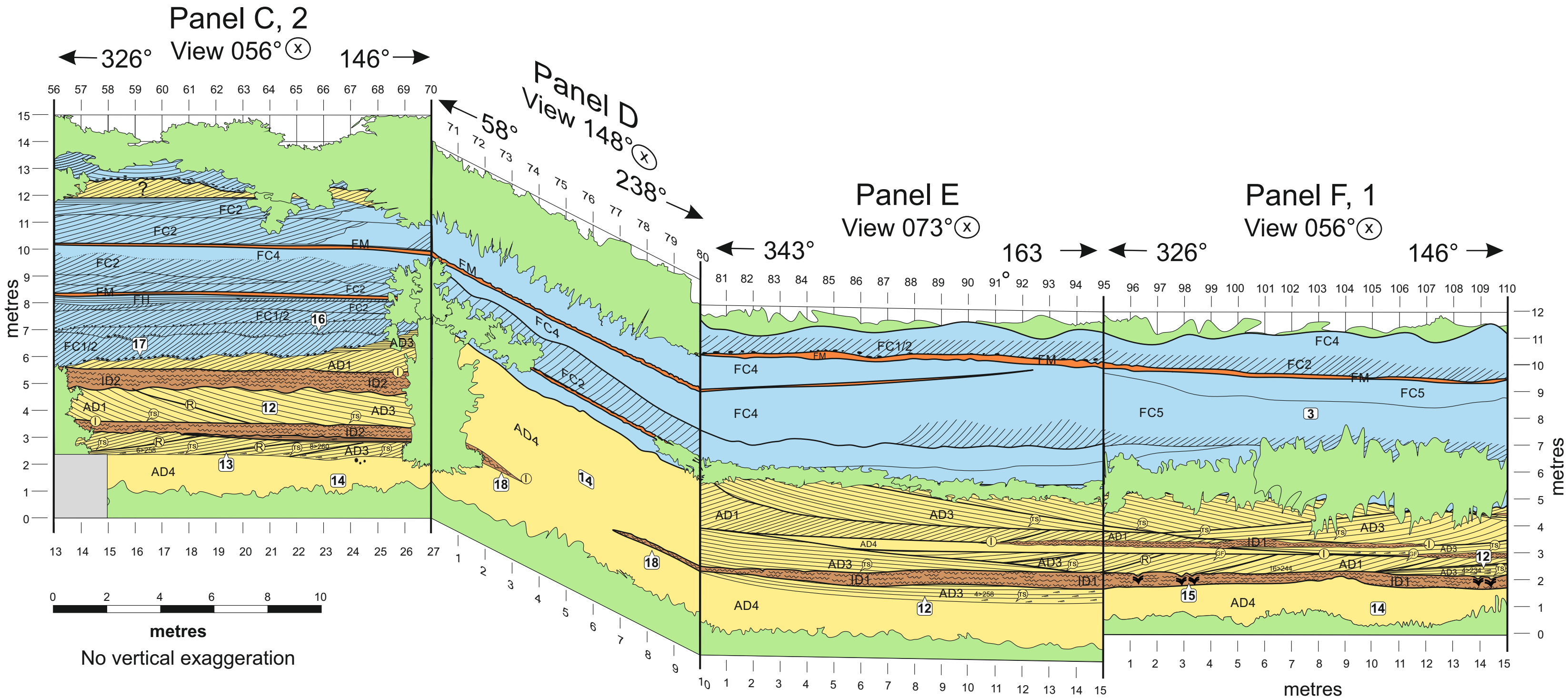
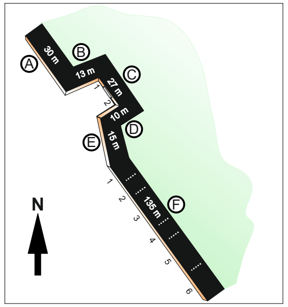


Figure 4.4: Cont.



19 Wavy to crinkly laminated sandstone, dominated by modified wind-ripple strata and adhesion-ripple strata; accumulated in interdune areas between merging aeolian dune forms.

20 Planar thinly (10-30 cm) laminated Silty mudstone.

21 Fluvial flood surface with erosive base exhibiting local topography and planar laminated silty mudstone fill.

22 Planar-tabular sets representing the migration of sandy bedforms within a fluvial channel system.

23 Sharp boundaries at the top of the channel-fill element record an abrupt abandonment phase.

24 Small-scale water escape (flame) structures in upper part of damp interdune units.

25 Fluvial trough cross-beds, representing the migration and accumulation of sandy mesoform within a fluvial channel system.

Panel F, 2-4

View 056°

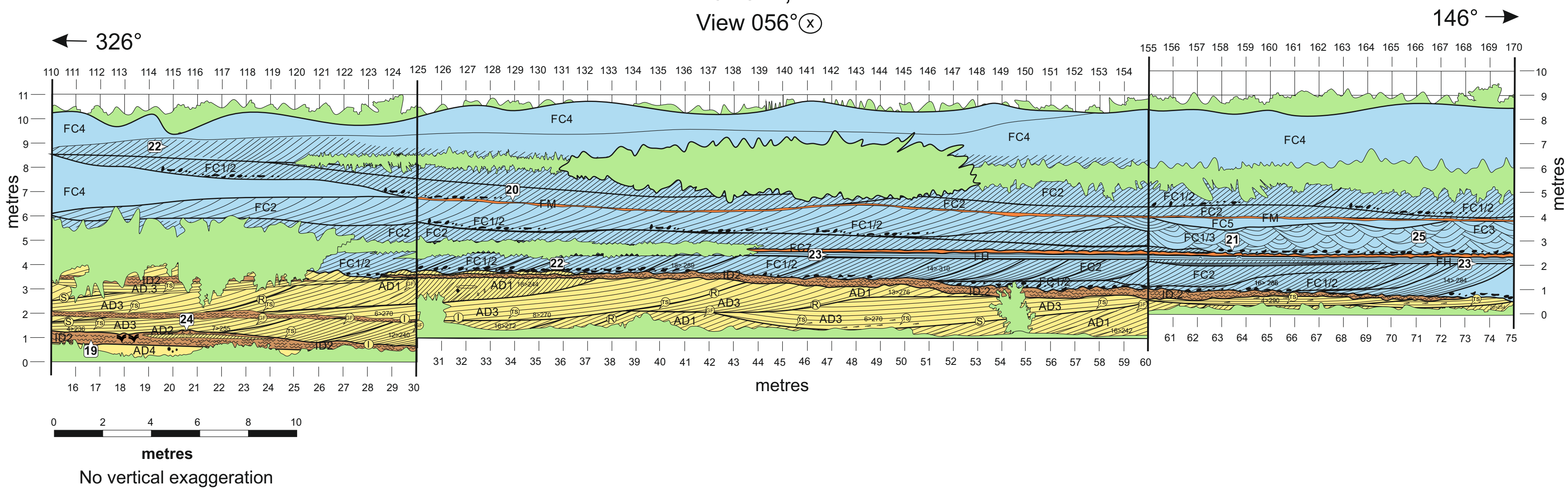
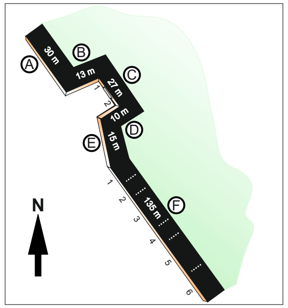


Figure 4.4: Cont.

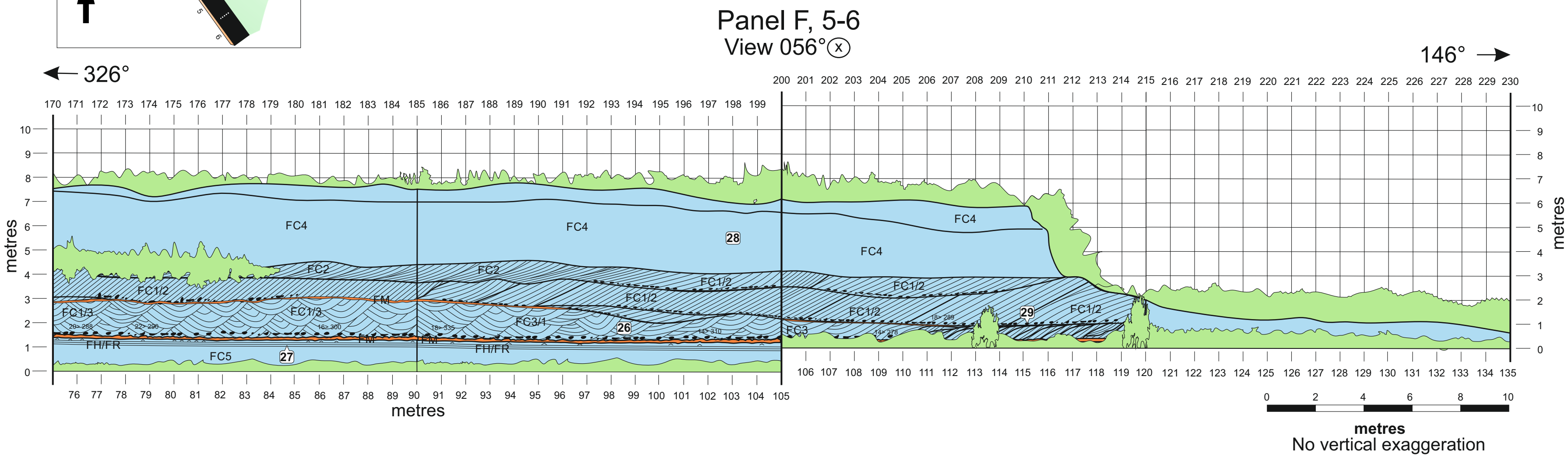


26 Trough cross bedding of fluvial origin, characterised by very coarse- to gravel grade lag deposits of locally derived materials (mud clasts).

27 Planar (horizontal), discontinuous, thinly laminated fine- to very fine-grained sandstone, with current lamination and mud-drapped climbing-ripple stratification toward the top.

28 Massive sandstone; lack of organised internal lamination; reflects the homogeneity of sandstone grain size.

29 Fluvial channel scour surface.



Facies Summary

Aeolian Facies	Fluvial Facies
AD1: Aeolian dune, moderate to high-angle bedded, grainflow dominated.	FC1: Fluvial channel-lag deposits, intraformational mud clasts.
AD2: Aeolian dune, moderate to high-angle bedded, grainfall dominated.	FC2: Fluvial channel fill, planar tabular bedding, straight crested dunes.
AD3: Aeolian dune, Low to moderate-angle bedded. Windripples dominated.	FC3: Fluvial channel fill, trough cross-beds, sinuous crested dunes.
AD4: Aeolian dune, massive, lacks clear stratification.	FC4: Fluvial channel fill, planar thinly laminated, upper flow regimes.
AD5: Aeolian dune, dune toe set soft-sediment deformation.	FC5: Fluvial channel fill, massive sandstone, rapid deposition.
ID1: Aeolian interdune, horizontally laminated, wind-ripple dominated.	FC6: Fluvial ch. fill, Ripple cross laminated sandstone with mud drapes.
ID2: Aeolian interdune, crinkly laminated, modified wind-ripple and adhesion-ripple strata.	FC7: Fluvial ch. fill or flood plain, mud-siltstone, suspension deposits.

Figure 4.4: Cont.

architectural element models have been proposed to illustrate the main architectural components of each formation; these elements serve as the building blocks required to reconstruct the larger-scale palaeo-depositional system and palaeoenvironment. To investigate the relationship between preserved aeolian set thicknesses, grainflow thicknesses and original aeolian dune bedform size, a further suite of data have been collected from the preserved aeolian bedsets, including measurements of dune set thicknesses (11 selected preserved sets) and thicknesses of individual grainflow deposits (130 readings).

4.6 Sedimentary facies analysis

Although the Wilmslow Sandstone Formation and Helsby Sandstone Formation have been referred to in several previous studies (e.g., Thompson, 1969; 1970a; Warrington et al., 1980; Planet et al., 1999; Thompson and Meadows, 1997; Mountney and Thompson, 2002), to date, a detailed sedimentological analysis of the Runcorn Expressway cutting has not been published, and no detailed studies have been published previously relating to the outcrops discussed herein. This study presents a formal and detailed facies scheme for the Runcorn Expressway road-cut, based on primary observations at the outcrop. Fourteen lithofacies types are recognised and these are grouped into four facies associations: aeolian dune, aeolian interdune, channelised fluvial and non-channelised fluvial (Table 4.1).

This extensive road-cut section studied provides a valuable example of the preserved remnants of an ancient water-table influenced aeolian system – a so-called wet aeolian system (sensu Kocurek and Havholm, 1993). Interdune units preserved between aeolian dune units record the impact of the water table on aeolian system construction and development and sediment accumulation and preservation (Kocurek et al, 1992; Carr-Crabaugh and Kocurek, 1998).

Table 4.1: Summary Lithofacies observed in the Runcorn road cut (outcrop): Helsby and Wilmslow Sandstone formations.

Code	Facies	Description	Interpretation	Occurrence	Facies Association
AD1	Moderate- to high-angle grainflow strata; sandstone	Reddish-brown, orange, fine to coarse, well- sorted, rounded and frosted quartz grains, forms planar, inversely-graded sand laminae between 10-60 mm thick, up to 23° inclined. Set thickness 10cm to 1 m.	Packages of grainflow strata, deposited by successive dry sandflows down the dune slipface.	Dune leeside foresets, very common.	Aeolian Dune
AD2	Moderate-angle grainfall strata; sandstone	Reddish-brown, orange, fine to medium, well-sorted, rounded grains, small dunes, preserved set thickness ranges from 10 to 30 cm, foresets inclined up to 15°.	Thin parallel cross-stratified grainfall strata, settled out of suspension. Recorded as draping over grainflow or wind-ripples strata.	Dune leeside foresets, rare	Aeolian Dune
AD3	Low- to moderate-angle wind-ripple strata; sandstone	Reddish-brown, orange, medium to very coarse-sand, well sorted, rounded to subrounded grains, lamina rarely exceeded 10mm in thickness, foreset inclined 12°. Inverse graded, pinstripe lamination evident.	Cross-stratified strata, deposited by the migration of aeolian ripples either down or across the lower part of the dune slipface.	Dune leeside toesets, common	Aeolian Dune
AD4	Massive to weakly cross-bedded sandstone	Reddish-brown, brown, orange, medium to coarse, homogenised sandstone, very well sorted, rounded to well-rounded grains, very faint, high-angle cross stratification.	Lack of internal structure is attributed to the accumulation of uniform grain size along the dune foresets, in a stable wind velocity conditions.	Dune foresets, common	Aeolian Dune
AD5	Soft-sediment deformed sandstone	Reddish-brown, orange, fine to coarse, moderate to well sorted, rounded grains. 10-50 mm-thick foreset laminae. Deformed small cross-stratified strata (soft-sediment deformation) intertongues with wavy laminated sandstone.	Intercalated packages of grainflow and wind-ripple lithofacies at the lower part of aeolian dunes, deposited under damp surface conditions on interdune flats between dunes.	Dune plinth or apron	Aeolian Dune
ID1	Horizontal to sub-horizontal laminated sandstone	Reddish-brown, brown, fine to very-coarse, moderately to well sorted, subrounded grains. Forming sets that are 10cm to 1 m thick. Near horizontal, discontinuous translatent wind-ripple strata.	Deposited through the migration of aeolian ripples across a dry interdune surface. The texture of sand grains indicates accumulation under a high wind velocity.	Areas low water table	Aeolian interdune
ID2	Wavy laminated sandstone	Reddish-brown, orange, whitish-grey, fine to very coarse sand. Wavy to crinkly laminated sandstone, sets are 10 cm to 1 m thick, adhesion ripples and small-scale flame structure are evident. This facies also records a down-folded lamination at set tops.	This facies accumulated in interdune areas between migrating aeolian dune forms. This facies records a vertical change to dry interdunes, reflects the passage to drier edge of interdune. The down-folded laminations are formed by vertebrate footprints (indenters).	Areas of high water table, common	Aeolian interdune

Table 4.1: Cont.

FC1	Intraformational mud-clast conglomerate	Reddish-brown, formed mainly of thinly laminated mudstone and silty mudstone as a rip-up clasts, vary in size (1 to 20 cm long).	Fluvial channel-lag deposits, composed of dispersed pebbles of reworked local sediments (mud clasts) sourced from in-channel and floodplain areas.	Channel lag/common	Fluvial Channel
FC2	Planar cross-bedded sandstone	Reddish-brown or brown, fine to very-coarse (fining upward), moderately-well sorted, sub-rounded sandstone grains. In cross-stratified sets (20cm to 1 m thick, 10 to 22° inclined), forming sets up to 2.5 m thick that may be stacked, sharp set boundaries, mud clasts present as lag deposits.	Fluvial channel-fill deposits, represents the migration of confined flow within a channelized setting, sediment deposited by straight-crested dune-scale bedforms.	Channel fill/common	Fluvial Channel
FC3	Trough cross-bedded sandstone	Reddish-brown or brown, fine to very-coarse sand, moderately to poorly sorted, sub-rounded sandstone grains. Medium-scale trough cross-strata. Occurs in sets of varying thickness, forming cosets up to 1.5 m. rip-up mudclasts of variable size are common.	Fluvial channel-fill deposits, representing the migration and accumulation of sandy mesoforms within a confined fluvial channel system, deposited by sinuous crested dune-scale bedforms.	Channel fill/common	Fluvial Channel
FC4	Massive (structureless) sandstone	Brown or light-brown, fine to medium grained hard sandstone, homogeneous, staked sets (10 cm to 3 m thick).	Lacks organised internal lamination, reflects the homogeneity of the sandstone grain size and suggest the accumulation through rapid deposition during flood events.	Channel fill/common	Fluvial Channel
FH	Horizontally laminated sandstone	Reddish-brown, pink, fine to medium sand, commonly normally graded, moderate to well sorted, rounded – sub-rounded grains. Thinly laminated (1-5 mm) in sets 20-60cm thick. Current lamination is evident in the upper parts of sets.	This facies records deposition under the influence of upper flow-regime conditions, either in confined channel flow or non-confined fluvial sheet-like flow.	Upper channel fill/common	Fluvial Channel/ Floodplain
FR	Ripple cross-laminated sandstone	Brown or light-brown, very-fine to coarse, moderate to poorly sorted, sub-rounded sandstone grains. Mud-draped climbing ripples and lenticular bedding with complete preservation of ripple forms are evident. Sets range in thickness from 1 to 30 cm.	Represent unidirectional migration of small-scale ripple forms during low flow regimes within the upper part of the fluvial fill or in a non-confined floodplain setting. Preserved ripple foresets commonly exhibit mud drapes indicating pulsed migration between episodes when the water formed standing ponds.	Top channel fill	Fluvial Channel/ Floodplain
FM	Horizontally thin-laminated silty mudstone	Reddish-brown, planar, thinly-laminated, laterally extensive, sets rarely exceed 30 cm, commonly adopts a white or motled greenish-white appearance, preserves desiccation cracks at the top of the mudstone sets.	Channel abandonment or overbank deposits, alternation of siltstone and mudstone gives rise to thin laminae, represents deposition from suspension settling in floodplains or the final stage of channel filling and abandonment.	Top channel fill/ overbank areas/common	Fluvial Channel/ Floodplain

4.6.1 Aeolian facies

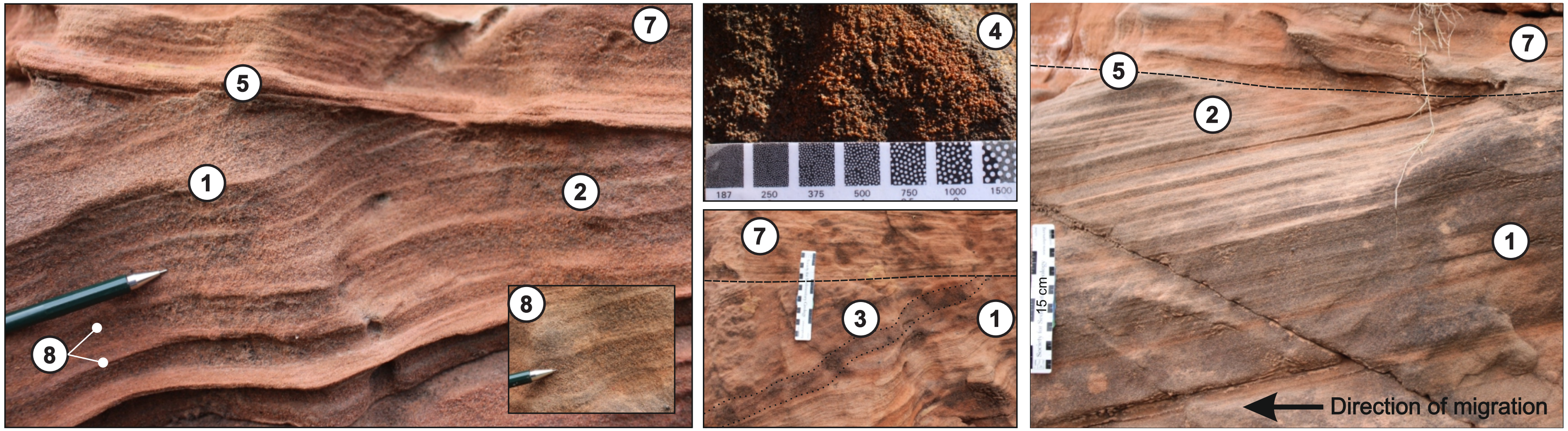
Seven aeolian lithofacies have been identified and collectively these represent the preserved products of aeolian dune and interdune deposition (Table 1).

4.6.1.1 Moderate to high angle grainflow strata facies (AD1)

Description: This lithofacies is red to brown-orange, characterised by well- to very well-sorted and rounded predominantly fine-grained (rarely medium- to coarse-grained) quartzarenite sandstone grains with millet seed texture (mature-supermature) (cf. Folk, 1951; Livingstone and Warren, 1996). Deposits are moderately indurated (cf. Lancaster, 1993), arranged in moderate to high angle cross-bedded sets characterised by packages of grainflow avalanche strata (cf. Hunter, 1977). Individual grainflow stratum form 5 to 50 mm-thick wedge-shaped tongues of massive or weakly inversely graded sandstone (cf. Sallenger Jr, 1979). Cross bedding within sets is inclined at angles up to 23°. Discrete grainflow strata thin in an up-dip direction toward the top of sets (cf. Lancaster, 1995), in most cases truncated by the overlaying set (Figure 4.5), and pass down-dip into interdune or dune plinth deposits. This facies represent approximately 35% of the succession. This facies also preserves distinctive bi-modal grain size lamination in places, especially in lower parts of sets. Individual grainflow packages stack together to form planar tabular cross-strata themselves arranged into sets that are up to 1 m thick. Foresets have a mean dip azimuth direction toward 255° (n = 130). In several places, sets of facies AD1 are truncated downwind by overlying sets to form reactivation surfaces (Brookfield, 1977 and Fryberger, 1993). These reactivation surfaces are ubiquitous and appear in cyclic patterns (mean lateral downwind spacing = 2.1 m) (Figure 4.4, Panels A-C; cf. Hunter and Rubin, 1983). Individual foresets produced by packets of grainflow strata are separated by either wind-ripple strata of facies AD3 (packages of which rarely exceed 10mm thick) or, more rarely, by grainfall laminations of facies AD 2 (1 to 3 mm-thick foreset laminations) (Figure 4.5).

Interpretation: The texture of the sandstone described above and the lack of any clay drapes (Figure 4.5), all are indications of an aeolian grain

Facies AD1 : Moderate-high angle grainflow (avalanche) strata



Facies Information

Colour	Reddish-brown, orange
Grain Size	Fine to coarse grained sand
Set Thickness	10 cm to 1m sets 10mm-60 mm foreset thickness
Sorting & Texture	Well to very-well sorted Rounded grains
Facies Association	Aeolian Dune
Architectural Elements	A1

Facies Characteristics

1. Aeolian dune high-angle cross strata, up to 23 degrees
2. Aeolian dune foreset grianflow dominated
3. Grainflow strata pinching out toward the base lee-slope
4. Grains have "millet-seed" texture
5. Dune strata truncations (erosional truncation surface)
6. Bi-modal grain size (pinstripe lamination).
7. Wind-ripples dominated strata
8. Concordant cyclic cross beds, thick alternating grainflow deposits separated by thin wind-ripple laminae

Interpretation

Moderate- to high-angle cross-stratified aeolian sand dune facies, with packages of grainflow strata, deposited by successive dry sandflows down the slipface of the dune. Lamina colouration is due to the variation of grainsize and/or fluid bleaching through porous and permeable strata. Bi-modal grain size resulted from kinetic sieving during avalanche movement down the dune slipface. Planar lamination indicates that the deposits record the migration and accumulation of aeolian dunes that had straight crest-line morphologies.

Figure 4.5: Aeolian dune, moderate-high angle grainflow lithofacies (AD1).

transport mechanism (Glennie, 1970). This facies is also characterised by its style of cross-bedding (inclination up to 23°), which is a typical and common characteristic feature of sedimentation on the lee of aeolian dunes (cf. McKee, 1965). Aeolian dunes typically show an angle of lee-slope inclination that is steeper than that of most dunes formed by fluvial processes (e.g. Selley, 1996; Hsü, 2004). Facies AD1 represents the preservation of aeolian sand dunes slipface (lee side) deposits that accumulated through the migration of dunes through repeated movement (repetitive redistribution) of friable weak inverse graded sand grains via avalanching (sandflow) over the dune slipface (Kocurek, 1996, Mountney, 2006a), through gravity-driven failure, that occurs when sand accumulates high on the dune lee slope surface until a critical angle of repose is exceeded, which is typically 32° to 34° for well-sorted, loose, dry sand, resulting in sand avalanches that carry grains down the slope (Figure 4.6; Allen, 1970; Carrigy, 1970; Hunter, 1977; Loope et al., 2012). The abundant occurrence of grainflow strata in this type of sandstone indicates that the dunes possessed well-developed slipfaces (cf. Scherer, 2000). The bi-modal grain-size distribution reflects the accumulation of thin units of finer grained wind-ripple and grainfall strata between thicker packages of grainflow strata (cf. Hunter, 1981, Mountney and Howell, 2000; Mountney, 2006b). It is also possible for such separation to form via the following processes: (i) as a result of the movement of grains of different sizes by different run-out distances down a dune slipface; (ii) by the downward gravitational settling of smaller grains through the cavities between larger grains due to kinetic sieving or gravitational sorting (Middleton, 1970, Fryberger and Schenk, 1988; Jullien et al., 2002). The planar nature of the foresets in directions along-strike indicates that the deposits record the migration and accumulation of aeolian dunes that had straight crest-line morphologies (cf. Mountney, 2006a). The presence of regularly spaced cyclic reactivation surfaces truncating the sets of grainflow strata indicates periodic erosion and reworking of the dune sediments on lee face of the migrating bedforms in response to a change in one or more of the following: (i) the aeolian bedform migration direction as a result of a change in wind direction (e.g. reversed airflow); (ii) change in wind velocity that erodes sand from one part and depositing it in another part of the dune; (iii)

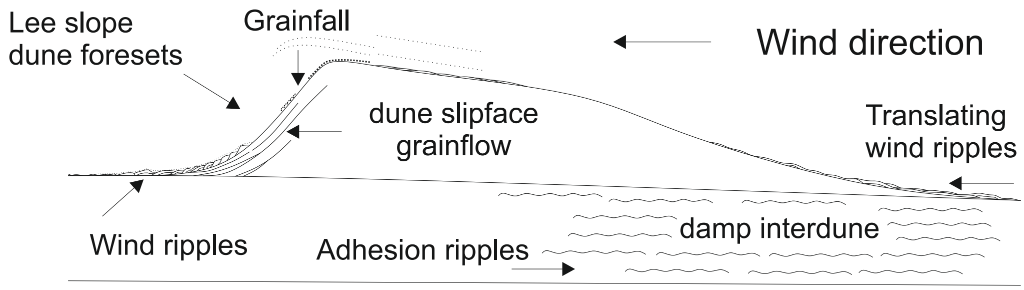


Figure 4.6: Aeolian dune and interdune cross section illustrating the main deposit types and its preferred location of accumulation. In this schematic example, the dune is migrating across a damp interdune sand flat without climbing .

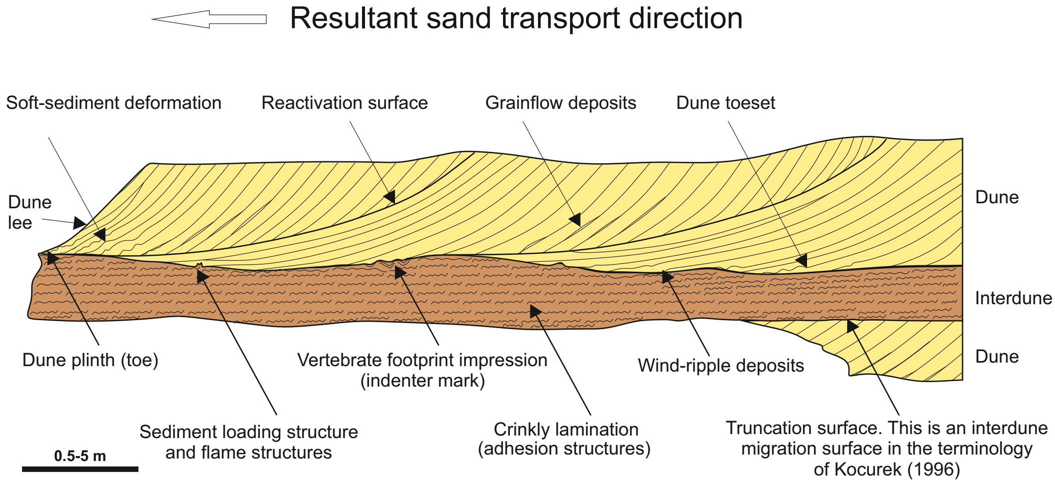


Figure 4.7: Sedimentary features of aeolian dune and damp interdune strata.

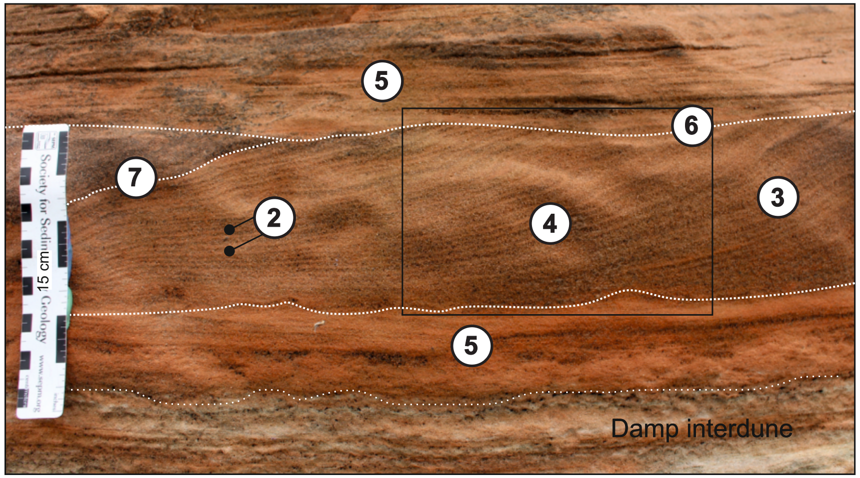
changes in dune asymmetry, lee slope steepness or dune height (Rubin, 1987; Kocurek, 1996; Clemmensen et al., 1997; Bristow and Mountney, 2013). The sub-horizontal and sub-parallel erosional bounding surfaces which separate the accumulation of the aeolian dune sets are interdune migration surfaces, and they overlie interdune deposits (see below) (Figure 4.7; Brookfield, 1977; Kocurek, 1981a, 1996).

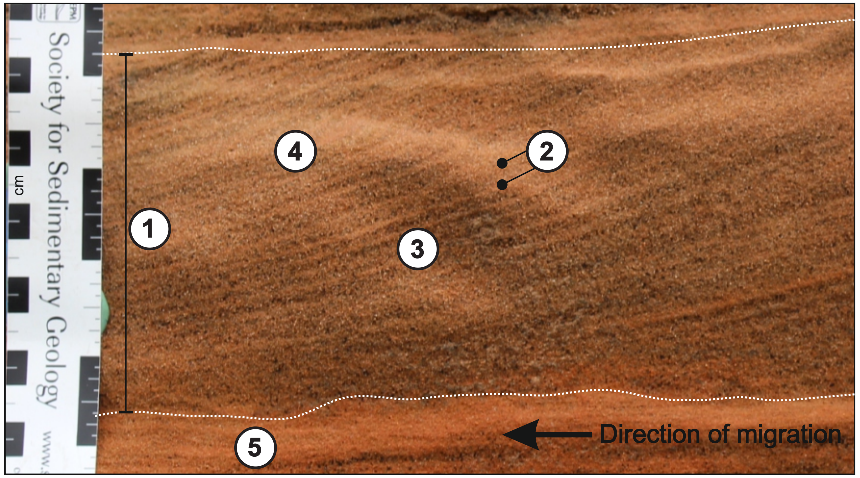
4.6.1.2 Moderate angle grainfall strata facies (AD2)

Description: This fine- to medium-grained, well- to very well-sorted, rounded, reddish-brown quartzarenite is characterised by thinly laminated (1 to 3 mm) cross-bedded sandstone, with foresets inclined at a moderate angle (typically 15°; rarely steeper). Laminations are commonly distinct due to clear grain-size segregation (bi-modal) forming cyclic cross beds of different grain sizes (fine and medium). Sets are 0.1 to 0.3 m thick. Individual foreset are separated by wedge-shaped grainflow laminations in cross strata that may extend to the base of the sets. This facies represents 5% of the succession. Examples of lithofacies AD2 pinch out in a down-dip direction where they merge with horizontal and low-angle-inclined wind-ripple deposits. This facies commonly preserves deflated ridge-and-swale structures on set upper bounding surfaces (cf. Simpson and Loope, 1985). Sets are truncated downwind by overlying sets to generate reactivation surfaces. Foreset dips decrease gradually towards the dune plinth (set base), and downlap with a near-asymptotic relationship to horizontal bounding surfaces that define the set base (interdune migration surface). The cross-strata dip toward the south-west, with a mean value of 255° (n = 130) (Figure 4.8).

Interpretation: Grainfall and grainflow strata are two major components of aeolian dune slipface foreset deposits (Hunter, 1985; Brookfield and Silvestro, 2010). Grainfall deposits represent airfall deposition and commonly accumulate in zones downwind of points of air flow separation at the dune crestline or brinkline (Collinson, 1986); grains transported by saltation processes during high-intensity winds pass over the dune brink, then fall through the separated flow zone where the wind speed is locally reduced, to deposit directly on the upper part of the aeolian dune lee slope. In exceptional cases, grainfall deposits may extend over the entire lee face

Facies AD2: Moderate angle grainfall strata





Facies Information

Colour	Reddish-brown, brown
Grain Size	Fine to medium sand
Set Thickness	10 to 30 cm sets 1 to 3 mm foreset thickness
Sorting & Texture	Well to very-well sorted Rounded grains
Facies Association	Aeolian Dune
Architectural Elements	A1

Facies Characteristics

1. Small scale dune cross strata
2. Bi-modal grain size
3. Alternating aeolian lithofacies deposits (grainflow, grain fall)
4. Concordant cyclic cross beds
5. Wind-ripple-dominated strata
6. Deflated ridges
7. Reactivation surface

Interpretation

Moderate-angle, thin parallel cross-stratified grainfall influenced aeolian sand dune facies. Grainfall is a distinctive feature of aeolian strata that records settling out of temporary suspension of previously saltating grains (during episodes of reduced wind velocity) on the lee slope of the bedform. Recorded as draping over grainflow or wind-ripple strata. Thickness depends on intensity and duration of the wind.

Figure 4.8: Aeolian dune, moderate angle grainfall lithofacies (AD2).

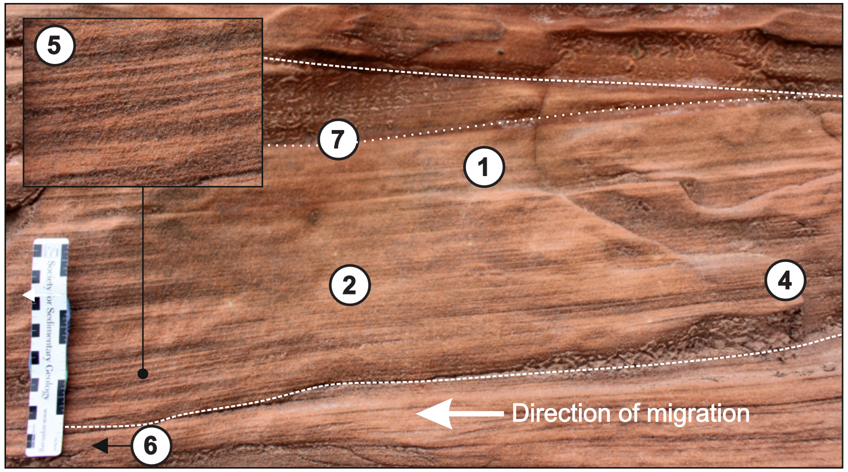
of small dunes (Hunter, 1977; Collinson, 1986; Nickling et al., 2002). In most modern settings grainfall deposits form thin (1 to 3 mm thick) laminations (Figure 4.6; Bristow and Mountney, 2013). The preservation of this lithofacies implies that the dune lee face slope was less than the angle of repose (32° to 34°, Allen, 1970) and the wind intensity was relatively low (at least locally and temporarily in the lee of the dune) such that it was insufficient to rework the sediment into wind-ripples (cf. Kocurek and Dott,1981). The alternation of grainfall and grainflow laminae suggests an occurrence near the base of a dune; this indicates that the sets likely represented relatively small dunes (cf. Kocurek and Dott, 1981; Mader, 1985a).

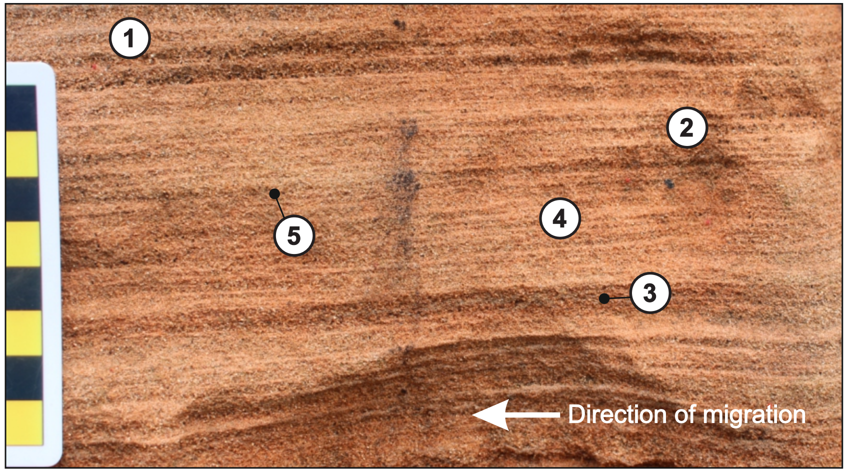
4.6.1.3 Low to moderate angle wind-ripple strata facies (AD3)

Description: This facies is characterised by orange-brown, medium- to very coarse-grained quartzarenite that is moderately to well-sorted, and possess grains that are sub-rounded to rounded. This facies is organised into low-to-moderate angle (4°-12°) inclined cross-stratified sandstone sets. Sets are dominated internally by thin (<10 mm) laminations produced primarily by climbing wind-ripple translatent strata, in some cases these are separated by thin (<3 mm) laminae of grainfall origin (cf. Hunter, 1977, 1981; Kocurek, 1991). Laminations are generally discontinuous (pinch-out laterally), rarely exceed 10 mm thick and are inclined at angles up to 12°. Laminations exhibit weak inverse grading in some sets (Figure 4.9). Laminations usually merge in an up-dip direction with more steeply inclined grainflow dominated facies (AD1), whereas they commonly transition down-dip into irregular-crinkly and/or planar laminations of facies ID1 and ID2 (Figure 4.4, Panels B and C, 2). Grain size homogeneity was observed in some sets of this lithofacies, which limits the ability to distinguish the lamination boundaries and also makes the inverse grading hard to discern. This facies represents 20% of the succession.

Interpretation: Wind ripples are the smallest scale of aeolian bedform (Goudie, 2013). They are very common in aeolian systems, and are usually the first structure to develop as a consequence of sand transport and deposition on a dry sand surface. Wind ripples develop in response to saltation processes, in conditions where wind shear decreases and the wind

Facies AD3: Low-moderate angle wind-ripples strata





Facies Information

Colour	Orange, brown
Grain Size	Medium to very-coarse
Set Thickness	10 cm to 60 cm sets laminae rarely exceeded 10 mm in thickness.
Sorting & Texture	Moderate to well-sorted Sub-rounded to rounded grains
Facies Association	Aeolian Dune
Architectural Elements	A1

Facies Characteristics

1. Aeolian dune low-angle cross strata, up to 12 degrees
2. Aeolian dune foreset wind-ripples dominated (translatent wind-ripple strata common)
3. Packages of translatent wind-ripple strata
4. Bi-modal grain size (pinstripe lamination)
5. Grain-size segregation; lamina pinch out laterally
6. Wind-ripple strata merge with interdune facies
7. Reactivation surface

Interpretation

Low- to moderate-angle cross-stratified aeolian sand dune facies, dominated by translatent and climbing wind-ripple strata; deposited by the migration of aeolian ripples either down or across the lower part of the dune slipface. Inverse graded lamination and pinstripe lamination are evident; formed by migration of ballistic wind ripples over lower dune slope surface.

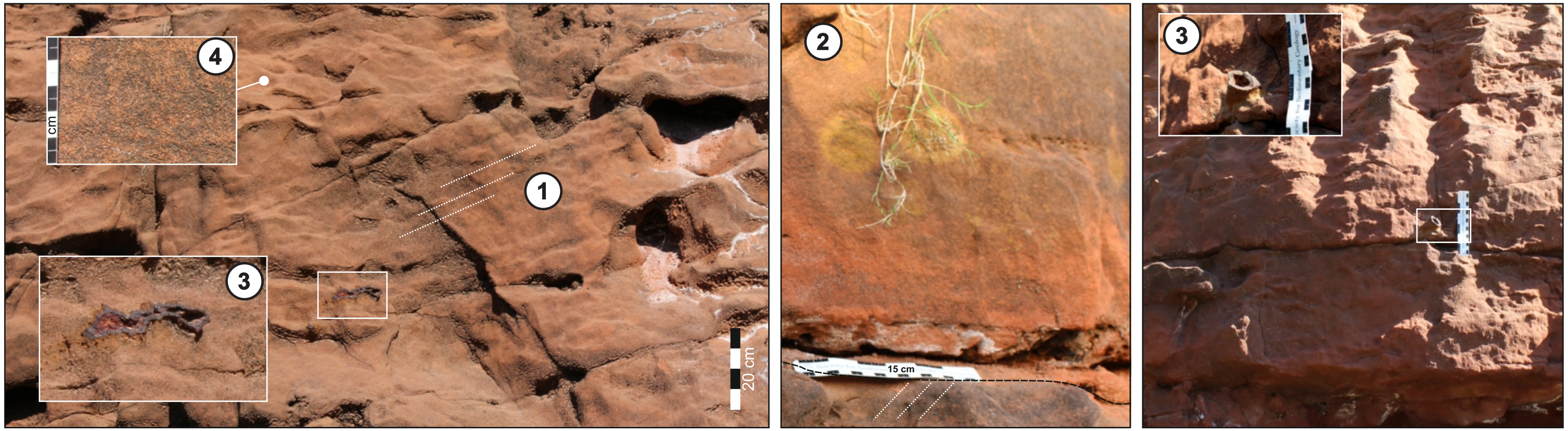
Figure 4.9: Aeolian dune, low-moderate angle wind-ripple lithofacies (AD3).

transport capacity declines resulting in the deposition of excess sediment to the bed forming wind-ripple lamiae (Lancaster, 1995; Kocurek, 1996). The development and migration of wind ripples is linked to the surface traction processes (Collinson, 1986; Anderson, 1987), and the predominance of this lithofacies (wind-rippled lamination) in the lower parts of the dune foresets (dune plinth areas) indicates the presence of winds that were sufficient to cause traction transport (Figure 4.6, and Figure 4.7; Kocurek, 1991; Sweet, 1992; Kocurek, 1996). Wind-ripple strata commonly reflect wind conditions that are intensive but insufficient for dune development (Hunter, 1977). This facies is generally restricted to the lower part of the preserved dune sets (dune plinth region) because its presence higher on the dune slipface makes it prone to reworking by grainflow avalanche processes, thereby preventing its widespread preservation in such settings. Aeolian wind ripples may develop in a variety of aeolian sub-environments including dry interdunes, aeolian sand sheets and on low- to moderately-inclined dune slopes (Clemmensen and Abrahamsen, 1983; Kocurek, 1991; Collinson, et al., 2006; Mountney, 2006b, Rodriguez-Lopez et al., 2012). The discontinuity of packages of wind-ripple strata observed in this study is attributed to episodic accumulation and deflation, which likely reflects local variations in wind intensity.

4.6.1.4 Massive to weakly cross-bedded sandstone facies (AD4)

Description: This facies is reddish-brown to orange, medium to coarse quartzarenite that is well-sorted and possesses well-rounded grains. Deposits occur in sets that are 1.5 to 2 m thick, and which possess a very faint (vague), high-angle-inclined cross lamination (in many cases massively bedded) (Figure 4.10). The faint cross-bedded sandstone sets are inclined up to 22°, and extend across almost the entire exposed section (Figure 4.4, Panels A-F2). In places, a variation on this facies type is characterised by a homogeneous sediment texture, whereby massive packages occur in sets up to 2 m thick and linked vertically and laterally with other aeolian lithofacies (AD1, AD3 and ID2). In places, small iron oxide cemented rhizoliths (5 to 8 cm in diameter) are evident, though are rare. No interaclasts have been observed in this facies (Figure 4.10). This facies represents 10% of the succession.

Facies AD4: Massive sandstone



Facies Information

Colour	Reddish-brown, orange
Grain Size	Medium to coarse sand
Set Thickness	15 cm to 2 m sets
Sorting & Texture	Well to very-well sorted Rounded to well rounded grains
Facies Association	Aeolian Dune
Architectural Elements	A1

Facies Characteristics

1. Very faint (vague) aeolian dune high-angle cross lamination, up to 22 degrees visible in a few places
2. Homogeneous sediment texture
3. Rhizoliths; iron oxide concretion; calcite cement
4. Well sorted, well rounded and frosted grains, aeolian origin

Interpretation

Stacked aeolian dune foreset deposits that lack clear stratification (internal structure absent) due to grain size homogeneity. The absence of identifiable grain-size segregation lamination is attributed to the accumulation of uniform grain size along the dune foresets, which suggests a stable wind velocity (effective sand-grain sorting mechanism). Rhizoliths Indicate the stabilization of aeolian sand dune by vegetation.

Figure 4.10: Aeolian dune, massive sandstone lithofacies (AD4).

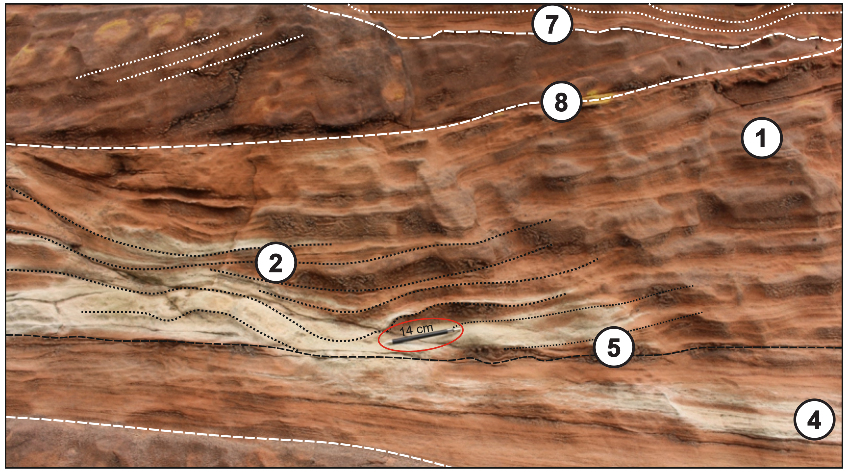
Interpretation: This facies represents aeolian dune foreset deposits that lack clearly developed stratification either because the internal structure is absent or because it has been obscured due to grain size homogeneity. The absence of identifiable grain size segregation lamination is attributed to the accumulation of uniform grain size along the dune foresets; this suggests an effective sand-grain sorting mechanism and weak tractional processes that are required to generate sedimentary structures and bed boundaries, at the time of deposition (cf. Chan, 1999). Structureless aeolian sandstone deposits have also been attributed to deposition from hyperconcentrated flows down the dune slipface (Simpson et al., 2002), for example in the aftermath of intense rain storms. The development of rhizoliths – a type of organo-sedimentary structure produced by plant roots (Loope, 1988) – is indicative of the episodic partial stabilisation of dunes to varying degrees. Lack of interaclasts in this lithofacies support the interpretation of an aeolian origin.

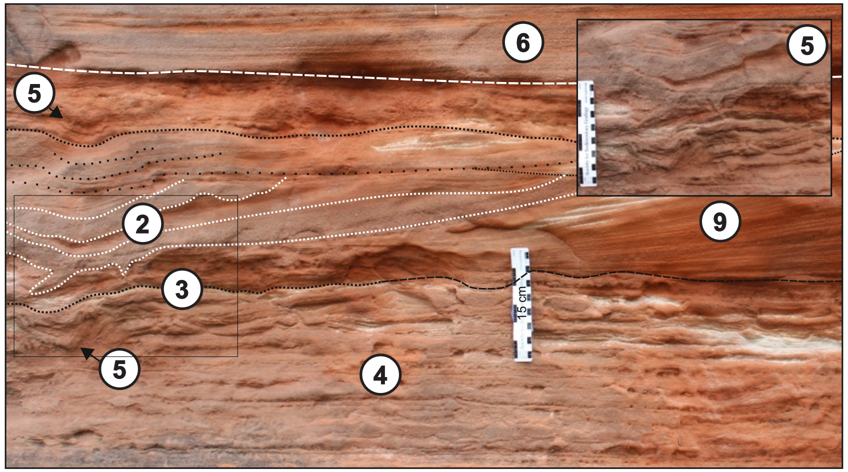
4.6.1.5 Soft sediment deformation facies (AD5)

Description: This facies is characterised by reddish-brown to orange coloured, fine- to coarse-grained quartzarenite that is moderately to well-sorted and characterised by rounded sandstone grains. This facies occurs as sets that range from 0.1 to 1 m thick, mainly in the lower part of the preserved aeolian dune sets. The internal structures of sets of this facies show small-scale deformed stratification, including small-scale liquefaction structures, flames (6 cm in height), and folding structures (8 to 27 cm in width). Examples of facies AD5 occur laterally and vertically intercalated with the facies AD1, AD3 and ID2. Slump and folding structures of mainly avalanche strata are especially common in the basalmost 25% of aeolian dune sets. The deformation is usually present at the contact with facies ID2, which is associated with accumulation under the influence of damp surface conditions (Figure 4.11). Facies AD5 represents 5% of the succession.

Interpretation: Soft-sediment deformation structures are the result of liquefaction in water-saturated sediments (Allen, 1982; Owen, 1987; Glennie, and Hurst, 2007; Topal, and Özkul, 2014). This facies potentially formed through several processes: 1) the re-sedimentation of aeolian

Facies AD5: Soft sediment deformation (dune toe sets)





Facies Information

Colour	Reddish-brown, orange
Grain Size	Fine to coarse sand
Set Thickness	10 cm to 1 m sets 10 mm to 50 mm foreset thickness
Sorting & Texture	moderate to well sorted Rounded grains
Facies Association	Aeolian Dune
Architectural Elements	A1

Facies Characteristics

1. Dune cross-stratified strata
2. Dune toe-set soft-sediment deformation
3. Intertonguing of aeolian toe-set strata with damp interdune
4. Crinkly laminated sandstone, damp interdune
5. Vertebrate footprint
6. Wind-ripple strata
7. Deflation surface
8. Reactivation surface
9. Grainflow tongue wedges out down-slope

Interpretation

Lower parts of aeolian dunes, influenced by soft-sediment deformed sandstone, composed of intercalated packages of grainflow (sandflow) and wind-ripple lithofacies. Evidence for localized slipface collapse of dunes. The presence of fine, wavy-laminated sandstone indicates the deposition under damp surface conditions directly upon interdune flats. Down-folded lamination in the uppermost layers of the damp interdune element are the cross-sectional view of vertebrate footprint structures; very common in the Triassic Wilmslow Sandstone and Helsby Sandstone formations of the Cheshire Basin.

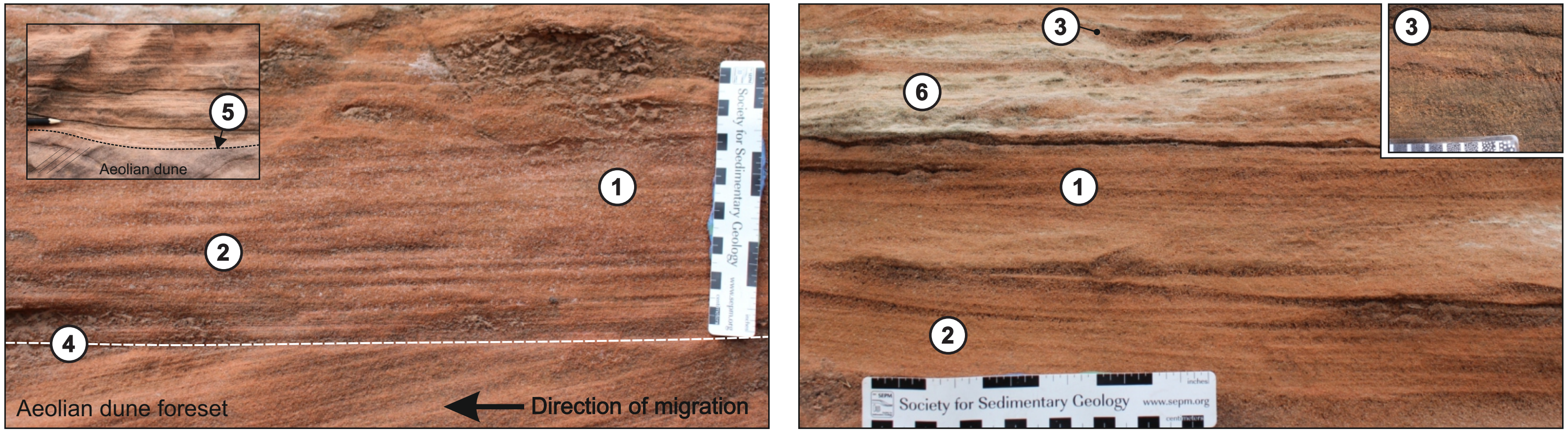
Figure 4.11: Aeolian dune, dune toe set facies, soft-sediment deformation (AD5).

sediment triggered by heavy rainfall (cf. Heward, 1991; Gareth and Berry, 1993); 2) inundation processes related to the change in water table, occurring as a result of fluvial flooding (Blaszczyk, 1981; Allen, 1982; Heward, 1991); 3) slumping of unconsolidated sand or moderately cohesive, moist sands on aeolian dune lee slopes (McKee et al., 1971; Glennie, 1972; Doe and Dott, 1980; Eschner and Kocurek, 1988), which is possibly the case herein. The part of the original dune influenced by soft sediment deformation is the lower part of the aeolian dune sets, which is composed of intercalated packages of grainflow (AD1) and wind-ripple (AD3) strata. The presence of this facies (AD5) above instances of wavy-laminated sandstone of lithofacies ID2 indicate the deposition in lower dune plinth area under the influence damp surface conditions where the dunes were constructed directly on interdune flats (cf. Kocurek, 1996). It is also possible for such sediment deformation to occur as a result of loading of the saturated sand by an advancing aeolian sand dune (Collinson, 1994; Horowitz, 1982). In some sets, the aeolian sand deposits display soft-sediment deformation (down-folding at the top of the damp interdune element) related to animal tracks (cast of vertebrate footprint). See description of facies ID2 for details of this structure (Figure 4.11; Mader, 1985; Loope, 1986; Rodriguez-Lopez et al., 2012).

4.6.1.6 Dry interdune facies (ID1)

Description: This facies is composed of fine- to very coarse-grained, reddish-brown to brown sandstone that is moderately to well-sorted; grains are sub-rounded. This facies occurs as 0.1 to 0.8 m-thick sets, examples of which have a maximum lateral extent of 20 m. This facies is less common than lithofacies ID2; it represents 5% of the entire aeolian succession. Rock composed of this facies has a friable nature. This facies is characterised internally by near-horizontal, discontinuous translatent strata (McKee, and Bigarella, 1979; Ahlbrandt and Fryberger, 1982). This facies is associated with small-scale (10 to 30 cm in length) deflation scours on the top surfaces of sets (Figure 4.12; Hunter, 1977, Kocurek, 1981). Many sets exhibit bimodal grain size segregation – so-called pinstripe laminations (Fryberger and Schenk, 1988). In other sets, such lamination is difficult to distinguish due to the well-sorted nature of the grains and the overall uniformity of the

Facies ID1: Horizontal-sub-horizontal laminated sandstone



Facies Information

Colour	Reddish-brown, brown
Grain Size	Fine to very-coarse
Set Thickness	10 cm to 1m sets
Sorting & Texture	moderate to well sorted Sub-rounded
Facies Association	Aeolian Interdune
Architectural Elements	A2

Facies Characteristics

1. Near-horizontal translatent wind-ripple strata
2. Discontinuous lamination (laminae pinch-out laterally)
3. Coarse grained Wind-ripple strata
4. Truncation of dune strata (erosional truncation surface)
5. Erosion surface (deflation surface)
6. Fluid bleached strata

Interpretation

Horizontal to sub-horizontal translatent wind-ripple strata. Deposited through the migration of aeolian ripples across a dry interdune surface. Texture of sand grains within the ripple strata indicate accumulation under high wind velocity, which was likely also responsible for the deflation of the underlaying dune accumulation. Dry interdune units record the nature of the substrate within aeolian dune fields.

Figure 4.12: Horizontal-sub-horizontal laminated sandstone lithofacies (ID1).

grain-size distribution. The inclination of these laminations rarely exceeds one degree; examples of sets of this facies commonly exhibit a gradual lateral or vertical transition into aeolian dune facies AD1, AD3.

Interpretation: Horizontal to sub-horizontal translatent wind-ripple strata accumulate through the migration of wind ripples across a dry interdune surface via climbing at very low angles (Hunter, 1977, Bristow and Mountney, 2013). The coarse texture of sand grains within the ripple strata, is indicative of the accumulation in interdune depressions, whereas finer sand grain fractions tend to be swept onto the dunes (Kocurek, 1996). The horizontally laminated appearance of this facies is may be related to wind velocity; high wind velocities can generate aeolian upper-stage plane bed lamination (Hunter, 1977, Allen and Leeder, 1980, Bridge, 2003), and this may occur intercalated with wind-ripple strata. The development of wind-ripple strata may also indicate a restriction in sediment availability or sediment supply, which is also possibly inferred if the underlying dune accumulation demonstrates evidence for partial deflation on its upper surface (Figure 4.12; Kocurek, 1981a; Kocurek and Nielson, 1986; Kocurek and Lancaster, 1999).

The grain-size distribution was generated as a consequence of the transport processes (Bagnold, 1941; Hunter, 1977). Tractional processes for example, forms plane bed lamination at high wind velocities, and the deposition in dune-fields by processes other than tractional deposition is normally restricted to the lee side of dune crests as a result of air flow separation from the surface (Hunter, 1977). Transport processes also responsible for the generation of bimodal grain size sorting, bimodal grain size sorting being a function of wind ripple development that occurs under conditions of constant wind velocity (Sharp, 1963; Glennie, 1970; Fryberger and Schenk, 1988). The sedimentary structure of the wind ripples is characterised by an arrangement whereby laminae generated by ripples that have coarser grains on their crests; this is due to the trapping of the sand finer grains in the lee side troughs in front of migrating ripples (Sharp, 1963; Mountney, 2006; Bristow and Mountney, 2013). This generates distinctive pinstripe lamination because of differences in grain size distribution. Such pin-striping is highlighted in the outcrop by the changes in colour: the finer grains tend to be slightly darker than the coarser grains; different diagenetic processes can also highlight such a type of lamination (Allen, 1984, Fryberger and Schenk, 1988; Anderson and Bunas, 1993; Makse, 2000). Pinstripe lamination is a distinctive feature of both modern and ancient aeolian sediments (Fryberger and Schenk, 1988; Cowan, 1993).

Interdune deposits accumulated between dunes formed in bimodal or complex wind regimes, such as linear and star dunes, tend to be thicker and more areally extensive than those associated with unimodal dune systems, which produce lenticular, diachronous and relatively thin (<2 m) interdune accumulations (e.g., (McKee and Moiola, 1975; Ahlbrandt and Fryberger, 1981). The limited thickness of the interdune facies observed in the studied outcrop (sets < 1 m) is typical of sedimentation in dry interdunes of limited lateral extent between dunes developed under the influence of a unimodal wind regime.

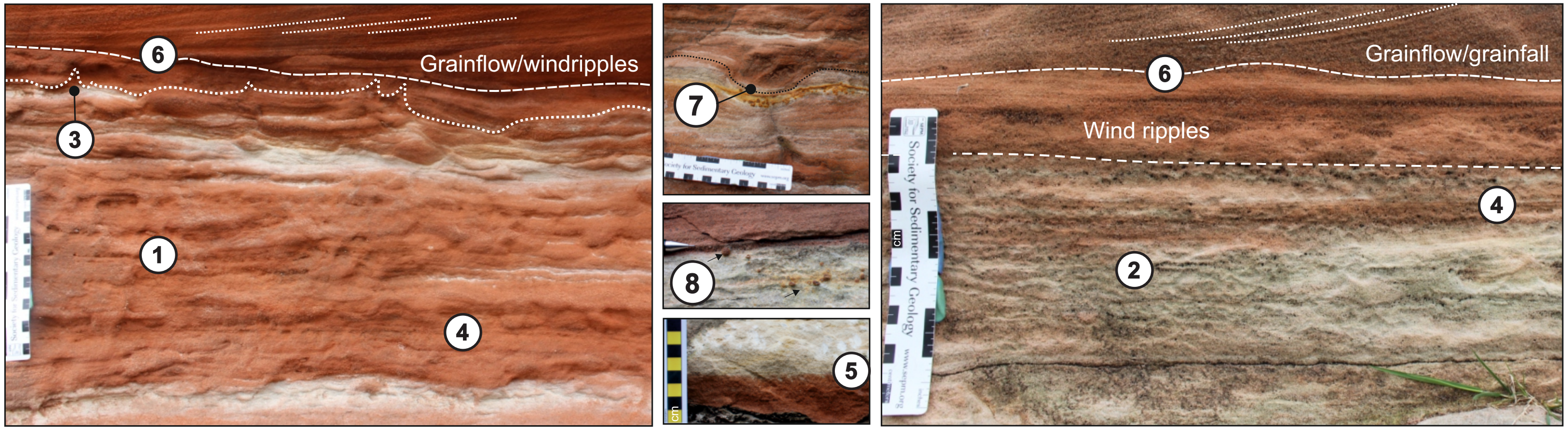
Dry interdune units reflect the nature of the substrate within the parts of the aeolian dune-field between the aeolian dune forms. They demonstrate a dry accumulation surface in interdune areas because of the lack of evidence that the sedimentation was influenced by any type of moisture and because wind ripple generally required dry, loose, non-cohesive, sand-grade sediment to form (Kocurek, 1981a; Mountney, 2006a).

4.6.1.7 Damp interdune facies (ID2)

Description: Facies ID2 is most the common type of interdune deposit; it represents 20% of the succession. This facies is composed of reddish-brown to orange, fine- to medium (rarely up to very-coarse) quartz-arenite; it is moderately to well-sorted; grains are sub-rounded. It occurs as sets that range in thickness from 0.1 to 1 m and that are typically several metres in lateral extent (up to 57 m). Internally, sets are composed of wavy and crinkly lamination dominated by modified wind-ripples (modification as a result of aqueous action) and adhesion structures (Glennie et al., 1978; Kocurek and Fielder, 1982). The characteristics of the sedimentary structures result in a marked textural contrast between this facies (damp interdune, ID2) and other juxtaposed aeolian facies (AD1, AD2, AD3) and interdune facies ID1. Toward the top of some sets of facies ID2, pseudo-lamination (Glennie, 1970, I972) is present and minor convolution may be present in the form of small, 30 to 40 mm-high flame structures (cf. Collinson,1994). This facies also exhibits several examples of down-folding structures up to 10 cm in width and 5 cm in depth (Figure 4.11, and Figure 4.13). The strata of some sets show gradual upward change from crinkly type lamination to more flat wind-ripple dominated strata (Figure 4.13).

Interpretation: This facies records deposition under the influence of a short- or long-lived, elevated or fluctuating water table (Langford, 1989; Ahlbrandt and Fryberger, 1981) that was at or close to the accumulation surface such that its capillary fringe met and interacted with the surface (cf. Mountney, 2006a). Interdune strata record accumulation in the depressions between aeolian dunes, and the variation of water table level plays a significant role in the nature of interdune sedimentary processes, such that dry, damp and wet interdune facies may be recognised. Facies ID2 represents a damp or wet interdune setting, whereas facies ID1 represents a dry interdune or lower dune plinth setting (cf. Mountney and Jagger, 2004). The adhesion structures observed in this lithofacies indicate deposition on a damp surface, where dry, wind-blown sediment migrated across a wet or damp surface and adhered to it (Figure 4.13; Glennie et al., 1978; Clemmensen, 1979; Kocurek, and Fielder, 1982). The irregular appearance of the strata that comprise this facies internally may reflect the repeated occurrence of local deflation and deposition processes, which is also a record of moisture and wind intensity variations over time, whereby deflation was locally limited by the water-table level (Simpson and Loope, 1985). The small-scale flame structures in this facies are a type of soft-sediment deformation related to fluid escape (Collinson, 1994). This structure indicates the presence of a damp interdune surface during the advance of an aeolian dune bedform, whereby the loading effect of the advancing dune squeezes moisture out of the interdune sediments in front of it, thereby resulting in deformation due to water escape (Doe and Dott, 1980; Horowitz, 1982). The down-folded structures are interpreted as vertebrate indentation marks in the damp interdunes; they are poorly defined animal footprint trackways and are common in many damp interdune settings (e.g. Lewis and Titheridge, 1978; Pollard, 1981; Loope, 1986; 2006; Allen, 1989, Rodriguez-Lopez et al.,

Facies ID2: Crinkly laminated sandstone



Facies Information

Colour	Brown, whitish grey.
Grain Size	Fine to very-coarse
Set Thickness	10 cm to 1 m sets
Sorting & Texture	moderate to well sorted Subrounded
Facies Association	Aeolian Interdune
Architectural Elements	A2

Facies Characteristics

1. Wavy or crinkly lamination
2. Adhesion ripple pseudo-lamination
3. Small-scale flame structures
4. Modified wind-ripple lamination
5. Area of fluid bleaching
6. Textural contrast between interdune and overlying dune toesets
7. Vertebrate footprint
8. Iron concretions

Interpretation

Wavy and crinkly laminated sandstone, dominated by modified wind-ripple and adhesion-ripple strata; small flame structures. This facies accumulated in interdune areas between migrating aeolian dune forms. Damp interdune units record the nature of the substrate within an aeolian dune field. The gradual upward change to dry interdune dominated by translatent wind-ripple strata records a passages to drier edge of interdune. The down-foldded structures are of vertebrate footprints; they are common in the Triassic Wilmslow Sandstone and Helsby Sandstone.

Figure 4.13: Aeolian interdune, wavy to crinkly laminated sandstone lithofacies (ID2).

2012). Similar features are common elsewhere in the Triassic Wilmslow Sandstone and Helsby Sandstone formations of the Cheshire Basin (Tresise, 1993; 1994; Tresise and Sarjeant, 1997, King and Thompson, 2000). The gradual upward change to dry depositional units dominated by wind-ripple lamination, either dry interdune (ID1) or dune plinths (AD3), are indications of the decrease in the relative level of the water table (Figure 4.13; Kocurek, 1981).

4.6.2 Aeolian facies association

From the interpretation of preserved characteristics, the various depositional facies seen in the Wilmslow Sandstone Formation exposed in the lower part of the Runcorn Expressway road-cut section (Figures 4.3, Figure, 4.14, and Figure 4.15) can be assigned to a mix of arid to semi-arid aeolian depositional settings. Collectively, this mixed facies association is interpreted to represent the preserved deposits of a water-table influenced aeolian dune and interdune system (*sensu* Glennie, 1970; Brookfield and Ahlbrandt, 1983; Kocurek, 1991; Cooke et al., 1993; Lancaster, 1995; Mountney and Howell, 2000; Mountney, 2006a).

The aeolian facies association (AD) in this study records the preserved expression of accumulation through the migration of aeolian dunes and associated interdunes. Collectively, the aeolian dune lithofacies (AD1-AD5) discussed above represent the preserved products of downwind and oblique migrating aeolian dunes inferred from measured mean foreset dip azimuth directions which range between 200° and 300° (cf. Mountney and Thompson 2002). The repetitive grainflow processes across the dune slipface of these dunes generated packages of advancing cross strata in facies AD1. The common presence of cyclic reactivation surfaces truncating the avalanches sets indicates periodic localised erosion and reworking of the sediments on dune slipfaces in response to minor changes in aeolian bedform migration direction or lee slope steepness, or dune slope asymmetry (cf. Rubin, 1987). The horizontal aeolian interdune facies (ID1, ID2; (Figure 4.15) likely record deposition in areas between migrating active aeolian dunes (cf. Ahlbrandt and Fryberger, 1981); see modern examples in Al-Masrahy and Mountney (2013, 2015).

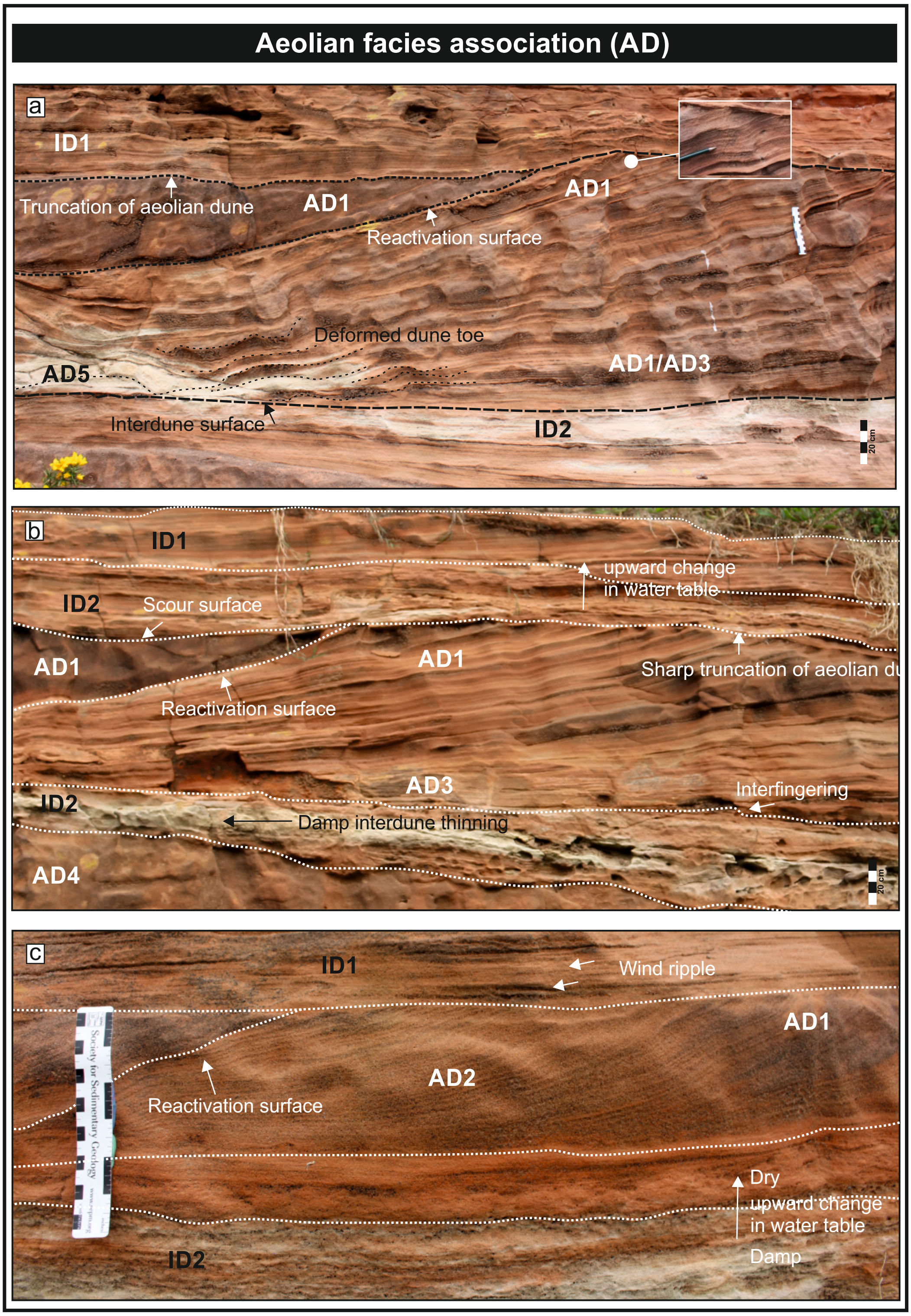


Figure 4.14: Aeolian facies association (AD). a) example of aeolian dune soft-sediment deformation at the plinth of aeolian dune; b) example of truncation surface observed in the aeolian dune sets, it also demonstrate evidence of upward change in water table; c) example of aeolian dune set dominated by grainfall and grain flow strata, truncated by wind ripples strata form dry interdune set, the lower part is showing evidence of drying up resulted from a change in water table.

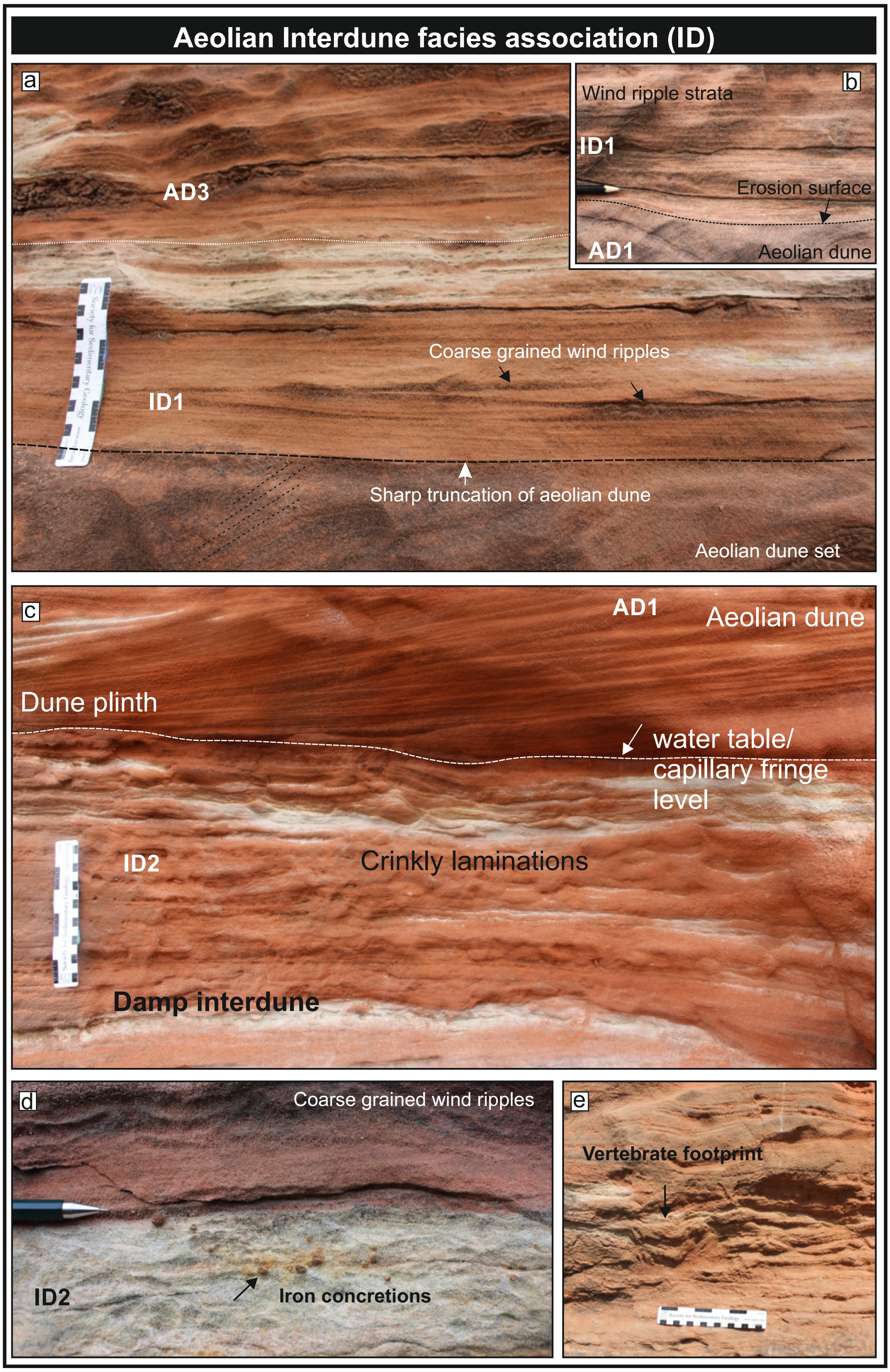


Figure 4.15: Aeolian interdune facies association (ID). a and b) example of aeolian dune truncated by dry interdune surface; c) example of aeolian dune set overlying but also interfingering with the uppermost deposits of the underlying damp interdune unit; d) example or iron concretions observed in damp interdune; e) example of vertebrate indenter mark observed at the top of damp interdune surface.

The simplest explanation is that these interdune accumulations separate accumulations of aeolian dune facies and therefore the lower bounding surface that defines the base of these interdune units represents an interdune migration surface that separates the accumulations of what would have once been laterally adjacent aeolian dune bodies (cf. Brookfield, 1977; Kocurek, 1981a, b, 1996; Ahlbrandt and Fryberger, 1981; Fryberger, 1993). These damp or dry interdune areas formed hollows or corridors between the aeolian dunes. Such interdune configuration could potentially be exploited by fluvial systems if open to the edge of the dune-field (see Chapter 3; cf. Cain and Mountney, 2011; Al-Masrahy and Mountney, 2015). In the case of closed, elliptical hollows between dune forms, interdunes may have collected local run-off from rain water and sedimentation would therefore have been associated with rainfall events within the dune-field (erg) system (Brookfield and Ahlbrandt, 1983; Svendsen et al., 2003; Pye and Tsoar, 2009).

The shift between the interdune and dune facies is represented by a gradual upward transition from the wavy to crinkly laminations of the damp interdune into wind rippled dominated strata of the dry interdune (ID1) and overlying aeolian dune toeset (AD3), and up into grainflow dominated strata sets of the dune lee slope (AD1). This gradational change demonstrates synchronous development and emplacement of the damp interdune, the dry interdune and adjacent dune facies (Pulvertaft, 1985; Mountney and Thompson, 2002; Mountney, 2006b). This relationship is most obviously explained by an interdune flat in which the central part of a slightly topographically depressed intedune hollow was closer to the water table, whereas the fringes of the interdune were slightly topographically higher and drier, especially where they passed into the plinth areas of any adjoining dunes.

Implicit in the interpretation provided above is the notion that the aeolian dunes and their adjoining interdunes migrated and accumulated over earlier deposits via bedform climbing whereby a succeeding dune migrated over a preceding interdune surface (cf. Mountney and Jagger, 2004). This general interpretation is only one of the possible models of aeolian dune and interdune interactions. A second possible interpretation is that sedimentation on damp or wet interdune flats was not on-going during the dune advance; rather, aeolian dune bedforms migrated across damp or wet interdune flats where angle of climb is fluctuating around zero, so-called non-climbing model (cf. Langford and Chan, 1988; Simpson and Loope, 1985, Mountney and Thompson, 2002; Jagger and Mountney, 2004). Interaction between aeolian dunes and adjoining interdunes is discussed further in the discussion part of this chapter (Section 4.7.2).

4.6.3 Aeolian architectural elements

Two aeolian architectural elements are identified and these are illustrated in three-dimensional models (Figures 4.16, and Figure 4.17): aeolian dune element A1 and aeolian interdune element A2, which comprise 75% and 25% of the studied section of the Wilmslow Sandstone Formation, respectively. These geometrical bodies stack together to form compound assemblages of multiple stacked elements that themselves extend laterally for 170 m as element groups or compound packages of strata that are up to ~8 m thick (Figures 4.16, and Figure 4.17).

4.6.3.1 Aeolian dune architectural elements (A1)

Description: The complex architecture of the aeolian dune elements within the outcrop is shown in Figure 4.4. The facies associated with aeolian dune architectural elements include the following: moderate- to high-angle grainflow strata (AD1); moderate- to high-angle grainflall strata (AD2), low to moderate angle wind-ripple strata facies (AD3), massive aeolian sandstone (AD4) and soft sediment deformed facies (AD5). Aeolian dune elements are the most prevalent element in the aeolian succession. Single elements laterally extend between 3 and 8 m, and reach a maximum thickness of 1.5 m. Each set is cut out by the succeeding set in a downwind direction (Figure 4.4, Panels A-F4). Aeolian dune elements occur vertically above interdune elements and together these two element types attain a maximum thickness of 8 m. Cosets of strata that form stacked aeolian dune elements rarely exceed 4 m, with cosets delimited by interdune migration bounding surfaces. Palaeocurrent measurements from facies that comprise A1 elements indicate an overall south-westward dune migration direction, inferred from

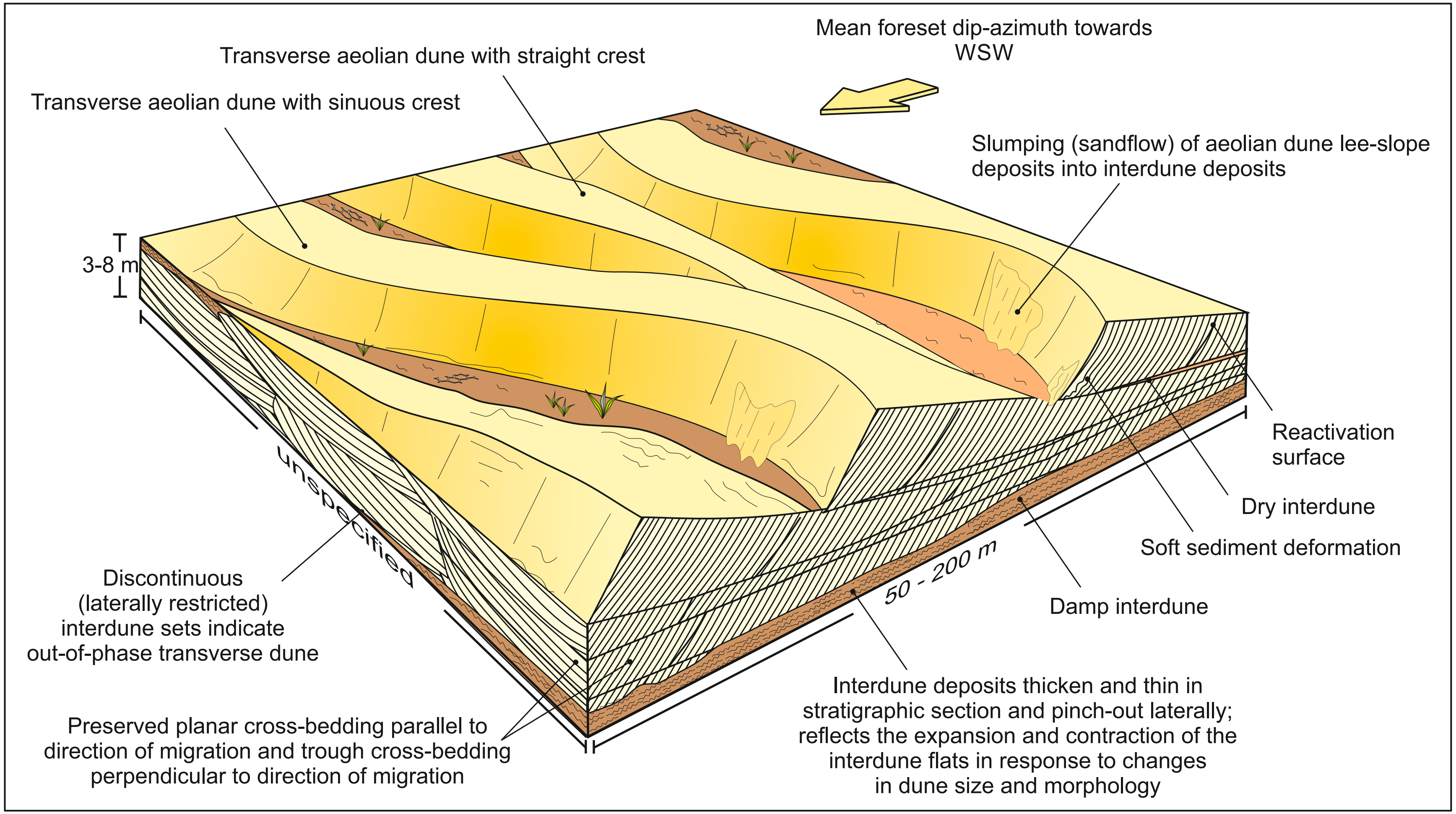


Figure 4.16: Three-dimensional summary model depicting the aeolian dune element A1 and associated aeolian interdune element (A2).

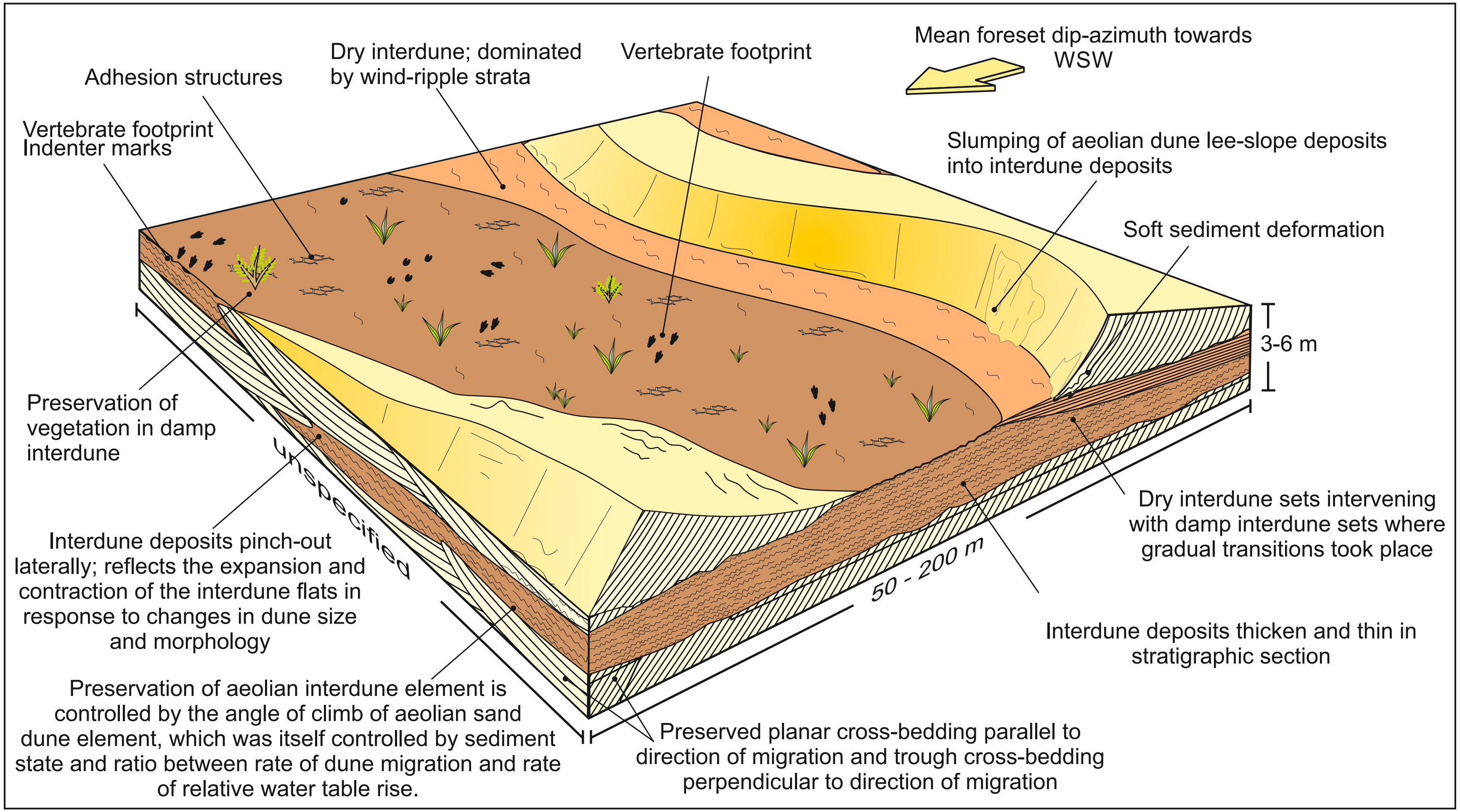


Figure 4.17: Three-dimensional summary model depicting the aeolian interdune elements A2 and associated aeolian dune elements (A1).

measured mean foreset dip azimuth directions; mean = 255°, angular deviation = 14, n = 130 (Figure 4.16).

Interpretation. The lateral extent of the aeolian elements reflects persistent episodes of aeolian activity. Vertically stacked aeolian dune elements represent deposits of an aeolian dune-field system for which conditions were favourable for bedform construction and accumulation (cf. Glennie, 1970; Lancaster, 1983; Kocurek, 1996, 1999). The observed interbedding of grainflow strata with wind-ripple strata represent accumulation on dry, sloping surfaces of dune lee slopes (Hunter, 1977; Kocurek, 1981; 1991; 1996). The upward transition from less steeply inclined AD3 facies representative of the lower dune plinth region to more steeply inclined AD1 facies representative of the dune lee slope within preserved sets is indicative of the transition from interdune sedimentation to dune sedimentation (Kocurek, 1991; Mountney, 2006a; Mountney and Thompson, 2002). Soft-sediment deformation observed within this element occurs as a result of sand accumulation failure at a high-angle dune slipface, possibly in response to surface precipitation, wet conditions (Allen, 1970; Hunter, 1977; Doe and Dott, 1980; Mountney, 2006a). Mean foreset azimuth direction (Palaeocurrent direction) suggest slipface orientation for aeolian dune element to be consistently to southwest (Figure 4.3), the spread of cross bed orientation throughout the measured sets likely reflect barchanoid transverse to oblique dune forms (Fryberger, 1979; Nurmi, 1985; Nichols, 2009; Mountney and Thompson, 2002; cf. Hunter et al., 1983; Rubin, 1987; Rubin and Careter 2006; Al-Masrahy et al., 2012). The relationship between this element and aeolian element A2 suggests that the preservation of the aeolian dune element occurred within a wet dune-field settings (cf. Mountney and Russell, 2009).

4.6.3.2 Aeolian interdune architectural elements (A2)

Description: Three distinct scales of aeolian interdune element are observed in the studied succession. First, thin (10 to 20 cm) interdune elements occur as lenses (5-15 m in lateral extent) of mainly ID2 facies that occur directly beneath aeolian dune elements A1 (Figure 4.4, Panel A). These examples are laterally restricted and are rare, representing only 1% of the succession. Second, units occur as less than 0.8 m-thick, 10 to 20 m-wide bodies, which represent about 5% of the succession (Figure 4.4, Panels A and B) and are composed internally of facies ID1. Third, units composed of facies ID2 occur that are up to 1 m thick of laterally discontinuous interdune sets, this later scale of interdune element is the dominant interdune scale in the studied outcrop. These element types range in length from 12 to 57 m, and represent 19% of the succession; they pinch out laterally, and examples of this element can typically be traced laterally for 20 to 30 m (Figure 4.4, Panels A-F4). Some instances of this element internally exhibit a gradual upward change from facies 1D2 to ID1 and then pass gradationally upwards into A1 elements. This type of interdune element occurs in close association with aeolian dune element A1 (Figure 4.16). A2 elements occur typically as lense-shaped bodies between aeolian dunes elements (Figure 4.17). The contact boundary between A1 and A2 elements is either sharp with evidence of erosion at the base of A2, or laterally transitional into overlying aeolian dune element A1 (Figure 4.17).

Interpretation: All the three recognised varieties of aeolian interdune elements described above are representative of the variation in the nature of the substrate at the time of accumulation on the interdune surface, which is typically related to the surface moisture conditions (Kocurek, 1981; Kocurek and Nielson, 1986; Pulvertaft, 1985; Mountney, 2006a; Mountney and Cain, 2009). Interdune elements composed of mainly of wind-ripple strata represent dry interdune elements formed during dry conditions and a water table that lay significantly beneath the accumulation surface (Kocurek, 1981; Mountney, 2006a). Irregular (wavy) lamination, with evidence of small-scale flame structures and indenter marks indicate the presence of an elevated water-table (Doe and Dott, 1980; Kocurek, 1981; Horowitz, 1982; Mountney and Thompson, 2002; Kocurek and Fielding, 1982). The gradual upward change from facies ID2 to ID1 and to A1 records a drying up in the sequence, which can most obviously be linked to a change in relative water table (Figure 4, Panels A and B; Kocurek, 1981). Interdunes occur between dunes in most dune-fields (Glennie, 1970; Ahlbrandt and Fryberger, 1981, Talbot, 1985; Mountney, 2006; Masrahy and Mountney, 2015). The lateral pinch-out of these elements reflects the expansion and contraction of the interdune flats in response to changes in dune size and morphology. Some relationships observed between A1 and A2 elements record coeval development and climb of dunes and adjacent interdunes (Figure 4.4, Panel F2 and Figure 4.15b; cf. Pulvertaft, 1985; Loope and Simon, 1992; Mountney and Thompson, 2002). The sharp contact boundary between A1 and A2 elements, which shows evidence of erosion at the base of A2, indicates an interdune erosional surface and overlying lateral transition into overlying aeolian dune element A1. This relationship indicates a coeval (i.e., synchronous) development of two laterally adjoining sub-environments: aeolian interdune and dune (cf. Pulvertaft, 1985; Mountney, 2006b; Mountney and Thompson, 2002). Other relationships record dune migration over an eroded interdune surface and therefore signify a break in deposition whereby the interdune is assigned to one sequence and the overlying dune represents the onset of a later sequence (Figure 4.4, Panel A, and Figure 4.14a and b; Loope, 1985; Jagger and Mountney, 2004). The presence of these elements in the hollow depressions and flat areas between successive aeolian dune elements, and the mechanism of alternation between aeolian dune facies and related aeolian interdune facies all confirm their association with aeolian dune element (A1) (Ahlbrandt and Fryberger, 1981; Mountney and Thompson, 2002; Mountney, 2012, Al-Masrahy and Mountney, 2013).

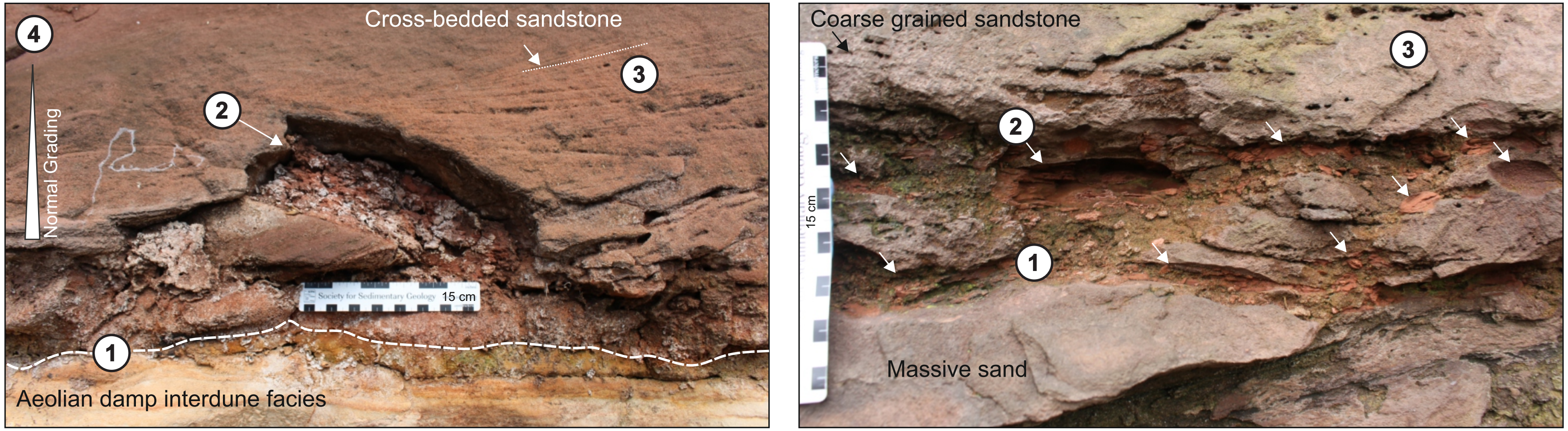
4.6.4 Fluvial facies

Seven fluvial lithofacies have been identified, and collectively these represent the preserved product of deposition from both channelised and non-channelised aqueous flow deposition (Table 1).

4.6.4.1 Intraformational mud clasts facies (FC1)

Description: This facies is composed of reddish-brown, poorly-sorted quartzarenite; grains in the matrix are angular to sub-angular and sub-rounded; intraformational mud clast pebbles are also present. This facies represents 5% of the succession (Figure 4.18). The sandstone matrix exhibits an upward decrease in grain size within sets (fining upward), from coarse or very coarse sandstone to fine sandstone, this lithofacies change in vertical profile from FC1 lithofacies to FC2 lithofacies. The intraformational rip-up clasts are composed of red to reddish-brown laminated (original

Facies FC1: Intraformational mudclast



Facies Information

Colour	Reddish-brown
Grain Size	1 cm to 20 cm
Set Thickness	10 cm to 30 cm sets
Sorting & Texture	Angular to rounded-bladed poorly sorted
Facies Association	Fluvial Channel
Architectural Elements	F1

Facies Characteristics

1. Channel base; erosional bounding surface
2. Intraformational rip-up clasts composed of red laminated (original depositional structure) silty mudstone. Clasts are of highly variable size and shape, generally randomly oriented indicating very short transport distance and rapid deposition
3. Channel-fill deposits
4. Fining-up sequence (normal grading)
5. Vertical abrupt change from facies FC1 to FC2, indicates rapid changes in flow strength

Interpretation

Intraformational mud clasts preserved as basal lag deposits associated with a fluvial channel-fill element. The channel base exhibits irregular relief; presence of locally derived mud clasts record the erosion of a muddy floodplain substrate by erosional flows. The great variation of clast sizes and shapes indicates a local source for the derived materials and a short transport distance. The absence of order and structure in the lag deposits indicates abrupt deposition from a rapidly waning current.

Figure 4.18: Fluvial channel, interaformational mudclast lithofacies (FC1).

depositional structure) silty mudstone. Clasts are of highly variable size and shape; they attain lengths that range from 1 to 25 cm and are especially abundant in the lower parts of sets of this facies. Sets range in thicknesses from 0.1 to 1.0 m. This facies is closely associated with other channel-fill facies (FC2 and FC3). Sets of this facies most commonly occur directly above prominent erosional bounding surfaces and some examples of this facies fill erosional scours present in the upper parts of previously deposited facies of aeolian and fluvial origin, elements, F1 and F2, (Figure 4.4, Panels B-F, see section 4.6.6.1, and 4.6.6.2).

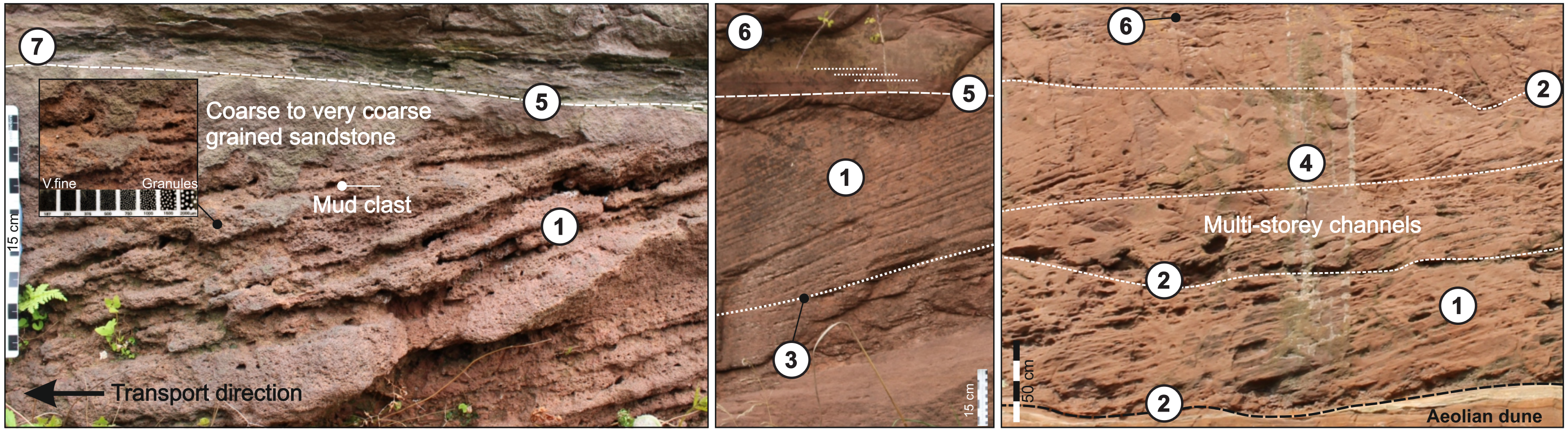
Interpretation: This intraformational mud-clast lithofacies is preserved principally as a channel lag deposit associated with fluvial channel-fill elements (Collinson and Lewin, 1983; Miall, 1996; Selley, 1996). This facies directly overlies the erosional bounding surface that defines the fluvial channel base (Miall, 1978, 1996; Steel and Thompson, 1983). The mud clasts in this facies most-likely represent the localised reworking of fluvial floodplain deposits, or possibly late-stage channel-fill deposits (Miall, 1996; Collinson, 1996; Medici et al., 2015). The presence of the large cobble and boulder-size mud clasts (up to 25 cm in diameter) embedded in a sandstone matrix likely indicates introduction by collapse of unstable channel banks. The angular nature of clasts of this friable material signify a very short transport distance and rapid deposition, probably from a rapidly waning current, an interpretation supported by the associated absence of order and structure in the lag deposits and the overall fining upward trend within sets (Mader, 1985, Jones et al., 2001; Bridge, 2003; Collinson et al., 2006; Scasso, et al., 2012). Facies similar to this have been shown to originate via erosion of fine-grained sediments of aeolian wet interdune origin during major floods (Mountney and Howell, 2000; Svendsen et al., 2003; Cain and Mountney, 2009), as well as via erosion of fluvial floodplains (Miall, 1978, 1996; Collinson, 1996; Medici et al., 2015). The observed rapid facies change from facies FC1 to FC2 suggests a rapid decrease in flow velocity and associated decrease in flow competency and sediment transport capacity (Miall, 2010a. Bridge, 2003; also see below).

4.6.4.2 Planar cross-bedded sandstone facies (FC2)

Description: Facies FC2 is composed of reddish-brown to light-brown, fine- to very coarse-grained sandstone that is poorly to moderately sorted, with intraformational mud clasts that occur primarily above a basal a erosional surface to form lag deposits (as in FC1) and also scattered throughout the facies, especially lining many foresets (Figure 4.19, and Figure 4.4, Panels B-F). This facies represents 25% of the succession. Facies FC2 is arranged into planar cross-bedded sets that are 0.2 to 1 m thick. The lower surface of each set can be either sharp but non-erosional or erosional. Internally, cross strata within sets vary from relatively low- to high- angle inclined (10° to 22°) cross-bedded foresets. Reactivation surfaces are observed although are rare or absent in some sets. The cross-bedding of this facies only rarely exhibits asymptotic bottom sets (Fielding, 2006; Tucker, 2011). Inclined cross-bedded foresets have azimuths oriented in a direction toward northwest with mean vector of 295° (n = 60). Multiple individual sets are commonly stacked into cosets, which themselves vary in thickness from 1.5 to 2.5 m (coset thickness decreases upwards through the succession (see 4.4, Panel C2).

Interpretation: This planar cross-bedded sandstone facies records the migration, accumulation and preservation of subaqueous mesoforms (i.e., dune scale bedforms) within a fluvial channel system (Miall, 1996; Collinson et al., 2006; Reesink et al., 2015). The dunes were straight-crested bedforms (Miall, 1977; Cant and Walker, 1978; Cant, 1982; Allen, 1982; Best, 2005; Leclair, 2011; Soltan and Mountney, 2016). The development of such 2D mesoforms indicates deposition during lower flow-regime conditions that gave rise to a unidirectional palaeoflow, as signified by the development of planar-tabular cross-bedding (Miall, 1996). The style of planar cross-bedding in some sets is tabular whereas in a small number of sets it is tangential (asymptotic) to the set base, reflecting deposition during high water stage and an increase in flow strength that produces a reverse flow component toward dune lee slope due to the presence of a strong separation eddy (Collinson et al., 2006, Miall, 2006, Bridge and Demicco, 2008). Sets characterised by sharp boundaries at their top record an abrupt abandonment phase. The preservation of only toesets and foresets of the

Facies FC2: Planar cross-bedded sandstone



Facies Information

Colour	Reddish-brown- light brown
Grain Size	Fine to very-course
Set Thickness	20 cm to 1 m sets- 2.5 m stacked cosets
Sorting & Texture	moderate-poorly sorted Sub-rounded
Facies Association	Fluvial Channel
Architectural Elements	F1

Facies Characteristics

1. Planar cross-bedded sandstone
2. Channel element defined by basal erosional surface
3. Fluvial reactivation surface
4. Cross bedding: asymptotic foresets and toesets
5. Sharp facies boundary
6. Upper-stage plane beds
7. Abrupt channel abandonment

Interpretation

Planar cross-beds of subaqueous origin, representing the migration of sandy mesoforms within a fluvial channel system (confined flow) deposited by straight crested dune-scale bedforms. This facies is further defined by planar-tabular sets that are tangential to the set base and which commonly are characterized by very coarse to gravel grade lag deposits of locally derived materials. Sharp boundaries at the top of the channel-fill element record an abrupt abandonment phase.

Figure 4.19: Fluvial channel, planar cross-bedded sandstone lithofacies (FC2).

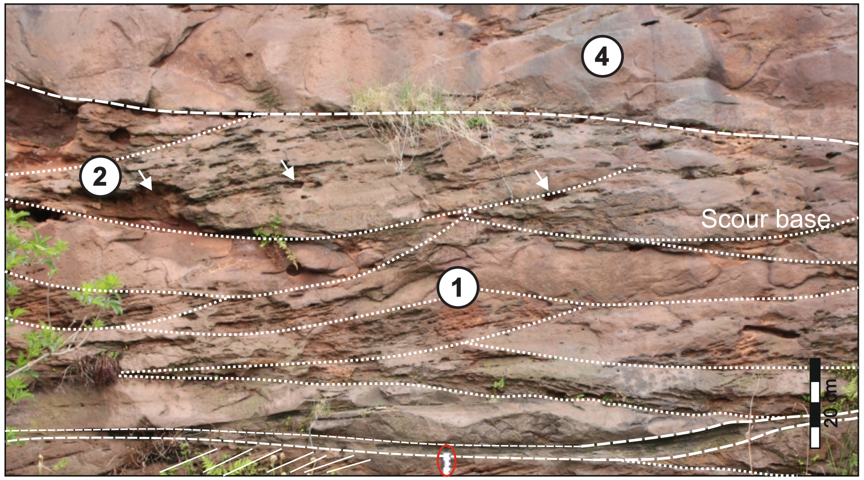
mesoforms indicates a subcritical angle of bedform climb, where the preservation of dune stoss-side deposits of the bedform did not occur (Rubin, 1987, Collinson et al., 2006). The reactivation surfaces observed in some sets (Figure 4.19) likely indicate fluctuations in the palaeoflow direction or current speed or intensity (Miall, 2010b), or flow-stage oscillations that varied between low and high stage (Rust and Jones, 1987; Collinson, 1970; Ashworth et al., 2011). This fluctuation of the flow may also account for the presence of the scattered intraformational mud clasts in the cross-bedded sets.

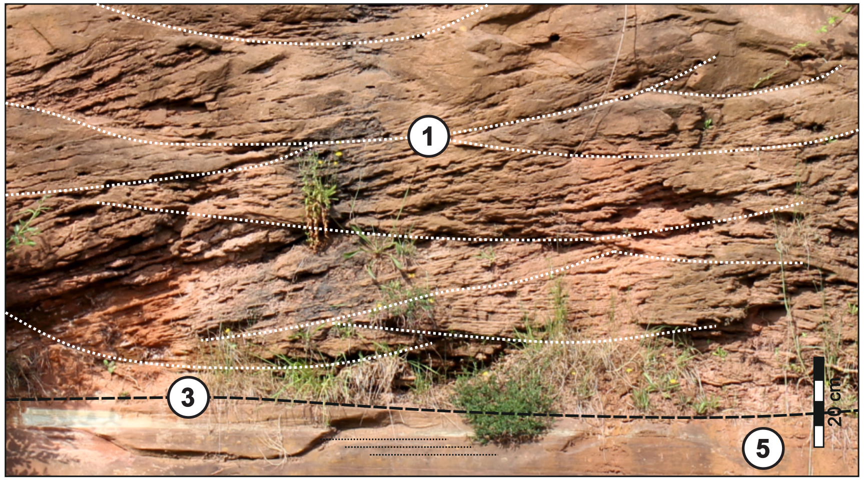
4.6.4.3 Trough cross-bedded sandstone facies (FC3)

Description: Facies FC3 is a trough cross-bedded sandstone, composed of reddish-brown to light-brown, fine- to very coarse-grained sandstone; sorting is poor to moderate, with intraformational mud clasts occurring primarily in the basal parts of individual troughs such that they form lag deposits, though they also occur scattered throughout the facies in some instances (Figure 4.20, and Figure 4.4, Panels F4-5). This facies occurs in sets that are typically 0.1 to 0.3 m thick, and which stack together to form cosets that are each up to 2 m thick. Troughs within the cross-bedded sets vary in width from < 1 m to ~2 m, the direction of palaeoflow is indicated by the orientation of the trough axes towards northwest with mean vector of 304° (n=10). This facies represents 15% of the succession.

Interpretation: These trough cross-bedded deposits are of subaqueous origin and represent the migration and accumulation of sandy mesoforms within a fluvial channel system. The sets represent the preserved deposits of sinuous-crested (i.e., three-dimensional) dune-scale bedforms (Miall, 1977; Allen, 1982; Rubin, 1987; Collinson et al., 2006; Banham and Mountney, 2014), the development of which is typical of channels that experienced turbulent flows (Allen, 1982; Collinson et al., 2006), under low to moderate flow regimes (Miall, 1996; Stikes, 2007; Nichols, 2009). Sharp boundaries at the bottom of the channel-fill element records a scouring and eroding of the underlying substrate (channel base) by flow circulation in front of the migrating bedforms (Friend et al., 1979; Jones, 2002; Gibling, 2006). The presence of intraformational rip-up mud clasts on the bases of troughs

Facies FC3: Trough cross-bedded sandstone





Facies Information

Colour	Reddish-brown- light brown
Grain Size	Fine to very-coarse
Set Thickness	10 cm to 30 cm sets 1 to 2m cosets.
Sorting & Texture	Moderate to poorly sorted sub-rounded to angular
Facies Association	Fluvial Channel
Architectural Elements	F2

Facies Characteristics

1. Fluvial trough cross-beds
2. Mud clasts (lag deposits)
3. Sharp boundary, representing transition from upper- to lower-flow regimes.
4. Faint lamination; massive sandstone
5. Flat laminated sandstone, waning flow stage

Interpretation

Trough cross-beds of subaqueous origin, representing the migration and accumulation of sandy mesoforms within a fluvial channel system (confined flow) deposited by sinuous-crested dune-scale bedforms. This facies is also characterised by very coarse-to gravel grade lag deposits of locally derived materials (mud clasts). Sharp boundaries at the bottom of the channel-fill element record an flash-flood deposits directly overlying older channel abandonment facies.

Figure 4.20: Fluvial channel, trough cross-bedded sandstone lithofacies (FC3).

indicates the re-working and entrainment of previously deposited sediments during periods of high-energy flow (DeCelles et al., 1983; Rubin, 1987).

4.6.4.4 Massive (structureless) sandstone facies (FC4)

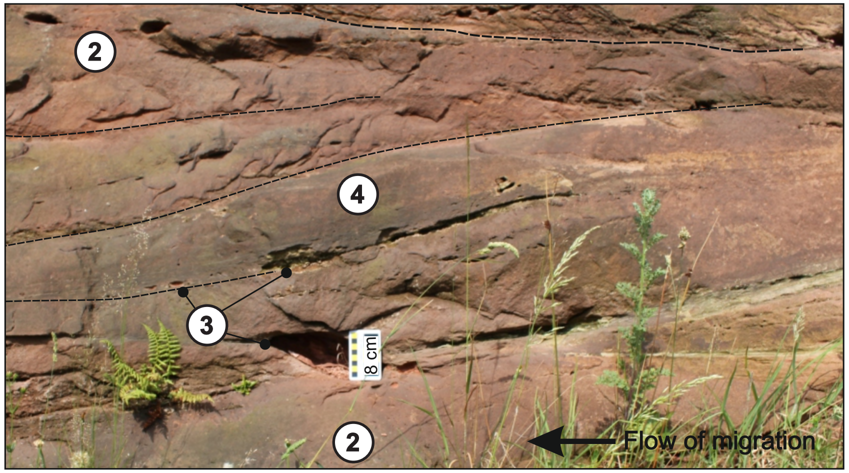
Description: This well-lithified, brown to light-brown coloured sandstone is devoid of prominent internal sedimentary structures. This facies occurs in many places across the studied outcrop (Figure 4.4, Panels B-F). This lithofacies represents 35% of the succession. It is characterised by fine to medium, moderately to well-sorted sandstone with sub-rounded to sub-angular grains but is structureless. This facies occurs in sets that are 0.1 to 3 m thick. Although massive (i.e. structureless) in the majority of occurrences, rarely a very faint and patchy lamination may be just discernable in small areas of some sets. This facies commonly occurs above erosional surfaces. Intraformational mud clasts up to 15 cm in length are observed in this facies, though are rare (Figure 4.21, and Figure 4.4, Panel F6).

Interpretation: The predominantly massive nature of this lithofacies is indicative of rapid deposition of sediment from hyper-concentrated or gravity-driven, possibly during single or multiple flood events (Olsen, 1987; Miall, 1996; Svendsen et al., 2003; Collinson et al., 2006). Hyperconcentrated flows are common in semi-arid ephemeral fluvial systems and stream subject to intermittent flow (Svendsen et al., 2003) where long periods between floods or fluvial activity (e.g. during drought episodes) promotes the accumulation of non-cohesive, sand-rich aeolian sediment in fluvial pathways that is prone to rapid fluvial reworking during high-magnitude flood events (Glennie, 1970; Al-Masrahy and Mountney, 2015). The lack of clear internal structure may also attributed to both the grain size homogeneity and to rapid deposition in a waning flow. The observed mud clasts reflect the high energy of the fluvial flow that was able to transport such clasts (cf. Miall, 1977).

4.6.4.5 Horizontally laminated sandstone facies (FH)

Description: This facies is a horizontally laminated sandstone, composed of brown to pinkish-brown, very fine- to medium-grained, moderately to well-sorted, sandstone composed of sub-rounded to sub-angular grains. The facies is laminated on an mm scale (1 to 5 mm thick); laminations are

Facies FC4: Massive (Structureless) sandstone



Facies Information

Colour	Brown, light-brown
Grain Size	Fine to medium
Set Thickness	10 cm to 3 m sets
Sorting & Texture	Moderate to well sorted, rounded to sub-rounded
Facies Association	Fluvial Channel
Architectural Elements	F1 and F2

Facies Characteristics

1. Homogeneous stacked sets
2. Absence of internal structure (massively bedded)
3. Mud-clasts lag deposits
4. Sigmoidal foresets

Interpretation

Massive sandstones (lack of organised internal lamination) represent accumulation through rapid deposition during flood events. The lack of organised internal lamination may also reflect the homogeneity of the sandstone grain size.

Figure 4.21: Fluvial channel, massive sandstone lithofacies (FC4).

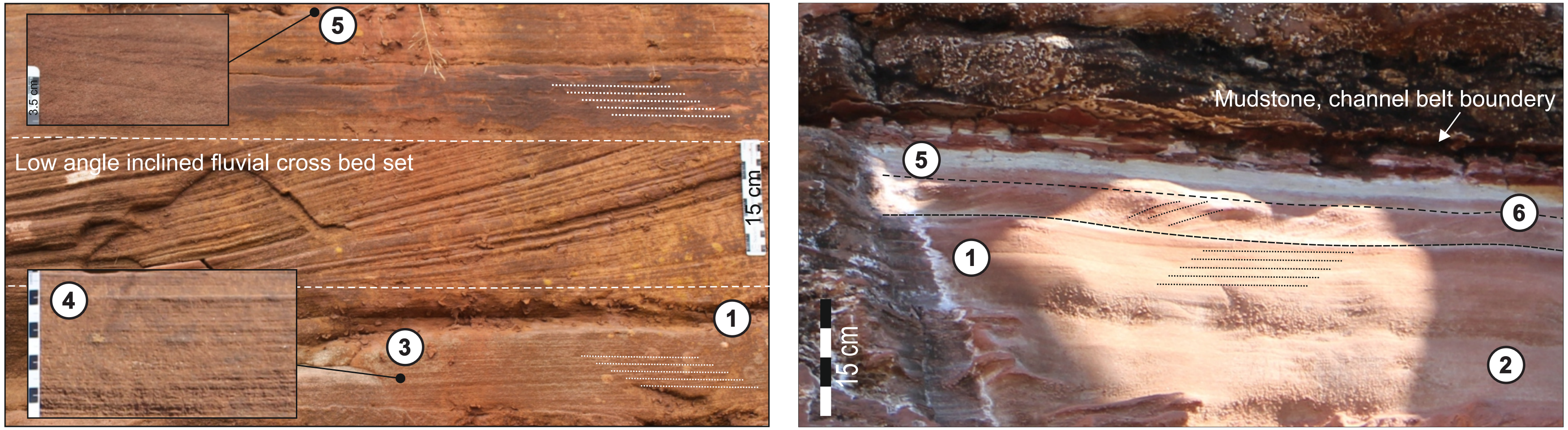
discontinuous in some sets extend laterally between 30 to 100 m. Sets are 0.2 to 0.6 m thick and commonly exhibit normally grading. Some set boundaries are difficult to distinguish due to grain size homogeneity. This facies commonly passes upwards gradationally into current ripple-laminated sandstones facies (FR) within sets (Figure 4.22). No reworked rip-up clasts are present in this facies. This facies represents 12% of the succession.

Interpretation: The fine-grained planar laminated sandstone facies represents deposition under the influence of upper flow regimes conditions (Allen, 1982; Picard and High, 1973; Cheel, 1990, Bridge, 2003), indicated by horizontal parallel lamination. This facies is indicative of either channelised fluvial flow or sheet-like flow (Stear, 1985; Bridge, 2003; Picard and High, 1973; Miall, 1978; 1996), the normal grading and presence of current ripple-lamination at the top or above this lithofacies indicate deposition during decelerating flow; it also evidence of falling stage in channel deposits (Bridge, 2006; Bridge and Demicco, 2008). Grain size homogeneity (though with vague lamination in some units) possibly reflects the localised reworking and re-deposition of a compositionally and texturally mature aeolian sediment during floods (Glennie, 1970; Good and Bryant, 1985; Langford, 1989).

4.6.4.6 Ripple cross-laminated sandstone facies (FR)

Description: This lithofacies is characterised dominantly by cross-laminated sandstone that is light brown-brown to reddish-brown, very fine- to coarse-grained, and which possesses dark mud drapes on some lamination surfaces. This facies represents 3% of the succession. This facies is also characterised in places by ripple-form stratification for which ripples climbed at subcritical angles of 5 to 10°; preserved ripples heights are <3.5 cm. Ripple forms may occur in trains. In places, this facies also rarely exhibits examples of supercritical climbing (Figure 4.23). This facies occurs in the upper parts of sets as packages of ripple cross-lamination and ripple-form strata at are 1 to 30 cm thick. Ripple forms are especially well developed on upper bedding surfaces of sets. The mud drapes, where present, are thin (< 2mm thick). This facies is commonly closely associated with horizontally laminated sandstone facies (FH).

Facies FH: Horizontally laminated sandstone



Facies Information

Colour	Brown, pinkish-brown.
Grain Size	Very-fine to medium
Set Thickness	20 cm to 60 cm sets 1-5 mm lamination
Sorting & Texture	moderate to well sorted sub-rounded- to sub-angular
Facies Association	Fluvial Channel
Architectural Elements	F3

Facies Characteristics

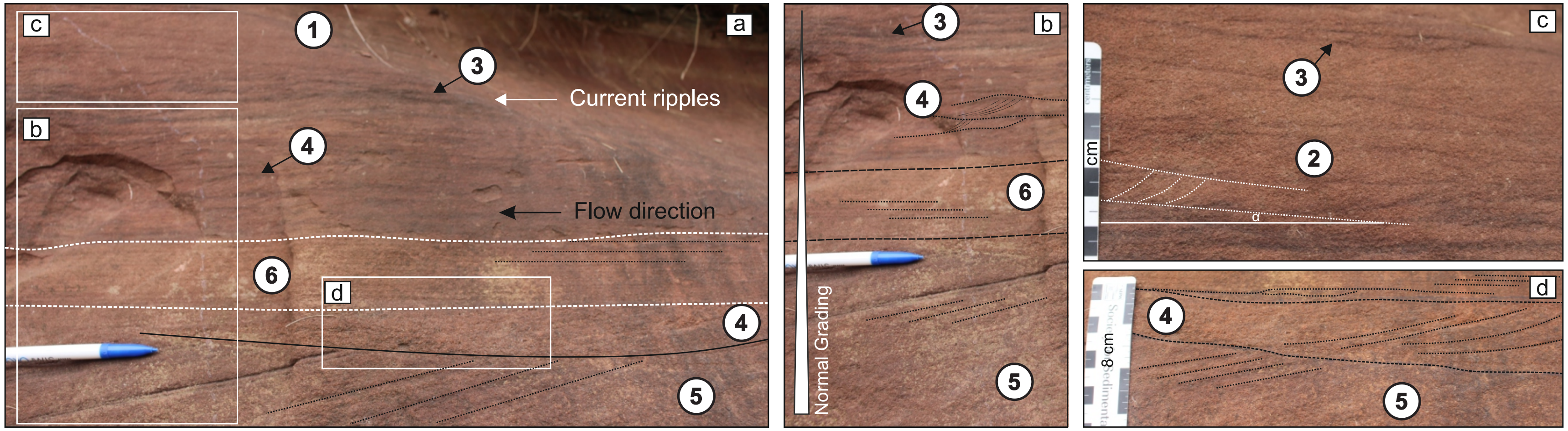
1. Planar (horizontal) laminated sandstone
2. Homogonous grain size; faint flat lamination due to well-sorted nature of grains
3. Bi-modally sorted sandstone
4. Bi-modally sorted sandstone
5. Sets exhibit normal grading, with current (ripple) laminated sandstone toward the top
6. Lower flow regime conditions

Interpretation

Horizontal, discontinuous, thinly laminated sandstone composed mainly of very-fine to fine sandstone grains, with current lamination toward the top. This facies records deposition under the influence of upper flow regimes (waning upward) conditions, either in channelised fluvial flow or sheet-like flows. Grain-size homogeneity (vague lamination) in some units possibly reflects the localised reworking of a compositionally and texturally mature aeolian sediment source.

Figure 4.22: Fluvial channel, horizontally laminated sandstone lithofacies (FH).

Facies FR: Ripple cross laminated sandstone



Facies Information

Colour	Brown to reddish-brown
Grain Size	Very-fine to coarse
Set Thickness	1 cm to 30 cm sets
Sorting & Texture	moderate to poorly sorted Sub-rounded to sub-angular
Facies Association	Fluvial Channel
Architectural Elements	F4

Facies Characteristics

1. Fine sand; low-stage flow; waning-flow mud drapes
2. Subcritical climbing-ripple stratification
3. Mud-drapes
4. Critical to supercritical climbing ripples; indicative of episodes of high sediment flux and waning flow to enable rapid accumulation
5. Inclined fluvial cross-bed set
6. Planar fluvial bed set, showing transition to current ripple lamination

Interpretation

Deposited by sub-aqueous system (fluvial); represents the unidirectional migration of small-scale ripple forms (microforms), during waning flow regimes within the upper part of the fluvial channel-fill or an a non-confined floodplain. Thin sets of current-ripple stratification implies sluggish and shallow flow. Preserved ripple foresets commonly exhibit mud drapes indicating pulsed migration between episodes when the water formed standing ponds, allowing suspension settling of mud fraction.

Figure 4.23: Fluvial channel, ripple cross-laminated sandstone lithofacies (FR).

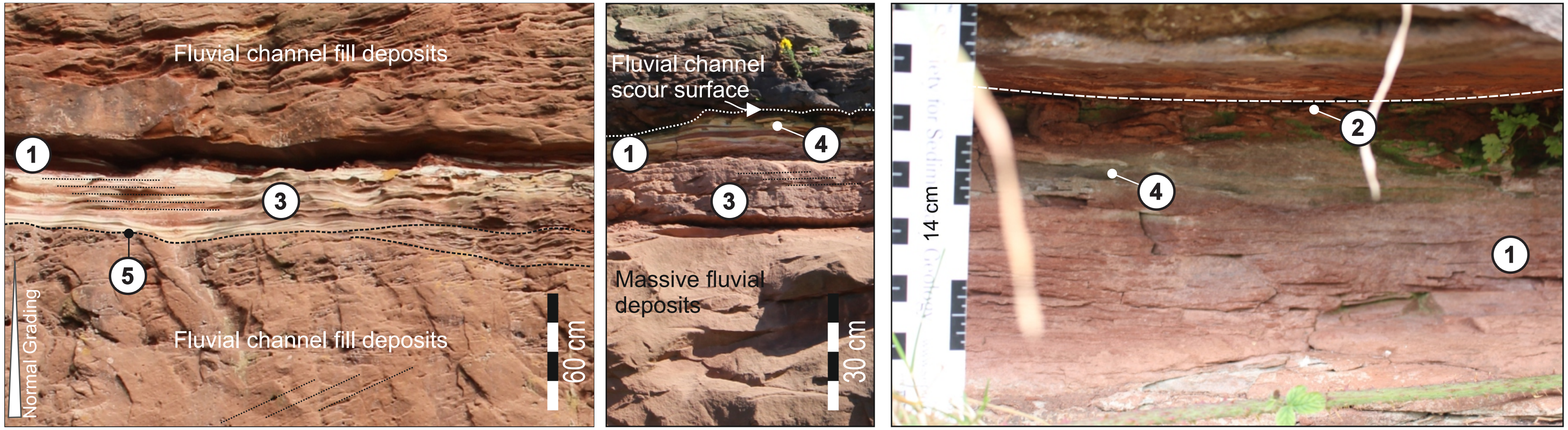
Interpretation: This facies accumulated in subaqueous conditions (fluvial); and represents the unidirectional migration of small-scale ripple forms (microforms) during waning flow regimes (low flow speed) within either the upper part of the fluvial channel-fill or in non-channelised fluvial settings (Stear, 1985). The sets of current-ripple stratification imply sluggish and shallow flow in a waning flow regime (Allen, 1968; Nichols, 2009; Banham and Mountney, 2014). The preserved ripple foresets that possess mud drapes indicate weak pulsing currents possibly moving into area where the water formed standing ponds, allowing suspension settling of the mud fraction (Allen, 1968, 1985; Reineck and Singh, 1980; Stear, 1985; Collinson et al., 2006). The ripple cross-laminated sandstone facies (FR) is commonly associated with horizontally laminated sandstone facies (FH); its occurrence on a horizontally laminated surface (FH) results from turbulent variation in near-surface flow velocity and variation in rate of sediment supply; these variations are also responsible for the differences in the ripple internal structures (Jopling and Walker, 1968; Allen, 1982; 1984; Bridge and Demicco, 2008).

4.6.4.7 Horizontally thin-laminated silty mudstone facies (FM)

Description: This lithofacies is characterised by red-brown (rarely bleached white) silty mudstone that typically occurs as beds of thinly laminated siltstone and massive or thinly laminated mudstone. Beds are commonly laterally extensive, extending for 192 m across the entire section of the fluvial part of the outcrop (Figure 4.4, Panels B-F), although some sets show limited lateral extent (Figure 4.4, Panel F). Grain size is almost entirely of clay and silt grades. This facies occurs in sets that are 0.1 to 0.3 m thick. Desiccation cracks have been observed in similar examples of this facies seen elsewhere in the local area (cf. Mountney and Thompson, 2002), (Figure 4.24). This facies represents 5% of the succession.

Interpretation: This facies represents deposition from very weak traction currents and suspension settling in floodplains or during the final stage of channel filling and abandonment; it records waning flow to form ponds and their subsequent desiccation (Miall, 1977; 1996; Olsen, 1987; Collinson, 1966; Jones et al., 2001; Bourke, 2003; Bridge, 2006; Fisher et al., 2008).

Facies FM: Horizontally thin-laminated silty mudstone



Facies Information

Colour	Red- brown-white
Grain Size	Mudstone to siltstone
Set Thickness	10 cm to 30 cm sets extend laterally 230 m
Sorting & Texture	Moderate to well sorted Sub-rounded
Facies Association	Fluvial non-confined flow
Architectural Elements	F5

Facies Characteristics

1. Planar thinly laminated laterally extensive silty mudstone
2. Top of mudstone horizon contains desiccation cracks
3. Channel-fill waning flow deposits
4. Colour mottling (streaks)
5. Abrupt channel abandonment

Interpretation

Red-brown colour (rarely bleached white); alternations of siltstone and mudstone laminae. Represents deposition from suspension settling in floodplains or during the final stage of channel filling and abandonment. Records waning flow to form ponds and their subsequent desiccation. Colour mottling may reflect weakly developed pedoturbation over relic fluvial bedforms or fluid bleaching.

Figure 4.24: Fluvial channel, horizontally laminated silty mudstone lithofacies (FM).

The presence of this facies may record the uppermost part of a fluvial channel-fill, which is a common feature of idealised channel-fill successions, especially in ephemeral fluvial systems in semi-arid settings (Abdullatif, 1989; Collinson, 2006; Bridge, 2006). Colour mottling may reflect weakly developed pedoturbation over relic fluvial bedforms or fluid bleaching (Wright, 1986). Mudstone lamination indicates the deposition from very slow flow or standing water (Picard and High, 1973; Fisher et al., 2008). This facies is the parent material of the intraformational mud clasts of facies FC1, FC2, FC3 and FC4. Thus, this facies was prone to extensive re-working by succeeding fluvial activity.

4.6.5 Fluvial facies association

The seven recognised lithofacies have been grouped into two associations representative of 1) channelised fluvial deposits (CF), composed of the facies (FC1, FC2, FC3, FC4, FH, FR and FM), and 2) non-channelised sheet-like fluvial deposits (UCF), composed of the facies (FH, FR and FM) (Figures 4.3, and Figure 4.25).

The channelised fluvial facies association (CF) represents the preserved deposits that formed by the accumulation under a variable set of processes that operated in a channelised fluvial system (Miall, 1996; 2014; Bourquin et al., 2009). Successions in Figure 4.25 (a and b) show a typical upward-fining succession, from pebbly lag deposits of intraformational origin at the base of the succession to finer sediment at the top (Figure 4.3, and Figure 4.25), which marks the culmination of confined channel filling (Collinson and Lewin, 1983; Miall, 1978, 1996; Steel and Thompson, 1983). A general upward thinning of bed-set thickness is also noted in channel-fill facies successions (Figure 4.25 a). Collectively, these trends reflect a decrease in the fluvial flow (Miall, 2014). Facies FC1 indicates the rapid-reworking of parts of the preserved adjoining overbank units to generate locally derived rip-up mud clasts, that were carried and moved through the fluvial channel as bed load sediment, and which subsequently were deposited in the lower parts of the channel during the waning stage of the flow to form lag deposits at the base of each fining-up-ward depositional cycle (Miall, 2010a., 1996, Bridge, 2003).

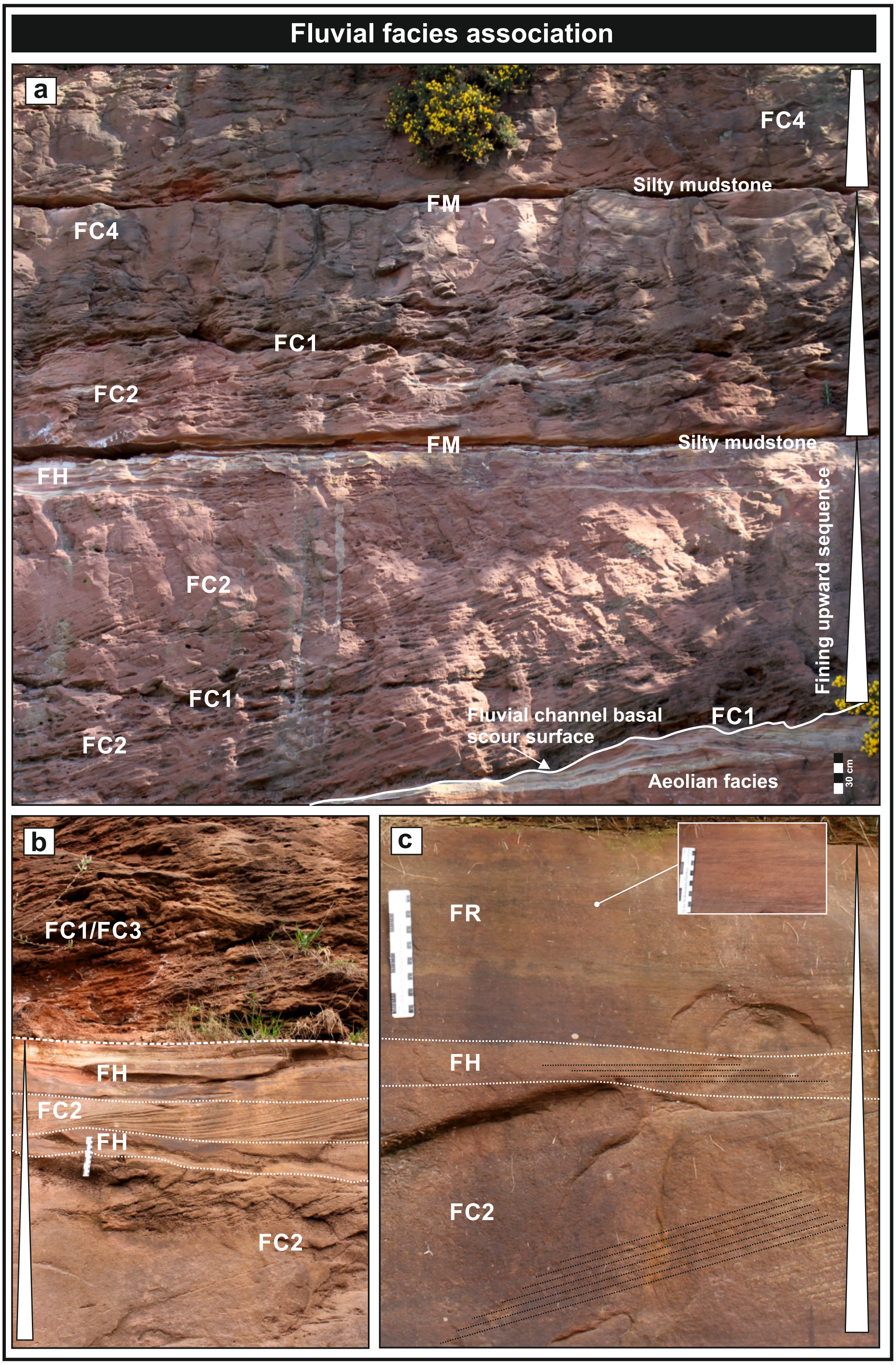


Figure 4.25: Fluvial facies association, Channelised (CF), facies (FC1,FC2, FC3 and FC4), Non-channelised (UCF), facies (FH, FR and FM).

Facies FC2, FC3 and FC4, which range in grain size from very coarse to fine sandstone, represent accumulation of mesoforms and dunes that were migrating downstream within channels under the influence of episodic flow of high-energy-fluvial-discharge during fluvial-flood events (Miall, 1977; Cant and Walker, 1978; Soltan and Mountney, 2016). The horizontally laminated silty mudstone Lithofacies (FM) may reflect final channel filling and abandonment as standing ponds of water left in abandoned channels (Miall, 1977; 1996).

The non-confined facies association (UCF), composed of facies FH, FR and FM, is best observed in the upper parts of fluvial depositional cycles (cf. Cheel, 1990), examples of which are shown in (Figure 4.25). Facies in this association are characterised principally by the absence of the erosional bounding surfaces, a generally finer grain size, sheet-like lamination and, in some cases, small-scale cross bedding (Stear, 1985). Collectively, these characteristics are indicative of deposition under waning-flow conditions in response to channel over-topping and flow in non-confined floodplain overbank settings (Bridge, 2003; 2006; Picard and High, 1973). The presence of desiccation cracks on top surfaces of facies FM (Figure 4.24) in the study area confirms deposition of suspension-load either in fluvial overbank settings or in standing pools of water within a drying-out channel (Collinson, 1996; Miall, 1966; Mountney and Thompson, 2002).

From the previous discussion, the various depositional facies seen in the upper part of the Runcorn expressway road-cut rock exposure (Figure 4.3) can be assigned to a mix of fluvial depositional sub-environments (Figure 4.25). These mixed facies associations are interpreted to be product of deposition via fluvial system sedimentation in both channelised and non-channelised settings (*sensu* Picard and High, 1973; Cant and Walker, 1978, Cant, 1982; Allen, 1985; Miall, 1977, 1996; Mader, 1985; Bridge, 2006; Collinson, 2006; Mountney & Thompson, 2002; Cain and Mountney, 2009).

4.6.6 Fluvial facies architectural elements

Four fluvial architectural elements representative of the preserved remnants of fluvial sub-environments are identified and illustrated in three-dimensional models. The fluvial architectural elements are as follows: fluvial architectural elements representing channelised fluvial flow processes (F1, F2, F3 and F4), and fluvial architectural elements representing non-channelised fluvial processes (F3 and F4) (Figure 4.26, and Figure 4.27).

4.6.6.1 Multi-storey, multi-lateral amalgamated channel elements (F1)

Description: Laterally and vertically amalgamated channel-fill elements are composed of a series of amalgamated channelised strata. F1 elements are 2.5 to 6 m thick and 5 to16 m wide. They have the following geometrical characteristics: (i) this element defined at its base by an erosional surface, commonly with a lag deposits of intraformational clasts (facies, FC1), and overlain by preserved stacked of dune-scale bedforms (mesoforms) to form multiple storeys (Bridge and Mackey 1993; Miall, 1996, 2014; Gibling, 2006); (ii) the element is composed internally of cross-bedded sandstone and some sets some pass gradationally upward into sets of horizontally laminated sandstone overlain by silty mudstone; (iii) this element is also characterized by a sheet-like geometry (Friend et al., 1979; Veiga et al., 2007) (Figure 4.26, and Figure 4.25 a). Multiple instances of these elements are not necessarily fully preserved because the erosional bases of overlying elements have commonly cut-out and removed the upper parts of the underlying elements.

Internally F1 elements are composed of vertically stacked and laterally overlapping storey that are made up of 1.5 to 2.5 m thick cosets that are each themselves composed of 0.2 to 1 m thick sets. Cosets stack together to form up to 6 m of stacked channel deposits, separated by horizontal thin silty mudstone units (FM). Individual storeys can be traced laterally for 5 to 16 m in directions parallel to regional palaeoflow to a point where typically one storey is cut-out laterally by an adjacent storey (Figure 4.4, Panel C). The lower set of the fluvial storey defined by basal erosional surface preserves evidence of scour into facies of aeolian origin at the top of the underlying Wilmslow Sandstone Formation (Figure 4.4, Panels B-F4, and Figure 4.26). The fill of each storey comprises facies FC1 and FC2, with FC2 being the major component of this element. Cross-bedded sandstone sets of facies FC2 may rarely pass gradationally upward into sets of FH, FR or FM in a single channel storey. This arrangement is only rarely observed

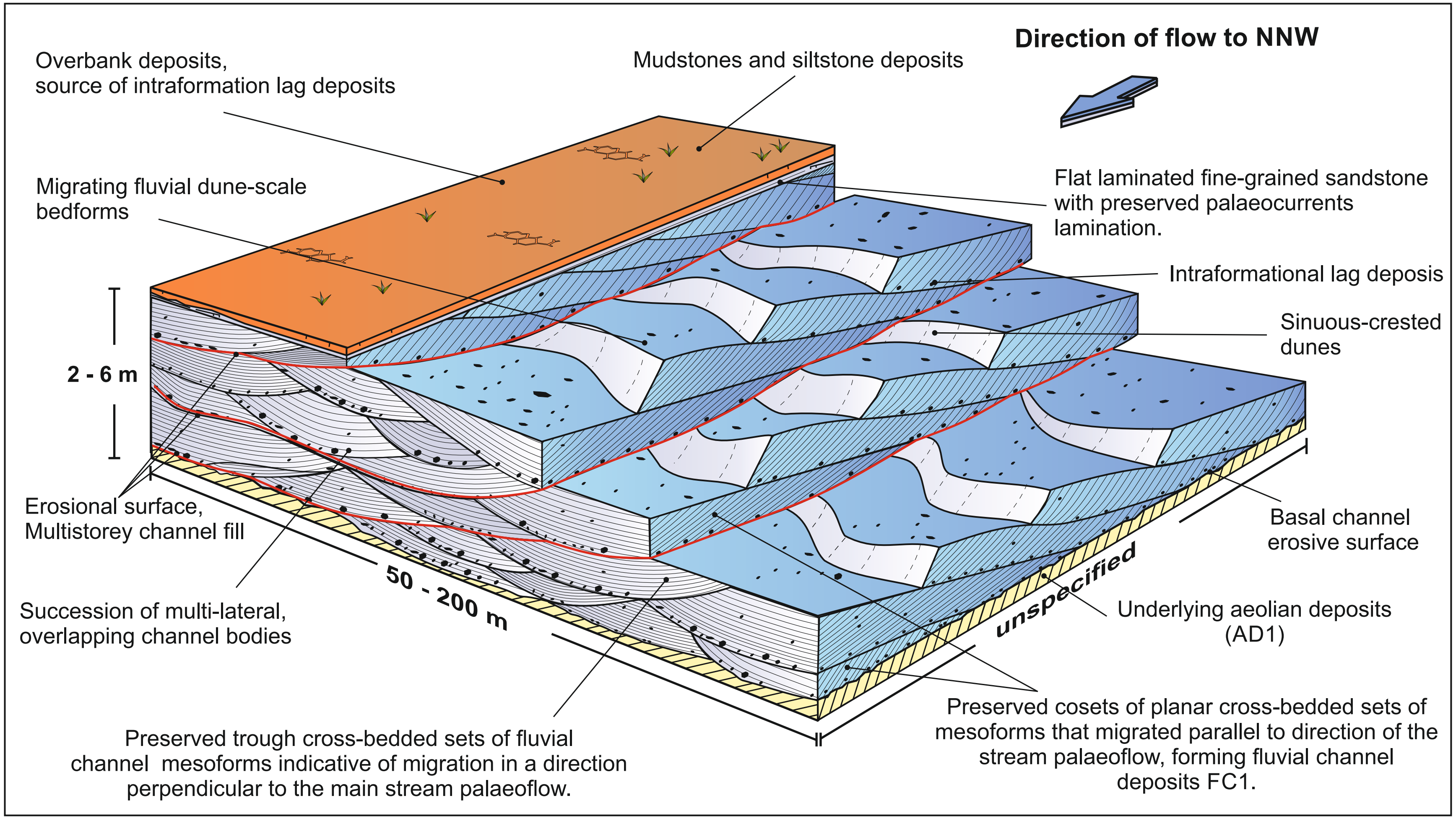


Figure 4.26: Multi-storey, multi-lateral amalgamated channel-fill complex elements (F1).

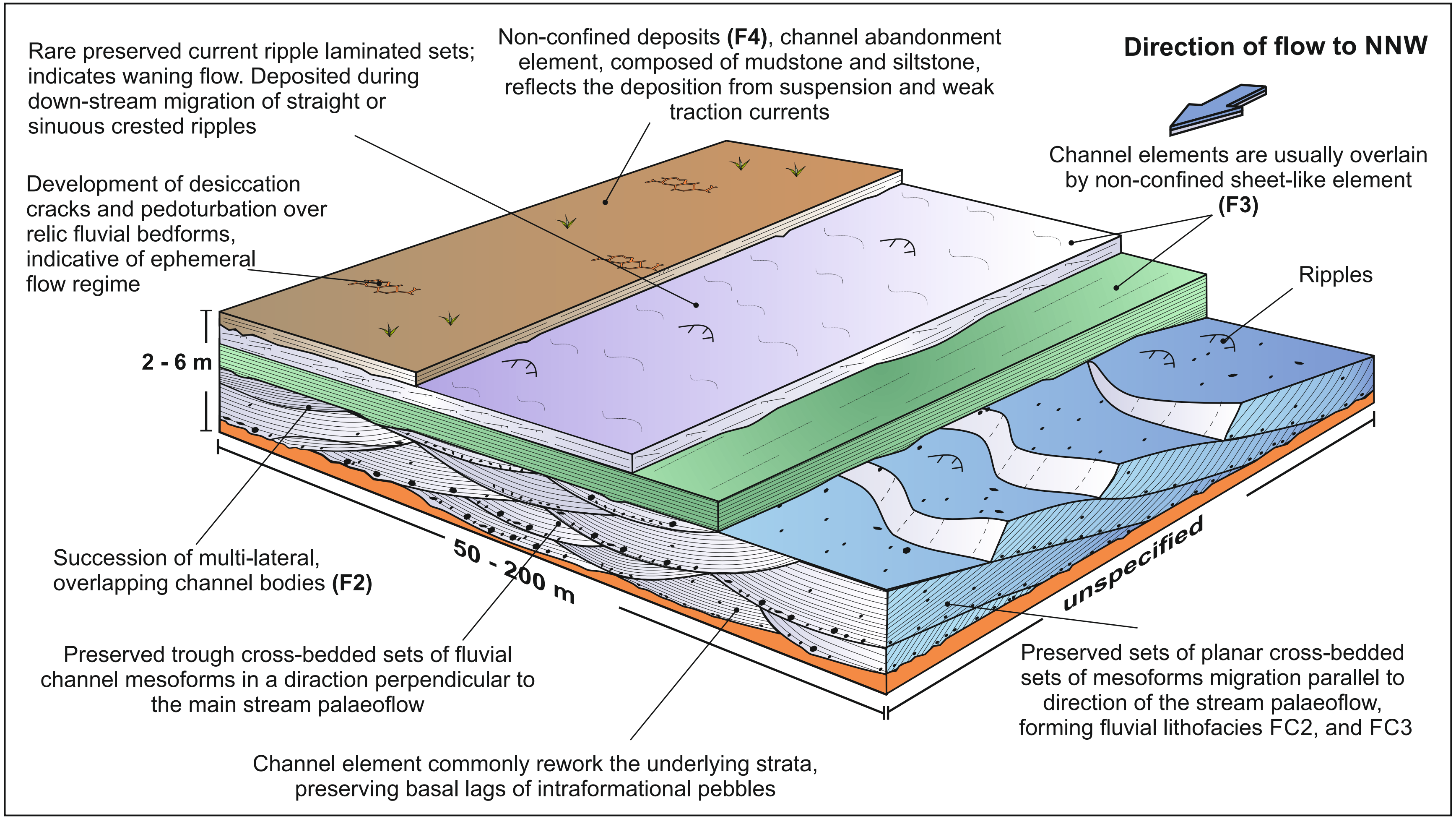


Figure 4.27: Multi-lateral amalgamated channel-fill complex elements (F2), and Non-confined elements (F3, and F4).

because it is more typical for the overlying element to erode down and cut-out (i.e. erode) the upper part of underlying F1 elements. Where the complete facies succession of a well-developed F1 element is preserved, a clear fining-up succession capped by facies FM, is evident (Figure 4.25 a).

Interpretation: Laterally extensive, amalgamated channel-fill elements of type F1 (Figures 4.26) represent the deposits of aggrading braided-channel-belts where intense fluvial erosion processes were active during major flood events (Flores and Pillmore, 1987; Bridge, 1993; Gibling, 2006). The occurrence of the intraformational mud clasts in association with erosively based channel storeys indicates a high velocity flow that is sufficiently powerful to erode the underlying substrate, followed by the transport of the locally derived materials as clasts prior to their deposition as a basal lag during the waning-flow stage (Collinson and Lewin, 1983; Miall, 1978, 1996; Steel and Thompson, 1983; Bridge, 2003; Cain and Mountney 2009). The multi-storey and multilateral nature of these channel-fill elements indicates a long-lived episode of fluvial depositional system activity in a localised area to allow the accumulation of the stacked and overlapping storeys of these elements (Cain and Mountney, 2009).

4.6.6.2 Single-storey, multi-lateral amalgamated channel elements (F2)

Description: Type F2 amalgamated channel-fill elements are composed of a series of laterally amalgamated channelised strata. F2 elements are 2 m thick and it can be traced laterally for 110 m. The have the following geometrical characteristics: (i) The basal surfaces of these element are sharp and erosional in other places with lag deposits of intraformational mud clasts (FC1) toward at the bottom, overlain by sets of inclined planar cross-bedded strata (FC2) or trough cross-bedded strata (FC3); (ii) the upper bounding surface is generally either sharp, cut by the overlaying set of horizontally laminated sandstone which itself overlaying by silty mudstone element (F4) or erosional, upper surface scoured by the overlying single storey element; (iii) this element have an overall sheet-like geometry.

Internally, F2 elements facies successions are encapsulated within a single coset of cross strata that forms a storey, though storeys are stacked laterally adjacent to each other, with their bounding surfaces typically showing local lateral incision into neighbouring storeys (Figure 4.27). Storey are composed of sets of facies FC1, FC2 and FC3. Cosets of this element which represent channel storeys are defined by a basal erosion surface that commonly has locally derived intraformational mudstone clasts (of reworked facies FM) associated with it. These lags are overlain by sets of trough cross-bedded strata (FC3).

Interpretation: This element represents a laterally overlapping and extensive but none aggrading braided fluvial channel-belt accumulation (Figure 4.25 b; Bridge, 1993; Gibling, 2006; Banham and Mountney, 2014). These elements are dominated by bedload transport processes, with trough cross-bedded sets representing the migration of dune-scale mesoforms within the braided channels (Miall, 1977; 1996). The nature of the lateral extent of multi-lateral channel elements arose from repeated avulsion of active and unstable shallow channels at a single stratigraphic surface to form a channel belt (Martinsen et al., 1999; Gibling 2006; Banham and Mountney, 2014). These types of elements and the fluvial systems that generate them are common in dryland systems (Tooth, 2000b, Tooth and Nanson, 2011).

4.6.6.3 Horizontally sheet-like elements (F3)

Description: Type F3 sheet-like elements are composed of lithofacies FH and FR. F2 elements are 0.9 m thick and 30 to 60 m wide, rarely 80 to 100 m, where obscured parts of poorly exposed sections are included (Figure 4.4, Panels B-F). They have the following geometrical characteristics: (i) sharp lower surface; (ii) sharp upper bounding surface overlain by element F4; (iii) this elements have an overall sheet-like geometry (Figure 4.27).

Internally, F3 elements comprise sets of horizontally laminated sandstone up to 0.6 m thick (FH; Figure 4.22). Top of this facies is associated with mud draped current ripple-laminated sandstones (FR).

Interpretation: The presence of facies FH and FR in this element could indicate either channelised fluvial flow or sheet-like, non-confined flow, (Bridge, 2003; Picard and High, 1973; Miall, 1996; Cain and Mountney, 2009; Banham and Mountney, 2014). Current ripple-lamination and associated thin mud films at the top or above this lithofacies are indicative of progressively decreasing current velocity This could be indicative of falling stage in-channel sedimentation processes or waning flow in a post-flood non-confined setting (Picard and High, 1973; Bridge, 2006; Bridge and Demicco, 2008).

4.6.6.4 Channel abandonment elements (F4)

Description: Examples of this element are composed entirely of lithofacies FM, attain a thickness of 10 to 30 cm and extend laterally for at least 192 m, although in places show limited lateral extent 60 to 80 m. F4 elements have the following geometrical characteristics: (i) sharp lower surface; (ii) sharp upper surface; (iii) this elements have an overall sheet-like geometry. The upper surfaces of F4 elements are characterised by desiccation cracks (Figure 4.4, Panels B-F, Figure 4.25 a, and Figure 4.27).

Interpretation: As the flow velocity decreased in the aftermath of flood events, the grain size of the sediment load also decreased until, during the last stage of flow, the suspended sediment load component (mud and silt) settled and was deposited, forming blanket of fine sediment on top of the channel fill. Two possible sub-environments are envisaged for F4 elements: (i) accumulation in shallow, sluggish or stagnant pools, thereby marking the channel abandonment phase; (ii) accumulation in non-confined overbanks floodplain settings. The presence of desiccation cracks on the upper surfaces of these elements is a diagnostic feature of ephemeral fluvial stream activity (Figure 4.27; Picard and High, 1973; Miall, 1977; 1996; Olsen, 1987; Swiecicki et al., 1995; Reid and Frostick, 1997; Bourke, 2003; Bridge, 2006; Fisher et al., 2008).

4.7 Discussion

4.7.1 Aeolian depositional system

Based on the detailed sedimentological analysis of Triassic deposits represented by Wilmslow Sandstone and Helsby Sandstone formations of the Sherwood Sandstone Group in the Runcorn Expressway road cut, two main continental depositional environments have been identified: (i) an aeolian dune-field system and (ii) a braided fluvial system; both formed in an arid to semiarid setting (Figure 4.28).

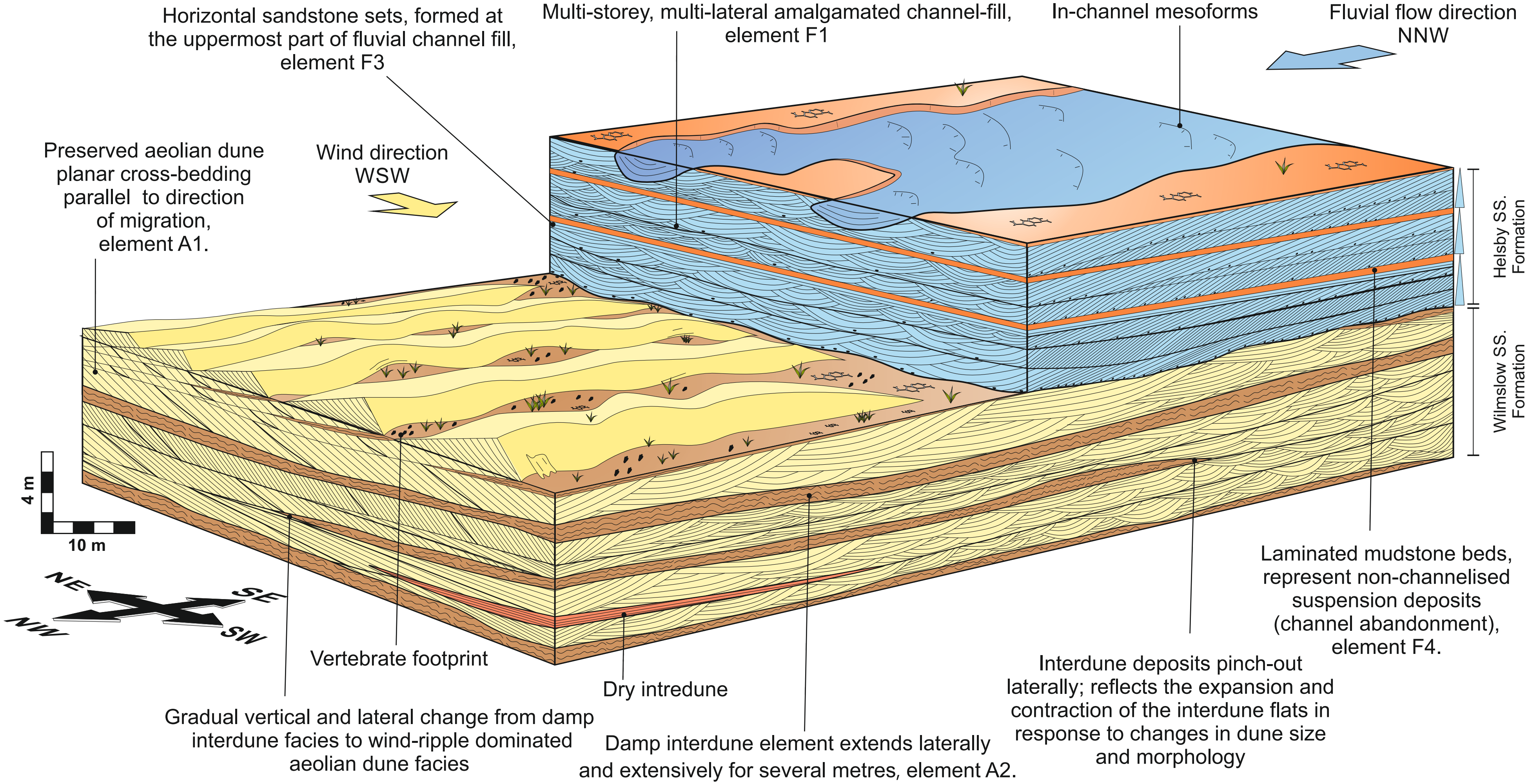


Figure 4.28: Depositional model for the Wilmslow Sandstone and Helsby Sandstone Formations in the studied area.

Although aeolian deposition is not limited to arid settings, it is most common in such settings (Breed et al. 1979; Glennie, 1970). However, even in arid climatic deserts, water remains a significant agent that influences deposition in most aeolian systems (Fryberger, 1990a; Kocurek and Havholm, 1993; Mountney, 2006a), The common association of heat and dryness with modern dune-fields means that it is possible to overlook the many ways that water directly or indirectly impacts the aeolian sedimentation process, including provenance, deposition, burial and diagenesis (Fryberger, 1990b). Water commonly impacts aeolian sedimentation because many dune-fields develop in proximity to standing or ephemeral lakes and streams, lagoons or the ocean. Water erosion processes modifies the surface of stabilised dunes, particularly in more humid regions (Jungerius and Meulen, 1988; Bridge and Rose, 1983; Pay and Tsoar, 2009).

In dry aeolian systems, where aeolian sedimentation is fundamentally controlled by aerodynamic configurations and the water table or its capillary fringe has no effect on sedimentation, migrating aeolian dunes will only commence climbing one over one another after they have been constructed to a size whereby interdune flats have been eliminated (Kocurek and Havholm, 1993). As a result, dry aeolian successions tend to lack thick accumulations of interdune elements, and are generally dominated by aeolian dune elements (Mountney, 2006a). By contrast, in wet aeolian systems, where aeolian accumulation is controlled by both aerodynamic configuration and the presence of moisture on the accumulation surface, which acts to restrict the availability of aeolian grains for transport (Kocurek, 1981a,b, Kocurek, 1996, Mounteny, 2012), aeolian accumulation can occur in response to a rise in the relative level of the water table (Kocurek and Havholm, 1993; Carr-Crabaugh and Kocurek, 1998; Mountney, 2012). The presence of an elevated water table in so-called wet aeolian systems means that the construction of aeolian bedforms in such systems is limited, whereas the potential for the accumulation of thick interdune flats (damp or wet) separating the downwind climbing dune strata is high (Hummel and Kocurek, 1984; Mountney and Jagger, 2004, Mountney, 2006a).

The level of the water table during or after the accumulation of aeolian sediment plays a significant role in the preservation mechanism of the deposited aeolian sediment (Kocurek, 1981a; Kocurek, 1996; Carr-Crabaugh and Kocurek, 1998). In this study area, the preserved aeolian succession is characterised by an abundance of damp interdune elements that separate aeolian dune elements (Figure 4.7). The interaction between the dune forms and the damp interdune strata exhibits an intertonguing (interfingering) and transitional relationship (Figure 4.11, Figure 4.13, and Figure 4.15). These relationships demonstrate that the dunes accumulated synchronously with the interdunes, and represent sub-environments that were laterally adjacent to each other. The interdunes were damp and water-table-influenced within a wet aeolian system for much of their development (cf. Kocurek, 1981a; Pulvertaft, 1985; Simpson and Loope, 1985; Mountney and Thompson, 2002, Mountney, 2006a and b). The water table acted to control the accumulation of this system (Øxnevad, 1991; Herries and Cowan, 1997; Mountney and Thompson, 2002).

4.7.2 Aeolian dune-interdune relationship, controls on dunes and interdune accumulation

Interdune deposits are the key indicators of specific palaeoenvironmental conditions in wet aeolian systems, where the water table lies at or close to the accumulation surface. The development of damp or wet interdunes will act to control the level to which aeolian deflation can occur and, hence, will govern the supply of sediment suitable for aeolian dune construction, and the availability of that sediment for aeolian transport (Kocurek, 1981; Hotta et al., 1984; Granja et al., 2008; Mountney and Russell, 2009; Mountney, 2012).

The accumulation and preservation of damp or wet interdune strata in wet aeolian systems is known to arise via two competing processes: climbing and non-climbing styles of interaction between dune bedforms and interdunes.

Climbing model: In wet aeolian systems, migrating aeolian dunes can accumulate and their deposits can be preserved in response to a gradual rise in water table (Kocurek and Havholm, 1993; Crabaugh and Kocurek, 1993; Carr-Crabaugh and Kocurek, 1998). The ratio between the rate of water table rise and the rate of aeolian dune bedform migration determines the angle of climb of dunes and adjoining interdunes. Changes in this balance between rate of water table rise and rate of aeolian dune bedform migration result in changes to the angle of climb. The expansion and contraction of interdune flats between dune forms are determined by the availability of sediment suitable for aeolian dune construction (Crabaugh and Kocurek, 1993; Kocurek and Havholm, 1993; Mountney, 2012). This process of accumulation and preservation requires high external sediment supply to maintain dune migration and climbing while the water table is rising. In response to changes in sediment availability for dune construction, the size and extent of dunes and adjoining wet or damp interdunes varies. An increase in the rate of water table rise tends to restrict the supply and availability of sediment for aeolian construction, thereby favouring a decrease in dune size and an associated increase in wet interdune size (Kocurek and Havholm, 1993; Kocurek, 1999; Mountney and Thompson, 2002; Mountney and Jagger, 2004; Mountney, 2012).

The Jurassic Entrada Sandstone, which was described by Kocurek (1981a, b), is a well-documented example of a wet aeolian system in which the dune bedforms and adjoining interdunes undertook climb (i.e. net accumulation) in response to contemporaneous dune migration with a progressive and gradual rise in the water table (Figure 4.29a).

Non-climbing model: This model was proposed by Simpson and Loope (1985) and Loope and Simpson (1992) – amongst others – from studies of preserved aeolian dune and interdune strata in the White Sands dune-field, New Mexico, and the Jurassic Wingate Sandstone, Utah. The premise of the model is that preserved sequences of aeolian dune and interdune deposits are primarily controlled by the periodicity of climatic change (an allocylic control) superimposed on a background of slow and possibly punctuated subsidence that generated accommodation (Figure 4.29b). These authors related the damp or wet conditions of the interdunes in modern ergs and the evidence of high water tables in ancient eolian sequences as an indications for low sand supply and small, typically spatially isolated dunes (Figure 4.29b).

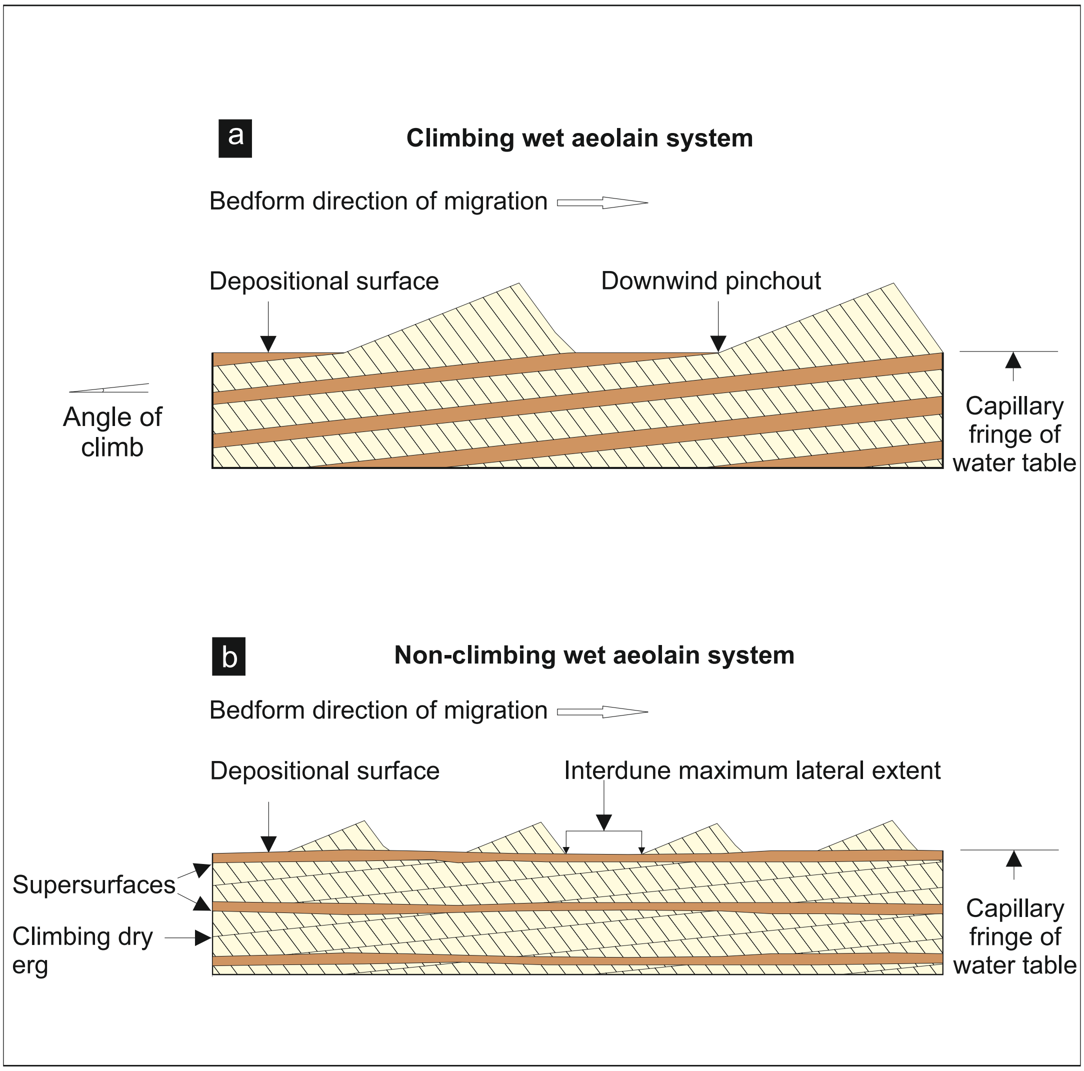


Figure 4.29: Models illustrating the two types of accumulation of aeolian system under the infulance of rising water table (wet system): a) Dune and interdunes climbing under the influence of a steadily rising water table. b) Non-climbing dunes migrating across interdune flats. Modified from (Crabaugh and Kocurek, 1993 and Mountney and Thompson, 2002).

Preservation of aeolian dunes in this model is attributed to the punctuated episodes of rising water table and availability of accommodation space. Hence, sediment bypassing is dominant, and the rock record therefore represents a series of amalgamated wet interdune elements that may be separated from time to time by the fortuitous accumulation of thin sets of aeolian dunes when accommodation and climate allows (Simpson and Loope, 1985). This style of preserved stratigraphic architecture reflects the migration of isolated aeolian dunes across wet interdune flats where the angle of climb was essentially zero for prolonged periods. Thus, preserved successions may contain numerous long-lived bypass supersurfaces (cf. Crabaugh and Kocurek, 1993). However, such surfaces are difficult to identify since they are effectively disastems (paraconformities) that separate bundles (sequences) of similar strata that become vertically stacked over long time periods.

The differentiation between climbing and non-climbing competing models from evidence present in outcrop data is not an easy task, particularly in areas that are tectonically not stable, where tilting of the originally deposited beds is possible, hence, leading to a mis-interpretation of the original accumulation and preservation mechanism (Figure 4.30). From an applied perspective, in hydrocarbon field development, differentiating between climbing and non-climbing damp/wet interdune units is important because it influences how interpretations and correlations of subsurface intervals between wells are undertaken. Climbing interdune units examined in this studied aeolian section are generally laterally restricted (5-20m) (cf. Kocurek, 1981; Mountney and Thompson, 2002; Mountney and Jagger, 2004). For such systems, it is difficult to undertake inter-well correlations when well spacing is more than few hundred metres. By contrast, non-climbing system interdune units generally tend to be more laterally extensive and possibly amalgamated (cf. Simpson and Loope, 1985; Loope and Simpson, 1992). Thus, it may be possible to confidently correlate such non-reservoir units for several hundred metres. Consequently, gaining an understanding of the mechanisms by which wet aeolian systems accumulate and become preserved is important to construct more reliable hydrocarbon depositional models, especially for such complex system. In the studied

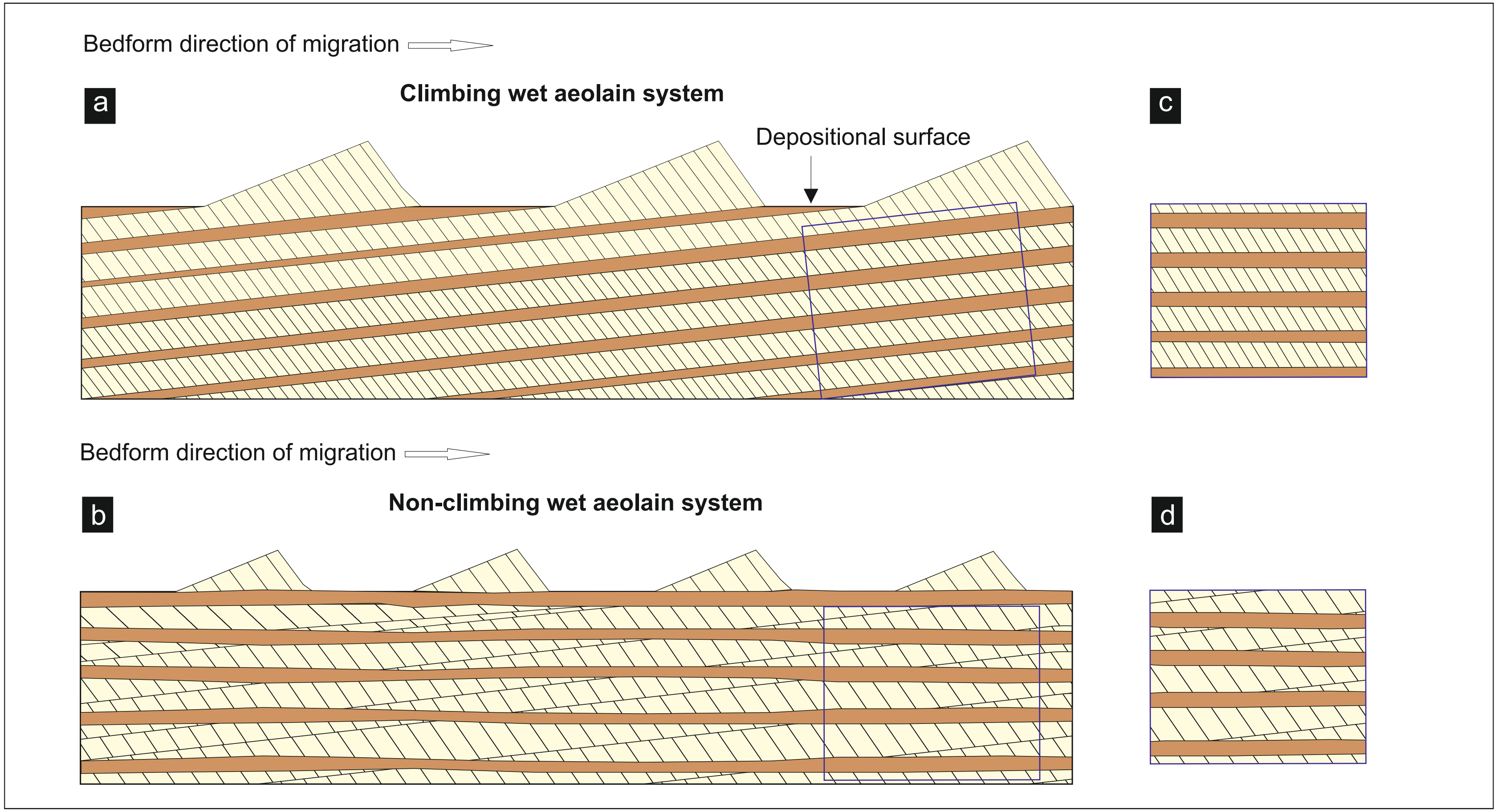


Figure 4.30: Models illustrating the accumulation and preservation of damp or wet interdune strata in wet aeolian system. a) Dunes migrate contemporaneously with a progressive and gradual rise in the water table. b) Migration of isolated aeolian dunes across wet interdune flats where the angle of climb is zero. c and d) Examples of preserved sections. Note: the similarity between the two preserved successions.

outcrop, the interdune basal surfaces are commonly sharp, abruptly terminating the underlying dune units (Figure 4.15). The contact between interdune and overlain dune sets are either sharp erosional (non-climbing model) (Figure 4.4, Panel A and B, and Figure 4.14) as a consequence of a slight fall in water table giving chance for wind reworking process to take place, or sharp non-erosional contact (Figure 4.4, Panels A-F) or gradational, where interdune facies pass gradationally upward to aeolian dune toe sets facies (Figure 4.7, Figure 4.8, Figure 4.13, and Figure 4.15). Additionally, in Figure 4.4, Panel B, an interdune composed of facies ID2 observed stacked against cross-strata of dune composed of facies AD1 and AD3. In some places, the intertonguing of dry interdune strata with overlying aeolian dune strata indicates interdune sedimentation that was synchronous with dune migration and accumulation (climbing model) (Pulvertaft, 1985; Loope and Simon, 1992; Kocurek and Havholm, 1993; Mountney and Thompson, 2002; Mountney, 2006b).

Lens-shaped interdune geometries which are dictated by the morphologic arrangement and spacing of adjacent dunes are observed in the studied sections (Figure 4.4, Panels A, B and D, and Figure 4.14b). These types of interdune elements are common in preserved aeolian dune successions that comprise the deposits of relatively small, rapidly downwind migrating transverse or oblique aeolian dunes between which isolated interdune depressions were present (Rubin, 1987; Carruthers, 1987; Lancaster, 1995). The style of lateral transitions and the nature of boundary surfaces between dune and interdune elements suggest changes in the level of water table, possibly in response to factors such as the amount of precipitation or longer-term changes in climate (cf. Crabaugh and Kocurek, 1993; Luna et al., 2012). In wet aeolian systems, the angle-of-climb of aeolian bedforms and laterally adjoining interdunes is known to change to reflect the fluctuation in the rate of water table rise and the rate of aeolian-migration (cf. Mountney and Thompson, 2002).

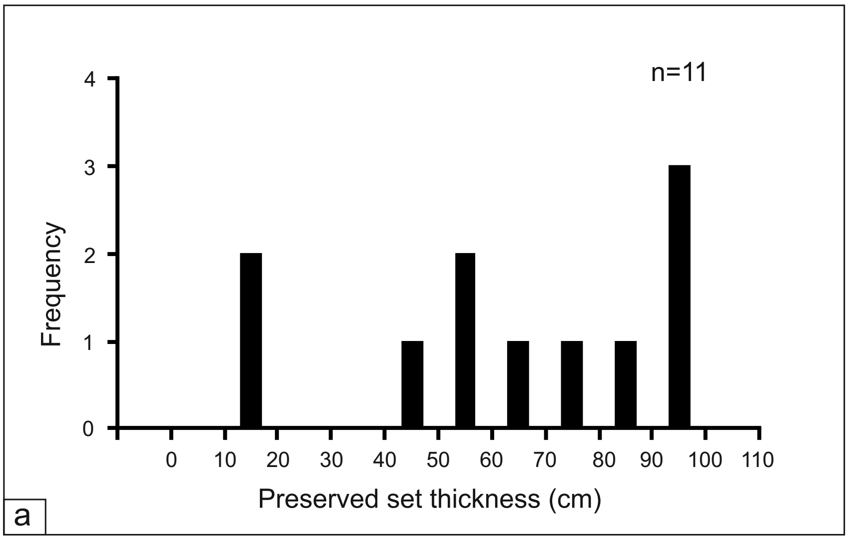
Damp interdune elements (A2) in the studied outcrop (Figure 4.4, Panels A and F) are abundant and are characterised by deposits that occur interfingered at a small scale with overlying aeolian dune toesets of A1 elements (e.g. Figure 4.4, Panel F2, and Figure 4.14b).

4.7.3 Relationship between grainflow dominated sets thickness and grain flow lithofacies thickness

Direct measurements of parameters relating to aspects of architecture of aeolian elements has enabled the establishment of empirical relationships, that can be used as a tool for predicting subsurface architectural parameters and arrangements (e.g., Romain and Mountney, 2014). The style of preserved grainflow lithofacies types and preserved set architectures enables prediction to be made regarding the relationship between preserved individual grainflow thickness and set thickness.

The amount of an aeolian succession that comprises packages of grainflow strata is an important consideration in reservoir geology; such packages represent bodies of high permeability, relative to packages of wind-ripple and grainfall strata. Thick and laterally continuous and connected accumulations of grainflow strata typically make for excellent reservoirs for hydrocarbons and water (Galloway and Hobday,1996; Howell and Mountney, 2001; Shephered, 2009; Loope et al., 2012; Ringrose and Bentley, 2014).

Aeolian sets, representing aeolian dune deposits and which are dominated by grainflow lithofacies (AD1) but also composed in part of grainfall and wind-ripple facies (AD2 and AD3), represent 35% of the aeolian succession observed with the Wilmslow Sandstone Formation exposed in the Runcorn Expressway road cut. Small-scale facies relationships from 11 preserved sets have been examined in detail (Figure 4.31a). For these sets, the thicknesses and type of 130 individual gainflow units have been determined, with each grainflow deposit having been defined based on sediment texture and grain size trends, and relationship to adjoining lithofacies AD2 and AD3 (see description and interpretation of lithofacies AD1 section 4.6.1.1, and aeolian element A1 section 4.6.3.1). The mean thickness of preserved sets is 0.7 m (Standard deviation 0.24 m); the mean grainflow thickness is 1.8 cm (Standard deviation 0.98 cm) (Figure 4.30). The data population exhibit a negative skewed distribution (Figure 4.31b). The relationship between set thicknesses and grainflow thicknesses data shows considerable variation



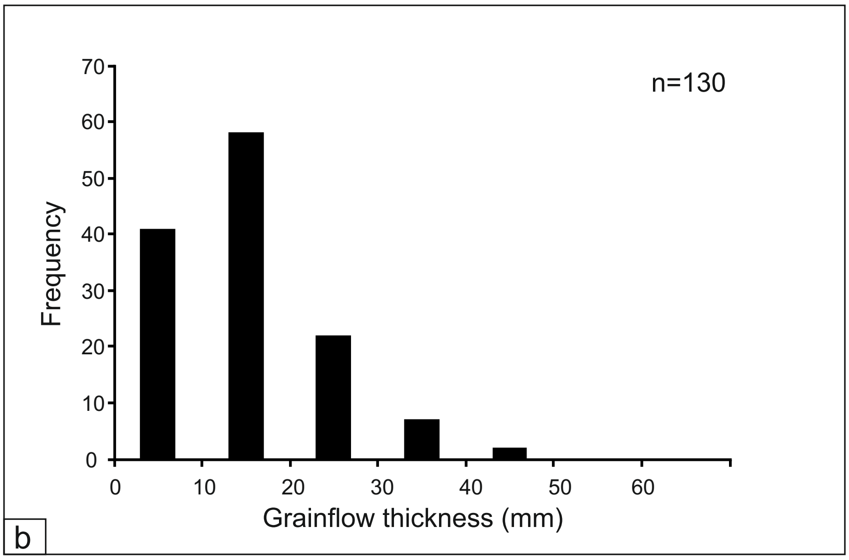
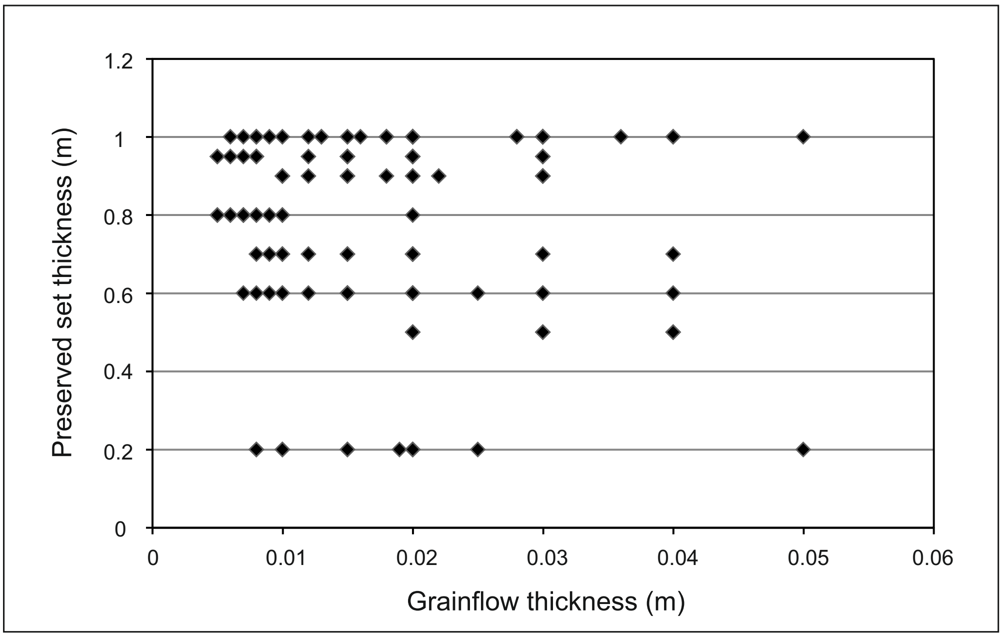


Figure 4.31: Frequency distribution of a) measured preserved bed-set thickness within which grainflow laminae occur, and b) measured grainflow thickness.

with no apparent trend and no clear correlation (Figure 4.32). The thickness of grainflow strata is directly linked to the volume of sand entrained at the front of the dune lee slope, and this in turn is governed by lee-slope length, which itself related to the dune height (Kocurek and Dott, 1981). Thicker grainflow deposits tend to develop on the lee slopes of larger dunes, whereas thinner grainflow deposits that are separated by thin grainfall laminae are typical of smaller dunes (Figure 4.14c; Hunter, 1981; Kocurek and Dott, 1981; Mader, 1985a). From a study of currently active dunes in the Little Sahara dune-field, Utah, Kocurek and Dott (1981) demonstrated a positive correlation between grainflow thickness and bedform height. The accumulation of aeolian dunes via bedform climbing is considered as the main mechanism of accumulation for the majority of preserved aeolian successions. For water-table-influenced examples of such systems, the angle of climb is controlled by changes in the ratio between the rates of water-table rise and dune migration, and the net aeolian sediment budget, which controls dune size (Mountney, 2004; Mountney, 2012). Bedform climbing processes typically allow only the lowermost part of the migrating aeolian bedform to be preserved (Rubin and Carter, 2006). Thus, preserved set thickness does not necessarily reflect the original dune bedform size; instead, thickness of preserved grainflows may represent a better proxy for dune height (Romain and Mountney, 2014).

For migrating dunes of a fixed size, a fixed angle of climb will result in generation of sets of dune strata with constant thickness (Mountney, 2012). Thus, for successions that have accumulated under the influence of a fixed angle of climb, there should be a strong positive correlation between preserved grainflow thickness and preserved set thickness. However, the data set herein highlights the absence of correlation between the grainflow strata thickness and set thickness. This could imply that the system was not climbing at a fixed angle but was rather climbing at variable angles during accumulation; else, it could indicate that successive dunes in a train were of variable size and/or spacing.

Examples of similar correlations were also demonstrated by Howell and Mountney (2001), and Romain and Mountney (2014). Results from both these studies concluded no significant relationship between preserved set



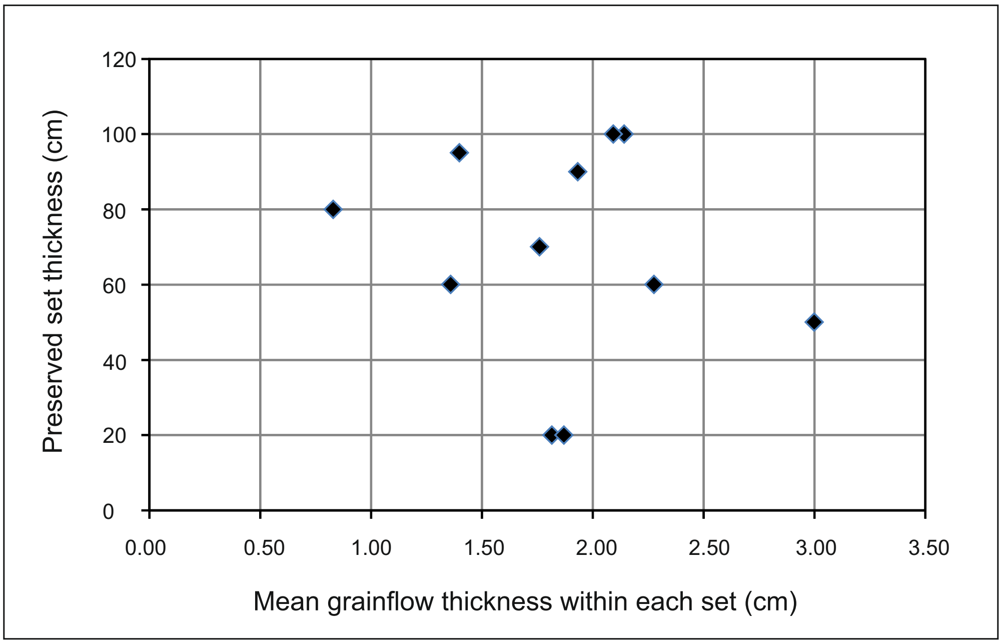


Figure 4.32: Relationship between grainflow thickness and preserved bed-set thickness within which grainflow laminae occur.

thickness and grainflow thickness. This variation could be a result of several reasons (Figure 4.33): 1) dunes of different bedform size but with the same angle of climb, the different original dune bedform sizes responsible for generating the sets preserved in different parts of the succession; 2) dunes of the same size but different angle of climb; 3) the exposed part of the dune trough could be only the edge, and the actual maximum preserved set thickness is more toward the trough central part that is not exposed; 4) Punctuated episodes of generation of accommodation. The preservation of an aeolian succession requires that the accumulation is placed beneath the level of deflation to protect it from later reworking. This can be achieved by: 1) placing the aeolian accumulation beneath the level of the water table, which acts to protect the accumulated sediment from deflation and reworking processes (Carr-Crabaugh and Kocurek, 1998; Mountney, 2012; Bristow and Mountney, 2013); 2) accumulation of aeolian sediment to a level beneath the baseline of erosion, thereby preventing aeolian sediment from becoming susceptible to deflation and reworking processes (Kocurek and Havholm, 1993). The thickness of accumulated and preserved aeolian deposits in the rock record therefore, is directly dependent on the availability of accommodation space.

4.7.4 Fluvial depositional model

Water can play a significant role in both erosion and deposition within dryland regions (Reid and Frostik, 2011, 1997; Tooth and Nanson, 2011). Dryland rivers tend to experience high –magnitude, low-frequency flood events (Graf, 1988, Tooth, 2000b), and due to the dependency of channel pattern on stream power provide an explanation of the common occurrence of braided rivers in dryland environments (Powell, 2009).

The deposits of braided rivers in dryland environments possess a set of sedimentary characterise by which they can be identified. These sedimentary characteristics are exhibited by the facies assemblages present in this study and are as follows: (i) a generally coarse-grained sediment composition (Figure 4.3 and Figure 4.18; Cant, 1982; Bristow, 1988, Soltan and Mountney, 2016); (ii) facies that are indicative of stream flow with both upper and lower flow regimes (Glennie, 1970; Miall, 1996); (iii) abundant and

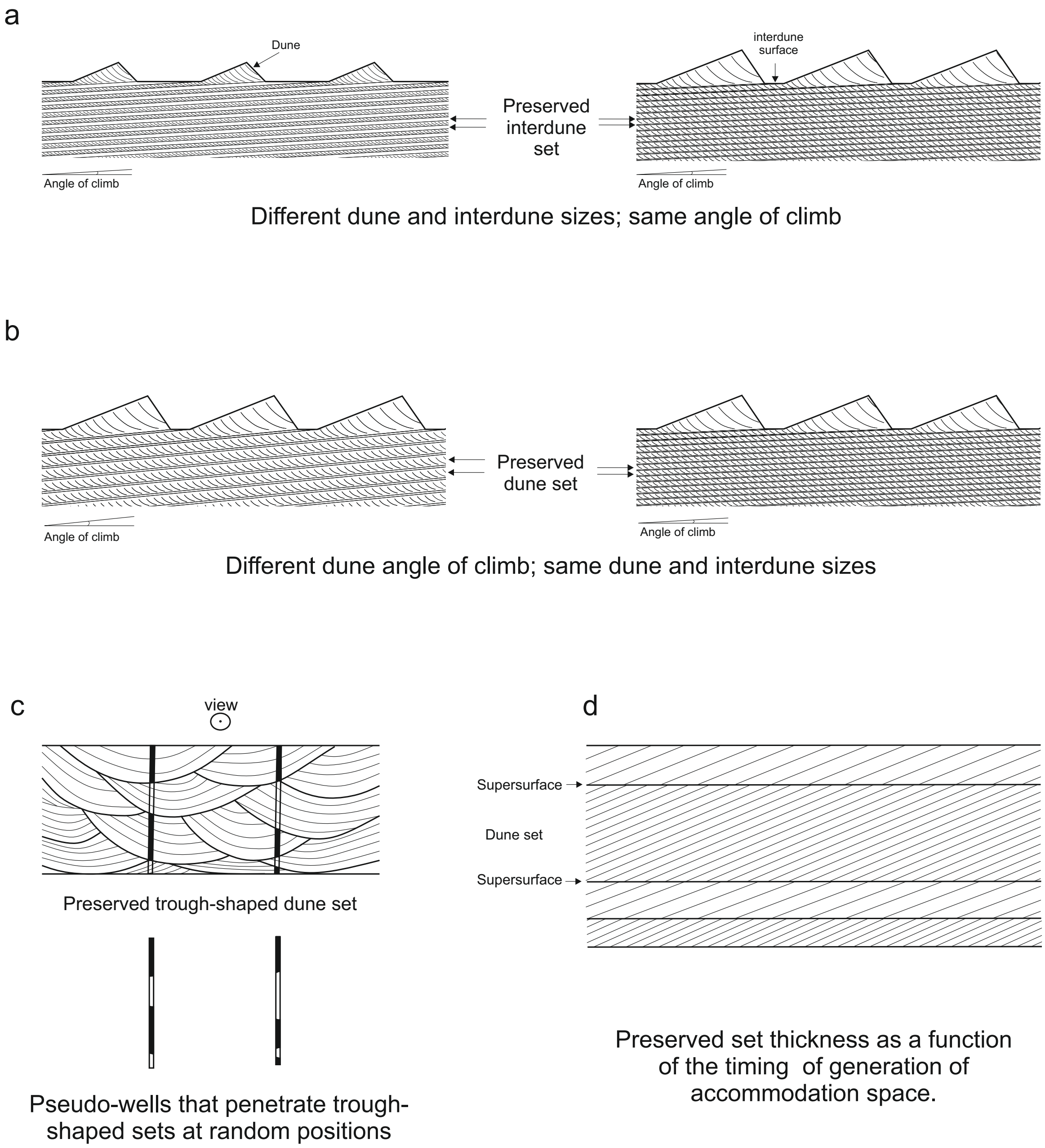


Figure 4.33: Stratigraphic explanation for the lack of clear trend in the relationship between grainflow thickness and preserved set thickness. a) different original dune bedform sizes responsible for generating the sets preserved in different parts of the succession, note the difference in the preserved interdune set thicknesses; b) variable angle of climb responsible for generating the sets preserved in different parts of the succession, note the difference in the preserved dune set thicknesses; c) clipping the edges of troughs; the pseudo-wells that penetrate trough-shaped sets at random positions reveal sets that are apparently of variable thickness. However, the true maximum thickness of the trough-shaped sets is actually very similar in all cases; d) punctuated episodes of generation of accommodation resulting in the episodic accumulation and preservation of aeolian dune sets.

repeated channel incision and fluvial erosion (Figure 4.4, Panel C); (iv) an abundance of planar cross-bedding (Figure 4.3, Figure 4.4, Panels B-F, and Figure 4.19; Smith, 1972; Swiecicki et al., 1995; Miall, 1996); (v) evidence for lateral migration of channels that with banks composed of poorly consolidated sand (Figure 4.4, Panel F; Cant and Walker, 1976; Cant, 1982); (vi) the presence of rip-up clasts (Figure 4.3, and Figure 4.18; Rubin, 1987; Miall, 2010); and (vii) general fining upward in grain size to form depositional cycles that occur vertically superimposed (Figure 4.3; Cant, 1982; Jones et al., 2001; Ashworth et al., 2011).

The presence of clay to silt grained deposited at the top of the channels in this study succession (Figure 4.25 a) implies that the fluvial system was probably related to a distal part of fluvial setting where the stream power was decreased, thereby promoting the accumulation of suspended-load sediments and the generation of such siltstone and mudstone at the top of each preserved cycle (cf. Miall, 1996; Collinson, 1996). Associations of facies and their occurrence as a distinctive suite of architectural elements record accumulation within a braided fluvial system.

Terminal fluvial systems, which are common fluvial system types in semiarid to arid regions represent a style of fluvial drainage network in which fluvial discharge does not drain to a significant standing body of water (e.g., a sea or lake), but instead terminates on an alluvial plain or in marginal aeolian dune-field setting (Figure 4.34; Friend, 1978; Kelly and Olsen, 1993; Cain and Mountney 2011, Al-Masrahy and Mountney, 2015). Desert ephemeral rivers or wadi channels carry large amounts of sediment as both suspension-load and bed-load (cf. Edgell, 2006). Terminal fluvial systems in such settings typically transport large volumes of sediment into the inland basins, thereby providing a significant source of sediment for later aeolian construction (Lancaster, 1995; Blair and McPherson, 2009; Belnap et al., 2011).

4.8 Conclusions

Results from this work form a valuable tool for comparative study of analogous aeolian and fluvial systems and preserved successions, in both

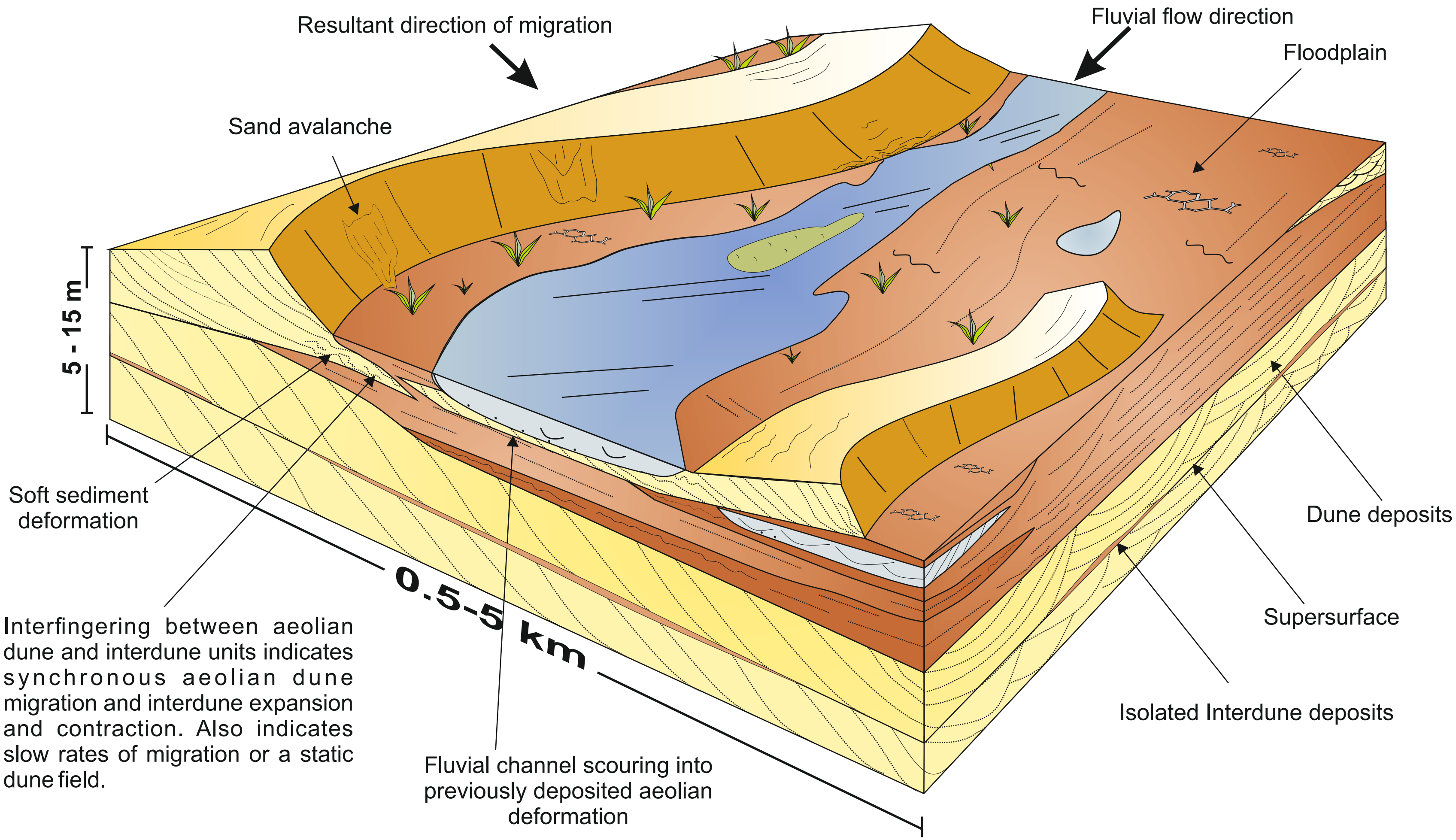


Figure 4.34: Schematic model illustrating the facies relationships arising from the termination of a fluvial system within an aeolian interdune corridor in a dune-field margin setting.

modern and ancient settings, which do not possess the three-dimensional characteristics of the outcrop succession of the Runcorn Expressway road cut. Sedimentological characteristics provide details with regard to the systems' architectural elements, internal lithofacies components and facies associations, within the framework of a generalised depositional model that depicts the arrangement and relationships of fundamental components of an ancient aeolian and braided fluvial succession.

The construction of aeolian desert sedimentary systems, their accumulation and their long-term preservation does not necessarily require extreme aridity. Desert aeolian systems may also be constructed and accumulate deposits in areas influenced by an elevated water table level that is close to, at or above the surface, such that the accumulation surface may be damp or even wet. This study has documented the preserved record of a wet aeolian system and an associated fluvial succession, and has developed further our understanding of processes that operate in aeolian and fluvial systems in arid and semi-arid depositional settings.

The Runcorn Expressway road-cut of northern Cheshire, England, provides an extensive section that is 13 m high and 230 m long. This section exposes strata of both aeolian and fluvial origin within the upper part of the Wilmslow Sandstone Formation and the lower part of the overlying Helsby Sandstone Formation (Figure 4.28). Aeolian and fluvial lithofacies, facies associations and architectural elements within this studied outcrop succession have been characterised in detail; the attributes of the aeolian and fluvial deposits (predominantly sandstones) are diverse. The latest techniques in lithofacies analysis and architectural-element analysis have been used to propose a novel depositional model for the aeolian and fluvial environments represented by these formations.

The studied section provides a valuable example of the preserved deposits of an ancient wet aeolian system. Interdune units preserved between aeolian dune units record the impact of the water table on the development and preservation of a water-table-influenced aeolian system. Also this studied section provides example of dryland fluvial system, and understanding the lateral and vertical arrangements of the ancient river system architectural elements is important to build more accurate models with which to advance the understanding of fluvial behaviour and the factors controlling the gross-scale of architecture of preserved fluvial successions.

This study provides a thorough and comprehensive description and interpretation for a well-exposed locality. The study concentrates on the overall understanding of an aeolian dune-field margin setting, where sedimentation was influenced by interaction with a dryland fluvial system. The outcomes of this study reveal the following: 1) the aeolian succession represents the preserved deposits of a wet aeolian system; 2) the system preservation mechanism was mainly controlled by the presence and change of water table level; 3) the aeolian system in the studied section preserves examples of both climbing and non-climbing dune-interdune behaviour, with examples of bypass and deflation supersurfaces preserved; 4) from observations in studied outcrop, the fluvial system developed as an ephemeral braided river in a dryland environment.

Results from this study are of importance in assessing the role of heterogeneity in partitioning hydrocarbon reservoirs and water aquifers, and for predicting lithofacies lateral distributions between isolated wells, and the likely arrangement of non-reservoirs units such as interdunes and fluvial channel-abandonment and floodplain lithofacies.

Preserved desert accumulations exhibit complex transitions on a variety of scales between the various sub-environments that comprise the arid climate depositional system. Typically, subsurface aeolian hydrocarbon reservoirs and groundwater aquifers are complex. Complexities arise from the stratigraphic anatomy and arrangement of lithofacies, including the possible associated occurrence of related fluvial lithofacies. Therefore, understanding the controls that govern the arrangement of aeolian and fluvial in arid-climate depositional systems, and the arrangement of architectural elements in erg-margin settings is important in assessing reservoir heterogeneity and predicting how low permeability interdune units may act as baffles to the fluid flow.

Chapter Five
Modelling system interactions in aeolian dune-field margin successions

This chapter employs the ten types of aeolian-fluvial system interactions described from modern desert systems (chapters Two and Three), together with the ancient outcrop case-study example (Chapter Four), and a suite of literature-derived data, to generate a series of ten semi-quantitative geological facies models with which to account for the nature and origin of stratigraphic complexity present in aeolian dune-field margin successions that arise from the interplay of both autogenic and allogenic controls.

5.1 Abstract

Mixed fluvial and aeolian successions form several major reservoirs for hydrocarbons, including the Permian Unayzah Formation of Saudi Arabia, the Permian Rotliegend Group of the North Sea, and the Jurassic Norphlet Sandstone of the Gulf of Mexico. Such reservoir successions typically exhibit stratigraphic heterogeneity at a number of scales. Quantitative stratigraphic prediction of the three-dimensional form of heterogeneities arising from fluvial and aeolian interaction is notoriously difficult: (i) the preserved products of system interactions observed in one-dimensional core and well-log data typically do not yield information regarding the likely lateral extent of sand bodies; (ii) stratigraphic heterogeneities typically occur on a scale below seismic resolution and cannot be imaged using such techniques. A database recording the temporal and spatial scales over which aeolian and fluvial events operate and interact in a range of present-day and ancient desert-margin settings has been collated using high-resolution satellite imagery, aerial photography and field observation. Together, these data have been used to develop a series of dynamic facies models to predict the arrangement of architectural elements that define gross-scale system

architecture. Case-study examples have enabled the construction of a series of depositional models to account for the diversity of styles of fluvial and aeolian system interactions.

Several styles of fluvial-aeolian interaction have been documented and the type and distribution preserved deposits can now be predicted through quantitative geological models that account for spatial and temporal changes in system dominance. For example, the preserved architectural elements of fluvially flooded interdunes tend to expand laterally as successive flood deposits develop in front of advancing aeolian dunes. In non-climbing aeolian systems, such behaviour favours the development of sheet-like bypass surpersurfaces. In aeolian systems that climb at low angles and for which fluvial incursions are episodic, thin and laterally impersistent fluvial elements tend to accumulate. The scale and connectivity of fluvial flood deposits tends to diminish with increasing distance toward the aeolian dune-field centre. Results from this study have implications for how reservoir models are constructed for subsurface plays developed in mixed aeolian and fluvial successions.

5.2 Introduction

Accumulation, deflation (erosion), supersurface development and sequence preservation are variable over both space and time in aeolian successions. As a result, the possible configurations of stratigraphic architectures that can result from the operation of these controls are many and it is not feasible to propose a single summary model to account for the possible range of aeolian stratigraphic complexity.

Aeolian sandstone deposits can form excellent reservoirs. The wind is a highly effective agent for sorting sediment meaning that the grain-size distribution of aeolian sand deposits tends to be restricted to a narrow range of sizes. Mean and median grain size and the standard deviation of their distribution may vary from the upwind to the downwind parts of aeolian dune fields and their preserved successions (e.g. Langford and Chan, 1993). At any given aeolian depositional site, wind-blown deposits tend to be relatively well-sorted by grain size compared to sediment deposited via other processes, such as fluvial activity. Effective sediment grain sorting processes generally favour the development of good reservoir quality through the preservation of high primary porosity (Slatt et al., 1993). Aeolian dune deposits commonly (though not always) comprise thick, cross-bedded intervals of sandstones with well-rounded and well-sorted grain textures and that are typically compositionally mature (e.g. quartz arenites). This gives rise to well preserved primary porosities (e.g. Weber, 1987). The primary porosity and permeability of sandstones depends largely upon primary depositional processes and sediment texture. However, diagenetic processes that occurred after deposition commonly exert a strong secondary influence on the final porosity and permeability of preserved rock successions (Bloch, 1991, Ali et al., 2010; Slatt, 2013). Aeolian dune grainflow-dominated slipface deposits, which comprise loosely packed, well-sorted sand grains, are normally the most productive lithofacies in aeolian reservoir systems, whereas dune-apron deposits tend to exhibit slightly reduced quality reservoir, and interdune units can form relatively low-porosity and low-permeability barriers and baffles within aeolian successions (Hunter, 1977; Weber, 1987; Heward, 1991; Herries, 1993; Shepherd, 2009, Mountney, 2006a).

Fluvial depositional environments, by contrast, tend to produce more heterogeneous reservoir successions for which porosity and permeability may vary considerably according to primary lithofacies type (North and Prosser, 1993). The facies associations of different architectural elements of fluvial successions can have markedly different porosity and permeability characteristics that chiefly depend on several factors: the nature of the rock matrix, lithologic heterogeneity, compaction, cementation, original sand sorting and grain size distribution character ((Morse, 1994; McKinley et al., 2011). Relatively poorly sorted sandstone grain textures tend to exhibit lower permeabilities than better-sorted sandstone (Slatt, 2013). Within fluvial successions, the original clay content trapped in between the framework grains commonly lead to the development of secondary cement formation during diagenesis and the presence of such clays typically detrimentally affects fluvial reservoir quality (Ramon and Cross, 1997). For example, sand-prone channel-fill elements may be composed of apparently well sorted sandstone but the presence of clays between framework-forming grains tends to reduce reservoir potential via permeability reduction (e.g., De Ros and Scherer, 2012). Clay content is commonly sufficient to choke pore throats. Moreover, clays may alter to form pore-blocking types such as hairy illite that contributes further to reservoir impairment (Woodward and Curtis, 1987; Ahmed, 2008; Greensmith, 2012). Examples of successions where illitization is noted include parts of the Southern North Sea Permian Rotliegend Group (Glennie and Provan, 1990).

In arid and semi-arid environments it is common for fluvial and aeolian systems to interact over a variety of spatial and temporal scales (see Chapter Three). Such interactions generate sequences that are composed internally of complex arrangements of aeolian and fluvial architectural elements, which are themselves composed of various facies associations (Chapter 4). The common types of interplay recorded from present-day aeolian and fluvial systems have been classified as ten distinctive types of interactions (Chapter Three, Figure 3.16; Al-Masrahy and Mountney, 2015). The accumulated and preserved lithofacies expressions of these different types of interactions typically result in different and possibly contrasting geometries and architectures (e.g., Herries, 1993; Mountney, 2012). Thus, preserved examples of such systems may form complex hydrocarbon reservoirs that are stratigraphically heterogeneous at several spatial and temporal scales.

From an applied perspective, aeolian dune and interdune successions form important reservoirs for hydrocarbons, including the Permian Rotliegened Group of North Sea (Glennie, 1990; Howell and Mountney, 1997; Sweet, 1999), the Triassic Ormskirk Sandstone Formation of the East Irish Sea (Herries and Cowan, 1997; Meadows, 2006), the Jurassic Norphlet Sandstone of the Gulf of Mexico (Kugler and Mink, 1999) and the Permian Unayzah Formation of Saudi Arabia (Melvin et al., 2010; Al-Masrahy et al., 2012). Dune facies and elements are typically the most productive lithofacies in aeolian reservoir systems, whereas interdune facies and elements tend to have lower porosities and permeabilities, and fluvial components may form non-net reservoir flow barriers. The ability to predict the geometry and degree of interconnectivity of these basic element types and changes from central to margin palaeo-dune-field environments is essential in assessing likely reservoir quality and the distribution of potential baffles and barriers to fluid flow (Weber, 1987; Chandler et al., 1989; Stanistreet and Stollhofen, 2002; Taggart et al., 2010; Mountney, 2012).

The complex distribution of aeolian dunes and interdunes in any dune field defines their spatial heterogeneity distribution which will affect reservoir behaviour, the impact of which often increases later in the life of a hydrocarbon field (Sweet et al., 1996). By mapping and modelling the 2D and 3D distribution of dune and interdune architecture, it is possible to capture details of types of facies interactions, the spatial extent of bounding surfaces, and the distribution of stratification types, all of which serve as fundamental controls on reservoir heterogeneity.

The preserved sedimentary signatures that arise from the interaction of fluvial and aeolian processes have been widely recognised in stratigraphic record (e.g. Andrews, 1981; Loope, 1985; Langford and Chan, 1988; 1989; Glennie, 1990; Trewin, 1993; Heries, 1993, Jones and Blakey, 1997; Howell and Mountney, 1997; Mountney et al. 1998; Sweet, 1999; Stanistreet and Stollhofen, 2002; Bullard and McTainsh, 2003; Mountney and Jagger, 2004; Scherer and Lvina, 2005; Veiga and Spalletti, 2007; Simpson, et al., 2008; Rodriguez-Lopez et al., 2010; Jordan and Mountney, 2010; Spalletti et al., 2010; Bongiolo and Scherer, 2010; Cain and Mountney, 2011; Mountney, 2012; East et al., 2015). However, specialised and sophisticated facies models have yet to be developed which account for the lateral facies changes known to occur in aeolian dune-field margins where competing aeolian and fluvial processes take place. This shortcoming is addressed by this work.

Three-dimensional geological models (i.e. facies models) are still considered one of the important tools for reservoir development purposes. The aim of this chapter is to propose a series of facies models, the internal anatomy and facies composition of which reflect the preserved record of different types of aeolian-fluvial interaction documented from the ancient rock record. A specific research objective is to predict the likely arrangement of architectural elements for both the aeolian and the fluvial components of different types of desert-margin successions. Such elements form the building blocks of hydrocarbon reservoirs in such successions.

5.3 Fluvial-aeolian system interaction types

This section discusses the approach taken to developing a series of facies models for the different types of interaction known to occur between competing fluvial and aeolian systems that are present both within and at the margins of aeolian dune-field systems (Chapter Three; summary figure 3.16). Ten distinct types of interaction are illustrated by ten different three-dimensional geological models.

5.3.1 Fluvial incursions oriented parallel to trend of aeolian dune forms

In this type of interaction, dune forms are arranged in a configuration of elongate ridges with crestlines aligned parallel or close to parallel to the direction of fluvial flow (Figure 5.1). The elongate shape of the dune forms (e.g., linear dunes) in this case enables the fluvial flow to pass through between dunes and flooding the open interdune areas. The potential extent of fluvial incursion into the dune field is variable; analysis of examples from modern systems (Chapter Three, Table, 3.1) reveals a range incursion distances of 13 to 114 km along open interdune corridors that provide access for fluvial systems within aeolian dune fields. Fluvial processes that operate during flood incursions may rework and erode aeolian sediment, and can also introduce new sediment into the aeolian system. A modern example of this type of interaction is the northern Simpson Desert, Australia (Figure 3.4; Nanson et al., 1995; Al-Masrahy and Mountney, 2015).

This setting generates either ribbon-like fluvial deposits between non-climbing or stabilised dune forms or sheet-like fluvial depositional elements in cases where the aeolian dunes progressively migrate between flood events (cf. Langford and Chan, 1988). Where fluvial channels are developed, their fills may be composed predominantly of fluvially reworked aeolian sand. The fluvial flow at the termination point within aeolian system

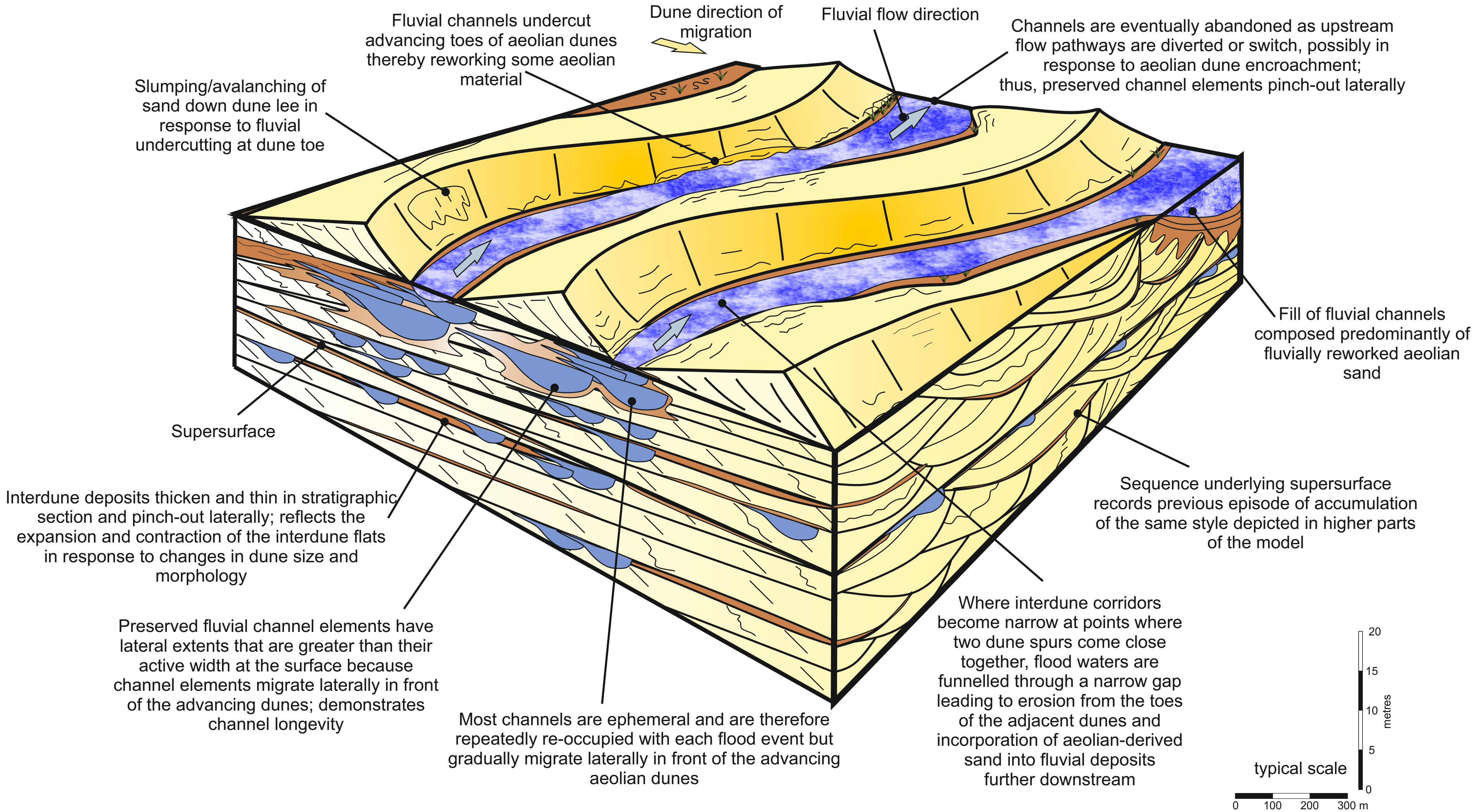


Figure 5.1: Fluvial incursions oriented parallel to trend of aeolian dune forms.

may form temporary ponds from which fine-grained suspended-load sediments are deposited to generate thin layers of mud that may act to stabilise the interdune substrate once the flood waters have receded (e.g., Stanistreet and Stollhofen, 2002). Such layers may act to restrict aeolian reworking thereby preserving underlying fluvial strata.

Although the majority of channels developed as a result of this type of interaction are ephemeral, they may be repeatedly re-occupied with each subsequent flood event. Where a slow rate of aeolian bedform migration occurs, such fluvial channels will migrate gradually laterally over time in front of the advancing aeolian dunes, and in doing so will form sheet-like fluvial elements of either extra-dune-field sediment or locally fluvially reworked aeolian sediment. Such deposits may accumulate between stacked aeolian dune deposits. The preserved fluvial deposits present in this type of aeolian setting can potentially be characterised by a lateral extent that is significantly greater than the width of a single flooded interdune at an instant in time because channel elements migrate laterally in front of the advancing dunes and thereby become grow through lateral translation and may merge with one another in non-climbing dune systems.

The interdune corridor itself will comprise a series of facies associations and nested architectural elements. 1) Fluvial channels that, in some places, may have erosional bases filled with a lag of coarse-grained clasts; some pebble or cobble clasts may be of intraformational origin in cases where sediment has been reworked from the margins of aeolian dunes as blocks of semi-consolidated sand; other pebble clasts may be of extraformational origin if the desert system is adjacent to a mountain front (cf. Anderson and Anderson, 1990). 2) Fluvial channels in some places may be erosionally based and have a fill of massive (structureless) sand or crudely cross-bedded sand that reflects a rapid accumulation processes. 3) The fill of fluvial channel elements can be largely of fluvially reworked aeolian sand. 4) The fill of fluvial channel elements may alternatively be aeolian sand that was blown into an under-filled abandoned channel, thereby filling it via aeolian processes. 5) Other parts of interdunes might be characterised by sheet-like (i.e. non-channelised) fluvial deposits. 6) Some parts of interdunes might not be subject to fluvial flow but might be characterised by damp interdune adhesion structures due to a locally elevated water table associated with the flooding taking place elsewhere in the dune-field margin. several ancient examples of such behaviours have been documented: the Permian Cutler Formation and Cedar Mesa Sandstone on the Colorado Plateau (Langford and Chan, 1989); the Middle Jurassic Page Sandstone in south central, USA (Jones and Blakey, 1997); the Triassic Helsby Sandstone Formation, Cheshire Basin, UK (Mountney and Thompson 2002); much of the foreland region of Paradox Basin during the Early Permian (Wolfcampian) time (Condon, 1997); the Cretaceous Troncoso Member, Neuquen Basin, Argentina (Stromback et al., 2005); and the Permian Organ Rock Formation, southeast Utah, USA (Cain and Mountney, 2011).

5.3.2 Fluvial incursions oriented perpendicular to the trend of aeolian dune forms

In this type of interaction, the dune forms are arranged in a configuration of elongate ridges with crestlines aligned perpendicular or close to perpendicular to the direction of fluvial flow (Figure 5.2). The occurrence of this configuration at the outer margin of an aeolian dune field, gives rise to flood events that may be prevented from passing into the dune field and may instead become ponded or be diverted in orientations parallel to the trend of the dunes at the outer dune-field margin.

This setting generates a sharp boundary between aeolian and adjoining fluvial environments. The persistent or repeated presence of fluvial systems in this configuration prevents aeolian bedform migration or growth beyond the boundary of the fluvial system; aeolian sediment may temporarily accumulate in the fluvial channels but will be reworked by and incorporated into fluvial flows to be transported and deposited further downstream. By contrast, if the aeolian system undertakes progressive retreat (retrogradation), the aeolian deposits may become overlain by flood deposits. Complex and repeated jostling of the respective systems may produce an alternating cycles of aeolian and fluvial deposits.

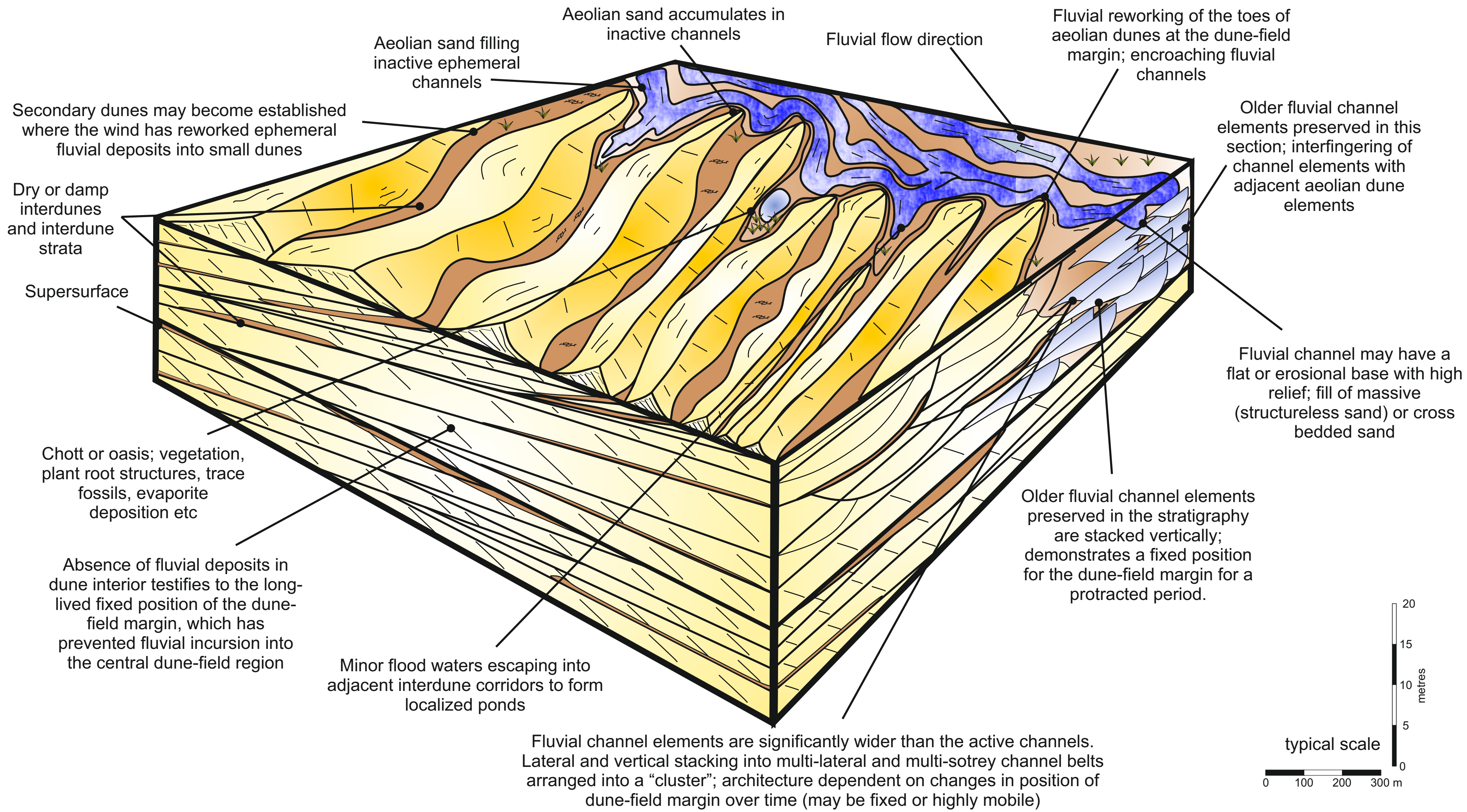


Figure 5.2: Fluvial incursions oriented perpendicular to trend of aeolian dune forms.
This model is analogous to the modern scenario at the eastern margin of Wahiba Sand Sea, Oman.

The preservation of stacked fluvial channel elements in this setting will occur at the aeolian system outer margin. The intercalation of fluvial and aeolian facies will tend to be limited to the narrow zone of interaction between the two systems in cases where the position of the dune-field margin is stable for a protracted period. Thus, preserved fluvial channel elements in the stratigraphic record may be stacked vertically in the area where the interaction is taking place, thereby demonstrating a fixed position for the aeolian dune-field margin for a protracted period. A known outcrop example of this type of behaviour is the lateral erg margin of the Permian Cedar Mesa Sandstone, Utah (Mountney and Jagger, 2004); a second example is the pre-White Rim sandstone (Permian) that is characterised by intertonguing fluvial and aeolian environments within the undifferentiated Cutler Group, also of the Paradox Basin, Utah (Chan, 1989).

5.3.3 Bifurcation of fluvial flow between isolated aeolian dune forms

Some aeolian systems are characterised by the presence of isolated aeolian dune bedforms (e.g., the star dunes of the southeastern part of the Rub' Al-Khali sand sea; Chapter Two, section 2.7; Figure 2.6d; Figure 3.7a). In such settings, fluvial systems may flow around these isolated dunes in the outer parts of aeolian dune fields and flow may bifurcate around the topographic obstacles on both sides, eroding and reworking the dune flanks, and then redepositing the aeolian sediment downstream in terminal splays within the aeolian dune field (Figure 5.3). The tortuous pathways of fluvial channels between and around spatially isolated aeolian dunes are dictated by dune morphology and spatial density. The distance of penetration of these fluvial systems into the marginal parts of aeolian systems may be high (7 to 161 km; Table 3.1), indicating that the zone within which aeolian and fluvial sediment intermixing might occur can be broad. Sediment of mixed aeolian and fluvial origin that is deposited at the point of termination of fluvial flow within an aeolian system results in preserved sediment accumulations of mixed affinity, which can make the task of discerning original sedimentary process and environmental settings difficult, particularly in situations where fluvial systems have travelled long distances into aeolian dune-field margins.

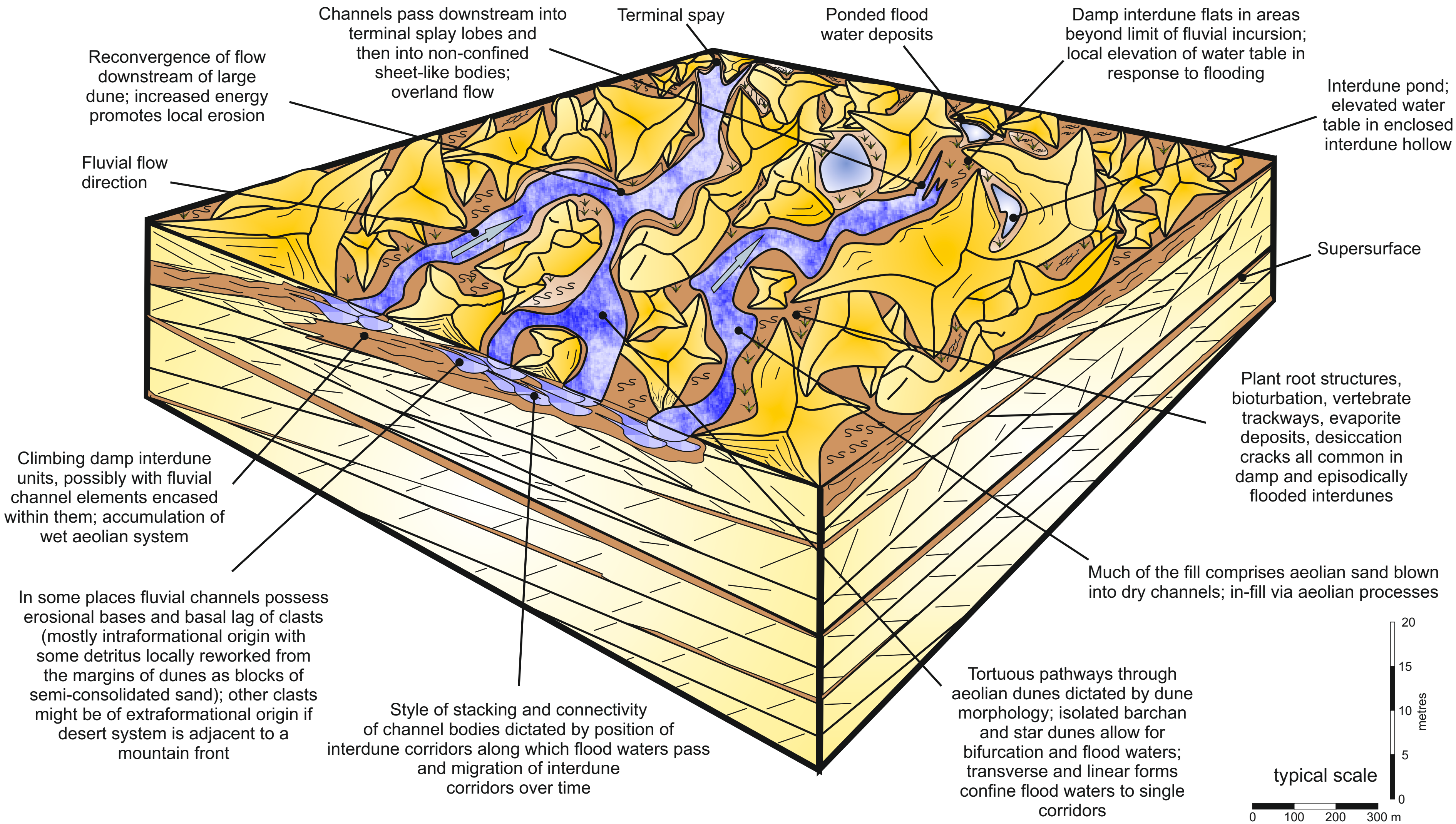


Figure 5.3: Bifurcation of fluvial flow between isolated aeolian dune forms.
This model is analogous to the modern scenario at the southeastern margin of the Rub' Al-Khali Sand Sea, Saudi Arabia.

In some fluvial systems, bedload-deposited sediments may be composed almost exclusively of reworked aeolian sediments and the sediment textural characteristics will reflect this.

Fluvial flow into the aeolian dune fields may locally charge the groundwater reservoir and this can influence the water-table level. Where the water table remains close to the surface for long periods, a wet aeolian system may develop (*sensu* Kocurek and Havholm, 1993). The long-term presence of a high water table will also enhance the preservation potential of the lowermost parts of migrating but spatially isolated aeolian bedforms (cf. Mountney and Russell, 2009).

Preserved fluvial deposits within aeolian systems of this type tend to vary in character spatially. In areas close to the outer dune-field margin, fluvial stream flow will tend to have higher velocity and be able to generate confined fluvial channels. Further downstream within the dune-field margin, the velocity of fluvial flow will typically decrease and flow will tend to be less confined, generating sheet-like elements that comprise some aeolian-reworked sediment. Interdune flats in areas beyond the limit of fluvial incursion will tend to become progressively “drier” with increasing distance from the flood termination point; interdune elements will change their internal facies composition as they change laterally from bodies that indicate a gradual shift from damp to dry interdune substrate conditions.

Documented ancient examples of such behaviour are as follows: the Jurassic Kayenta-Navajo transition, northern Arizona, USA (Herries, 1993); the lower part of the Ormskirk Sandstone Formation, Triassic, England (Meadows and Beach, 1993); the Vallecito Formation, Lower Miocene in the Andean foreland basins of the Precordillera, Argentina (Tripaldi and Limarino, 2005); and the Paleoproterozoic Baker Lake Group, Nunavut, Canada (Hadlari et al., 2006).

5.3.4 Through-going fluvial channel networks that cross entire dune-fields

In this type of interaction, fluvial systems pass through entire aeolian dune-fields. Such through-going fluvial channel fairways are established where major drainage courses emerge from mountain catchments and pass into desert dune fields (Figure 3.16, and Figure 5.4). The presence of a fluvial course passing through an entire dune field may act to effectively partition the dune field, thereby limiting aeolian sediment transport pathways. One example of a permanent, perennial fluvial desert system is the Nile River (Figure 3.8a). The continuous presence of water will enhance the level of the groundwater table, potentially resulting in moisture at the accumulation surface in areas of the dune field adjacent to the river course. This will reduce aeolian sediment mobility, and therefore the availability of that sediment for aeolian transport (Ward, 1987; Krapf et al., 2003). During wet seasons when floods occur, the channel bank-full capacity of through-going rivers may be exceeded, resulting in the dispersal of fluvial sediment laterally along aeolian interdune corridors adjacent to the river. This is a mechanism to allow fluvial sediment dispersal beyond the fluvial channel margins (i.e. in floodplain settings). Such fluvial overbank sediments will typically become overrun by aeolian dunes in post-flood times after the water level has receded and the river once again has become confined. The type example where such processes occur is the River Nile, where 1000 kilometres of riverbank are lined with active aeolian dunes separating the Nubian Desert from the Libyan Desert (Spencer et al., 2012; Woodward et al., 2015).

Fluvial flooding events may also form localised ponding along interdune corridors adjacent to the main river course. Here, thin mud beds typically accumulate in the aftermath of suspended-load sedimentation. The repeated reoccurrence of this process will favour the generation of a zone of accumulation of wet interdune strata adjacent to the main fluvial channel fairway. This zone is characterised by a progressive accumulation of mudstone units.

The configuration of aeolian dunes along the river channel bank may act to fix the position of the channel fairway. The vertical facies succession at the site of the main fluvial fairway will typically be characterised by a vertically stacked cluster of amalgamated fluvial channel elements that testifies to long-lived fluvial activity and a relatively fixed channel position (Figure 5.4).

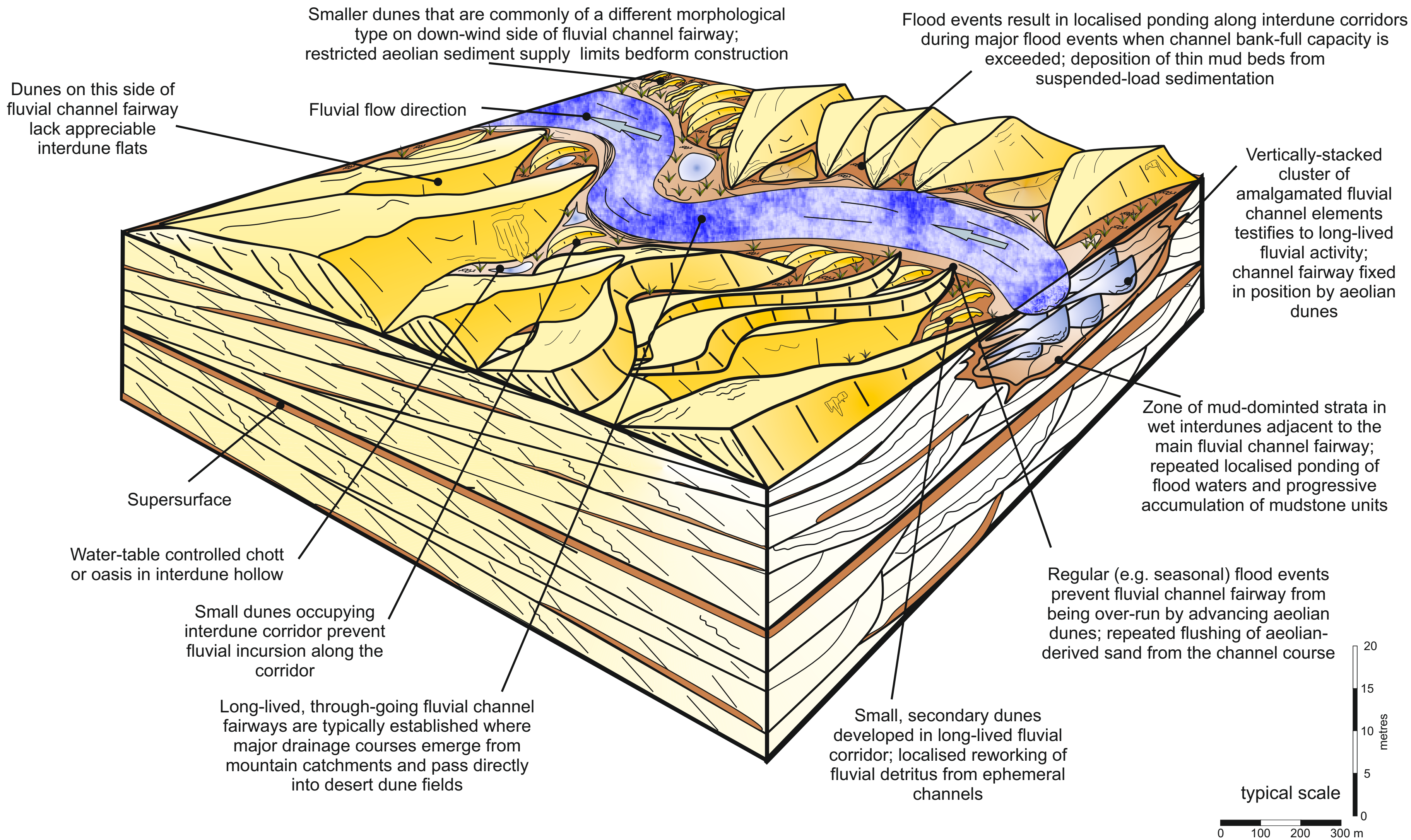


Figure 5.4: Through-going fluvial channel networks that cross entire dune-fields.

During dry seasons, in areas characterised by the presence of ephemeral rivers, aeolian activity will tend to rework fluvial detritus and small dunes may migrate into and along the fluvial channel path and across floodplain areas. In such situations, the preservation of aeolian sediment is unlikely as such aeolian deposits will likely be reworked by fluvial flow during next flood event. Documented ancient examples of such behaviour are as follows: the eastern part of the northern Permian Basin, Karl Formation, Offshore Denmark (Stemmerik et al., 2000), the Pennsylvanian to Permian lower Cutler beds, southeast Utah, USA (Jordan and Mountney, 2010); the Permian Slochteren Formation in the Netherlands (McKie, 2011b); and the Upper Triassic Tadrart Ouadou Sandstone Member, Argana Valley; South-west Morocco (Mader and Redfern, 2011).

5.3.5 Fluvial flooding of aeolian dune-fields associated with elevated water table level

In this type of interaction, aeolian dune fields are characterised by a high water-table level that may inundate interdune areas during wet episodes (Figure 3.9, and Figure 5.5). The permanent presence of a high water table will generate wet or damp interdune flats between aeolian dunes for protracted episodes, thereby allowing the development of a wet aeolian systems (*sensu* Kocurek and Havholm, 1993). A high water table is also important as it increases the likelihood of the long-term preservation of the aeolian deposits (see the example from Al Jafurah Desert, Saudi Arabia; Figure, 3.9b; Mountney and Russell, 2009). In wet aeolian systems, sediments that comprise the floors of damp or wet interdune flats are relatively cohesive, and sediment availability for aeolian transport is restricted (Hotta et al., 1984; Good and Bryant, 1985; Fryberger et al., 1988; Crabaugh and Kocurek,1993; McKenna and Scott, 1998; Mountney and Russell, 2009). Such water-table controlled aeolian systems tend to be characterised by reduced rates of aeolian dune migration due to damp or wet surface conditions that prevent aeolian bedform advancement. One ancient example is part of the Jurassic Wingate Sandstone, Utah (Simpson and Loope, 1985; Loope and Simpson, 1992).

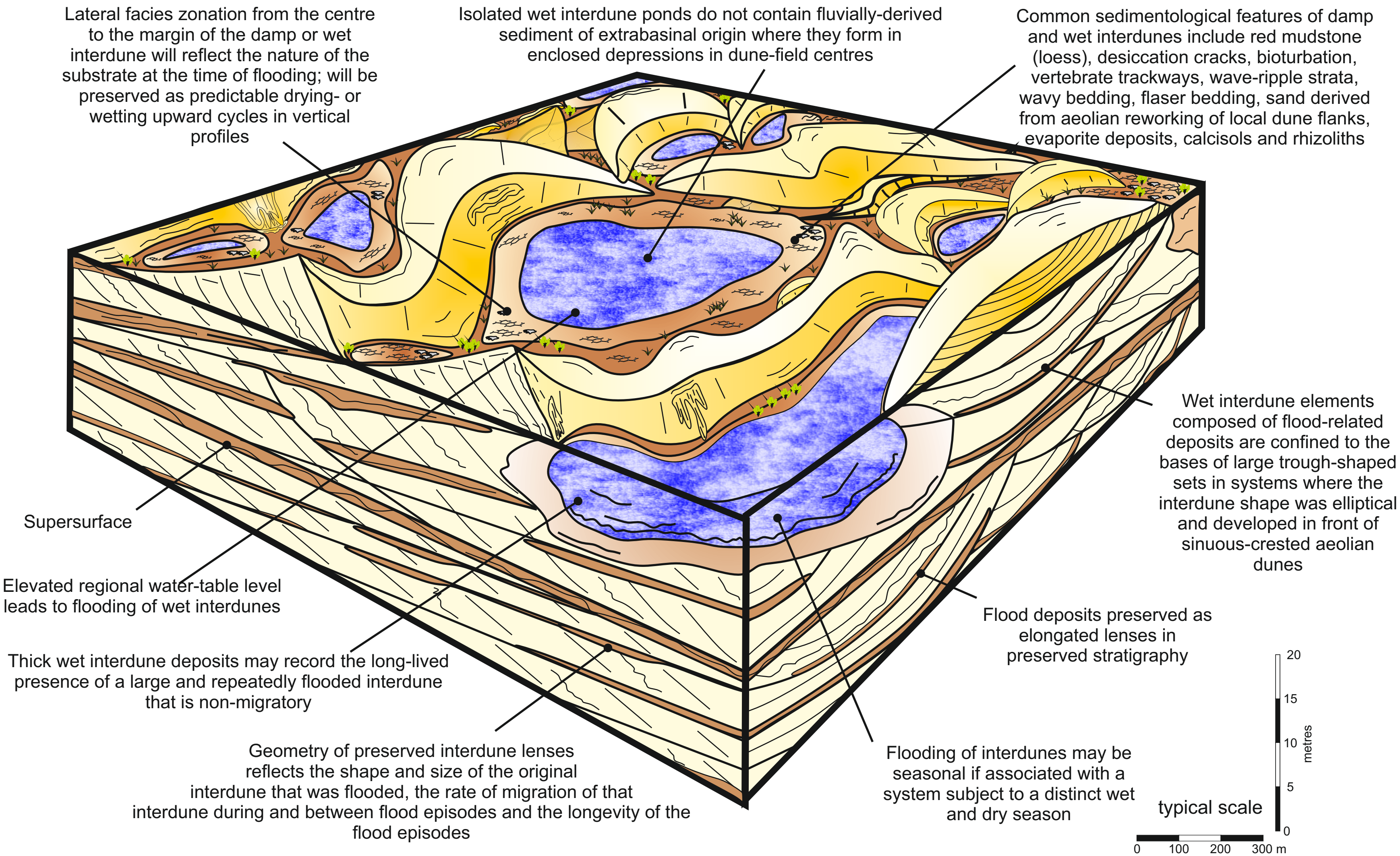


Figure 5.5: Fluvial flooding of aeolian dune-fields associated with elevated water table level.

Common sedimentological features of damp and wet interdune deposits typical of this type of interaction include red mudstone (loess), desiccation cracks, adhesion structures (adhesion ripples, adhesion warts and adhesion plane beds), aqueous-ripple structures, wavy laminations, contorted structures and brecciated laminae, bioturbation, vertebrate trackways, sand derived from aeolian reworking of local dunes, calcisols and rhizoliths (Chapter Four, section 4.6.1.7; Kocurek, 1981; Kocurek and Fielder, 1982).

The long-lived presence of extensive and repeatedly flooded interdune areas that are non-migratory favours the accumulation of thick wet interdune deposits, and the preserved geometry of the interdune (i.e. lenses) reflects the shape and size of the original interdune that was flooded, the rate of migration of that interdune during and between flood episodes and the longevity of the flood episodes (Figure 5.5). Regional elevated water-table level leads to flooding of interdune areas between aeolian dune forms; this may generate isolated and enclosed wet depressions that lack any fluvially-derived sediment of extrabasinal origin. Such damp or wet interdune elements may show a lateral facies zonation from their centre to their margin, which reflects the nature of the substrate at the time of flooding and will be preserved as predictable drying or wetting upward cycles in vertical profiles of interdune elements. Temporary rise in the groundwater table brought about by fluvial floods is also documented by Ahlbrandt and Fryberger (1981), Petit-Maire et al., (1980), and Ward (1988).

Further documented ancient examples where such behaviour is recorded are as follows: the Cutler Group and the Cedar Mesa Sandstone (Permian), southeastern Utah (Langford and Chan, 1988; 2008); the Ormskirk Sandstone Formation (Lower Triassic), East Irish Sea Basin (Herries and Cowan, 1997); the Permian Whitehorse Group of south-central Oklahoma, Permian Delaware basin region (Kocurek and Kirland, 1998); Triassic to Lower Jurassic continental red beds of the Argana Valley, Morocco (Hofmann et al., 2000); the Late Permian Dawlish Sandstone Formation, Wessex Basin, South west UK (Newell, 2001); the Avile Member of the Agrio Formation (Lower Cretaceous), central Neuquen Basin, Argentina (Veiga et al., 2002); the upper Karoo aeolian strata (Early Jurassic), Tuli Basin, South

Africa (Bordy and Catuneanu, 2002); the Ordovician Guaritas Rift, southernmost Brazil (Paim and Scherer, 2007); and mid-Cretaceous western Tethyan margin, Iberian Basin, Spain (Rodríguez-López et al., 2006).

5.3.6 Single point source fluvial incursion into aeolian dune-fields

In this type of interaction, the fluvial channels emanate from basin-bounding highland areas to pass as single-thread systems to intersect aeolian dune-field systems at specific points along their margins (Figure 3.10, and Figure 5.6). These types of channels commonly do not migrate laterally and are not long-lived but are transient features. Therefore, they are preserved as isolated features in the preserved stratigraphy, with size, frequency and degree of interconnectedness of fluvial channel elements typically decreasing toward the dune-field centre. The distance of fluvial incursion along interdune corridors is controlled by the magnitude of the flood and the length of open interdune corridors. Modern examples show a range of incursion distance of 2.91 to 135 km (Chapter Three, Table 3.1).

Preferential vertical stacking of fluvial channel elements at one position in the outer margin of the aeolian dune field indicates a dune-field margin that has maintained a fixed position for a protracted period and may be associated with a limited distance of penetration of the fluvial system into an aeolian dune field (Figure 5.6). Consequently, the preserved sedimentary record might be characterised by limited lateral variations. The presence of relatively small and isolated fluvial channel elements in the central parts of otherwise aeolian-dominated successions may indicate a single thread fluvial channel that was characterised by a high-magnitude flow history, with associated floods that were able to penetrate a considerable distance into a dune-field centre setting; these types of incursion are relatively uncommon in modern systems.

Documented ancient examples of this type of behaviour are as follows: the Proterozoic Mancheral Quartzite, Sullavai Group, Pranhita-Godavari Valley, India (Chakraborty and Chaudhuri, 1993); the Lower to Middle Triassic Buntsandstein of north west Sardinia, Italy (Costamagna, 2012); the early

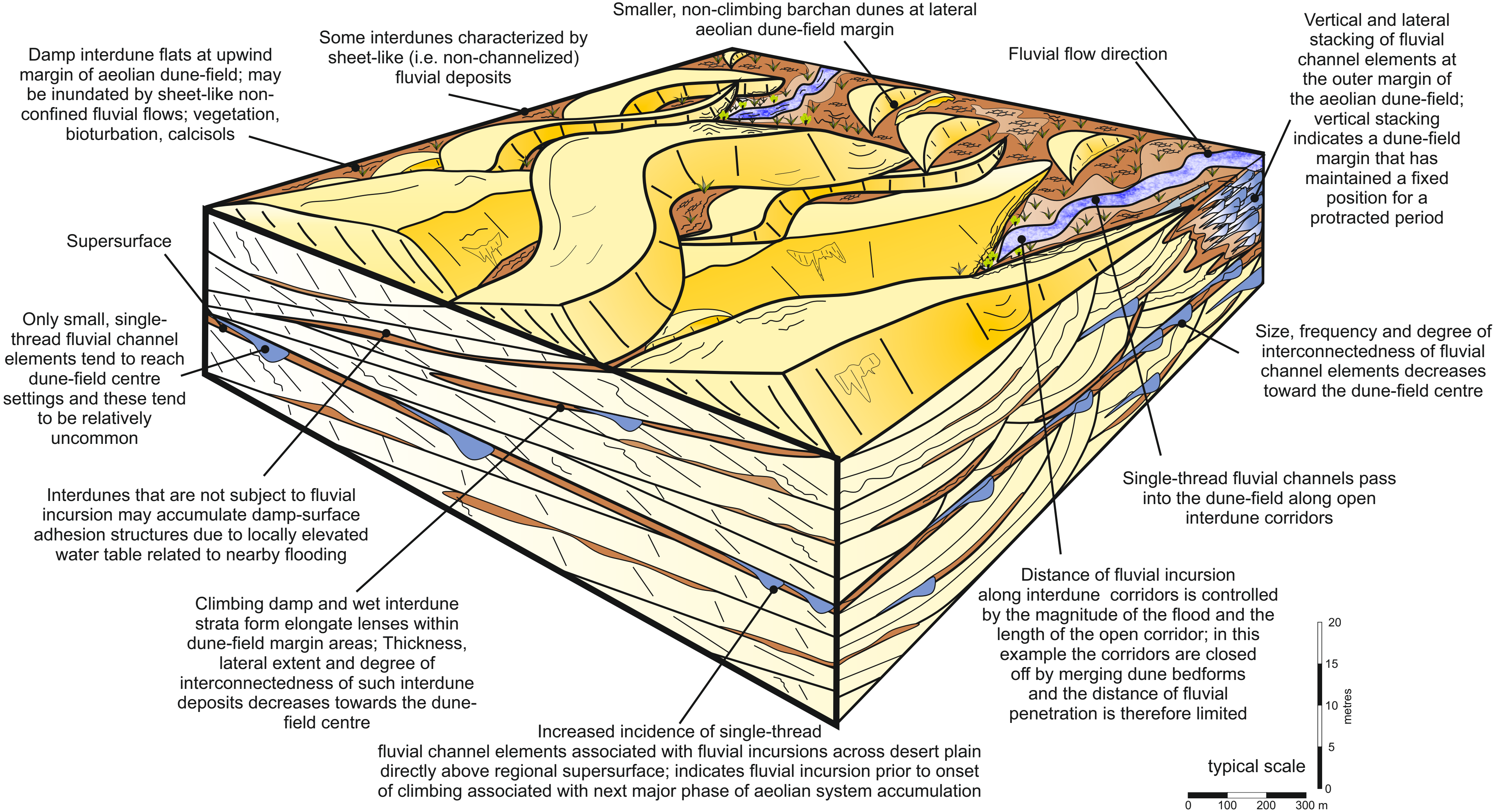


Figure 5.6: Single point source fluvial incursions into aeolian dune-fields.

Proterozoic Whitworth Formation, Mount Isa Inlier, Australia (Simpson and Eriksson, 1993); the Middle Devonian Lower Eday Sandstone, Orkney (Astin, 1985); and The Lower Triassic, western German Basin (Bourquin et al. (2006).

5.3.7 Fluvial incursion into aeolian dune-fields associated with a multiple sheet source

Distributive fluvial systems form networks of channels, commonly arranged into broad areas occupied by poorly-defined channels flow over low-gradient surfaces, where they pass out onto low relief desert plains (Cain and Mountney, 2009; Hartley et al., 2010; Weissmann et al., 2011). Along their width of intersection with aeolian dune-field margins, non-confined fluvial sheet-like bodies are characterised by shallow and poorly defined braided channel networks. These fluvial systems may pass between aeolian dunes along multiple adjacent interdune corridors, in some cases for distances of many tens of kilometres (Figure 3.1, Figure 3.2, and Figure 3.11). Such non-confined flows typically pass into dune-fields penecontemporaneously along multiple open interdune corridors with access gained from multiple points along the dune-field margin, spreading fluvial-derived sediment at several areas within dune fields (cf. Cain and Mountney, 2009; Figure 5.7).

Interdune flats at the upwind margins of aeolian dune-fields may be inundated by sheet-like non-confined fluvial flows generally at the fluvial termination point where the power of the stream flow is less efficient and less erosive. Interdune flats that are not subject to fluvial incursion may accumulate damp-surface adhesion structure due to locally elevated water table related to nearby flooding. In the central part of aeolian dune fields, where the fluvial processes are absent, dry interdune elements will develop.

In areas where fluvial flow can pass further into the aeolian system, erosion of aeolian sand occurs by water flowing through narrow neck areas between dunes. Deposition of fluvial and fluvially reworked aeolian sediment via fluvial processes occurs at the termini of interdunes where flood waters finally pond. This type of fluvial system may cover large areas at dune-field margins. Such fluvial systems are commonly ephemeral; flood frequency

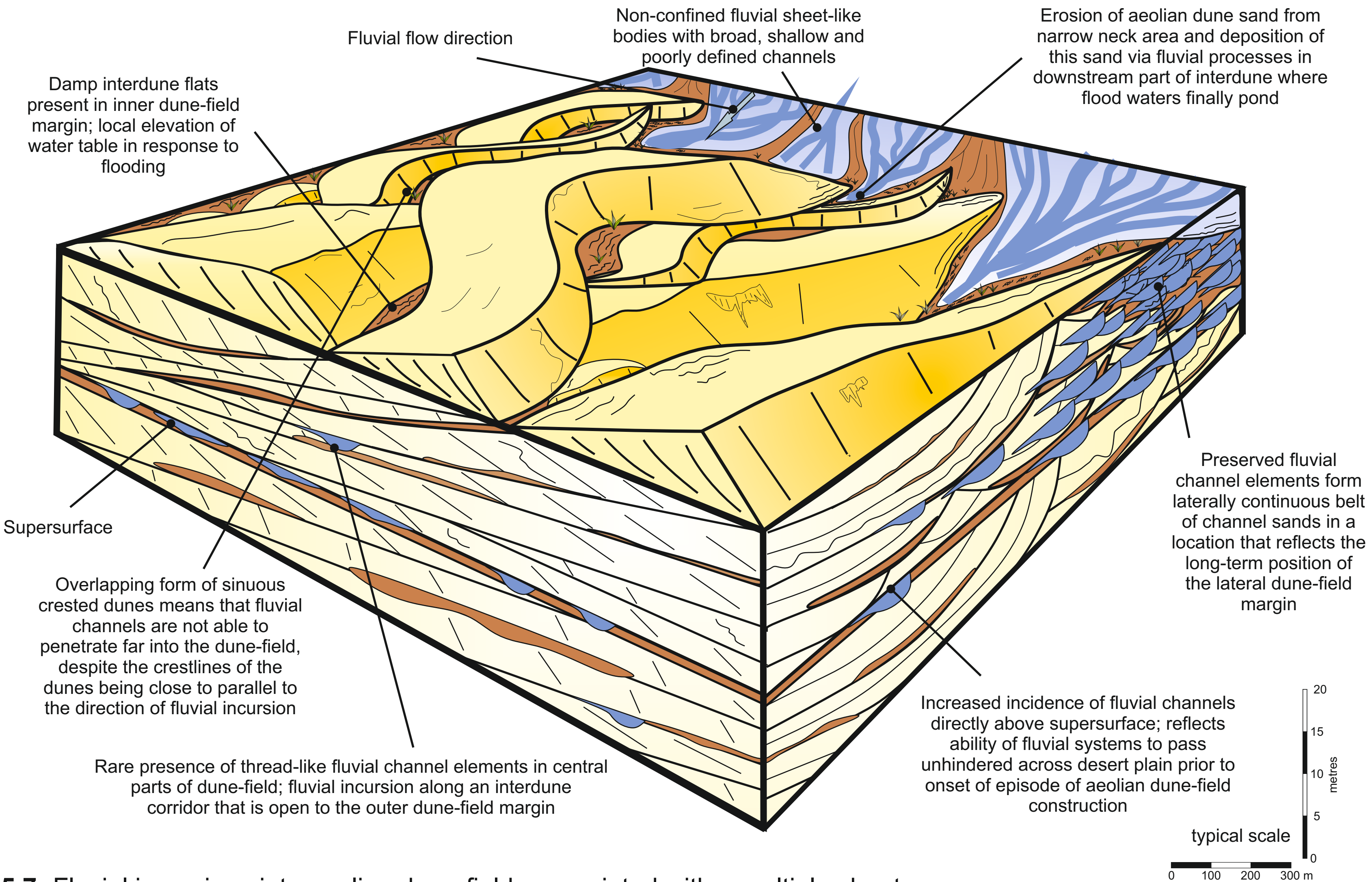


Figure 5.7: Fluvial incursions into aeolian dune-fields associated with a multiple sheet source.

and magnitude are directly linked to the changes in climate conditions, which may be seasonal. Thus, fluvial sedimentation may dominate in aeolian dune-field margin settings during wet seasons. The result is preserved braided fluvial channel networks that form a laterally continuous belt of sand. This reflects the long-term position of the lateral dune-field margin. In dry seasons, aeolian processes dominate.

Documented ancient examples of this type of behaviour are as follows: the Upper Rotliegend Group, UK south North Sea (Sweet, 1999); Cambro-Ordovician cratonic sheet sandstones of the northern Mississippi Valley, USA (Dott et al., 1986); the Tumblagooda Sandstone, Late Silurian, of west Australia (Trewin, 1993); the western part of the Germanic Basin, Olenekian, Early Triassic (Péron et al., 2005); the Siluro-Devonian Swanshaw Sandstone Formation, southwest Scotland (Smith et al., 2006); the Wolfville Formation, Late Triassic synrift succession of the Minas sub-basin, Bay of Fundy, Nova Scotia (Leleu and Hartley, 2010); and parts of the undifferentiated Cutler Group, Permian, Utah, USA (Venus et al., 2015).

5.3.8 Cessation of encroachment of aeolian dune-fields by fluvial systems

In this type of interaction, fluvial systems that experience sufficient flow discharge, either continuously or seasonally, are able to halt aeolian dune migration by flushing aeolian sand that blown into the channel system downstream. Thus, downwind encroachment of aeolian dune bedforms and entire aeolian dune-field margins is halted. One modern example of such behaviour is the Kunene River that defines the limit of the Skeleton Coast dune-field, Namibia (Figure 3.12, and Figure 5.8).

Where such fluvial systems define the downwind margin of entire aeolian dune fields, aeolian sediment will be carried away with each flood event to be re-deposited beyond the confines of the aeolian system. The final fate of such aeolian sediment may be a long-term sediment sink in some cases, for example, the Atlantic Ocean in the case of the Kuiseb River that defines the northern margin of the Namib Sand Sea. This type of interaction may leave

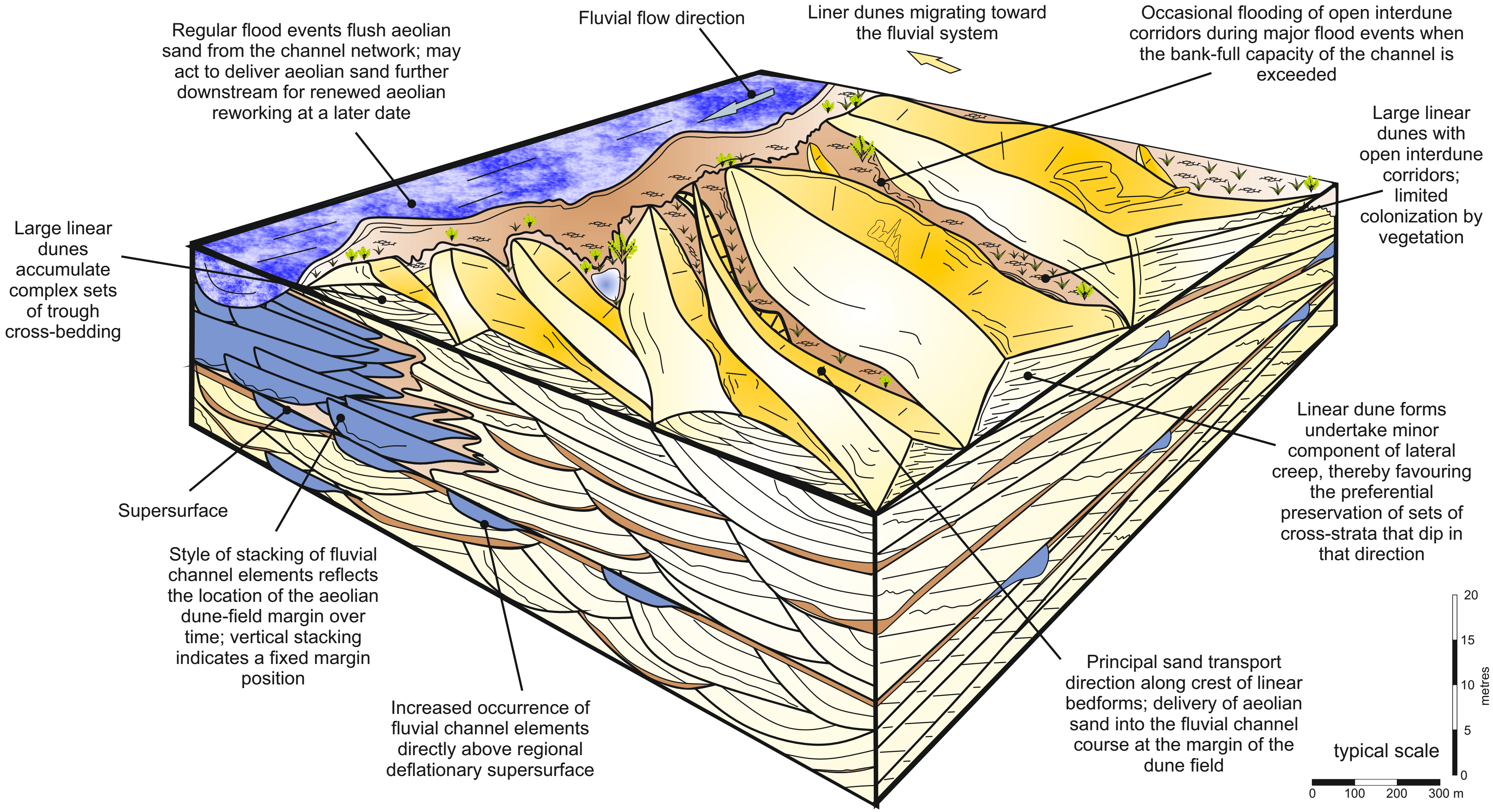


Figure 5.8: Cessation of encroachment of aeolian dune-fields by fluvial systems.
This model is analogous to the modern scenario at the northern margin of the main Namib Sand Sea where the Kuiseb River curtails the northward migration of large linear dunes.

no clear indication of interaction between competing aeolian and fluvial systems.

Rare major fluvial flood events that exceed the river bank-full capacity may flood the open interdune corridors, thereby generating a mixed and stacked deposits of fluvial and aeolian sediment within the aeolian dune-field, particularly in open interdune corridors developed adjacent to the fluvial system. The deposited fluvial sediment within the aeolian system will commonly be reworked later by aeolian processes. The aeolian and fluvial interaction in this configuration is limited to the site of juxtaposition of the aeolian and fluvial system. The vertical stacking of the fluvial channel elements at the boundary of the aeolian dune-field succession indicate a fixed erg margin position, and a limited interaction between the two systems.

Documented ancient examples of such behaviour include the Permian lower Culter beds, southeast Utah, USA, during episodes of humid climate conditions (Jordan and Mountney, 2010) and the Lower Jurassic Wingate Sandstone, northeastern Arizona (Clemmensen and Blakey, 1989).

5.3.9 Termination of fluvial channel networks in aeolian dune-fields

In this type of interaction, fluvial systems terminate within the inner parts of aeolian dune-fields (Figure 5.9). Fluvial channels terminate at points where interdune corridors narrow and close, for example where two adjacent dunes meet and overlap, or where aeolian dune forms become arranged perpendicular to the direction of fluvial incursion. Analysis of modern dune fields reveals that this type of interaction may occur within distances of 1.4 to 101 km from the outer margins of dune fields (Chapter Three, table 3.1).

The configuration of aeolian dunes in a dune field, along with the energy of the water flow each act to control the extent of fluvial flow into a dune field (Figure 3.13 and Figure 3.15b). This type of interaction is common in ephemeral fluvial systems (e.g., Trarza Desert, Mauritania; Figure 3.13c), and could occur in any part of an aeolian dune field depending on the energy of the flow.

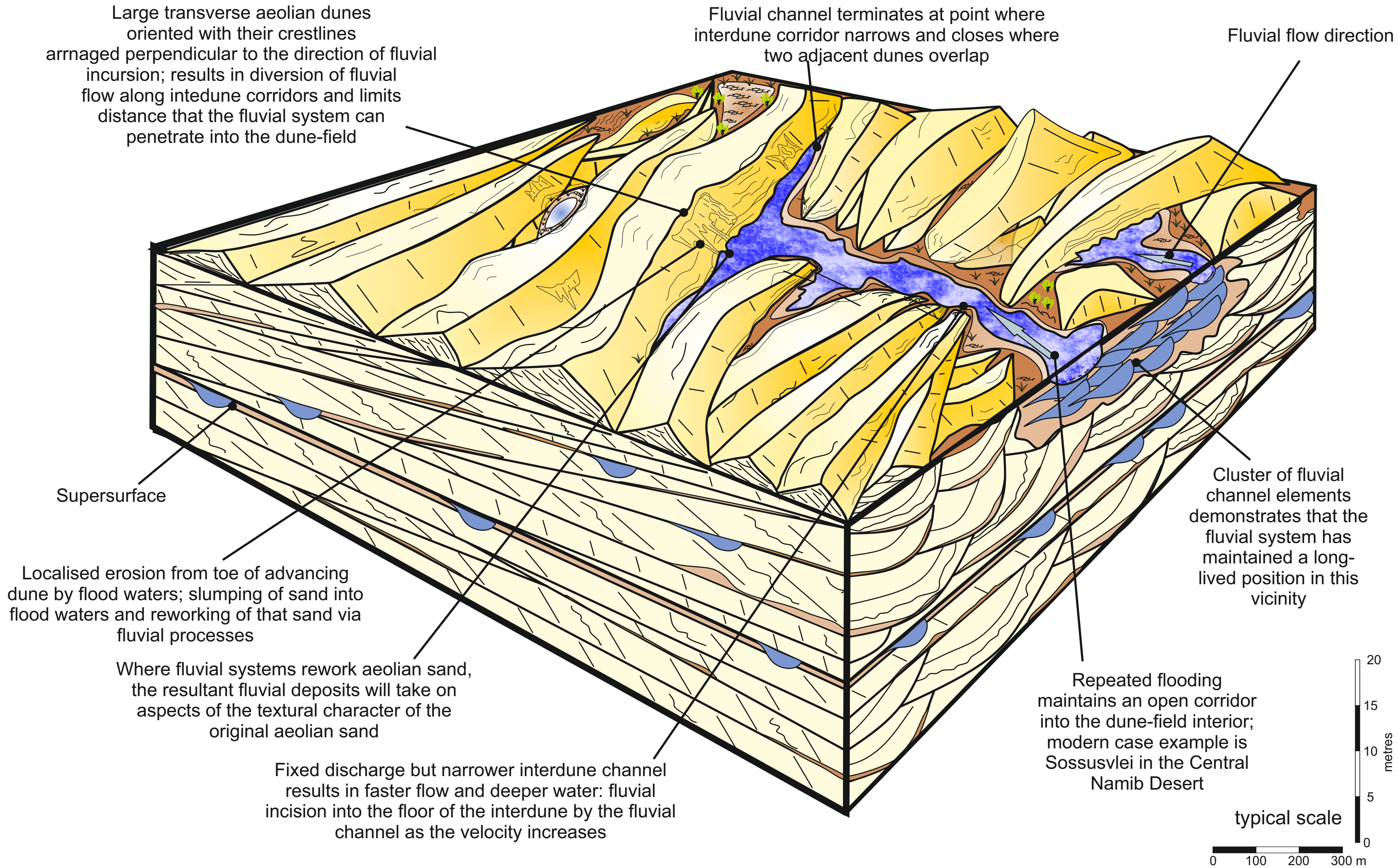


Figure 5.9: Termination of fluvial channel networks in aeolian dune-fields.

The sedimentary signature at the point of fluvial flow termination records the ponding of fluvial flow and the deposition of clay and fine silt from suspension in standing ponds in areas where fluvial flow has ceased (Reid and Frostick, 1987; Reid, 2002; Stanistreet and Stollhofen, 2002). This results in the accumulation of mud layers in interdune flat and playa areas. During dry seasons, aeolian systems become more active and the wind processes establish migrating aeolian dunes that may cover former flood deposits. Such active dunes may also form obstacles that curtail fluvial channel flow during later floods, thereby reducing the opportunity for future flood events to breach into the central parts of aeolian dune-fields during subsequent wet seasons (e.g., Figure 3.5; Mountney, 2006b). Fluvial incursions into aeolian dune fields may act to erode the lower flanks of aeolian dune bedforms, triggering aeolian processes such as dune sand avalanches (grainflows) and slumping of dune lee slopes. The slumping of aeolian sand into flood waters will lead to reworking of that sand by fluvial processes and later redeposition elsewhere within the aeolian dune field; the resultant fluvial deposits will take on aspects of the textural character of the original aeolian sand, making the differentiation of such sediment difficult based on its textural character alone. At the termination point where large-scale aeolian dune bedforms have acted to pond flood waters and limit the extent of fluvial incursion, playa deposits may result in the generation of a significant surface crust of calcite or gypcrete, especially where flood waters repeatedly pond (e.g, Sossusvlei, Namib Desert).

Documented ancient examples of such behaviour are as follows: aeolian-fluvial interaction in the Page Sandstone (Middle Jurassic) in south central, USA (Jones and Blakey, 1997); the late Cretaceous to middle Tertiary Nima Basin fill in the central Tibetan Plateau (DeCelles et al., 2007); and the Rush Springs Sandstone (Permian, Guadalupian) of western Oklahoma, USA (Poland and Simms, 2012).

3.5.10 Examples of short-term versus long-term fluvial-aeolian interaction

This type of interaction reflects the influence of climate on sedimentation in desert-margin settings. During relatively humid episodes, fluvial systems are more active and the level of the water table is high. This tends to generate successions that are fluvially dominated. In contrast, during relatively arid episodes the activity of aeolian processes is increased and accumulated fluvial deposits serve as a supply for aeolian construction of environments that are aeolian dominated (Figure 5.10).

In aeolian dune-field settings, interdune deposits reflect the nature of the substrate at the time of sediment accumulation: dry, damp and wet interdune types are all recognised (Mountney, 2006a). In the stratigraphic record, such dry, damp and wet interdune elements may thicken and thin slightly over space, and in some cases may pinch out laterally reflecting the expansion and contraction of the interdune flats in response to changes in dune size, morphology and type, reflecting changes in climate conditions, interdune size and water-table level. During humid episodes, aeolian preservation potential is enhanced by an elevated water-table level.

From a study of the response of an aeolian system to Holocene climate and hydrologic changes in the northern margin of Sahara Desert, Swezey et al. (1999) recorded cycles of aeolian deposits which reflect a phase of humid climatic conditions and high lake levels (a stabilised aeolian system), and an arid phase characterised by playa desiccation and deflation, and associated dune-field construction.

Documented ancient examples of such behaviour are as follows: interaction between aeolian and fluvial processes during accumulation of the Upper Cretaceous capping sandstone member of the Wahweap Formation, Kaiparowits Basin, USA (Simpson et al., 2008); the Jurassic Guara Formation, southern Brazil (Scherer and Lavina, 2005); the Sherwood Sandstone Group of East Irish Sea Basin, UK (Cowan, 1993); Copper Harbor Formation, Late Proterozoic, Lake Superior Basin, USA (Taylor and Middleton, 1990); the Middle Proterozoic Eriksfjord Formation, southwest Greenland (Tirsgaard and Øxnevad, 1998); the Triassic sandstones of Scrabo, County Down, Northern Ireland (Buckman et al., 1998); numerous Late Palaeozoic and Mesozoic examples from NW Europe and the Western Interior of the USA (Clemmensen et al., 1994); and the mid-Cretaceous

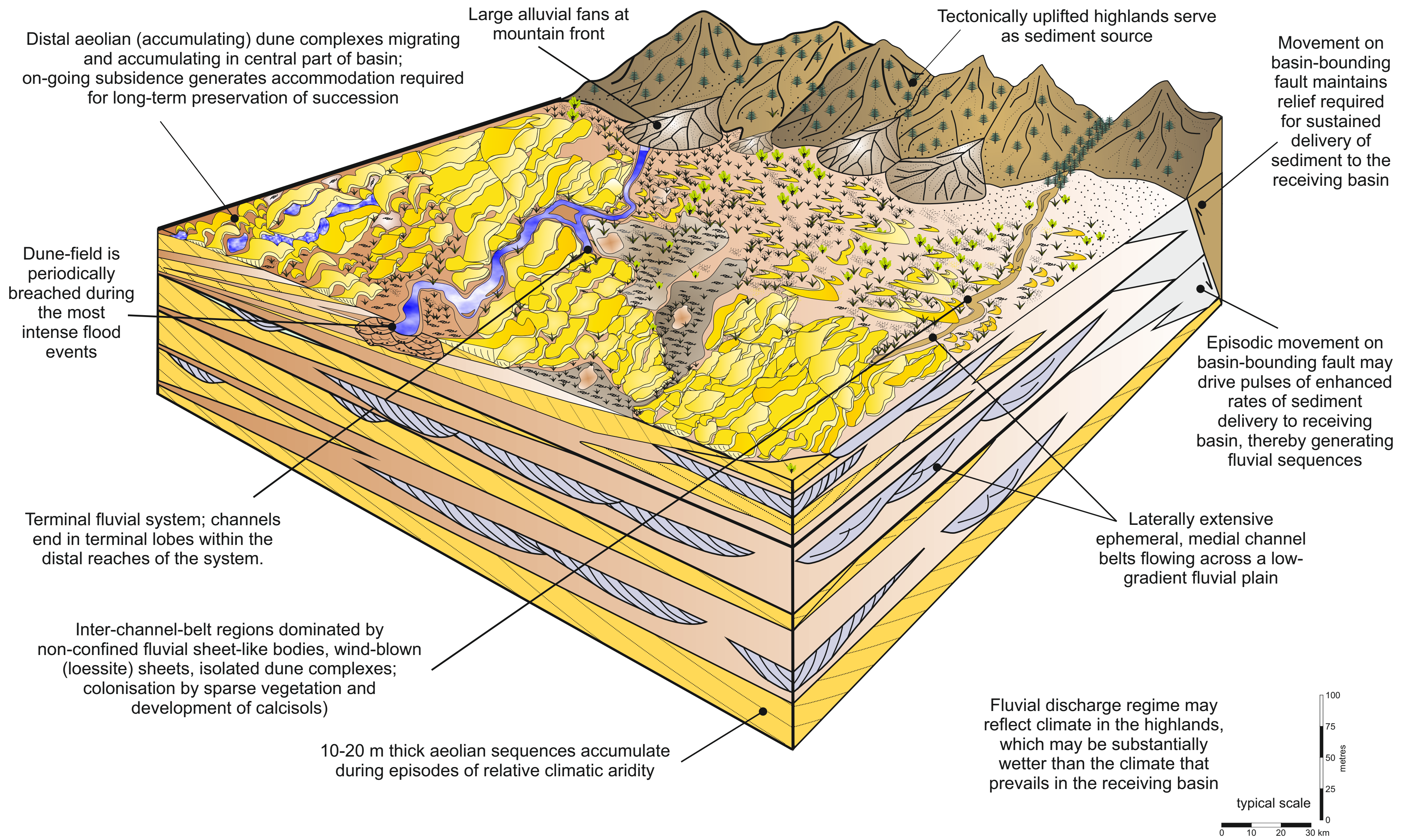


Figure 5.10: Long-term versus short-term styles of fluvial-aeolian interaction.

subtropical erg-margin system close to the Variscan Iberian Massif, Spain (Rodríguez-López et al., 2010).

5.4 Discussion

Types of fluvial-aeolian interactions and their resultant deposits are variable within the fluvial-aeolian systems investigated by this study. Characterisation of how this variability is expressed is important to gain an improved understanding of the processes involved in generating a preserved stratigraphic record in such settings and also in the evaluation of likely reservoir heterogeneity. Current models that describe different types of aeolian-fluvial interaction processes and their resultant accumulated products are still largely qualitative. Yet it is important to undertake quantitative studies to document the geometry of such interactions in ancient outcropping and subsurface successions to better predict reservoir behaviour (e.g., Bongiolo and Scherer, 2010). This study attempts to provide a series of semi-quantitative models for which the expected dimensions (and ranges thereof) of architectural elements are constrained.

The shape of interdune flats along which fluvial systems may pass is controlled by the morphology of the dune bedforms (Chapter Two, Section 2.7-2.9). At the dune-field margin, where much of the interaction between aeolian and fluvial processes occurs, dune and interdune size and shape exert an important control that defines the type of interaction between fluvial and aeolian systems. Linear dunes, for example, typically provide open corridors that promote the passage of fluvial flood water far into aeolian dune fields, especially when the flow is parallel to the trend of the straight crestlines of aeolian dune forms. In other settings, where the trend of elongate aeolian bedforms is arranged perpendicular to the trend of fluvial channels, damming of the fluvial flow acts to confine fluvial systems and their deposits solely to the outer margins of aeolian dune fields. Thus, the thickness and lateral extent of the mixed aeolian and fluvial deposits is affected by the morphology and arrangement of the aeolian dunes and associated interdunes in relation to the preferred fluvial flow direction.

One important effect of fluvial incursion into aeolian dune fields is the impact on the level of the water table. Water-table level determines the level to which deflation may occur. A rise in relative water table to a level close to the accumulation surface will lead to the generation damp interdunes and, where the water table level rises temporarily but repeatedly above the accumulation surface, wet interdune playas will form; the resultant deposits are characterised by sedimentary features such as adhesion structures and salt crusts. The thickness of these damp or wet interdune deposits is commonly a function of the length of time over which the interdune accumulation surface develops (Ahlbrandt and Fryberger, 1981).

5.5 Reservoir implications

Analysis of data acquired from a range of modern desert systems reveals the diversity of aeolian and fluvial interactions, which occur on a variety of scales (Chapter Three), and which are characterised by complex spatial variation in sedimentary architecture at the aeolian dune-field margin settings where such processes occur (cf. Mountney and Jagger, 2004). Aeolian-fluvial interaction processes and resultant sedimentological signatures give rise to preserved stratigraphic heterogeneities that influence reservoir properties and behaviour in such mixed depositional systems typically. Variable lateral and vertical facies arrangements characterise architectural elements composed of stratal units with markedly variable reservoir properties.

The limited volume of subsurface data available with which to recognise evidence for stratigraphic partitioning in mixed aeolian-fluvial reservoir successions is problematic (e.g., North and Boering, 1999). Therefore, more sophisticated facies models to better account for reservoir architecture and connectivity are required. Models for the characterisation of types of interaction between aeolian and fluvial processes and resultant deposits derived from both case-examples of modern (Chapter Two and Chapter Three) and ancient (Chapter four and this chapter) case studies can be applied to predict subsurface reservoir geology (cf. Langford, 1989; Langford and Chan, 1989). Consequently, prediction of the possible types of interaction serves as a useful tool when ranking targets within larger prospect areas.

The occurrence of fluvial elements embedded within successions that are otherwise dominated by aeolian elements generates successions that stratigraphically heterogeneous at a variety of lateral and vertical scales. Laterally, changes from a fluvial-dominated erg margin to an aeolian-dominated erg margin impact on regional variations in potential reservoir quality. Determination of types and length-scales of such interactions is required for play fairway mapping and the establishment of gross depositional environment (GDE) maps. Fluvial deposits within the aeolian system are significantly poorer in terms of porosity and permeability characteristics than the adjacent aeolian bodies and this influences reservoir quality (Chandler et al., 1989; Trewin, 1993; Bloomfield et al., 2006; Glennie, 2009). Coeval interactions between fluvial and aeolian systems (Chapter Three) take place within the aeolian system, where fluvial floods pass into interdune areas, thereby generating ribbon-like fluvial geometries (string-like sand bodies). Such ribbon-like geometries tend to act as baffles to lateral flow but will be less restrictive to vertical flow; the impact is generally local (e.g., Herries, 1993). For example, Figure 5.1 illustrates the parallel type of interaction between aeolian and fluvial systems, where fluvial channels travel in a direction parallel to the strike of the aeolian dune bedforms. The fluvial flow is confined by the aeolian dunes, thereby restricting the presence of the fluvial deposits to the interdune corridors. In some cases, thick accumulations of mud occur as a result of settling of sediment from suspension (clay and silt) at the termination point of flood waters within an erg setting (e.g., Krapf et al., 2003). There accumulation of such low permeability units that may act as baffles to fluid flow within reservoirs is significant.

In contrast, the expansion of a fluvial system at the expense of a retreating aeolian dune-field system in response to a change to wetter climatic conditions (Chapter Three, section 3.5.2), will tend to generate more laterally extensive sheet-like fluvial geometries that could significantly compartmentalise a reservoir (e.g., Fryberger, 1993; Herries, 1993).

Fluvial incursion into aeolian dunes usually interact with the lower part of the aeolian dunes (lee side avalanche deposits), which is the portion that most likely to be preserved in the rock record in a subsurface reservoir. This fluvial interaction leaves behind fine sediment that reduces the quality of the preserved aeolian facies. Therefore, this part will represent a significant permeability barrier that affect reservoir performance within an otherwise more permeable portion of an aeolian succession (e.g., Stanistreet and Stollhofen, 2002).

The juxtaposed occurrence of relatively permeable units of aeolian elements and relatively impermeable fluvial elements within a reservoir is significant; the aeolian elements will generally flow more effectively than neighbouring fluvial elements and this will typically lead to fast hydrocarbon depletion from the aeolian unit. Where aeolian elements are not in communication with other nearby aeolian elements, productivity problems may arise: a poorly swept reservoir, unpredictable flow rates, early water cut. The presence of relatively thin but highly permeable aeolian elements within a reservoir will have a significant impact on the flow rate during well testing: short-duration well testing may give inaccurate predictions of long-term reservoir performance (Cowan, 1993). Such factors will also affect reservoir gas or water injection: injected gas or water will preferentially flow through high-permeability aeolian layers, leaving the less permeable fluvial layers unswept (e.g., Wehr and Brasher, 1996). This can lead to early water cut. Thus, identifying the presence of aeolian facies deposited within otherwise fluvial-dominated successions is important.

5.6 Conclusions

Aeolian and fluvial processes interact in a complex variety of ways, both spatially and temporally, in most desert-margin settings. Such interactions generate a range of types of sedimentary interaction, the effects of which may be preserved in the ancient stratigraphic record. Interaction types combine to result in distinctive aeolian and fluvial deposits that accumulate in the marginal areas of aeolian dune-field systems and successions. Preserved sedimentologic and stratigraphic relationships provide evidence from which to establish the likely type of aeolian-fluvial interaction. This chapter has presented series of ten bespoke facies models that demonstrate different types of aeolian-fluvial interaction from dune-field margin settings. These ten semi-quantitative geological facies models have been developed based on analysis of modern systems from earlier chapters (Chapter Three and Four), and consideration of ancient literature-derived case-study examples. The facies models presented here account for the nature and origin of stratigraphic complexity present in aeolian dune-field margin successions that arise from the interplay of both autogenic and allogenic controls.

The effects of fluvial processes that operate in aeolian landscapes are significant. Floods modify interdune surfaces; dunes are partially eroded and finer sediments, such as mud, may be deposited in low-relief interdune flats. Muddy flood deposits in interdune-flat settings are resistant to deflation and increase the long-term preservation potential of underlying deposits, particularly when dune bedforms migrate over these interdune surfaces. Fluvial incursion can influence groundwater-table level; a high water table will promote aeolian system stabilisation and will limit the level to which aeolian deflation can proceed.

Understanding the nature and surface expression of various types of aeolian and fluvial interaction, and considering their resultant sedimentological expression, is important for prediction and interpretation of preserved deposits of such interactions that might be recognised in the ancient stratigraphic record. Assessment can be made of the spatial scale over which such interactions are likely to occur and this facilitates the prediction of net reservoir sandbody dimensions through models that constrain the geometry and lateral and vertical connectivity of sand bodies in reservoir successions. Assuming layer-cake correlations between neighbouring wells within stratigraphically complex reservoirs of mixed aeolian and fluvial facies is inappropriate; instead, a range of bespoke facies models should be utilised, each of which considers possible stratigraphic configurations and each of which has implications for likely reservoir performance.

Chapter Six
Conclusions and future work

This chapter summarises the generic outcomes of the research project. Further, it postulates possible future related research that could be undertaken to advance our present understanding of the interaction between aeolian and fluvial depositional systems.

6.1 Conclusions

Studies of modern desert dune fields allow geologists to draw conclusions about the controls that govern the development of spatial patterns of arrangement of desert landforms. This knowledge can be applied to predict the likely arrangement of architectural elements in preserved ancient desert successions. This serves as the basis for the development of more sophisticated facies, architectural-element and sequence stratigraphic models that can be applied in sedimentary geology generally and in reservoir geology specifically.

Geomorphological elements in desert settings tend to change systematically from the central parts of dune fields, where aeolian processes are dominant, to dune-field margins, where non-aeolian systems, including ephemeral fluvial streams dominate. Aeolian and fluvial processes operate coevally in most desert-margin settings to generate a range of types of sedimentary interaction that are documented from both modern arid systems and analogous ancient preserved outcrop successions. Such types of system interaction give rise to considerable complexity in terms of sedimentology and preserved stratigraphy. The physical boundary between geomorphic systems in hot deserts is dynamic such that facies belts undertake considerable lateral shift over time with the result that preserved sequence architectures exhibit complexity arising from system interactions that operate at a range of spatial and temporal scales from local to regional. An improved understanding of factors that govern these multiple scales of interaction is important for prediction of preserved stratigraphic architecture. Across desert margins where fluvial and aeolian systems interact, the location of assemblages of surface landforms may change gradationally or abruptly. An improved understanding of contemporary interactions serves as the basis for a database of modern analogues that can be used to account for types of aeolian-fluvial interactions preserved in the stratigraphic record.

Satellite imagery of dunes and interdunes in desert dune fields has provided the basis for an approach to qualitative and quantitative studies of patterns of arrangement of large-scale aeolian bedforms and adjoining interdunes in large aeolian sand seas. The collection and collation of data relating to primary landform morphology has enabled an improved understanding of modern desert sedimentary systems and the spatial arrangement of various sub-environments within these systems. In particular, the morphological changes and distributions of aeolian bedforms and interdunes across dune-field systems provides important information with which to improve our understanding of the likely arrangement of architectural elements in ancient aeolian preserved successions, several of which form important reservoirs for hydrocarbons.

Observations from modern dune-field margins have enabled the spatial rate of change of morphology of aeolian sub-environments to be characterised and described through empirical relationships. Results are enabling the proposition and development of a range of dynamic facies models for aeolian systems that can be used as predictive tools for subsurface reservoir characterisation. A combination of morphological and architectural data from a range of modern dune fields and their ancient counterparts preserved as successions in the geologic record can be used to constrain forward stratigraphic models for the prediction of aeolian reservoir heterogeneity. Such heterogeneity is likely to vary in three-dimensions within a reservoir volume.

The Rub'Al-Khali of south-eastern Saudi Arabia is covered by the latest generation of public-release satellite imagery, which reveals a varied range of dune types, the morphology of which changes systematically from central dune-field areas to marginal areas where aeolian interdunes, sand sheets, and ephemeral fluvial systems dominate. Analysis of geomorphic relationships between dune and interdune sub-environments within a series of modern dunes fields of the Rub' Al-Khali has been undertaken to document how the morphology, geometry, internal facies arrangement and relationship of the various depositional architectural elements produced by these geomorphic features vary over space from dune-field-centre to dune-field-margin settings. Analysis of this active modern dune-field system shows a characteristic reduction in aeolian dune size and degree of connectivity and a corresponding increase in interdune size and degree of connectivity towards outer dune-field margins.

A series of quantitative approaches have been employed to characterise the complexity present in a range of dune-field settings where large, morphologically complex and compound bedforms gradually give way to smaller and simpler bedform types at dune-field margins. The following parameters describing aspects of morphology have been measured: dune bedform height; elevation of interdune flats; along-crest length of a dune segment; bedform spacing; mean dune wavelength; maximum and minimum dune wavelength; amplitude of along-crest sinuosity; bedform long-axis orientation; distance of dune forms from a fixed point at the centre of the studied dune field to its outer margin. Additionally, attributes recorded for interdunes are as follows: interdune length; interdune width; interdune long-axis orientation; elevation; the relationship between the geometry of interdune depressions and their distance from a fixed point at the centre of the studied dune field in a direction toward its outer margin.

The analysis undertaken as part of this study regarding the geomorphic relationships between dune and interdune sub-environments within the modern active dune fields of the Rub' Al-Khali documents how the morphology, geometry, internal facies arrangement and relationship of the various depositional architectural elements produced by these geomorphic features vary over space from dune-field-centre to dune-field-margin settings. The data collected from the Rub' Al-Khali desert reveals a reduction in aeolian dune size and degree of connectivity and corresponding increase in interdune size and degree of connectivity toward the dune-field margins. The dunes changes from large compound and complex barchanoid bedforms (>155 m dune height) at the dune-field centre, to spatially isolated star dune forms and small barchan dunes at the dune-field margin, separated by water-table controlled interdunes that are >40 km in length. This has implications for understanding reservoir heterogeneity and predicting how low-porosity and permeability interdune areas may be distributed and may vary in geometry (shape and size) across ancient dune-field-margin successions.

The complex distribution of aeolian dunes and interdunes in any dune field defines their spatial heterogeneity distribution (which will affect reservoir behaviour), the impact of which typically becomes more significant later in the life of a hydrocarbon field (Sweet et al., 1996). By mapping and modelling the 2D and 3D distribution of dune and interdune architecture, it is possible to capture details of types of facies interactions, the extent of bounding surfaces, and the distribution of stratification types, all of which serve as fundamental controls on reservoir heterogeneity.

The physical boundaries between many desert geomorphic systems are dynamic. Along desert dune-field margins where aeolian and fluvial processes interact, the location of the boundary and the assemblage of surface landforms present may change either gradually or sharply over both space and time. Short-term shifts in the positions and form of such boundaries are controlled by the competition between fluvial flash-flood events and on-going aeolian dune construction. Medium- and long-term changes in boundary position and form are governed by changes in climate and tectonic basin evolution, respectively.

Aeolian and fluvial processes in desert-margin settings rarely operate independently: they are usually dynamically linked and exhibit a range of styles of sedimentary interaction documented from modern arid systems. Interactions between aeolian and fluvial systems are important and widespread in modern deserts, as revealed by analysis of global satellite imagery. A diverse range of styles of system interaction gives rise to considerable complexity in terms of geomorphology, sedimentology and preserved stratigraphy. Ten distinct styles of fluvial-aeolian interaction are recognized: fluvial incursions aligned parallel to trend of linear chains of aeolian dune forms; fluvial incursions oriented perpendicular trend of aeolian dunes; bifurcation of fluvial systems around the noses of aeolian dunes; through-going fluvial channel networks that cross entire aeolian dune fields; flooding of dune fields due to regionally elevated water-table levels associated with fluvial floods; fluvial incursions emanating from a single point source into dune fields; incursions emanating from multiple sheet sources; cessation of the encroachment of entire aeolian dune fields by fluvial systems; termination of fluvial channel networks into playas within aeolian dune fields; long-lived versus short-lived styles of fluvial incursion. Recognition of these interaction types forms the basis for a classification scheme that can be applied to desert dune-field systems generally.

Across desert margins where fluvial and aeolian systems interact, the location of assemblages of surface landforms may change gradationally or abruptly. The varied range of temporal and spatial scales over which aeolian-fluvial processes are known to interact means that simple generalised models for the classification of styles of interaction must be applied with caution when interpreting ancient preserved successions, especially those known only from the subsurface. By understanding the nature and surface expression of various types of aeolian and fluvial interaction, and by considering their resultant sedimentological expression, prediction can be made regarding how the of preserved deposits of such interactions might be predicted in the ancient stratigraphic record and assessment can be made of the spatial scale over which such interactions are likely to occur.

Given the economic importance and complex stratigraphic and sedimentologic nature of aeolian and fluvial successions, it is important to maximise subsurface hydrocarbon reservoir potential. Reservoir successions that exhibit marked lateral and vertical facies changes are complex such that flow behaviour is not easy to predict. Therefore, it has become essential to develop both qualitative and quantitative models with which to account for dynamic spatial and temporal aspects of aeolian-fluvial system behaviour at the dune-field and basin scales. The modelling-based approach and associated classification framework is a key objective of this wider research project, and it has potential applications in the development of predictive models with which to account for reservoir heterogeneity in aeolian reservoirs targeted for the production of hydrocarbons. Results from this project have been applied to generate a range of synthetic three-dimensional stratigraphic architectural models (cf. Mountney, 2012) with which to illustrate the range of possible sedimentological complexity likely to be present in preserved dune-field-margin successions (Al-Masrahy and Mountney, 2013 and 2015). Appreciation of this complexity has significant applied implications because interdune and dune-plinth elements typically act as principal and subordinate baffles to flow, respectively, in aeolian hydrocarbon reservoirs, whereas dune lee-slope elements typically represent effective net reservoir.

Reservoir anisotropy in aeolian successions profoundly affects reservoir performance throughout the producing life of a field. Although aeolian reservoirs are internally complex, they are predictable and can be managed efficiently once their three-dimensional internal architecture has been accurately characterised and modelled. Temporal and spatial variations in original dune and interdune morphology act as primary controls on resultant preserved set architecture. This study has quantified how aeolian dune, interdune and dryland fluvial morphological arrangements can be expressed in a variety of styles, in many cases predictably, across the zone transition from a dune-field centre to its margin. This represents an important step in the development of generic quantitative models with which to account for aeolian and mixed aeolian-fluvial reservoir architectural variability where changes are considered to occur spatially across a play, or within a single field. Each development project should be carefully characterised prior to initiating a more extensive drilling programme.

This study has utilised modern outcrop analogue data for the development of a suite of models designed to develop a bridging link between data provided by sedimentological studies and its appropriate application in the construction of reservoir models. The internal reservoir sedimentary architecture, together with the smaller-scale fabric and sedimentary structure of component lithofacies, ultimately control the path of fluid migration during oil and gas emplacement and subsequent extraction. This architecture is, in turn, the product of the depositional and diagenetic processes that created the sediment body. If an understanding of the sedimentological origin of the reservoir is developed, reservoir architecture, and hence fluid flow paths, become predictable (North and Prosser, 1993). In arid regions, it is common for fluvial and aeolian processes, and resultant strata to occur inter-mixed, with the result that overall preserved successions exhibit marked complexity (Glennie, 1990). Thus, to understand the fluid flow properties of mixed fluvial-aeolian reservoirs it is important to determine the geometry and the relationship of sedimentary bodies of fluvial and aeolian origin (Newell, 2001). The presence of stratigraphic complexity and heterogeneity at a scale below seismic resolution, coupled with stratigraphic architectures characterised by notable lateral facies changes, means that prediction of 3D stratigraphic architecture in subsurface reservoirs is challenging (e.g., Sweet, 1999). Therefore, studying appropriate outcrops and modern analogues is imperative to provide insight into reservoir heterogeneity and potential variability in geological models (e.g., Herrise, 1993; Mountney et al., 1998; North and Boering, 1999; Visser and Chessa, 2000; Newell, 2001; Bongiolo and Scherer, 2010).

6.2 Principal research findings of this study

1) The collection of data relating to primary desert dune-field landform morphology has enabled an improved understanding of the sediment system state of the modern Rub' Al-Khali Desert sedimentary system. Observed trends arise as a function of spatial changes in the sediment state of the system whereby sediment supply, the availability of that supply for transport and the sediment transporting capacity of the wind each combine to dictate the geomorphology of dune and interdune forms, which vary from thick accumulations of sands in the form of coalesced compound and complex barchanoid

bedforms in dune-field centre settings, to spatially discrete star dunes and small, spatially isolated barchan dunes separated by extensive water-table-controlled interdune flats in dune-field margin settings.

2) Observations from the modern Rub' Al-Khali Desert have enabled the spatial rate of change of morphology of aeolian sub-environments to be characterised and described through a series of empirical relationships.
3) Results of the study of the modern Rub' Al-Khali Desert have implications for developing an improved understanding of the likely controls on the detailed sedimentary architecture of preserved aeolian successions by enabling the proposition and development of a range of dynamic facies models for aeolian systems. This has wider applied implications and significance: for example, the morphological changes in the distribution of aeolian bedforms and interdunes across dune-field systems provides important information with which to improve our understanding of the likely arrangement of architectural elements in ancient aeolian preserved successions, several of which form important reservoirs for hydrocarbons. Results of this work represent an important step in the development of improved models for the characterisation of stratigraphic complexity and heterogeneity in aeolian reservoirs.
4) The physical boundaries between geomorphic systems are dynamic over short temporal time-scales. Across desert margins where fluvial and aeolian systems interact, the location of assemblages of surface landforms may change gradationally or abruptly. The varied range of temporal and spatial scales over which aeolian-fluvial processes interact means that simple generalised models for the classification of types of interaction must be applied with caution when interpreting ancient preserved successions, especially those known only from the subsurface. An improved understanding of interactions has been revealed by a database of modern systems and ancient preserved successions that can be used to account for types of aeolian-fluvial interactions preserved in subsurface reservoir intervals. From an applied standpoint, quantitative depositional models arising from this

database-driven approach serve to minimise uncertainties relating to stratigraphic heterogeneity in subsurface reservoir settings and aid inter-well correlation and prediction.

5) The study of aeolian and fluvial systems from the Triassic Wilmslow and Helsby Sandstone formation (Triassic, Cheshire Basins, UK) provide details of sedimentological characteristics of preserved aeolian and fluvial desert successions, including details on architectural elements and their internal lithofacies components and facies associations. Results are represented within the framework of a generalised depositional model that depicts the arrangement and relationships of fundamental components of an ancient aeolian and braided fluvial succession. This study has documented the preserved record of a wet aeolian system and an associated fluvial succession. It has developed further our understanding of processes that operate in aeolian dune-field margin settings, and more generally that operate in aeolian and fluvial systems under the influence of an arid to semi-arid climatic regime.
6) Outcrop observations of damp interdune elements preserved between aeolian dune elements record the impact of the water table on the development and preservation of a water-table-influenced aeolian system. Such arrangements have a direct impact on the vertical and lateral heterogeneity of subsurface aeolian reservoirs.
7) The aeolian system in the studied section of the Wilmslow Sandstone Formation preserves examples of both climbing and non-climbing dune-interdune behaviour of the aeolian system. This study, therefore, adds new examples of such system behaviours.
8) Results from this study are of importance in assessing the role of heterogeneity in partitioning hydrocarbon reservoirs and water aquifers, and for predicting lateral distributions of lithofacies between isolated wells, and the likely arrangement of non-reservoir units, such as interdunes and fluvial channel-abandonment and floodplain elements and associated lithofacies.

6.3 Future work

Although this research has addressed the originally stated aim and objectives, further investigation is required to address the different types of aeolian and fluvial system interaction from outcrop and subsurface data. Additional studies, particularly from subsurface successions, will provide evidence for and applications of the types of aeolian-fluvial interaction classification described here and, therefore, benefit hydrocarbon reservoir modelling by providing detailed information on reservoir heterogeneity and lateral connectivity within mixed aeolian-fluvial reservoirs. The following avenues of potential future research could be explored, results from which would complement this study.

1) Studies of modern desert dune-fields allow conclusions to be drawn regarding the spatial pattern of landforms in desert systems. This knowledge can be applied to predict the likely arrangement of elements in preserved ancient desert palaeoenvironments. This serves as the basis for the development of more sophisticated architectural-element and sequence stratigraphic models. Recent research has resulted in a rapid increase in the use of numerical modelling techniques in aeolian studies (e.g., Parsons, 2004; Rubin and Carter, 2006; Salles et al., 2011; Mountney, 2012) and fluvial studies (e.g., Geleynse et al., 2011; Geach et al., 2015). Building numerical forward stratigraphic models of aeolian dune and interdune architecture that can account for types of aeolian-fluvial interactions in 3D and 4D is important to better characterise and manage subsurface hydrocarbon reservoirs and groundwater aquifers. This especially important in aeolian reservoirs where subtle changes in porosity-permeability structure make the difference between net and non-net reservoir. A major outcome of this future research in this area will be the development of predictive, quantitative facies and sequence stratigraphic models, which will improve our understanding of the factors that control the distribution of aeolian and fluvial elements in subsurface reservoirs and aquifers.

2) A further avenue of future research should be the development of a quantitative database system that sample input parameters relating to aeolian depositional systems, architectural elements and lithofacies in order to construct reservoir models for development engineering purposes in petroleum industries. This type of database has already been established for fluvial depositional systems: the “Fluvial Architecture Knowledge Transfer System” or FAKTS (Colombera et al., 2012). FAKTS is a relational database that stores hard and soft data about fluvial sedimentary architecture and which has been populated with data derived from modern rivers, ancient outcrop successions and published literature data. This system can be used to construct a practical architectural model for reservoir planning based on limited input data available from preliminary exploration and interpretation of facies (Colombera et al., 2013; Miall, 2014). The development of a similar database for digital reproduction of all the essential features of aeolian sedimentary architecture that accounts for the style of internal organization of aeolian dune forms, their geometries and spatial distribution would be a useful tool to predict the dimensions of architectural elements and the arrangement of neighbouring elements away from the borehole, especially in areas where the subsurface data is limited.
3) Three-dimensional geological computer modelling is one of the important, most innovative and most widely applied tools for reservoir management purposes (e.g., Robinson et al., 2008). One potential future research avenue will be to utilise a forward stratigraphic modelling approach to predict preserved stratal architectures and facies distributions for aeolian and mixed aeolian-fluvial successions based on a combined process-based and geometrical modelling approach. Such models should ideally be integrated fully within industry standard reservoir modelling software tools such as Schlumberger Petrel, Landmark DecisionSpace and Bakker Hughes JewelSuite. Resultant reservoir models would be able to demonstrate the likely variation in the arrangement and geometry of mixed aeolian and fluvial geo-bodies that contribute to net versus non-net reservoir

units in a subsurface reservoir successions. This could be achieved using stochastic modelling techniques that make use of a mathematical model to incorporate pseudo-random possibilities in a given realisation or set of realisations (Zou, 2013). Stochastic models are useful in generating multiple realisations of the spatial distribution of depositional systems sediment properties in settings where data density is limited or insufficient to construct a unique deterministic facies architecture model (Geel and Nonselaar, 2007). Conceptual geological models (e.g., aeolian-fluvial interactions models, chapter 5), could be used to constrain numerical models that capture the expected range of possible spatial arrangements of elements in aeolian-fluvial systems, and the facies bodies that are likely to be present in the subsurface. Such an approach could be used to generate training images to integrate geological information into reservoir models (e.g., Strebelle and Levy, 2008).

References

Abd El Rahman, A. A., 1986. The deserts of the Arabian Peninsula. In: Evenari, M., Noy-Meir, I., Goodall, D.W (Eds.), *Ecosystems of the world 12B-hot deserts and arid Shrublands*. Elsevier, Amsterdam, pp. 29-54.

Abdullatif, O.M., 1989. Channel-fill and sheet-flood facies sequences in the ephemeral terminal River Gash, Kassala, Sudan. *Sedimentary Geology* 63, pp. 171-184.

Ahlbrandt, T.S., Fryberger, S.G., 1981. Sedimentary features and significance of interdune deposits. In: Ethridge, F.G., Flore, R.M. (Eds.), *Recent and ancient non-marine depositional environments: models for exploration.* Society of Economic Paleontologists and Mineralogists, Special Publication 31, pp. 293-314.

Ahlbrandt, T.S., Fryberger, S.G., 1982. Introduction to eolian deposits. In: Scholle, P.A., Spearing, D. (Eds*.), Sandstone depositional environments. American Association of Petroleum Geologist*, Tulsa, USA, pp. 11-47.

Ahmed, W., 2008. Contrast in clay mineralogy and their effect on reservoir properties in sandstone formations, *Bulletin of the Chemical Society of Ethiopia* 22, pp. 41-65.

Ali, A.S., Clark, W.J., Dribus, J.R., 2010. Diagenesis and reservoir quality. *Oilfield Review 22*, pp. 14-27.

Al Farraj A., Harvey, A.M., 2004. Late Quaternary interactions between aeolian and fluvial processes: a case study in the northern UAE. *Journal of Arid Environments* 56, pp. 235-248.

Allen, J.R.L., 1968. *Current Ripples: Their relation to patterns of water and sediment motion*. North-Holland Publishing Company, Amsterdam. 433 pp.

Allen, J.R.L., 1970. The avalanching of granular solids on dune and similar slopes. *Journal of Geology* 78, pp. 326-351.

Allen, J.R.L., 1982. *sedimentary structures: their character and physical basis*, Volume 1. Elsevier, Amsterdam. 611 p.

Allen, J. R. L., 1984. Sedimentary Structures: Their character and physics basis, Elsevier, Amsterdam. *Development in Sedimentology* 30, 593 p.

Allen, J.R., Leeder, M.R., 1980. Criteria for the instability of upper-stage plane beds. *Sedimentology* 27, pp. 209-217.

Al-Masrahy, M.A., Mountney, N.P, 2013. Remote sensing of spatial variability in aeolian dune and interdune morphology in Rub' Al-Khali, Saudi Arabia. *Aeolian Research* 11, pp. 155-170.

Al-Masrahy M. A., Mountney N.P., 2015. A classification scheme for fluvial–aeolian system interaction in desert-margin settings. *Aeolian Research* 17, pp. 67-88.

Al-Masrahy, M. A., Melvin, J., Abdulatif, O., Makkawi, M., 2012. Core and image log analysis of paleowind directions in the lower Permian Unayzah-A reservoir, subsurface central and eastern Saudi Arabia. *American Association of Petroleum Geologist, Search and Discovery,* Article 50734, pp. 1-4.

Alsharhan, A.S., Glennie, K.W., Whittle, G.L., Kendall, C.G., 1998. *Quaternary deserts and climate change*. Proceedings of the international conference on Quaternary deserts and climate change, Al Ain, UAE, December 9-11, 1995. A.A. Balkema. 621 p.

Alsharhan, A.S., Wood, W. W., Goudie, A.S., Fowler, A., Abdellatif, E.M., 2003. *Desertification in the third millennium*. Proceedings of an international conference, Dubai, February 12-15, 2000. A. A. Balkema. 490 p.

Andrews, 1981. Sedimentology of Great Sand Dunes National Monument. In: Etheridge, F.G., Flores, R.M. (Eds.), *Recent and ancient nonmarine depositional environments: models for exploration*. Society of Sedimentary Geology, Special Publication 31, pp. 279-292.

Anderson, S.P., Anderson, R.S., 1990. Debris-flow benches: dune-contact deposits record paleo-sand dune positions in north Panamint Valley, Inyo County, California. *Geology* 18, pp. 227-42.

Anderson R.S., Haff, P.K., 1991. Wind modification and bed response during saltation of sand in air. *Acta Mechanica* 1, pp. 21-52.

Anderson, R.S., Bunas, K.L., 1993. Grain size segregation and stratigraphy in aeolian ripples modeled with a cellular automaton. *Nature* 365, pp. 740-743.

Anton, D., 1984. Aspects of geomorphological evolution: paleosols and dunes in Saudi Arabia. In: Jado, R.A., Jotl, J. (Eds.), *Quaternary Period in Saudi Arabia.* Springer-Verlag, Wien, pp. 275-297.

Anton, D., 1985. Environmental changes and aeolian dynamics during the Quaternary in the Arabian Peninsula. In: El-Baz, F., El-Tayeb, I.A., Hassan, M.H. (Eds.), *Proceedings of the international workshop on sand transport and desertification in Arid Lands*, Khartoum, Sudan. World Scientific, Singapore, pp. 209- 234.

Arzani, H., 2005. The fluvial megafan of Abarkoh Basin (Central Iran): an example of flash-flood sedimentation in arid lands. In: Harvey, A.M., Mather, A.E., Stomps, M. (Eds.), *Alluvial fans: geomorphology, sedimentology, dynamics*. Geological Society of London, Special Publication 251, pp. 41-59.

Ashour, M.M., 2013. Sabkhas in Qatar Peninsula. landscape and geodiversity. *Studies of Integrated Geography* 1, pp. 10-35.

Ash, J.E., Wasson, R. J., 1983. Vegetation and sand mobility in the Australian desert dunefield. *Zeitschrift Für Geomorphologic* 45, pp. 7-25.

Ashworth, P.J., Sambrook Smith, G.H., Best, J.L., Bridge, J.S., Lane, S.N., Lunt, I.A., Reesink, A.J.H., Simpson, C.J., Thomas, R.E., 2011. Evolution and sedimentology of a channel fill in the sandy braided South Saskatchewan River and its comparison to the deposits of an adjacent compound bar. *Sedimentology* 58, pp. 1860-1883.

Astin, T. R., 1985. The palaeogeography of the Middle Devonian Lower Eday Sandstone, Orkney. *Scottish Journal of Geology* 21, pp. 353-375.

Atallah, M., Saqqa, W., 2004. Characterization of aeolian terrain facies in Wadi Araba desert, southwestern Jordan. *Geomorphology* 62, pp. 63-87.

Baas, Andreas C., 2007. Complex systems in aeolian Geomorphology. *Geomorphology* 91, pp. 311-331.

Bagnold, R.A., 1951. Sand formations in southern Arabia. *Geological Journal* 117, pp. 78-86.

Banham, S.G., Mountney, N.P., 2013. Evolution of fluvial systems in salt-walled mini basins: a review and new insights. *Sedimentary Geology* 296, pp. 142-166.

Banham, S.P., Mountney, N.P. 2014, Climatic versus halokinetic control on sedimentation in a dryland fluvial succession. *Sedimentology* 61, pp. 570-608.

Barth, H.J., 2001. Characteristics of the wind regime north of Jubail, Saudi Arabia, based on high resolution wind data. *Journal of Arid Environments* 47, pp. 387-402.

Basilici, G, Führ Dal' Bó, P.F, Bernades Ladeira, F.S., 2009. Climate-induced sediment-palaeosol cycles in a Late Cretaceous dry aeolian sand sheet: Marília Formation (North-West Bauru Basin, Brazil). *Sedimentology* 56, pp. 1876-1904.

Belnap,J., Munson, S.M., Field, J.P., 2011. Aeolian and fluvial processes in dryland regions: the need for integrated studies. *Ecohydrology* 4, pp. 615-622.

Benton, M.J., Warrington, G., Newell, A.J., Spencer, P.R., 1994. A review of the British Triassic tetrapod assemblages. In: Fraser, N.C., Sues, H.D. (Eds.), *In the Shadow of the Dinosaurs*, Cambridge University Press, Cambridge, pp.131-160.

Berg, R.R., 1986. *Reservoir sandstone*, Prentice-Hall, Inc. Englewood Cliff, New Jersey 07632. 481 p.

Best, 2005. The fluid dynamics of river dunes: A review and some future research directions. *Journal of Geophysical Research* 110, (F04S02), doi:10.1029/2004JF000218.

Beydoun, Z.R., 1966. Geology of the Arabian Peninsula – Eastern Aden protectorate and part of Dhufar, U. S. Geological Survey Professional Paper, 560. *Geological Society of London, Special Publication* 49, pp. 697-714.

Bishop, M.A., 2010. Nearest neighbor analysis of mega-barchanoid dunes, Ar Rub' Al- Khali, sand sea: The application of geographical indices to the understanding of dune field self-organization, maturity and environmental change. *Geomorphology* 120, pp. 186-194.

Blakey, R.C., 1988. Basin tectonics and erg response. *Sedimentary Geology* 56, pp. 127-151.

Blakey, R.C., Havholm, K.G. and Jones, L.S., 1996. Stratigraphic analysis of eolian interactions with marine and fluvial deposits, Middle Jurassic Page Sandstones and Carmel Formation, Colorado Plateau, U.S.A. *Journal of Sedimentary Research* 66, pp. 324-342.

Blair, T.C., 1999. Cause of dominance by sheetflood vs. debris-flow processes on two adjoining alluvial fans, Death Valley, California. *Sedimentology* 46, pp. 1015-1028.

Blair, T.C., McPherson, J.G., 1994. Alluvial fan processes and forms. In: Abrahams, A.D., Parsons, A. (Eds.), *Geomorphology* of desert environments. Chapman Hall, London, pp. 354-402.

Blair, T.C., McPherson, J.G., 2009. Alluvial fan process and forms, In: Parsons, A.J., Abrahams, A.D. (Eds.), *Geomorphology* of Desert Environments, 2nd edition. Springer, pp. 413-467.

Blaszczyk J. K., 1981. Palaeomorphology of Weissliegendes top as the control on facies variability in ore-bearing series of the Lubin Copperfield, southwestern Poland. *Geologia Sudetica* 16, pp. 195-217.

Bloch, S., 1991, Empirical prediction of porosity and permeability in sandstones. *American Association of Petroleum Geologists, Bulletin* 75, pp. 1157-1160.

Bloomfield, J.P., Moreau, M.F., Newell, A.J., 2006. Characterization of permeability distributions in six lithofacies from the Helsby and Wilmslow sandstone formations of the Cheshire Basin, UK. In: Barker, R. D., Tellam, J. H. (Eds.), *Fluid flow and solute movement in sandstones: the onshore UK Permo-Triassic Red Bed Sequence.* Geological Society of London, Special Publication 263, pp. 83-101.

Blum, M., Kocurek, G.A., Deynoux, M., Swezey, C., Lancaster, N., Price, D.M. Pion, J-C., 1998. Quaternary wadi, lacustrine, aeolian depositional cycles and sequences, Chotta Rharsa basin, southern Tunisia. In: Alsharhhan, A.S., Glennie, K.W.,Whittle, G.L., Kendall, C.G. (Eds.), *Quaternary deserts and climate change*. Balkema, Rotterdam, pp. 539-552.

Blumberg D.G., 2006. Analysis of large aeolian (wind-blown) bedforms using the Shuttle Radar Topography Mission (SRTM) digital elevation data. *Remote Sensing of Environment* 100, pp. 179-189.

Bongiolo, D.E., Scherer, C.M.S., 2010. Facies architecture and heterogeneity of the fluvial aeolian reservoirs of the Sergi formation (Upper Jurassic), Recôncavo Basin, NE Brazil. *Marine and Petroleum Geology* 27, pp. 1885-1897.

Bordy, E.M. and Catuneanu, O., 2002. Sedimentology and palaeontology of upper Karoo aeolian strata (Early Jurassic) in the Tuli Basin, South Africa. *Journal of African Earth Sciences* 35, pp. 301-314.

Bourke, 2003. A small-scale throughflow outburst in a heterolithic ephemeral channel. *Journal of Sedimentary Research* 73, pp. 2-4.

Bourquin, S., Guillocheau, F., Peron, S., 2009. Braided rivers within an arid alluvial plain (example from the Lower Triassic, western German Basin): recognition criteria and expression of stratigraphic cycles. *Sedimentology* 56, pp. 2235-2264.

Breed, C.S., Grow, T., 1979. Chapter J. Morphology and distribution of dunes in sand seas observed by remote sensing. In: McKee, E.D. (Ed.), *A Study of Global Sand Seas*. U.S Geological Survey, Washington, pp. 253-302.

Breed, C.S., Fryberger, S.G., Andrews, S., McCauley, C., Lennartz, F., Gebel, D. and Horstman, K.,1979. Chapter K. Regional studies of sand seas, using land sat (ERTS) imagery. In: McKee, E.D. (Ed.), *A study of global sand seas.* U. S. Geological Survey, Washington, pp. 305-397.

Bridge, J.S., 1993. The interaction between channel geometry, water flow, sediment transport and deposition in braided rivers. *Geological Society of London*, *Special Publications* 75, pp. 13-71.

Bridge, J.S., 2003. *Rivers and floodplains*. Oxford, U.K., Blackwell. 504 pp.

Bridge, J.S., 2006. Fluvial facies models: recent developments. In: H.W., Walker, R.G., (Ed.), *Facies models revisited*. International Association of Sedimentologists, Special Publication 84, pp. 85-170.

Bridge, J.S., Andmackey, S.D., 1993. A revised alluvial stratigraphy model. In: Marzo, M., Puigdefa´bregas, C., (Eds.), *Alluvial sedimentation.* International Association of Sedimentologist, Special Publication 17, pp. 319-336.

Bridge, J.S., Tye, R.S., 2000. Interpreting the dimensions of ancient fluvial channel bars, channels, and channel belts from wireline-logs and cores. *American Association of Petroleum Geologist*, *Bulletin* 84, pp. 1205-122.

Bridge, J.S., Demicco, 2008. *Earth surface processes, landforms and sediment deposits*. Cambridge, University Press. Cambridge, UK, 815 pp.

Bristow, C., Mountney, N.P., 2013. Aeolian stratigraphy. In: John F. S. (Ed.), *Treatise on geomorphology* 11, pp. 246-268.

Bristow, C.S., 1988. Controls on the sedimentation of the Rough Rock Group (Namurian) from the Pennine Basin of northern England. In: Besly, B.M., Kelling, G. (Eds.). *Sedimentation in a synorogenic basin complex; the Upper Carboniferous of Northwest Europe.* Blackie, Glasgow, pp. 114-131.

Brookfield, M.E., 1977. The origin of bounding surfaces in ancient aeolian sandstone. *Sedimentology* 24, pp. 303-332.

Brookfield, M.E., Ahlbrandt, T.S., 1983. *Eolian sediment and processes.* Development in Sedimentology 38, Elsevier, Amsterdam, 660 p.

Brookfield M.E., Silvestro, S. 2010. Eolian systems. In: James, N.P., Dalrymple, R.W. (Eds.), *Facies models*. Geological Association of Canada, GEOtxt 6. St Johns, Newfoundland, pp.139-166.

Brown, G.F., 1960. Geomorphology of western and central Saudi Arabia. *21st International geological Congress*, Copenhagen, pp. 150-159.

Brown, G.F., Layne, N., Goudarzi, G.H., MacLean, W.H., 1963. Geologic map of the northeastern Hijaz quadrangle, Kingdom of Saudi Arabia: *US Geological Survey Miscellaneous Geologic Investigation*. Map I-205A.

Buckman, J.O., Doughty, P.S., Benton, M.J., Jeram, A.J., 1998. Palaeoenvironmental interpretation of the Triassic sandstones of Scrabo, County Down, Northern Ireland; ichnological and sedimentological studies indicating a mixed fluviatile aeolian succession. *Irish Journal of Earth Sciences* 16, pp. 85-102.

Bull, L.J., Kirkby, M.J., 2002. Dryland river characteristics and concepts. In: Bull, L.J., Kirkby, M.J. (Eds.), *Dryland rivers: hydrogeology and geomorphology of semi-arid channels*. John Wiley and Sons, Chichester, pp. 3-15.

Bullard, J.E., Livingstone, I., 2002. Interactions between aeolian and fluvial systems in dryland environments. *Area* 34, pp. 8-16.

Bullard, J.E., McTainsh, G.H., 2003. Aeolian-fluvial interactions in dryland environments: scales, concepts and Australia case study. *Progress in Physical Geography* 27, pp. 471-501.

Bullard, J.E., White, K., Livingstone, I., 2011. Morphometric analysis of aeolian bedforms in the Namib sand sea using ASTER data*. Earth Surface Processes and Landforms* 36, pp. 1534-1549.

Bunker, D.G., 1953. The southwest borderlands of the Rub Al-Khali. *Geological Journal* 119, pp. 420-430.

Burckhardt, J.L., 1829. *Travels in Arabia* . Henry Colburn, London.

Burley, S.D., 1984. Patterns of diagenesis in the Sherwood Sandstone Group (Triassic), United Kingdom. *Clay Minerals* 19, pp. 403-440.

Butler, D., 2006. The web-wide world. *Nature* 439, pp. 776-778.

Cain, S.A., Mountney, N.P., 2009. Spatial and temporal evolution of a terminal fluvial fan system: the Permian Organ Rock Formation, Southeast Utah, USA. *Sedimentology* 56, pp. 1774-1800.

Cain, S.A., Mountney, N.P., 2011. Downstream changes and associated fluvial-aeolian interactions in an ancient terminal fluvial fan system: the Permian Organ Rock Formation, SE Utah. In: Davidson, S., Leleu S., North, C., (Eds.), *From river to rock record.* Society for Sedimentary Geology, Special Publication 97, pp. 165-187.

Calvache, M.L, Viseras, C., Ferrkndez, J., 1997. Controls on fan development – evidence from fan morphometry and Sedimentology; Sierra Nevada, SE Spain. *Geomorphology* 21, pp. 69-84.

Cant, D.J., 1982. Fluvial facies model and their application. In: Scholle, P.A., Spearing, D. (Eds.), *Sandstone depositional environments.* American Association of Petroleum Geologist, Tulsa, USA, pp.115-137.

Cant, D.J., Walker, R.G., 1978. Fluvial processes and facies sequences in the sandy braided South Saskatchewan River, Canada. *Sedimentology* 25, pp. 625-648.

Carr-Crabaugh, M. and Kocurek, G., 1998. Continental sequence stratigraphy of a wet eolian system: A key to relative sea-level change, In: Shanley, K., McCabe, P. (Eds.), *Relative roles of eustasy, climate, and tectonism in continental rocks*. Society of sedimentary Research, Special Publication 59, pp. 213-228.

Carrigy, M.A., 1970. Experiment on the angles of repose of granular materials. *Sedimentology* 14, pp. 147-158.

Carruthers, R. A., 1987. Aeolian sedimentation from the Galtymore Formation (Devonian), Ireland. In: Frostick, L., Reid, I. (Eds), *Desert sediments : ancient and modern*. Geological Society of London, Special Publication 35, pp. 251-268.

Chadwick, R.A., 1997. Fault analysis of the Cheshire Basin, NW England. In: Meadows, N.S., Trueblood, S.P., Hardman, M., Cowan, G. (Eds.), *Petroleum geology of the Irish Sea and adjacent areas.* Geological Society of London, Special Publication 124, pp. 297-313.

Chadwick, R.A., Evans, D.J., 1995. The timing and direction of Permo-Triassic extension in southern Britain. In: Boldy, S.A.R. (Ed.), *Permian and Triassic rifting in northwest Europe*. Geological Society of London, Special Publication 91, pp.161-192.

Chakraborty, T., Chaudhuri, A.K., 1993. Fluvial-aeolian interaction in a Proterozoic alluvial plain: example from the Mancheral Quartzite, Sullavai Group, Pranhita-Godavari Valley, India. *Geological Society of London*, *Special Publication* 72, pp.127-141.

Chan, M.A., 1989. Erg margin of the Permian White Rim Sandstone, SE Utah. *Sedimentology* 36, pp. 235-251.

Chan, M.A., 1999. Triassic loessite of North-Central Utah: stratigraphy, petrophysical character, and paleoclimate implications. *Journal of Sedimentary Research* 69, pp. 477-485.

Chandler, M.A., Kocurek, G., Goggin, D.J., Lake, L.W., 1989. Effects of stratigraphic heterogeneity on permeability in eolian sandstone sequence, Page Sandstone, northern Arizona. *American Association of Petroleum Geologist Bulletin* 73, pp. 658-668.

Cheel, R.J., 1990. Horizontal lamination and the sequence of bed phases and stratification under upper-flow-regime conditions. *Sedimentology* 37, pp. 517-529.

Clark, A., 1989. Lakes of Rub Al- Khali . *Aramco World* 40, pp. 28-33.

Clarke, M.L., Rendell, H.M., 1998. Climate change impacts on sand supply and the formation of desert sand dunes in the southwest USA. *Journal of Arid Environments* 39, pp. 517-531.

Clemmensen, L.B., 1979. Triassic lacustine red-beds and paloeoclimate: the "Buntsandstein" of I-Ielgokand and the Malmros Klint Member of east Greenland. *Geologische Rundschau* 68, pp. 748-774.

Clemmensen, L.C., Abrahamsen, K.,1983. Aeolian stratification and facies association in desert sediments, Arran basin (Permian), Scotland. *Sedimentology* 30, pp. 311-339.

Clemmensen, L.B., Blakey, R.C., 1989. Erg deposits in the Lower Jurassic Wingate Sandstone, Northeastern Arizona: Oblique dune sedimentation. *Sedimentology* 36, pp. 449-470.

Clemmensen, L.B., Øxnevad, I.E.I., De Boer, P.L., 1994. Climatic controls on ancient desert sedimentation: some late Palaeozoic examples from NW Europe and the western interior of the USA. In: De Boer, P.L., Smith, D.G. (Eds.), Orbital Forcing and Cyclic Sequences. *International Association of Sedimentologists, special publication* 19, pp. 439-457.

Clemmensen, L.C., Fornos, J.J., Rodriguez-Perea, A., 1997. Morphology and architecture of a late Pleistocene cliff-front dune, Mallorca, Western Mediterranean. *Terra Nova* 9, 251-254.

Collinson, J. D., 1986. *Deserts.* In: Reading, H.G., (Ed.), *Sedimentary Environments and Facies*, 2nd edition, Oxford: Blackwell, pp. 95-112.

Collinson, J.D., 1994. *Sedimentary deformational structures.* In: Maltman, A., (Ed.) *The geological deformation of sediments*. London, Chapman and Hall, pp. 95-125.

Collinson, J.D., 1996. Alluvial sediments. In: H.G. Reading, (Ed.), *Sedimentary Environments*. Blackwell- Science, Oxford, pp. 37-82.

Collinson, J.D., Lewin, J., 1983. Modern and Ancient Fluvial Systems: an introduction. In: Collinson, J.D., Lewin, J., (Eds.), *Modern and ancient fluvial systems.* The International Association of Sedimentologist, Special publication 6, pp. 1-2.

Collinson J., Mountney N., Thompson D., 2006. *Sedimentary structures*. Terra Publishing, England, 292 p.

Colombera, L., Mountney, N.P. and McCaffrey, W.D., 2012. A relational database for the digitization of fluvial architecture: concepts and example applications. *Petroleum Geosciences* 18, pp. 129-140.

Colombera, L., Mountney, N.P. and McCaffrey, W.D., 2013. A quantitative approach to fluvial facies models: methods and example results. *Sedimentology* 60, pp. 1526-1558.

Colter, V.S., Barr, K.W., 1975. Recent developments in the geology of the Irish Sea and Cheshire Basins. In: Woodland, A.W. (Ed.), *Petroleum and the continental shelf of north west Europe.* Applied Science Publishers, London, pp. 61-75.

Condon, S.M., 1997. Geology of the Pennsylvanian and Permian Cutler Group and Permain Kaibab Limestone in the Paradox Basin southeastern Utah and southwestern Colorado. *U.S. Geological Survey Bulletin* 2000. pp. 1-44.

Cooke, R.U., Warren, A., Goudie, A.S., 1993. *Desert geomorphology*. UCL Press, London. 526 p.

Costamagna, L.G., 2012. Alluvial, aeolian and tidal deposits in the Lower to Middle Triassic “Buntsandstein” of NW Sardinia (Italy): a new interpretation of Neo-Tethys transgression. *German Journal of Geology* ,pp. 165-183.

Cowan, G., 1993. Identification and significance of aeolian deposits within the dominantly fluvial Sherwood Sandstone Group of the East Irish Sea Basin UK. *Geological Society of London*, *Special Publication* 73, pp. 231-245.

Crabaugh, M., Kocurek, G., 1993. Entrada Sandstone: an example of a wet aeolian system. In: Pye, K. (Ed.), *The Dynamics and Environmental Context of Aeolian Sedimentary Systems*. Geological Society of London, Special Publication 72, pp. 103-126.

DeCelles, P.G., Langford, R.P., Schwartz, R.K., 1983. Two new methods of paleocurrent determination from trough cross-stratification. *Journal of Sedimentary Petrology* 53, pp. 629-642.

DeCelles, P.G., Kapp, P., Ding, L., Gehrels, G.E., 2007. Late Cretaceous to middle Tertiary basin evolution in the central Tibetan Plateau: Changing environments in response to tectonic partitioning, aridification, and regional elevation gain. *Geological Society of America Bulletin* 119, pp. 654-680.

De Ros, L.F., Scherer, C.M., 2012. Stratigraphic controls on the distribution of the diagenetic processes, quality and heterogeneity. In: Morad, S., Ketzer, M., F. de Ros, L. (Eds.), *Linking Diagenesis to Sequence*

Stratigraphy. International Association of Sedimentologists, Special Publication 45, pp. 105-132.

Doe, T.W., Dott, R.H., 1980. Genetic significance of deformed cross bedding – with examples from the Navajo and Weber Sandstones of Utah. *Journal of Sedimentary Petrology* 50, pp. 793-811.

Doughty, C.M., 1888. *Travels in Arabia*, Cambridge University press, London, 690 p.

Dott, R.H., Byers, C.W., Fielder, G.W., Stenzel, S.R., Winfree, K.E., 1986. Aeolian to marine transition in Cambro-Ordovician cratonic sheet sandstones of the northern Mississippi valley, U.S.A. *Sedimentology* 33, pp. 345-367.

Draut, A.E., 2012. Effects of river regulation on aeolian landscapes, Colorado River, southwestern USA. *Journal of Geophysical Research* 117, pp. 1-22.

East, A.E., Clift, P.D., Carter, A., Alizai, A., Vanlaningham, S., 2015. Fluvial-eolian interactions in sediment routing and sedimentary signal buffering: an example from the indus basin and thar desert. *Journal of Sedimentary Research* 85, pp. 715-728.

Edgell, H.S., 1989. Evolution of Rub' Al- Khali desert. King Abdul Aziz University, *Journal of Earth Sciences*, *Special Issue*, 1st Saudi Arabia Symposium on Earth Sciences, Jeddah Saudi Arabia, pp. 109-126.

Edgell, H.S., 2006. *Arabian deserts: nature, origin and evolution*. Springer. 664 p.

El-Sayed, M.I., 1999. Sedimentological characteristics and morphology of the aeolian sand dunes in the eastern part of the UAE: a case study from Ar Rub' Al- Khali. *Sedimentary Geology* 123, pp. 219-238.

El-Sayed, M.I., 2000. The Nature and possible origin of mega-dunes in Liwa, Ar Rub' Al- Khali. *Sedimentary Geology* 134, pp. 305-330.

Eschner, T.B., Kocurek, G., 1986. Marine destruction of eolian sand seas – origin of mass flows. *Journal of Sedimentary Petrology* 56, pp. 401-411.

Eschner, T.B., Kocurek, G., 1988. Origin of relief along contacts between eolian sandstone and overlying marine strata. *American Association of Petroleum Geologists, Bulletin* 72, pp. 932-943.

Evans, D.J., Rees, J.G., Holloway, S., 1993. The Permian to Jurassic stratigraphy and structural evolution of the central Cheshire Basin. *Geological Society of London* 150, pp. 857-870.

Ewing, R.C., Kocurek, G. 2010a. Aeolian dune-field pattern boundary conditions, *Geomorphology* 114, pp. 175-187.

Ewing, R.C, Kocurek, G., 2010b. Aeolian dune interactions and dune field and dune field pattern formation: White Sand Dunefield, New Mexico, *Sedimentology* 57, pp. 1119-1219.

Ewing, R.C., Kocurek, G., Lake, L.W., 2006. Pattern analysis of dune-field parameters. *Earth Surface Process and Landforms* 31, pp. 1176-1191.

Falorni, G., Teles, V., Vivoni, E.R., Bras,R.L., Amaratunga, K.S., 2005. Analysis and characterization of the vertical accuracy of digital elevation models from the Shuttle Radar Topography Mission. *Journal of Geophysical Research* 110, pp. 1-20.

Fielding, C.R., 2006. Upper flow regime sheets, lenses and scour fills: Extending the range of architectural elements for fluvial sediment bodies. *Sedimentary Geology* 190, pp. 227-240.

Fischer, C., Gaupp, R. Dimke, M., Sill, O., 2007. A 3D high resolution model of bounding surfaces in aeolian-fluvial deposits: An outcrop analogue study from the Permian Rotliegend, northern Germany. *Journal of Petroleum Geology* 30, pp. 257–274.

Fisher, J.A., Krapf, C.B.E., Lang, S.C., Nichols, G.J. and Payenberg, T.H.D., 2008. Sedimentology and architecture of the Douglas Creek terminal splay, Lake Eyre, central Australia. *Sedimentology* 55, pp. 1915-1930.

Fisher, G.B., Amos, C.B., Bookhagen, B., Burbank, D.W., Godard, V., 2012. Channel widths, landslides, faults, and beyond: The new world order of high-spatial resolution Google Earth imagery in the study of earth surface processes. In: Whitmeyer, S.J., Bailey, J.E., De Paor, D.G., Ornduff, T., (Eds.), *Google Earth and virtual visualizations in geoscience education and research*. Geological Society of America, Special Paper 492, pp. 1-22.

Flores, R.M., Pillmore, C.L., 1987. Tectonic Control on Alluvial Paleoarchitecture and of the Cretaceous and Tertiary Ration Basin, Colorado and New Mexico. In: *Recent developments in fluvial* *sedimentology.* International Association of Sedimentologists, Special Publication 39, pp. 311-320.

Folk, R. L., 1951. Stages of textural maturity in sedimentary rocks. *Journal of Sedimentary Petrology* 21, pp. 127-130.

Frostick, L.E., Reid, I., 1987. *Desert sediment: ancient and modern.* Geological Society of London, Special Publication 35, Blackwell Scientific, Oxford. 401 p.

Friend, P.F., Slater, M.J., Williams, R.C., 1979. Vertical and lateral building of river sandstone bodies, Ebro Basin, Spain. *Geological Society of London*136, pp. 39-46.

Fryberger, S.G., 1990a. Role of water in eolian deposition. In: Fryberger, S.G. Krystinik, L.F., Schenk, C.J., (Eds.), *Modern and ancient eolian deposits: petroleum exploration and production.* Society of sedimentary Research, Denver, pp. 5.1-5.11

Fryberger, S.G., 1990b. Bounding surface in eolian sediment. In: Fryberger, S.G. Krystinik, L.F., Schenk, C.J. (Eds.), *Modern and ancient eolian deposits: petroleum exploration and production*. Society of sedimentary Research, Denver, pp. 7.1-7.15.

Fryberger, S.G., 1990c. Coastal eolian deposits of Oregon, USA, Guererro Negro, Mexico, and Jafurah Sand Sea, Saudi Arabia. In: Fryberger, S.G. Krystinik, L.F., Schenk, C.J. (Eds.), *modern and ancient eolian deposits: petroleum exploration and production.* Society of sedimentary Research, Denver, pp. 11.1-11.15

Fryberger, S.G., 1993. A review of aeolian bounding surface, with examples from the Permian Minnelusa Formation, USA. In: North, C.P., Prosser, J.D. (Eds.), *Characterization of fluvial and aeolian reservoirs*. Geological Society of London, Special Publication 73, pp. 167-197.

Fryberger, S.G., Schenk, C.J., 1988. Pin stripe lamination-a distinctive feature of modern and ancient eolian sediments. *Sedimentary Geology* 55, pp. 1-15.

Fryberger, S.G., Ahlbrandt, T.S., 1979, Mechanisms for the formation of aeolian sand seas. *Geomorphology* 23, pp. 440-460.

Fryberger, S.G., Dean, G., 1979. Dune forms and wind regime. In: McKee, E.D. (Ed.), *A study of global sand seas*. US Geological Survey, Washington, pp. 137-170.

Fryberger, S.G, Al-sari, A.M., Clisham, T.J., 1983. Eolian dune, interdune, sand sheet, and Eolian dune, interdune, sand sheet, and siliciclastic sabkha sediments of an offshore prograding Sand Sea, Dhahran Area, Saudi Arabia. *American Association of Petroleum Geologist*s, *Bulletin* 67, pp. 280-312.

Fryberger, S.G., Schenk, C.J., Krystinik, L.F., 1988. Stokes surfaces and the effect of near-surface groundwater table on aeolian deposition. *Sedimentology* 35, pp. 21-41.

Fryberger, S.G., Krystinik, L.F., Schenk, C.J, 1990. Tidally flooded back-barrier dunefield, Guerrero Negro area, Baja California, Mexico. *Sedimentology* 37, pp. 23-43.

Galloway, W.E., Hobday, D.K., 1996. *Terrigenous clastic depositional system, applications of fossil fuel and groundwater resources,* 2nd edition. Springer-Verlag, Berlin Heidelberg. 489 p.

Gareth, T.G., Berry, K.J., 1993. A new lithostratigraphy and depositional model for the upper Rotliegend of the UK Sector of the Southern North Sea. In: North, C.P., Prosser, D.J. (Eds.), *Characterization of fluvial and aeolian reservoirs*. Geological Society of London, Special Publication 73, pp. 291-319.

Geach, M.R., Viveen, W., Mather, A.E., Telfer, M.W., Fletcher, W.J., Stokes, M., Peyron, O., 2015. An integrated field and numerical modelling study of controls on Late Quaternary fluvial landscape development (Tabernas, southeast Spain). *Earth Surface Processes and Landforms* 40, pp. 1907-1926.

Geel, C.R. Nonselaar, M.E., 2007. Reservoir modelling of heterolithic tidal deposits: sensitivity analysis of an object-based stochastic model. *Netherlands Journal of Geosciences* 86 (4), pp. 403-411.

Geleynse, N., Storms, J.E., Walstra, D.R., Jagers, H.R., Wang, Z.B., Stive, M.J., Controls on river delta formation; insights from numerical modelling. *Earth and Planetary Science Letters* 302, pp. 217-226.

Gibling, M.R., 2006. Width and thickness of fluvial channel bodies and valley fills in the geological record: A literature compilation and classification. *Journal of Sedimentary Research* 76, pp. 731-770.

Gillette, D. A., Chen, W., 2001. Particle production and aeolian transport from a ‘‘supply-limited’’ source area in the Chihuahuan Desert, United States, *Journal of Geophysical Research* 106, pp. 5267-5278.

Glennie, K.W., 1970. *Desert sedimentary environments*. Development in Sedimentology 14. Elsevier, Amsterdam, 222 p.

Glennie, K.W., 1972, Permian Rotliegendes of north-west Europe interpreted in light of modern desert sedimentation studies: *American Association of Petroleum Geologist*s, *Bulletin* 56, pp. 1048-1071.

Glennie, K.W., 1983. Early Permian (Rotliegend) palaeowind of the North Sea. *Sedimentary Geology* 34, pp. 120-152.

Glennie, K.W., 1987. Desert sedimentary environments, present and past , a summary. *Sedimentary Geology* 50, pp. 135-165.

Glennie, K.W., 1990. Lower Permian – Rotliegend. In: Glennie, K.W. (Ed.), *Introduction to the petroleum geology of the North Sea*. Blackwell Science, Oxford, pp. 120-152.

Glennie, K.W., 1995. Permian and Triassic rifting in northwest Europe. *Geological Society of London, Special Publication* 91, pp. 1-5.

Glennie, K.W., 1998. The desert of southeast Arabia: a product of Quaternary climatic change. In: Alsharhan, K.W., Whittle, G.L., Kendall, G.G. (Eds.), *Quaternary deserts and climatic change*. Balkema, Rotterdam, pp. 315-325.

Glennie, K.W., 2005. *The desert of southeast Arabia: Desert Environments and Sediments*. Gulf Petrolink, Bahrain. 215 p.

Glennie, K.W, 2009. Lower Permian-Rotliegend. In: Glennie, K.W, (Ed.) *Petroleum Geology of the North Sea: Basic Concepts and Recent Advances*. John Wiley and Sons, pp- 137-173.

Glennie, K.W., Provan, DM., 1990. Lower permian rotliegend reservoir of the southern north sea gas province. In: Brooks, J. (Ed.), *Classic Petroleum Provinces*, Geological Society Special Publication 50, pp 399-416.

Glennie, K.W., Singhvi, A.K., 2002. Event stratigraphy, paleoenvironment and chronology of SE Arabian deserts. *Quaternary Science Reviews* 21, pp. 853-869.

Glennie, K., A. Hurst, 2007., Fluidization and associated soft-sediment deformation in eolian sandstones: Hopeman Sandstone (Permian), Scotland, and Rotliegend, North Sea. In: Hurst, A., Cartwright, J. (Eds.), *Sand injectites: Implications for hydrocarbon exploration and production.* American Association of Petroleum Geologist, Memoir 87, pp. 245-252.

Glennie, K.W., Mudd, G.C., Nagtegaal, P.J., 1978. Depositional environment and diagenesis of Permian Rotliegendes Sandstones in Lemon Bank and Sole Pit areas of the UK southern North Sea. *Geological Society of London*, *Quarterly Journal* 135, pp. 25-34.

Good, T.R., Bryant, I.D., 1985. Fluvio-aeolian sedimentation – an example from Banks Island, N. W. T., Canada. Geografiska Annaler, Series A, *Physical Geography* 67A, pp. 33-46.

Goodall, T.M., North, C.P., Glennie, K.W., 2000. Surface and subsurface sedimentary structures produced by salt crusts. *Sedimentology* 47, pp. 99-118.

Goudie, A.S., 1972. Climate, weathering, crust formation, dunes and fluvial features of central Namib Desert near Gobabeb, south west Africa. *Madoqua* 2, pp. 15-31.

Goudie, A.S., 2013. *Arid and semi-arid Geomorphology*. Cambridge University Press, New York, 454 p.

Goudie, A.S, Colls, A., Stokes, S., Parker, a., White, K., Al-Farraj, A., 2000. Latest Pleistocene and Holocene dune construction at the north-eastern edge of the Rub Al Khali, United Arab Emirates, *Sedimentology* 47, pp. 1011-1021.

Granja, H.M., De Groot, T.A.M. and Costa, A.L., 2008. Evidence for Pleistocene wet aeolian dune and interdune accumulation, S. Pedro da Maceda, north-west Portugal. *Sedimentology*, 55, pp. 1203-1226.

Greensmith, J.T., 2012. *Petrology of the sedimentary rocks*, 7th edition. Springer Science & Business Media. London, 244 p.

Griffiths, K J, Shand, P, Ingram, I., 2002. *Baseline reports series: 2. The Permo-Triassic Sandstones of west Cheshire and the Wirral*. British Geological Survey Commissioned Report CR/02/109N.

Griffiths, K J, Shand, P, Ingram, I., 2003. *Baseline report series: 8. The Permo-Triassic sandstones of Manchester and east Cheshire.* British Geological Survey Commissioned Report CR/03/265N.

Hadlari, T., Rainbird, R.H., Donaldson, J.A., 2006. Alluvial, eolian and lacustrine sedimentology of a Paleoproterozoic half graben, Baker Lake Basin, Nunavut. *Sedimentary Geol*ogy 190, pp. 47-70.

Haig, D.B., Pickering, S.C., Probert, R., 1997. The Lennox oil and gas field. In: Meadows, N.S., Trueblood, S.P., Hardman, M., Cowan, G. (Eds.), *Petroleum geology of the Irish Sea and adjacent areas*. Geological Society of London, Special Publication 124, pp. 417-436.

Harland, W.B., Armstrong, R.L., Cox, A.V., Craig, L.E., Smith, A.G., Smith, D.G., 1990. *A geologic time scale 1989*. Cambridge University Press, Cambridge. 279 p.

Hartley, A., Weissmann, G.S., Nichols, G.J., Warwick, G.L., 2010. Large Distributive Fluvial Systems: Characteristics, distribution, and controls on development. *Journal of Sedimentary Geology* 80, pp. 167- 183.

Hampton, B.R., Horton, B.K. 2007. Sheetflow fluvial processes in a rapidly subsiding basin, Altiplano plateau, Bolivia. *Sedimentology* 54, pp. 1121-1147.

Harvey, A., 2011. Dryland alluvial fans. In: Thomas, D. S. (Ed.), *Arid zone geomorphology, process, form and change in drylands*, 3rd edition. John Wiley and Sons, pp. 333-401.

Havholm, K.G., Kocurek, G., 1994. Factors controlling aeolian sequence stratigraphy: clues from super bounding surface features in the Middle Jurassic Page Sandstone. *Sedimentology* 41, pp. 913-934.

Herries, R.D., 1993. Contrasting style of fluvial-aeolian interaction at a downwind erg margIn: Jurasic Kayenta-Navajo transition, northeastern Arizona, USA.. In: North, C.P., Prosser, D.J. (Eds.), *Characterization of fluvial and aeolian reservoirs.* Geological Society of London, Special Publication 73, pp. 199-218.

Herries, R.D., Cowan, G., 1997. Challenging the 'sheetflood' myth: the role of water-table-controlled sabkha deposits in redefining the depositional model for the Ormskirk Sandstone Formation (Lower Triassic), East Irish

Sea Basin. In: Meadows, N.S., Trueblood, S.P., Hardman, M., Cowan, G. (Eds.), *Petroleum geology of the Irish Sea and adjacent areas.* Geological Society of London, Special Publication 124, pp. 253-276.

Heward, A. 1991. Auk - the anatomy of an eolian oil reservoir. In: Miall, A.D. and Tyler, N. (Eds.), *The three-dimensional facies architecture of terrigenous clastic sediments and its implications for hydrocarbon discovery and recovery*. Society of Sedimentary Geology, Concepts in Sedimentology and Palaeontology 3, pp. 44-56

Hofmann, A., Tourani, A., Gaupp, R., 2000. Cyclicity of Triassic to Lower Jurassic continental red beds of the Argana Valley, Morocco: implications for palaeoclimate and basin evolution. *Palaeogeography, Palaeoclimatology, Palaeoecology* 161, pp. 229-266.

Hollands, C. B., Nansona, G.C., Jonesa, B. G., Bristow, C.S., Pricea, D.M., Pietsch, T.J., 2006. Aeolian–fluvial interaction: evidence for Late Quaternary channel change and wind-rift linear dune formation in the northwestern Simpson Desert, Australia. *Quaternary Science Reviews* 25, pp. 142-162.

Holm, D.A., 1960. Desert geomorphology in the Arabian Peninsula. *Science* 132, pp. 1369-1379.

Holm, D.A., 1968. Sand dunes. In: R.W. Fairbridge, (Ed.) *The encyclopaedia of geomorphology.* Reinhold Book Corporation, New York, pp. 973-979.

Horne, R., 1975. The association of alluvial fan, aeolian and fluviatile facies in the Caherbla Group (Devonian), Dingle Peninsula, Ireland. *Journal of Sedimentary Petrology* 45, (2), pp. 535-540

Horowitz, D.H., 1982. Geometry and origin of large-scale deformation structures in some ancient wind-blown sand deposits: *Sedimentology*. 29, pp. 155-180.

Hotta, S., Kubota, S., Katori, S., Horikawa, K., 1984. Sand transport by wind on sand surface. *Coastal Engineering* 2, pp. 1265-1281.

Hotzl, H., Kramer, F., Maurin, V., 1978. Quaternary sediments. In: Al-Sayari S.S, Zotel, J.G. (Eds.), *Quaternary period in Saudi Arabia 1*. Springer-Verlag, Vienna, pp. 264-301.

Howard, A S, Hough, E, Crofts, R G, Reeves, H J, Evans, D J. 2007. Geology of the Liverpool district. *Sheet Explanation of the British Geological Survey, Sheet 96* (England and Wales).

Howell, J.A., Mountney, N.P., 2001. Aeolian grainflow architecture: Hard data for reservoir models and implications for red bed sequence stratigraphy. *Petroleum Geoscience* 7, pp. 51-56.

Höyng, D., D'Affonseca, F.M., Bayer, P., Gomes de Oliveira, E., Perinotto, J.A., Reis, F., Weiß, H., Grathwohl, P., 2014. High-resolution aquifer analogy of fluvial-aeolian sediments of the Guarani aquifer system. *Journal of Environmental Earth Science* 71, Springer Berlin Heidelberg, pp.3081-3094.

Hsü, K.J., 2004. *Physics of sedimentology*. 2nd edition. Springer, Germany. 240 p.

Hugenholtz, C.H., Barchyn, T. E., 2010. Spatial analysis of sand dunes with a new global topographic dataset: new approaches and opportunities. *Earth Surface Processes and Landforms* 35, pp. 986-992.

Hummel, G., Kocurek, G., 1984, Interdune areas of the back-Island dune field, North Padre Island, Texas. *Sedimentary Geology* 39, pp. 1-26.

Hunter, R.E., 1977. Basic types of stratification in small eolian dunes. *Sedimentology* 24, pp. 361-387.

Hunter, R.E., 1981. Stratification styles in eolian sandstones: Some Pennsylvanian to Jurassic examples from the western interior USA. In: Ethridge, F.G., Flore, R.M. (Eds.), *Recent and ancient non-marine depositional environments: models for exploration.* International Association of Sedimentologists, Special Publication 31, pp. 315-329.

Hunter, R.E., 1985. A kinematic model for the structure of lee-side deposits. *Sedimentology* 32, pp. 409-422.

Hunter and Rubin, 1983. Interpreting cyclic cross-bedding, with an example from the Navajo Sandstone. In: Brookfield, M.E., Ahlbrandt, T.S. (Eds.), *Eolian sediment and processes*. Amsterdam, Elsevier, Development in Sedimentology 38, pp. 429-454.

Hunter, R.E, Richmond, B.M., Alpha, T.R., 1983. Storm-controlled oblique dunes of the Oregon Coast. *Geological Society of America* 94, pp. 1450-1465.

Jones, A.D., Auld, H.A., Carpenter, T.J., Fetkovich, E., Palmer, I.A., Rigatos, E.N. and Thompson, M.W., 2005. Jade Field: an innovative approach to high-pressure, high-temperature field development. In: Dore, A.G., Vining, B.A. (Eds.) *Petroleum geology: north-west Europe and global perspectives*. Petroleum Geology Conference, Series 6, pp. 269-283.

Jones, S. J., 2002. Transverse rivers draining the Spanish Pyrenees: large scale patterns of sediment erosion and deposition. In: Jones, S. J., Frostick, L.E., (Eds.) *Sediment Flux to Basins: Causes, Controls and Consequences.* Geological Society, London, Special Publications 191, pp. 171-185.

Jones, S. J. 2004. Tectonic controls on drainage evolution and development of terminal alluvial fans, southern Pyrenees, Spain. *Terra Nova 16*, pp. 121-127.

Jones, S.J., Frostick, L.E., 2008. Inferring bedload transport from stratigraphic successions: examples from Cenozoic and Pleistocene rivers, south central Pyrenees, Spain. In: Gallagher, K., Jones, S. J., Wainwright, J. (Eds.), *Landscape evolution: denudation, climate and tectonics over different time and space scales*. Geological Society of London, Special Publication, 296, pp. 129-145.

Jones, S.J., Frostick, L.E., Astin, T.R., 2001. Braided stream and flood plain architecture: The Rio Vero Formation, Spanish Pyrenees. *Sedimentary Geology* 139, pp. 229-260.

Jones, L.S., Blakey, R.C., 1997. Aeolian-fluvial interaction in the Page Sandstone (Middle Jurassic) in south-central, USA- a case study of erg-margin processes. *Sedimentary Geology* 109, pp. 181-198.

Jopling, A.V, Walker, R.G.,1968. Morphology and origin of ripple-drift cross lamination, with example from the Pleistocene of Massachusetts. *Journal of Sedimentary Petrology* 38, pp. 971-984.

Jordan O. D., Mountney N. P., 2010. Styles of interaction between aeolian, fluvial and shallow marine environments in the Pennsylvanian to Permian lower Cutler beds, south-east Utah, USA. *Sedimentology* 57, pp. 1357-1385.

Jullien, R., Meakin, P., Pavlovitch, A., 2002. Three-dimensional model for particle-size segregation by shaking. *Physical Review Letters* 69, pp. 640-643.

Jungerius, P.D., Meulen, F., 1988. Erosion processes in a dune landscape along the Dutch coast. *Catena* 15, pp. 217-228.

Kerr, D.R., Dott, R.H, 1988. Eolian dune types preserved in the Tensleep Sandstone (Pennsylvanian-Permian), north-central Wyoming. *Sedimentary Geology* 56, pp. 383-402.

King, M.J., Thompson, D.B., 2000. Triassic vertebrate footprints from the Sherwood Sandstone Group, Hilbre, Wirral, Northwest England. *Proceedings of the Geologists Association* 111, pp. 111-132.

Kocurek, G.A., 1981a. Significance of interdune deposits and bounding surfaces in eolian dune sands. *Sedimentology* 28, pp. 753-780.

Kocurek, G.A., 1981b. Erg reconstruction – the Entrada Sandstone (Jurassic) of northern Utah and Colorado. *Palaeogeography, Palaeoclimatology, Palaeoecology* 36, pp. 125-153.

Kocurek, G.A., 1991. Interpretation of ancient eolian sand dunes. *Annual Review of Earth and Planetary Science* 19, pp. 43-75.

Kocurek, G., 1996. Desert aeolian systems. In: Reading, H.G. (Ed.), *Sedimentary Environments: Processes, facies and stratigraphy*. 3rd edition, Blackwell, Science, Oxford. pp. 125-153.

Kocurek, G., 1998. Aeolian system response to external forcing- a sequence stratigraphic approach. In: Alsharhan, A.S., Glennie, K.W., Whittle, G.L., Kendall, G.G. (Eds.), *Quaternary desert and climatic change*, Balkema, Rotterdam, pp. 327-338.

Kocurek, G., 1999. The aeolian rock record. In: Goudie, A.S., Livingstone, I., Stokes, S. (Eds.), *Aeolian environments sediments and landforms.* Chichester, John Wiley and Sons Ltd., pp. 239-259.

Kocurek, G., Dott, R.H., 1981. Distinctions and uses of stratification types in the interpretation of eolian sand. *Journal of Sedimentary Petrology* 51, pp. 579-595.

Kocurek, G., Fielder, G., 1982. Adhesion structures. *Journal of Sedimentary Petrology* 52, pp. 1229-1241.

Kocurek, G., B.L. Kirkland, 1998. Getting to the source: Aeolian influx to the Permian Delaware basin region. *Sedimentary Geology* 117, pp. 143-149.

Kocurek, G.A., Nielson, J., 1986. Conditions favourable for formation of warm-climate aeolian sand sheet. *Sedimentology* 33, pp. 795-816.

Kocurek, G., Crabaugh, M., 1993. Significance of thin sets of eolian cross-strata – Discussion. *Journal of Sedimentary Petrology* 63, pp. 1165-1169.

Kocurek, G., Havholm, K.G., 1993. Eolian sequence stratigraphy - a conceptual framework. In: Weimer, P., Posamentier, H.W. (Eds.). *Siliciclastic sequence stratigraphy: recent developments and applications.* American Association of Petroleum Geologists, Memoir 58, pp. 393-409.

Kocurek, G., Lancaster, N., 1999. Aeolian system sediment state: theory and Mojave Desert Kelso dune field example. *Sedimentology* 46, pp. 505-515.

Kocurek, G., Ewing, R.C., 2005. Aeolian dune field self-organization implication for the formation of simple versus complex dune-field patterns. *Geomorphology* 72, pp. 94-105.

Kocurek, G., Ewing, R.C., Mohrig, D., 2010. How do bedform patterns arise? New views on the role of bedform interactions within a set of boundary conditions. *Earth Surface and Landforms* 35, pp. 51-63.

Kocurek, G., Townsley, M., Yeh, E., Havholm, K., Sweet, M.L., 1992. Dune and dune-field development on Padre Island, Texas, with implications for interdune deposition and water-table-controlled accumulation. *Journal of Sedimentary Petrology* 62, pp. 622-635.

Krapf, C.B., Stollhofen, H., Stanistreet, I.G., 2003. Contrasting styles of ephemeral river systems and their interaction with dunes of Skeleton Coast erg (Namibia). *Quaternary International* 104, pp. 41-52.

Krapf, C.B.E., Stanistreet, I.G., Stollhofen, H., 2005. Morphology and fluvio-aeolian interaction of the tropical latitude, ephemeral braided river dominated Koigab Fan, north-west Namibia. In: Blum, M.D., Marriott, S.B., Leclair, S.F. (Eds.), *Fluvial Sedimentology VII*. International Association of Sedimentologist, Special Publication 35, pp. 99-120.

Kugler, R.L., Mink, R.M., 1999. Depositional and diagenetic history and petroleum geology of the Jurassic Norphlet Formation of the Alabama coastal waters area and adjacent federal waters area*. Marine Georesources & Geotechnology* 17, pp. 215-232.

Kumar, A., Mahmoud, M.A., 2011. An overview of origin, morphology and distribution of desert forms, sabkhas and playas of the Rub' Al- Khali Desert of the southern Arabian Peninsula, *Earth Science India* 4, pp. 105-135.

Laity, J., 2009. *Desert and desert environments*. John Wiley and Sons, 360 p.

Lancaster, N., 1983. Controls of dune morphology in the Namib Sand Sea. In: Brookfield, M. E., Ahlbrandt, T. S., (Eds.), *Eolian sediments and processes.* Elsevier, Amsterdam, pp. 261-89.

Lancaster, N., 1988. Controls of eolian dune size and spacing, *Geology* 16, pp. 972-975.

Lancaster, N., 1989. *The Namib sand sea-dune forms, processes and sediments.* Rotterdam: Balkema, 200 p.

Lancaster, N., 1989. Star dunes. *Progress in Physical Geography* 13, pp. 67- 91.

Lancaster, N., 1994. Controls on aeolian activity: some new perspective from the Kelso Dunes, Mojave Desert, California, *Journal of Arid Environment* 27, pp. 113-125.

Lancaster, N., 1993. Origins and sedimentary features of supersurfaces in the northwestern Gran Desierto Sand Sea. In: Pye, K., Lancaster, N. (Eds.), *Aeolian sediments: ancient and modern*. International Association of Sedimentologist, Special publication 16, pp. 71-83.

Lancaster, N., 1995. Response of eolian geomorphic systems to minor climate change: examples from the southern Californian deserts. *Geomorphology* 19, pp. 333-347.

Lancaster, N., 1998. Dune Morphology, Chronology and Quaternary climatic changes. In: Alsharhan, K. W., Whittle, G. I., Kendall, G.G. (Eds.), *Quaternary deserts and climate change*. Balkema, Rotterdam, pp. 339-349.

Lancaster, N., Teller, J.T., 1988. Interdune deposits of the Namib Sand Sea. *Sedimentary Geology* 55, pp. 91-107.

Lancaster, N., Tchakerian, V.P., 2003. Late Quaternary eolian dynamics, Mojave Desert, California. In: Enzel, Y., Wells, S. G., Lancaster, N. (Eds.), *Paleoenvironments and paleohydrology of the Mojave and southern Great Basin Deserts.* Geological Society of America, pp. 231-249.

Langford, R.P., 1989. Fluvial-aeolian interactions: Part I, modern systems. *Sedimentology* 36, pp. 1023-1035.

Langford, R.P., Chan, M.A., 1988. Flood surfaces and deflation surfaces within the Cutler Formation and Cedar Mesa Sandstone (Permian), southeastern Utah. *Geological Society of America*, Bulletin 100, pp. 1541-1549.

Langford, R.P., Chan, M.A., 1989, Fluvial-aeolian interactions; Part II, Ancient systems. *Sedimentology* 36, pp. 1037-1051.

Langford, R.P., Pearson, K.M., Duncan, K.A., Tatum, D.M., Adams, L., Depret, P.A., 2008. Eolian topography as a control on deposition incorporating lessons from modern dune seas: Permian Cedar Mesa Sandstone, SE Utah, U.S.A. *Journal of Sedimentary Research* 78, pp. 410-422.

Laity, J.E., 1995. Wind abrasion and ventifact formation in California. In: Tchakerian, V.P. (Ed.), *Desert Aeolian Processes*. London, Chapman and Hall, pp. 295-321.

Leclair, S.F., Bridge, J.S., 2001. Quantitative interpretation of sedimentary structures formed by river dunes. *Journal of Sedimentary Research* 71, pp. 713-716.

Leleu, S., Hartley, A.J., 2010. Controls on the stratigraphic development of the Triassic Fundy Basin, Nova Scotia; implications for the tectonostratigraphic evolution of Triassic Atlantic rift basins. *Journal of the Geological Society of London* 167, pp. 437-454.

Levin, N., Tsoar, H., Herrmann, H.J., Maia, L.P., Claudino-Sales, V., 2009. Modelling the formation of residual dune ridges behind barchan dunes in north-east Brazil. *Sedimentology* 56, pp. 1623-1641.

Li, S., Sun, J., Li, B., 2012. Holocene environmental changes in central Inner Mongolia revealed by luminescence dating of sediments from the Sala Us River valley. *The Holocene* 22 (4), pp. 397-404.

Liu, B., Coulthard, T.J., 2015. Mapping the interactions between rivers and sand dunes: implications for fluvial and aeolian Geomorphology. *Geomorphology* 231, pp. 246-257.3

Liu, K.Y., Paterson, L., Wong, P., QI, D.S., 2002. A sedimentological approach to upscaling. *Transport in Porous Media* 46, pp. 285-310.

Livingstone, I. and Warren, A., 1996. *Aeolian geomorphology: an introduction.* Longman, Singapore, 204 p.

Loope, D.B., 1985. Episodic deposition and preservation of eolian sands – a Late Paleozoic example from southeastern Utah. *Geology* 13, pp. 73-76.

Loope, D.B., 1986. Recognizing and utilizing vertebrate tracks in cross section: Cenozoic hoofprints from Nebraska. *Palaios* 1, pp. 141-151.

Loope, D.B.,1988. Rhizoliths in ancient eolianites. *Sedimentary Geology* 56, pp. 301-314.

Loope, D.B., 2006. Dry-Season Tracks in dinosaur-triggered grainflows. *Palaios* 21, pp. 132-142.

Loope, D.B., Simpson, E.L., 1992. Significance of thin sets of eolian cross-strata.*Journal of Sedimentary Petrology* 62, pp. 849-859.

Loope, D.B. and Simpson, E.L., 1993. Significance of thin sets of eolian cross-strata – reply. *Journal of Sedimentary Petrology* 63, pp. 1170–1171.

Loope, D.B., Swinehart, J.B., Mason, J.P., 1995. Dune-dammed paleovalleys of the Nebraska Sand Hills: intrinsic versus climatic controls on the accumulation of lake and marsh sediments. *Geological Society of America, Bulletin* 107, pp. 396-406.

Loope, D.B., Rowe, C.M. and Joeckel, R.M., 2001. Annual monsoon rains recorded by Jurassic dunes. *Nature*, 412, pp. 64-66.

Loope, D.B., James, F.E., Sweeney.M.R., 2012. Downslope coarsening in aeolian grainflows of the Navajo Sandstone. *Sedimentary Geology* , pp. 156-162.

Luna, Marco C. M. de M., Parteli, Eric J. R., Herrmann, H.J., 2012. Model for a dune field with an exposed water table. *Geomorphology* 159-160, pp. 169-177.

Makse, H. A., 2000. Grain segregation mechanism in aeolian sand ripples. *European Physical Journal* E 1, pp.127-135.

Mader, D., 1985a. Aeolian sedimentation in the Middle Buntsandstein in the Eifel North-South Depression Zone. In: Mader, D. (Ed.), *Aspect of fluvial sedimentation in the Lower Triassic Buntsandstein of Europe*. Springer, Berlin. Lecture Note in Earth Science 4, pp. 90-126.

Mader, D., 1985b. Braidplain, floodplain and playa lake, alluvial-fan, aeolian and palaeosol facies composing a diversified lithogenetical sequence in the Permian and Triassic of South Devon (England). In: Mader, D. (Ed.), *Aspect of fluvial sedimentation in the Lower Triassic Buntsandstein of Europe.* Springer, Berlin. Lecture Note in Earth Science 4, pp.90-126.

Mader, N.K., Redfern, J., 2011. A sedimentological model for the continental Upper Triassic Tadrart Ouadou sandstone member: Recording an interplay of climate and tectonics (Argana valley; south-west Morocco). *Sedimentology* 58, pp. 1247-1282.

Martinsen, O.L., Ryseth, A., Helland-Hansen, W., Flesche, H., Torkildsen, G. and Idil, S.,1999. Stratigraphic base level and fluvial architecture: Ericson

Sandstone (Campanian), Rock Springs Uplift, SW Wyoming, USA. *Sedimentology* 46, pp. 235-259.

McClure, H.A., 1976. Radiocarbon chronology of the Late Quaternary Lakes in the Arabian Desert. *Nature* 263, pp. 755-756.

McClure, H.A., 1978. Ar Rub' Al- Khali. In: Al-sayari, S.S., Zölt, J.G. (Eds.), *Quaternary Period in Saudi Arabia*, Springer-Verlag, Vienna, pp. 252-263.

McKee, E.D., 1966. Structures of dunes at White Sands National Monument, New Mexico (and a comparison with structures of dunes from other selected areas). *Sedimentology* 7, pp. 1-69.

Mckee, E.D., 1979. Introduction to a study of global sand sea. In: McKee, E.D. (Ed.), *A study of global sand seas*. US Geological Survey Washington, pp.1-19.

Mckee, E.D., Douglass, J.R., Rittenhouse, S.,1971. Deformation of lee side laminae in eolian dunes: *Geological Society of America*, Bulletin 82, pp. 359-378.

McKee, E.D., Moiola, R.J., 1975. Geometry and growth of the White Sands New Mexico dune field. US Geological Survey, *Journal of Research* 3, pp. 59-66.

McKee, E.D., Breed, C.S., 1976. Sand seas of the world. In: Williams, R.S, Carter, W. D. (Eds.) ERTS-1, *A new window on our planet.* US Geological Survey, pp. 81-88.

McKee, E.D., Bigarella, J., 1979. Sedimentary structures in dunes. In: McKee, E. D. (Eds.), *A study of global sand seas.* US Geological Survey, Washington, pp. 83-134.

McKee, E. D., Crosby, E. J., Berryhill, H. L., 1967. Flood deposits, Bijou Creek, Colorado, June 1965. *Journal of Sedimentary Petrology* 37, pp. 829-851.

McKenna, C.N., Scott, M.M., 1998. A wind tunnel study of the influence of pore water on aeolian sediment transport. *Journal of Arid Environment* 39, pp. 403-419.

McKie, T., 2011a. Architecture and behaviour of dryland fluvial reservoirs, Triassic Skagerrak Formation, Central North Sea. In: Davidson, S.K., Leleu, S., North, C.P., (Eds.), *From rivers to rock record*, Society of Sedimentary Geology, Special Publication 97, pp. 189-214.

Mckie, T., Jolley,S.J., Kristensen, M.B., 2010. Stratigraphic and structural compartmentalization of dryland fluvial reservoirs: Triassic Heron Cluster, central North Sea. *Geological Society of London*, Special Publication 347, pp. 165-198.

McKie, T., 2011b. A comparison of modern dryland depositional systems with the Rotliegend Group in the Netherlands. In: Grötsch, G., Gaupp, R. (Eds.), *The Permian Rotliegend of the Netherlands*. Society for Sedimentary Geology, Special Publication 98, pp. 89-103.

McKinley, J. M., Atkinson, P. M., Lloyd, C. D., Ruffell, A. H., Worden, R. H., 2011. How Porosity and Permeability Vary Spatially With Grain Size, Sorting, Cement Volume, and Mineral Dissolution In Fluvial Triassic Sandstones: The Value of Geostatistics and Local Regression. *Journal of Sedimentary Research* 12, pp. 844-858.

Meadows, N.S., 2006. The correlation and sequence architecture of the Ormskirk Sandstone Formation in the Triassic Sherwood Sandstone Group of the east Irish sea basin, NW England. *Geological Journal* 41, pp. 93-122.

Meadows, N.S., Beach, A., 1993. Structural and climatic controls on facies distribution in a mixed fluvial and aeolian reservoir: the Triassic Sherwood Sandstone in the Irish Sea in: North, C.P., Prosser, D.J. (Eds.), *characterization of fluvial and aeolian reservoirs.* Geological Society of London, Special Publication 73, pp. 247-264.

Medicia, G., Boulesteixb, K. Mountney, N.P., Westa, L.J., Odlinga, N.E., 2015. Palaeoenvironment of braided fluvial systems in different tectonic realms of the Triassic Sherwood Sandstone Group, UK, *Sedimentary Geology* 329, pp. 188-210.

Melvin, J., Wallick, B.P., Heine, C.J., 2010. Advances in Arabian stratigraphy: Allostratigraphic layering related to paleo-water table fluctuations in eolian sandstones of the Permian Unayzah A reservoir, South Haradh, Saudi Arabia: *GeoArabia* 15, pp. 55-86.

Membery, D.A., 1983. Low-level wind profiles during the Gulf Shamal. *Weather* 38, pp. 18-24.Miall, A.D., 1977. A review of the braided river depositional environment. *Earth Science Reviews* 13, pp.1-32.

Miall, A.D., 1978. Fluvial sedimentology. *Canadian Society Petroleum Geologists,* Memoir 5, pp.105-127.

Miall, A.D., 1996. *The geology of fluvial deposits*. Springer, Berlin. 582 p.

Miall, A.D., 2014. *Fluvial depositional systems*. Springer Science & Business Media, 316 p.

Miall, A.D., 2010a. Alluvial deposits. In: James, N.P., Dalrymple, R.W. (Eds.), *Facies models.* Geological Association of Canada, GEOtxt 6. St Johns, Newfoundland, pp. 105-137.

Miall, A.D., 2010b. *The Geology of fluvial deposits: sedimentary facies, basin analysis and petroleum geology*. 4th Edition. Springer, 582 pp.

Middleton, G.V., 1970. Experimental studies related to problems of flysch sedimentation. In: Lajoie, J. (Ed.), *Flysch sedimentology of North America.* Geological Society of Canada, Special publication 7, pp. 253-272.

Mores, D.G., 1994. Siliciclastic reservoir rocks. In: Magoon, L. B, Dow, W. G. (Eds.), *The petroleum system-From source to trap*: American Association of Petroleum Geologists, Memoir 60, pp. 121-139.

Mountney, N.P., 2005. Deserts. In: Selley, C.R, Cocks, L.R., Plimer, R.I. (Eds.), *Encyclopedia of geology* 4. Elsevier, pp. 539-549.

Mountney, N.P., 2006a. Eolian Facies Models. In: Posamentier, H., Walker R.G. (Eds.), *Facies models revisited*. Society for Sedimentary Geology, Special Publication 84, pp. 19-83.

Mountney, N. P., 2006b. Periodic accumulation and destruction of aeolian erg sequences: the Cedar Mesa Sandstone, White Canyon, southern Utah. *Sedimentology* 53, pp. 789-823.

Mountney, N. P., 2012. A stratigraphic model to account for complexity in aeolian dune and interdune successions, *Sedimentology* 59, pp. 964-989.

Mountney, N.P., Howell, J., 2000. Aeolian architecture, bedform climbing and preservation space in the Cretaceous Etjo Formation, NW Namibia. *Sedimentology* 47, pp. 825-849.

Mountney, N.P., Thompson, D.B., 2002. Stratigraphic evolution and preservation of aeolian dune and damp/wet interdune strata: an example from the Triassic Helsby Sandstone Formation, Cheshire Basin, UK. *Sedimentology* 49, pp. 805-834.

Mountney, N.P., Jagger, A., 2004. Stratigraphic evolution of an aeolian erg margin system: the Permian Cedar Mesa Sandstone, SE Utah, USA. *Sedimentology* 51, pp. 713-743.

Mountney, N.P., Russell, A.J., 2006. Coastal aeolian dune-field development and response to periodic fluvial inundation, Sólheimasandur, southern Iceland. *Sedimentary Geology* 192, pp. 167-181.

Mountney, N.P., Russell, A.J., 2009. Aeolian dune field development in a water table-controlled system: Skeiđarársandur, southern Iceland. *Sedimentology* 56, pp. 2107-2131.

Mountney, N.P., Howell, J.A., Flint, S.S., Jerram, D.A., 1998, Aeolian and alluvial deposition within the Mesozoic Etjo Sandstone Formation, NW Namibia: *Journal of African Earth Sciences* 27, pp. 175-192.

Mountney, N.P., Howell, J.A., Flint, S., Jerram, D.A., 1999. Climate, sediment supply and tectonics as controls on the deposition and preservation of the aeolian-fluvial Etjo Sandstone Formation, Namibia. *Journal of the Geological Society* 156, pp. 771-777.

Muhs, D. R., Holliday, V.T., 1995. Evidence of active dune sand on the Great Plains in the 19th century from accounts of early explorers. *Quaternary Research* 43, pp. 198-208.

Nanson, G.C., Chen, X.Y., Price, D.M., 1995. Aeolian and fluvial evidence of changing climate and wind patterns during the past 100 ka in the western Simpson Desert, Australia. *Palaeogeography, Palaeoclimatology, Palaeoecology* 113, pp. 87-102.

Nanson, G.C, Tooth, S., Knighton, D., 2002. A global perspective on dryland rivers: Perceptions, misconceptions and distinctions. In: Bull, L.J., Kirkby, M.J. (Eds.), *Dryland rivers: hydrology and geomorphology of semi-arid channels.* John Wiley and Sons, Chichester Ltd., pp.17-54.

Nash, D.J., 2011. Ground water controls and process. In: Thomas, D. S. (Ed.) *Arid zone geomorphology, process, form and change in drylands,* 3rd edition. John Wiley and Sons, Ltd., pp. 402- 424.

Newell, A.J., 2001. Bounding surfaces in a mixed aeolian-fluvial system (Rotliegend, Wessex Basin, SW UK). *Marine and Petroleum Geology* 18, pp. 339-347.

Nichols, G.J.,2009. *sedimentology and stratigraphy*, 2nd edition. Wiley-Blackwell, UK. 419 p.

Nichols, G. J., Hirst, J. P., 1998. Alluvial fans and fluvial distributary systems, Oligo–Miocene, northern Spain: contrasting processes and products. *Journal of Sedimentary Research* 68, pp. 879-889.

Nickling, W.G., Mckenna Neuman, C., Lancaster, N., 2002. Grainfall processes in the lee of transverse dunes, Silver Peak, Nevada: *Sedimentology* 49, pp. 191-209.

North, C.P., Boering, M., 1999. Spectral gamma-ray logging for facies discrimination in mixed fluvial-eolian successions: A cautionary tale[1]. *American Association of Petroleum Geologists Bulletin* 83, pp. 155-169.

North, C.P., Prosser, J.D., 1993. Characterization of fluvial and aeolian reservoirs: problems and approaches, in: North, C.P., Prosser, J.D. (Eds.), Characterization of fluvial and aeolian reservoirs. Geological Society of London, Special Publication 73, pp. 1-6.

Nurmi, R.D., 1985. Eolian sandstone reservoirs: Bedding facies and production geology modeling. *Society of Petroleum Engineers,* paper 14172, pp. 1-8.

Newell, A.J., 2001. Bounding surfaces in a mixed eolian–fluvial system Rotliegend, Wessex Basin, SW UK. *Marine and Petroleum Geology* 18, pp. 339-347.

Newell, A.J., Boering, M. 1999. Spectral gamma-ray logging for facies discrimination in mixed fluvial-eolian successions: a cautionary tale1. *American Association of Petroleum Geologists*, *Bulletin* 83, pp. 155-169.

Olsen, H., Due, P.H., Clemmensen, L.B., 1989. Morphology and genesis of asymmetric adhesion warts- a new adhesion surface structure: *Sedimentary Geology* 61, pp. 277-285.

Owen, G.,1987. Deformation processes in unconsolidated sands. In: M. E. Jones, M.E., Preston, R.M. (Eds.), *Deformation of sediments and sedimentary rocks*, Geological Society of London, Special Publication 29, pp. 11-24.

Øxnevad, I.E.I., 1991. *Aeolian and mixed aeolian-subaqueous sedimentation in modern and ancient sub-tropical desert basins: examples from the Sahara and Permo-Triassic of NW Europe*. Unpubl. DSc Thesis, University of Bergen, Norway, 161 p.

Paim, P.S.G., Scherer, C.M.S., 2007. High-resolution stratigraphy and depositional model of wind- and water-laid deposits in the Ordovician Guaritas rift (Southernmost Brazil). *Sedimentary Geology* 202, pp. 776-795.

Painter, T.H., Deems, J.S., Belnap, J., Hamlet, A. F., Landry, C.C., Udall, B., 2010. Response of Colorado River runoff to dust radiative forcing in snow. *Proceedings of the National Academy of Sciences of the United State of America* 107, pp. 17125-17130.

Parsons, A.J., Abrahams, A.D., 2009. *Geomorphology of desert environment,* 2nd edition. Springer Science+Business Media B.V, Netherland, 831 p.

Parsons, D.R., Wiggs, G.F, Walker, I.J., Ferguson, R.I., Garvey, B.G., 2004. Numerical modelling of airflow over an idealised transverse dune. *Environmental Modelling and Software* 19, pp. 153-162.

Parteli, E.J.R., Schwammle, V., Herrmann, H.J., Monteiro, L.H.U., Maia, L.P., 2006. Profile measurement and simulation of a transverse dune field in the Lençóis Maranhenses. *Geomorphology*, 81, pp. 29- 42.

Péron, S., Bourquin, S., Fluteau, F., Guillocheau, F., 2005. Paleoenvironment reconstructions and climate simulations of the Early Triassic: impact of the water and sediment supply on the preservation of fluvial system. *Geodinamica Acta* 18, pp. 43-446.

Petit-Mairen, N., Casta,L., Delibriags, G. , Gavin, Ch., Testuda, ,M., 1980. Preliminary data on Quaternary Palaeolacustrine deposits in the Wadi ash Shati area Libya. In: Salem, M. J., Busrewil, M.T. (Eds.), *The Geology of Libya Volume III*. Academic Press, London, pp. 797-808.

Picard, M.D., High, L.R., Jr., 1973. *Sedimentary structures of ephemeral streams.* Developments in Sedimentology 17. Elsevier, Amsterdam. 223 p.

Pickup, G., Marks, A., Bourke, M., 2002. Paleoflood reconstruction on floodplains using geophysical survey data and hydraulic modelling. *Water Science and Application* 5, pp. 47-60.

Plant, J.A., Jones, D.G., Haslam, H.W. (Eds.), 1999. *The Cheshire Basin: Basin evolution, fluid movement and mineral resources in a Permo-Triassic rift setting*. British Geological Survey. 263 p.

Poland, Z.A., Simms, A. R., 2012. Sedimentology of an erg to an erg-margin depositional system, the Rush Springs Sandstone of western Oklahoma, U.S.A.: implications for paleowinds across northwestern Pangea during the Guadalupian (Middle Permian). *Journal of Sedimentary Research* 82, pp. 345-363.

Porter, M.L., 1986. Sedimentary record of erg migration. *Geological Society of America, Geology* 14, pp. 497-500.

Powell, D.M., 2009. Dryland rivers: process and forms. In: Parsons, A.J., Abrahams, A.D. (Eds.), *Geomorphology of desert environments,* 2nd edition. Springer Science+Business Media B.V, Netherland, pp. 333-373.

Powers, R.W., Ramirez, C.D., Redmond, C.D., Elberg Jr., E.L., 1966. Geology of the Arab Peninsula: Sedimentary geology of Saudi Arabia. *U.S Geological Survey, Professional Paper* 560-DUS, Washington, pp.1-147.

Preusser, F., 2009. Chronology of the impact of Quaternary climate change on continental environments in the Arabian Peninsula. *Comptes Rendus Geoscience* 341, pp. 621-632.

Preusser, F., Radies, D., Matter, A., 2002. A 160,000-year record of dune development and atmospheric circulation in Southern Arabia. *Science* 296, pp. 2018-2020.

Prospero, J.M., Lamb. P.J., 2003. African droughts and dust transport to the Caribbean. *Climate change implications Science 302*, pp. 1024 -1027.

Pye, K., Tsoar, H., 2009. *Aeolian sand and sand dunes*, Springer, 458 p.

Pulvertaft, T.C.R., 1985. Aeolian dune and wet interdune sedimentation in the Middle Proterozoic Dala-Sandstone, *Sweden Sedimentary Geology* 44, pp. 93-111.

Ramon, J.C., Cross, T.A., 1997. Characterization and prediction of reservoir architecture and petrophysical properties in fluvial channel sandstones, Middle Magdalena Basin, Columbia. *CT and F-Ciencia, Tecnologia y Futuro* 1 (3), pp. 19-45.

Reesink, A.J., Van den Berg, J.H., Parsons, D.R., Amsler, M.L., Best, J.L., Hardy, R.J., Orfeo, O., Szupiany , R.N., 2005. Extremes in dune reservation: Controls on the completeness of fluvial deposits. *Earth-Science Reviews* 150, pp. 652-665.

Reid, I., 2002. Sediment Dynamics of ephemeral channels. In: Bull, L.J., Kirkby, M.J. (Eds*.), Dryland Rivers: Hydrology and geomorphology* of semi-arid channels. John Wiley and Sons, Chichester, pp. 107-128.

Reid, I., Frostick, E.L., 1987. Flow dynamics and suspended sediment properties in arid zone flash floods. *Hydrological Processes* 1, pp. 239-253.

Reid, I., Frostick, L.E., 1997. Channel form, flows and sediments in deserts. In: Thomas, D.S.G., (Ed.) *Arid zone geomorphology*: *process, form and change in drylands,* 2nd edition, John Wiley and Sons Ltd, UK, pp. 205-229.

Reid, I., Frostick, E.L., 2011. Channel form, flows and sediments of endogenous ephemeral rivers in deserts. In: Thomas, D. S. (Ed.), *arid zone geomorphology, process, form and change in drylands,* 3rd edition. John Wiley and Sons, pp. 301-332.

Reineck, H.E., Singh, I.B., 1980. *Depositional sedimentary environments*, 2nd edition. Springer, Berlin. 439 p.

Ringrose, P., Bentley, M., 2015., *Reservoir model design*. Springer Dordrecht Heidelberg, 249 p.

Robinson, A. Griffiths, P., Price, S., Hegre, J. , Muggeridge, A., 2008. *The future of geological modelling in hydrocarbon development*. The Geological Society of London, Special publication 309, 226 p.

Robinson, C.A, El-Baza, F., Kusky, T.M, Mainguet, M., Dumay, F., Al Suleimani, Z., Al Marjeby, A., 2007. Role of fluvial and structural processes in the formation of the Wahiba Sands, Oman: A remote sensing perspective. *Journal of Arid Environments* 69, pp. 676-694.

Rodgers, D.W., Gunatilaka, A., 2002. Bajada formation by monsoonal erosion of a subaerial forebulge, Sultanate of Oman. *Sedimentary Geology* 154, pp. 127-146.

Rodríguez-López, J.P., Meléndez, N., De Boer, P.L., Soria, A.R., 2008. Aeolian sand sea development along the mid-Cretaceous western Tethyan margin (Spain): erg sedimentology and palaeoclimate implications. *Sedimentology* 55, pp. 1253-1292.

Rodríguez-López, J.P., Meléndez, N., De Boer, P.L., Soria, A.R., 2010. The action of wind and water in a mid-Cretaceous subtropical erg-margin system close to the Variscan Iberian Massif, Spain. *Sedimentology* 57, pp. 1315-1356.

Rodriguez-Lopez, J.P., Lies, C.L., Van Dam, J., La Fuente, P., Arlegui, Ezquerro, L. and De Bore, P., 2012. Aeolian construction and alluvial dismantling of a fault-bounded intracontinental aeolian dune field (Teruel Basin, Spain); a continental perspective on Late Pliocene climate change and variability. *Sedimentology* 59, pp. 1536-1567.

Romain, H., Mountney, N.P., 2014. Reconstruction of three-dimensional eolian dune architecture from one-dimensional core data through adoption of analog data from outcrop. *American Association of Petroleum Geologist*s, *Bulletin* 98, pp. 1-22.

Rubin, D.M., 1987. *Cross-bedding, bedforms and palaeocurrents*. Society of Economic Paleontologists and Mineralogists Concepts in Sedimentology and Paleontology 1, 187 p.

Rubin, D.M., Carter, C.L., 2006. Cross-bedding, bedforms, and paleocurrents. *Society of Sedimentary Geology,* Concepts Sedimentology and Paleontology 1, 2nd edition, 195 p.

Rubin, D. M., Hunter, R.E., 1982. Bedform climbing in theory and Nature. *Sedimentology* 29, pp. 121-138.

Rust, B.R., 1981. Sedimentation in an arid-zone anastomosing fluvial system: Cooper's Creek, Central Australia. *Journal of Sedimentary Petrology* 51, 745-755.

Rust, B.R., Jones, B.G., 1987. The Hawkesbury Sandstone south of Sydney, Australia: Triassic analogue for the deposits of a large braided river. *Journal of Sedimentary Petrology* 57, pp. 222-233.

Sallenger Jr, A.H., 1979. Inverse grading and hydraulic equivalence in grain flow deposits. *Journal of Sedimentary Petrology* 49, pp. 553-562.

Salles, T., Griffiths, C., Dyt, C., 2011. Aeolian sediment transport integration in general stratigraphic forward modeling. *Journal of Geological Research* 2011, pp. 1-2.

Sarnthein, M., 1978. Sand desert during glacial maximum and climatic optimum. *Nature* 272, pp. 43-46.

Scasso, Roberto A, Dozo, M.T., Cuitino, J.B.,2012. Meandering tidal-fluvial channels and lag concentration of terrestrial vertebrates in the fluvial-tidal transition of an ancient estuary in Patagonia. *Latin American Journal of Sedimentology and Basin Analysis* 19 (1), pp. 27-45.

Scherer, C.M.S., 2000. Eolian dunes of the Botucatu Formation (Cretaceous) in southernmost Brazil: morphology and origin: *Sedimentary Geology* 137, pp. 63-84.

Scherer, C.M.S., Lavina, E.L.C., 2005. Sedimentary cycles and facies architecture of aeoliane-fluvial strata of the Upper Jurassic Guará formation, Southern Brazil. *Sedimentology* 52, pp. 1323-1341.

Schenk, C.J., Fryberger, S.G., 1988. Early diagenesis of eolian dune and interdune sands at White Sands, New Mexico. *Sedimentary Geology* 55, pp. 109-120.

Selley, R.C., 1996. *Ancient Sedimentary Environments and their sub-surface diagnosis*. 4th edition. Chapman and Hall, London. 300 p.

Simpson, E.L., Loope, D.B.,1985. Amalgamated interdune deposits, White-Sands, New-Mexico. *Journal of Sedimentary Petrology* 55, pp. 361-365.

Simpson, E.L., Eriksson, K.A., 1993. Thin eolianites interbedded within a fluvial and marine succession: Early Proterozoic Whitworth Formation, Mount Isa Inlier, Australia. *Sedimentary Geology* 87, pp. 39-62.

Simpson, E.L., Eriksson, K.A., Eriksson, P.A., Bumby, A.J., 2002. Eolian dune degradation and generation of massive sandstone bodies in the Paleoproterozoic Makgabeng Formation, Waterberg Group, South Africa. *Journal of Sedimentary Research* 72, pp. 40-45.

Simpson, E.L., Hilbert-Wolf, H.L., Simpson, W.S., Tindall, S.E., Bernard, J.J., Jenesky, T.A. and Wizevich, M.C., 2008. The interaction of aeolian and fluvial processes during deposition of the Upper Cretaceous capping sandstone member, Wahweap Formation, Kaiparowits Basin, Utah U.S.A.. *Palaeogeography, Palaeoclimatology, Palaeoecology* 270, pp. 19-28.

Sharp, R.P., 1963. Wind ripples. *The Journal of Geology* 71, pp. 617-636.

Sharp, R.P., 1966. Kelso Dunes, Mojave Desert, California. *Geological Society of America*, Bulletin 77, pp. 1045-1074.

Shepherd, M., 2009, Eolian reservoirs. In: Shepherd, M. (Ed.), *Oil field production geology.* American Association of Petroleum Geologist, Memoir 91, pp. 255-259.

Slatt, R.M., 2013. *Stratigraphic reservoir characterization for petroleum geologist, geophysicists, and engineers, 2nd edition*. Development in petroleum science 61, Elsevier, Amsterdam, 670 p.

Slatt, R. M., Phillips, S., Boak, J.M., Lagoe, M.B., 1993. Scales of geological heterogeneity of a deep-water sand giant oil field, Long Beach unit, Wilmington field, California. In: Rhodes, E.G., Moslow, T.F. (Eds.), *Marine clastic reservoirs, examples and analogs*. Springer-Verlag, New York, pp. 263-292.

Smith, R.A., Jones, N.S., Monaghan, A.A., Arkley, S., 2006. Fluvial and aeolian deposition in the Siluro-Devonian Swanshaw Sandstone Formation, SW Scotland. *Scottish Journal of Geology* 42, pp. 161-177.

Soltan, R., Mountney, N.P., 2016. Interpreting complex fluvial channel and barform architecture: Carboniferous central Pennine Province, northern England. *Sedimentology* 63, pp. 207-252.

Spalletti, L.A., Veiga, G.D., 2007. Variability of continental depositional systems during lowstand sedimentation: an example from the Kimmeridgian of the Neuquen basin, Argentina. *Latin America Journal of Sedimentology and Basin Analysis* 14, pp. 85-104.

Spalletti, L.A., Limarino, C.O., Colombo, F., 2010. Internal anatomy of an Erg sequence from the aeolian-fluvial system of the De La Cuesta Formation (Paganzo Basin, northwestern Argentina). *Geologica Acta* 8, pp. 431-447.

Spencer, N., Macklin, M.G., Woodward, J.C., 2012. Re-assessing the abandonment of Amara West: the impact of a changing Nile?. *Sudan and Nubia 16*, pp. 37-43.

Stanistreet, I.G., Stollhofen, H., 2002. Hoanib River flood deposits of Namib Desert interdunes as analogues for thin permeability barrier mudstone layers in aeolianite reservoirs. *Sedimentology* 49, pp. 719-736.

Stear W. M., 1985. Comparison of the bedform distribution and dynamics of modern and ancient sandy ephemeral flood deposits in the southwestern Karoo Region, South Africa. *Sedimentary Geology* 45, pp. 209-230.

Steel, R.J., Thompson, D.B., 1983. Structures and textures in Triassic braided stream conglomerates ('Bunter' Pebble Beds) in the Sherwood Sandstone Group, North Staffordshire, England. *Sedimentology* 30, pp. 341-368.

Stemmerik, L., Ineson, J.R., Mitchell, J.G., 2000. Stratigraphy of the Rotliegend Group in the Danish part of the Northern Permian Basin, North Sea. *Journal of the Geological Society of London* 157, pp.1127-1136.

Stikes, M.W., 2007. Fluvial facies and architecture of the Poison Strip Sandstone lower Cretaceous Cedar Mountain Formation, Grand County, Utah. Miscellaneous Publication 06-2, Utah, *US Geological Survey*, 111 p.

Strebelle, S., Levy, M., 2008. Using multiple-point statistics to build geologically realistic reservoir models: the MPS/FDM workflow. In:

Robinson, A. Griffiths, P., Price, S., Hegre, J. , Muggeridge, A. (Eds.), *The future of geological modelling in hydrocarbon development*. The Geological Society of London, Special publication 309, pp. 67-74.

Strömbäck, A., Howell, J.A., Veiga, G.D., 2005. The transgression of an erg – sedimentation and reworking soft-sediment deformation of aeolian facies: the Cretaceous Troncoso Member, Neuquen Basin, Argentina. In: Veiga, G.D., Spalletti, L.A., Howell, J.A., Schwarz, E. (Eds.), *The Neuquén Basin, Argentina: A case study in sequence stratigraphy and basin dynamics.* Geological Society, London, Special Publications 252, pp. 163-183.

Svendsen, J., Stollhofen, H., Krapf, C.B., Stanistreet, I.G., 2003. Mass and hyperconcentrated flow deposits record dune damming and catastrophic breakthrough of ephemeral rivers, Skeleton Coast Erg, Namibia. *Sedimentary Geology,* 160, pp. 7-31.

Sweet, M.L., 1992. Lee-face airflow, surfaces processes, and stratification types: their significance for refining the use of eolian cross-strata as paleocurrent indicators. *Geological Society of America*, Bulletin 104, pp. 1528-1538.

Sweet, M.L., 1999. Interaction between aeolian, fluvial and playa environments in the Permian Upper Rotligend group, UK southern North Sea. *Sedimentology* 46, 171-187.

Sweet, M.L., Nielson, J., Havholm, K., Farrelly, J., 1988. Algodones dunefield of SE California: case history of a migrating modern dune field. *Sedimentology* 35, pp. 939-952.

Sweet, M.L., Blewden, C. J., Carter, A.M., Mills, C.A., 1996. Modeling heterogeneity in a low-permeability gas reservoir using geostatistical techniques, Hyde field, southern North Sea: *American Association of Petroleum Geologist, Bulletin* 80, pp .1709-1735.

Swezey, C., Lancaster, N., Kocurek, G., Deynoux, M., Blum, M., Price, D., Pion, J.C., 1999. Response of aeolian systems to Holocene climatic and hydrologic changes on the northern margin of the Sahara: A high resolution record from the Chott Rharsa basin, Tunisia. *The Holocene* 9, pp. 141-148.

Swiecicki, T., Wilcockson, P., Canham, A., Whelan, G., Homann, H., 1995. Dating, correlation and stratigraphy of Triassic sediment in west Shetlands area. In: Boldy, S.A. (Ed.), *Permian and Triassic rifting in northwest Europe*, Geological Society of London, Special Publication 91, pp. 57-85.

Taggart, S., Hampson, G.J., Jackson, M.D., 2010. High-resolution stratigraphic architecture and lithological heterogeneity within marginal aeolian reservoir analogues. *Sedimentology* 57, pp. 1246-1279.

Talbot, M.R., 1985. Major bounding surfaces in aeolian sandstones-a climatic model. *Sedimentology* 32, pp. 257-265.

Taylor, I.E., Middleton, G.V., 1990. Aeolian sandstones of the Copper Harbor Formation, Late Proterozoic, Lake Superior basin. *Canadian Journal of Earth Sciences* 27, pp.1339-1347.

Tirsgaard, H., Øxnevad, I.E.I., 1998. Preservation of pre-vegetational mixed fluvio-aeolian deposits in a humid climatic setting: an example from the Middle Proterozoic Eriksfjord Formation, Southwest Greenland. *Sedimentary Geology* 120, pp. 295-317.

Thesiger, W., 1949. A further journey across the Empty Quarter. *Geographical Journal* 113, pp. 21-46.

Thompson, D.B.,1969. Dome shaped aeolian dunes in the Frodsham Member of the so-called 'Keuper' Sandstone Formation [Scythian - ? Anisan: Triassic] at Frodsham, Cheshire (England). *Sedimentary Geology* 3, pp. 263-289.

Thompson, D.B. 1970a. The stratigraphy of the so-called Keuper Sandstone Formation (Scythian-?Anisian) in the Permo-Triassic Cheshire Basin. *Geological Society of London*, *Quarterly Journal* 126, pp. 151-181.

Thompson, D.B., 1970b. Sedimentation of the Triassic (Scythian) red pebbly sandstones in the Cheshire Basin and its margins. *Geological Journal* 7, pp. 183-216.

Thompson., J., Meadows, N.S., 1997. Clasticsabk has and diachroneity at the top of the Sherwood Sandstone Group: East Irish Sea Basin. *Geological Society of London, Special Publication* 124, pp. 237-251.

Thomas, D.S., Stokes, S., Shaw, P.A., 1997. Holocene aeolian activity in the southwestern Kalahari Desert, southern Africa: significance and relationships to late-Pleistocene dune building events. *The Holocene* 7, pp. 273-81.

Tooth, S., 2000a. Process, form and change in dryland rivers: a review of recent research. *Earth-Science Reviews* 51, pp. 67-107.

Tooth, S., 2000b. Downstream changes in dryland river channels: the Northern Plains of arid central Australia. *Geomorphology*, 34, pp. 33-54.

Tooth, S., 2006. Virtual globes: a catalyst for the re-enhancement of Geomorphology?. *Earth Surface Processes and Landforms* 31, pp. 1192-1194.

Tooth, S., Nanson, G.C., 2011. Distinctiveness and diversity of arid zone river systems. In: Thomas, D.S. (Ed*.), Arid zone geomorphology*, process, *form and change in drylands*, 3rd edition John Wiley and Sons Ltd. UK, pp. 269-300.

Topal, S., Özkul, M., 2014.Soft-Sediment Deformation Structures Interpreted as Seismites in the Kolankaya Formation, Denizli Basin (SW Turkey). The *Scientific World Journal*, Article 352654, pp. 1-13.

Tresise, G.R. 1993. Triassic vertebrate footprints from Cheshire, England; localities and lithologies. *Modern Geology* 18, pp. 407-417.

Tresise, G.R. 1994. The Runcorn Quarries and the footprint finds of the 1840s. *Proceedings of the Geologists Association* 105, pp. 125-140.

Tresise, G.R., Sarjeant, W. A. 1997. *The tracks of Triassic Vertebrates: fossil evidence from north-west England*. The Stationery Office, Norwich, 204 p.

Trewin, N.H., 1993. Controls on fluvial deposition in mixed fluvial and aeolian facies within the Tumblagooda Sandstone (Late Silurian) of Western Australia. *Sedimentary Geology* 85, pp. 387-400.

Tripaldi, A., Limarino, C.O., 2005. Vallecito Formation (Miocene): The evolution of an eolian system in an Andean foreland basin (northwestern Argentina). *Journal of South American Earth Sciences* 19, 343-357.

Tsoar, H., 1978. *The dynamics of longitudinal dunes*. Final technical report Da-ERO 76-G-072, European Research Office, US Army, London.

Tucker, 2011. *Sedimentary Rocks in the Field: A Practical Guide*, 4th edition. Wiley-Blackwell. 270 p.

Valyashko, M.G., 1972. Scientific works in the field of geochemistry and the genesis of salt deposits in the USSR. Earth Science (Paris), *Sciences de la Terre* 7, pp. 289-311.

Veiga, G.D., Luis, A. Spalletti, L.A., , Flint, S.S., 2002. Aeolian/fluvial interactions and high-resolution sequence stratigraphy of a non-marine (Lowerstand wedge: the Avile Member of the Agrio Formation (Lower

Cretaceous), central Neuquen Basin, Argentina. *Sedimentology* 49, pp. 1001-10149.

Veiga, G.D., Spalletti, L.A., , Flint, S.S., 2007. Anatomy of a fluvial lowstand wedge: the Avilé Member of the Agrio Formation (Hauterivian) in central Neuquén Basin (northwest Neuquén Province), Argentina. In: Nichols, G., Williams, E., Paola, C.(Eds.) *Sedimentary processes, environments and basins: a tribute to peter friend*. International Association of Sedimentologists, special publication 38, pp. 341-365.

Venus, J.H., Mountney, N.P., McCaffrey, W.D., 2015. Syn-sedimentary salt diapirismas a control on fluvial-system evolution: an example from the proximal Permian Cutler Group, SE Utah, USA. *Basin Research* 27, pp. 152-182.

Visser, C.A. Chessa, A.G., 2000. Estimation of length distributions from outcrop datasets–application to the Upper Permian Cutler Formation Utah. *Petroleum Geoscience* 6, pp. 29-36.

Visser, S.M., Sterk, G., Ribolzi, O., 2004. Techniques for simultaneous quantification of wind and water erosion in semi-arid regions. *Journal of Arid Environments* 59, pp. 699-717.

Wainwright, J., Bracken, L.J., 2011. Runoff generation, overland flow and erosion on hillslopes, In: Thomas, D. S. (Ed.), *Arid zone geomorphology, process, form and change in drylands*, 3rd edition. John Wiley and Sons, pp. 237-267.

Warrington, G.,1970. The stratigraphy and palaeontology of the 'Keuper' Series of the central Midlands of England. *Geological Society of London*, Quarterly Journal 12 (6), pp.183-223.

Warrington, G. and Ivimey-Cook, H.C., 1992. Triassic, in: J.C.W. Cope,J.K. Ingham and P.F. Rawson, (Eds.), *Atlas of Palaeogeography and Lithofacies.* Geological Society of London, Memoir 13, pp. 97-106.

Warrington, G., Audley-Charles, M.G., Elliott, R.E., Evans, W.B., Ivemey-Cook, H.C., Kent, P., Robinson, P.L., Shotton, F.W., Taylor, F.M., 1980. A correlation of Triassic rocks in the British Isles. *Geological Society of London*, Special Report 13, 78 p.

Ward, J.D., 1983. A reappraisal of the Cenozoic stratigraphy in the Kuiseb valley of the central Namib desert. In: Vogel, J.C. (Ed.), *Late Cainozoic palaeoclimates of the southern hemisphere*. Rotterdam, pp. 455-63.

Ward, J.D., 1987. The Cenozoic succession in the Kuiseb valley, Central Namib Desert. *Geological Survey of Namibia, Memoir* 9, 124 p.

Ward, J.D., 1988. Eolian, fluvial, and pan (playa) facies of the Tertiary Tsondab Sandstone in the central Namib Desert, Namibia. Sedimentary Geology 55, pp. 143-162.

Warren, A., 1988. The dunes of the Wahiba Sands. In: Dutton, R.W. (Ed.), *The Scientific results of the royal geographical society's Oman Wahiba Sands project 1985 -1987.* Journal of Oman Studies, special report 3. Ministry of National Heritage and Culture, Muscat, Sultanate of Oman, pp. 131-160.

Warren, A., Knott, P., 1983. Desert dunes: A short review of the needs in desert dune research and a recent study of micro-meteorological dune-initiation mechanisms. In: Brookfield, M.E., Ahlbrandt, T.S. (Eds). *Eolian sediments and processes developments*, Development in Sedimentology 38, pp. 343-352.

Weber, K.J., 1987. Computation of initial well productivities in aeolian sandstones on the basis of a geological model, Leman Gas Field, UK. In: Tillman, R.W., Weber, K.J. (Eds.), *Reservoir Sedimentology.* Society of Sedimentary Research, Special Publication 40, pp. 333–354.

Wehr, F., Brasher, I.D., 1996. Impact of sequence-based correlation style on reservoir model behaviour, lower Brent Group, North Cormorant Field, UK North Sea. In: Aitken, J.F., Howell, J.A., (Eds.), *High resolution sequence stratigraphy: Innovations and applications*. Geological Society of London, Special Publication 104, pp. 115-128.

Weissmann, G.S., Hartley, A.J., Nichols, G.J., Scuderi, L.A., Olson, M., Buehler, H., Massengill, L.C., 2011. Alluvial facies distributions in continental sedimentary basins Distributive fluvial systems. In: Davidson, S.K., Leleu, S., North, C. (Eds.), *From river to rock record: the preservation of fluvial sediments and their subsequent interpretation.* Society of Sedimentary Geology, Special Publication 97, pp. 327-355.

Werner, B.T., Kocurek, G., 1997. Bed-form dynamics: does the tail wag the dog. *Geology* 25, pp. 771-774.

White, K., Bullard, J., Livingstone, I., Moran, L., 2015. A morphometric comparison of the Namib and southwest Kalahari dunefields using ASTER GDEM data. *Aeolian Research* 19, pp. 87-95.

Wilkins, D.E., Ford, R.L., 2007. Nearest neighbour methods applied to dune field organization: the Coral Pink Sand Dunes, Kane County, Utah, USA. *Geomorphology* 83, pp. 48-57.

Wilson, I.G., 1971. Desert sandflow basins and a model for the development of ergs. *Geographical Journal* 137, pp. 180-199.

Wilson, I.G., 1972. Aeolian bedforms – Their development and origins: *Sedimentology* 19, pp. 173-210.

Wilson, I.G., 1973. Ergs, *Sedimentary Geology* 10, pp. 77-106.

Winspear, N.R., Pye, K., 1995. Sand supply to the Algodones dunefield, southeastern California, USA. *Sedimentology* 42, pp. 875-891.

Woodward, K., Curtis, C. D., 1987. Predictive modelling of the distribution of production constraining illites Morecambe Gas Field, Irish Sea, Offshore UK. In: Brooks, J., Glennie, K. (Eds.), *Petroleum Geology of North West Europe.* Graham and Trotman, pp. 205-215.

Woodward, J., Macklin, M., Fielding, L., Millar, I., Spencer, N., Welsby, W., Williams, M., 2015. Shifting sediment sources in the world's longest river: A strontium isotope record for the Holocene Nile. *Quaternary Science Reviews* 130, pp. 124-140.

Wright, V.P., 1986. *Paleosols their recognition and interpretation*. Oxford, Balkwell Scientific, 315 p.

Yan, P., Li, X. Ma, F., Wu, W., 2015. Yao Qian Morphological characteristics of interactions between deserts and rivers in northern China. *Aeolian Research*. In press, doi:10.1016/j.aeolia.2015.01.005.

Yang, X., Jaekelc, Z.D., Owend, L.A., Han, J., 2002. Late Quaternary palaeoenvironment change and landscape evolution along the Keriya River, Xinjiang, China: the relationship between high mountain glaciation and landscape evolution in foreland desert regions. *Quaternary International* 97-98, pp. 155-166.

Yang, L., Ding, Z., 2013. Expansion and contraction of Hulun Buir Dunefield in north-eastern China in the last late glacial and Holocene as revealed by OSL dating. *Environmental Earth Sciences* 68, pp. 1305-1312.

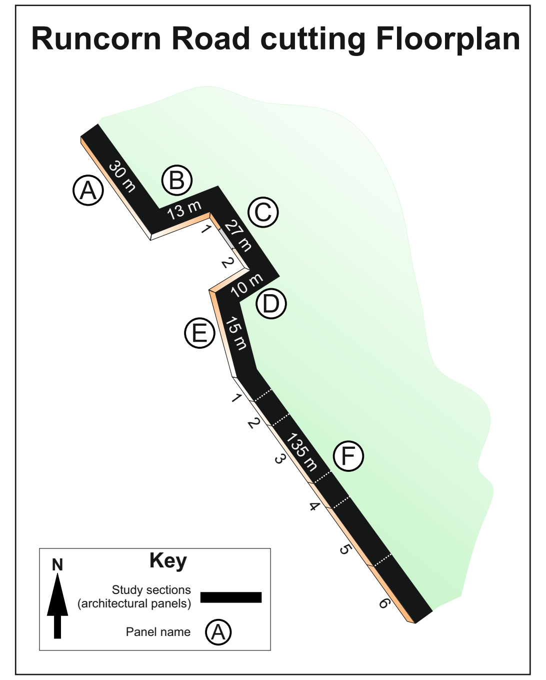
Yu, L., Gong, P., 2012. Google Earth as a virtual globe tool for earth science applications at the global scale: progress and perspectives. *International Journal of Remote Sensing* 33, pp. 3966-3986.

Ziegler, P.A, Dèzes, P., 2006. Crustal evolution of Western and Central Europe. *Geological Society of London*, *Memoirs* 32, pp. 43-56.

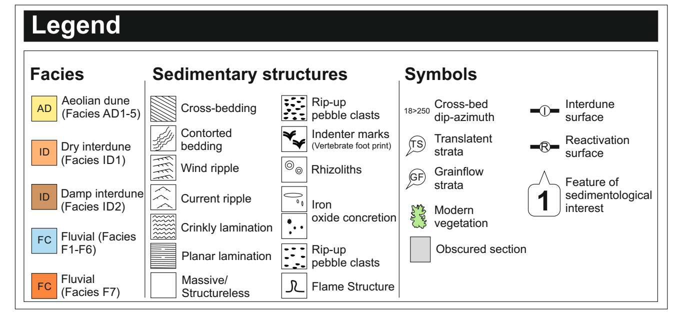
Zou, F., 2013. Geologic and engineering modelling. In: Slatt, R.M., *Stratigraphic reservoir characterization for petroleum geologist, geophysicists, and engineers*, 2nd edition. Development in petroleum science 61, Elsevier, pp. 621-656.

Appendix

Appendix 1: Composite architectural panels depicting the stratigraphic architecture of the Wilmslow Sandstone and Helsby Sandstone Formations as observed in Runcorn Expressway road cutting outcrop.

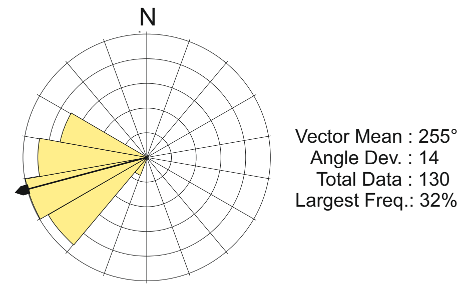


1 Stacked aeolian dune-strata separated by interdune surface.
2 Moderate-high angle aeolian dune foresets (grainflow dominated), pinching out toward the base off lee-slope.
3 Dune toe-set soft-sediment deformation, laterally intertonguing with damp interdune strata.
4 Fluvial faint flat lamination; homogonous grain size due to well-sorted nature of grains.
5 Dune strata with erosional truncation surface; deflation surface.
6 Downwind-dipping reactivation surfaces.
7 Spatially isolated damp interdune unit.
8 Damp interdune unit showing erosive topography on both upper and lower surface, pinches out laterally.
9 Medium-to coarse-grained sandstone dominated by planar and wind-ripple laminated dry interdune strata.
10 Lateral transition from crinckly-wavy laminated (damp) to wind-ripple (dry) interdune strata.
11 Lateral thickening of wavy laminated interdune horizon.
12 Low-angle inclined aeolian dune foresets; wind-ripple dominated (translatent wind-ripple strata common).
13 Near horizontal, bi-modal grain size (pinstripe lamination) forming clear grain size segregation unit; pinches out laterally and merge with interdune facies.
14 Stacked aeolian dune foreset deposits that lack clear stratification due to grain size homogeneity.
15 Down-folded lamination in the uppermost layers of the damp interdune represent cross-sectional view of vertebrate footprint structure.
16 Fluvial channel base; erosional bounding surface.
17 Intraformational rip-up clasts composed of red (laminated silty mudstone); highly variable size and shape. Basal lag deposits associated with channel fill-element.
18 Lense-shaped interdune element.
19 Wavy to crinkly laminated sandstone, dominated by modified wind-ripple strata and adhesion-ripple strata; accumulated in interdune areas between merging aeolian dune forms.
20 Planar thinly (10-30 cm) laminated Silty mudstone.
21 Fluvial flood surface with erosive base exhibiting local topography and planar laminated silty mudstone fill.
22 Planar-tabular sets representing the migration of sandy bedforms within a fluvial channel system.
23 Sharp boundaries at the top of the channel-fill element record an abrupt abandonment phase.
24 Small-scale water escape (flame) structures in upper part of damp interdune units.
25 Fluvial trough cross-beds, representing the migration and accumulation of sandy mesoforms within a fluvial channel system.
26 Trough cross bedding of fluvial origin, characterised by very coarse- to gravel grade lag deposits of locally derived materials(mud clasts).
27 Planar (horizontal) discontinuous , thinly laminated fine-to very fine-grained sandstone, with current lamination and mud-drapped climbing-ripple stratification toward the top.
28 Massive sandstone; lack of organised internal lamination; reflects the homogeneity of sandstone grain size.
29 Fluvial channel scour surface

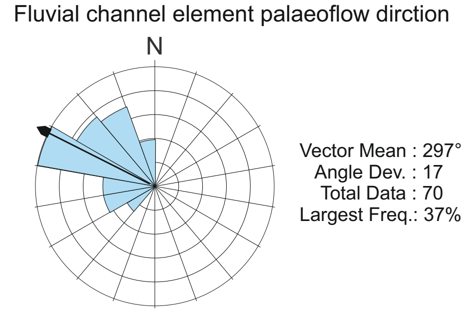


Runcorn Road cut architectural panels

Aeolian sand dune element flow direction



Fluvial channel element palaeoflow direction



Facies Summary

Aeolian Facies

AD1: Aeolian dune, moderate to high-angle bedded, grainflow dominated.
AD2: Aeolian dune, moderate to high-angle beeded, grainfall dominated.
AD3: Aeolian dune, Low to moderate-angle beede. Windripples dominated.
AD4: Aeolian dune, massive, lacks clear stratification.
AD5: Aeolian dune, dune toe set soft-sediment deformation.
ID1: Aeolian interdune, horizontally laminated, wind-ripple dominated.
ID2: Aeolian interdune, crinckly laminated, modified wind-ripple and adhision-ripple strata.

Fluvial Facies

FC1: Fluvial channel-lag deposits, intraformational mud clasts.
FC2: Fluvial channel fill, planar tabular bedding, straight crested dunes.
FC3: Fluvial channel fill, trough cross-beds, sinuous crested dunes.
FC4: Fluvial channel fill, planar thinly laminated, upper flow regimes.
FC5: Fluvial channel fill, massive sandstone, rapid deposition, also due to grain size homogeneity.
FC6: Fluvial ch. fill, Ripple cross laminated sandstone with mud drapes.
FC7: Fluvial ch. fill or flood plain, mud-siltstone, suspension deposits.