

**Modern and Ancient Fluvial-Aeolian Interactions in Dune-
Field Margin Systems: Analysis and Implications**

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Abstract

Fluvial drainage networks are common at the margins of desert basins. Some penetrate 10^1 to 10^2 km into the inner parts of aeolian dune fields; others are dammed and ponded at outer margins. Some systems occupy long-lived fluvial corridors that partition dune fields; others occupy transient interdune corridors that open and close as desert dunes migrate. Thus, a range of styles of fluvial-aeolian interaction arise. Results from two case studies are documented here: the presently active Skeleton Coast Erg, Namibia, and the Triassic Helsby Sandstone Formation, Cheshire Basin, UK. Fluvial incursions into desert basins are controlled by precipitation-event frequency and magnitude, fluvial runoff distance from catchment to receiving basin, sediment yield, changes to regional water-table level in response to flood events, and long-term climate change. These factors conspire to determine fluvial discharge to the receiving basin in the form of confined floods within channelized networks or as widespread non-confined sheet flows in dune-field margins. Modern system interactions include: (i) the establishment of long-lived major through-going open interdune corridors along which well-established rivers pass; (ii) the damming of river courses by active aeolian dunes that form barriers to flow, thereby resulting in the ponding of flood waters and the development of large, slowly draining flood basins; (iii) the passage of flood waters as sheet flows into the outer margins of aeolian dune fields. Ancient preserved examples of fluvial-aeolian interactions include: (i) relationships indicative of systematic temporal change from an aeolian dune field characterised by small, isolated, dry interdunes to one in which interdunes were large and interconnected such that they acted as conduits for fluvial flow whereby fluvial channels were able to penetrate into dune-field centre settings; (ii) evidence for fluvial reworking of aeolian dune deposits by erosive flows that resulted in temporary cessation in dune migration in the immediate aftermath of flood events.

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Chapter 1

Thesis Introduction

1.1 Introduction

Interactions between competing fluvial-aeolian sedimentary systems are common at the margins of desert regions where drainage networks penetrate into the inner parts of aeolian dune fields over distances of 10^0 - 10^2 km. Some fluvial systems occupy long-lived fluvial corridors that partition dune fields; others temporarily exploit transient interdune corridors that open and close as aeolian-dunes migrate. This study documents the effects of temporal and spatial variability on the preserved stratigraphic architecture of mixed fluvial-aeolian systems arising from: (i) morphological changes in aeolian dune configuration; (ii) variability in processes of sediment transport across desert basins for both fluvial and aeolian systems; (iii) variability in mechanisms of sediment accumulation and preservation.

Controls on fluvial incursions into dune fields include temporal and spatial changes in the frequency and magnitude of precipitation events, fluvial runoff distance from catchment to receiving basin, sediment yield, changes to regional water-table level in response to flood events, subtle variations in palaeotopography of the accumulation surface. These factors conspire to determine whether floods are confined within channelized networks or occur as non-confined sheet-like flows that inundate large areas of dune-field margins. Results from two case studies are presented: the present-day Skeleton Coast erg (dune field) system of northern Namibia and the ancient Triassic Helsby Sandstone Formation of the Cheshire Basin, England, UK.

Present-day active aeolian-fluvial system interactions in aeolian dune fields and at their margins include the following types: (i) the establishment of long-lived major through-going open interdune corridors along which well-established rivers pass; (ii) the damming of river courses by active aeolian dunes that form barriers to flow, thereby resulting in the ponding of flood waters and the development of large, slowly draining flood basins; (iii) the

passage of flood waters as non-confined sheet-like flows into the outer margins of aeolian dune fields.

The preserved stratigraphic expression of ancient fluvial-aeolian system interactions include the following: (i) relationships indicative of systematic temporal change from an aeolian dune field characterized by preserved interdune elements indicative of small, isolated dry interdunes at the time of sediment accumulation to one in which preserved interdune elements are indicative of large and interconnected interdunes at the time of accumulation that acted as conduits for fluvial flow whereby fluvial channels were able to penetrate into dune-field centre settings; (ii) evidence for fluvial reworking of aeolian dune deposits by erosive flows that resulted in temporary cessation in dune migration in the immediate aftermath of flood events. Such interactions influence the generation of a supply of sediment for aeolian system construction and its availability for aeolian transport. These factors govern how mixed aeolian and fluvial successions accumulate and how they ultimately become preserved.

1.2 Aim and Objectives

Sedimentological studies of the interactions between fluvial and aeolian systems have been widely documented from many settings (see Rodríguez-López et al., 2014 for review). The scientific understanding of the depositional mechanisms associated with such system interactions and criteria for the recognition of the accumulated and preserved lithofacies arrangements and stratigraphic relationships of bodies of strata arising from such interactions in both modern and ancient systems alike is here extended through the application of novel and innovative studies techniques.

The principal aim of this research is to develop a series of predictive facies models with which to relate processes associated with present-day fluvial and aeolian system interactions to analogous preserved successions preserved in the ancient rock record. Fulfilment of this aim will enable allogenic controls on sedimentation to be differentiated from autogenic controls. This aim is achieved through the implementation of a varied and novel set of data collection methods that integrates high-resolution image analyses of currently active systems and detailed lithofacies analyses of ancient preserved successions. The aim of this research is accomplished

through the following case studies: 1) analysis of present-day aeolian-fluvial system interactions in the Skeleton Coast erg (Chapter 3) using high-resolution satellite images collected from Google Earth Pro software; and 2) analysis of the facies and architectural-element arrangement of rock bodies that comprise a mixed fluvial and aeolian succession in the Triassic Helsby Sandstone Formation (Chapter 4) through analysis of outcrop data including detailed vertical sedimentary logging, palaeocurrent analysis, 2D and 3D architectural-element analysis and the development of 3D depositional and sequence stratigraphic models. Utilisation of facies models from the studied examples of the Skeleton Coast erg and the Helsby Sandstone Formation allows for discussion (Chapter 5) of the role played by the geometry and distribution of relatively low-permeability fluvial and damp/wet interdune elements within otherwise high-permeability, aeolian dune-dominated successions. Such low-permeability elements act as baffles and barriers to flow in analogous subsurface reservoir successions, especially in aeolian preserved dune-field margin reservoir intervals.

Specific objectives of this research are as follows:

- To describe the nature of identified styles of interaction that occur between aeolian dune, interdune, channelized fluvial and non-channelized floodplain sub-environments as a result of the following: 1) allogenic factors such as long-term climate change, significant changes in long-term sediment supply, and changes in accommodation generation; 2) autogenic factors arising through competition between coevally-active aeolian and fluvial systems. Fulfilment of this objective will utilize data acquired from both satellite imagery and outcrop analysis.
- To construct a series of high-resolution, three-dimensional models based on observations of complex modern and ancient aeolian and fluvial interactions to determine the wide variety of identified interactions known to occur and to present these interaction types within a framework of spatial and temporal relationships for erg-margin systems and their preserved successions.
- To develop generic sequence stratigraphic models with which to explain and account for the controls responsible for the observed interactions, and to investigate the nature of the observed

relationships in terms combined and interacting autogenic and allogenic factors.

- To demonstrate how careful field examination can be used to document detailed stratigraphic architectural relationships that can be used as the basis for the development of high-resolution, three-dimensional facies models, which can themselves be used to account for the style and mechanisms of accumulation and preservation of fluvial and aeolian successions. In fulfilment of this objective an explicit research outcome of this work will be to show the intimate role of water-table level as a control on so-called wet aeolian system accumulation and preservation for case examples that are influenced by fluvial flood events into the marginal parts of aeolian dune fields.
- To develop a discussion that considers the principal factors that influence sedimentation in erg-margin settings, including an examination of the role of water-table level, sediment supply, and sediment availability.
- To provide a summary of three-dimensional stratigraphic relationships present at sub-seismic scale that cannot be determined directly from subsurface reservoir datasets provided by core and well-log data which are essentially one-dimensional in form.
- To summarise the applied significance of this research and to consider future avenues of applied sedimentological research by demonstrating how the studied Helsby Sandstone Formation (Cheshire Basin, UK) can serve as a reservoir analogue.

1.3 Thesis Structure

To provide an outline of the framework of this thesis, a brief summary is here provided of the content of each of the chapters. Table 1.1 summarizes the structure of the thesis showing key information. Chapters 3 and 4 of this thesis have been written in a style such that they can be submitted as stand-alone pieces of publishable research. As such, these chapters both contain sections that serve as independent Introduction, Background, Geological

Setting, Methods, Data Analysis, and Discussion sections for the work presented within; they each represent stand-alone case studies.

Chapter Two

This chapter provides a general discussion on the sedimentology and stratigraphy associated with fluvial and aeolian systems interactions that are documented from fluvial-dominated and aeolian-dominated erg-margin environments. Controls on the accumulation and preservation of both dry and wet aeolian systems associated with dryland river incursions are also reviewed.

Chapter Three

This chapter, which comprises ~30% of the thesis, considers modern types of fluvial and aeolian system interaction through a detailed analysis of the present-day Skeleton Coast erg (dune field), which has been studied using high-resolution satellite imagery data collected from Google Earth Pro software. The 6 to 22 km-wide study area spans ~300 km of the Atlantic coastline of NW Namibia, across which several ephemeral rivers drain south-southwest towards the Atlantic Ocean. During major flood-events, a 2000 km² region of active aeolian construction and accumulation acts as a major obstacle to fluvial flow at the eastern erg margin. Fluvial systems are subject to ephemeral or intermittent flow but undergo marked changes in discharge in response to seasonal monsoonal rainfall events in their continental-interior catchments. The analysis of a time-series of satellite images reveals the history of recent evolution the Hoanib, Hunkab and Uniab rivers that intersect the erg and reveals evidence for interaction between the competing fluvial versus aeolian processes. The results of this chapter serve as the basis for the development of a series of models that can be used to predict likely preserved sedimentological and stratigraphic arrangements in aeolian dune-field margins that are subject to repeated fluvial incursions.

Chapter Four

This chapter, which comprises ~50% of the thesis, characterizes a series of outcrops of the Triassic Helsby Sandstone Formation (HSF; Anisian) in the

Cheshire Basin of NW England. Studied deposits constitute a variety of types of sediment accumulation of mostly mixed aeolian and fluvial origin. In this study, the HSF has been the subject of detailed lithofacies and architectural-element analysis to establish the dominant factors that governed the nature of interactions between the competing fluvial and aeolian depositional systems that were apparently active coevally in many parts of the basin during its development. The collected outcrop data in this chapter provide a detailed dataset with which to observe and document the sedimentary signatures that reveal evidence for the nature of interactions between competing fluvial and aeolian processes in the HSF. The data collected from these areas include lithological graphic logs, architectural panels of the selected exposures and palaeocurrent data from both fluvial and aeolian facies. The facies and architectural-element analysis undertaken as part of this work have been combined with other published and unpublished data collected from the region of study to produce a series of predictive sedimentological models for the inferred palaeoenvironment and its preserved stratigraphic expression. In this chapter, the interpretation of the ancient succession allows for a more general discussion of the main forcing factors responsible for determining facies distributions (both laterally and vertically) within both the aeolian and fluvial parts of the studied succession.

Chapter Five

This chapter discusses the potential applications of the findings of this research. Specifically, this chapter discusses how the proposed models and results from chapters 3 and 4 can be used to assess likely subsurface prospectively in mixed fluvial and aeolian reservoir successions known from both aeolian-dominated inner erg-margin settings and fluvial-dominated outer erg-margin settings. A general discussion is also provided that considers further developments in our understanding of processes that operate in mixed fluvial and aeolian systems in arid and semi-arid environmental settings. This is enabled by the development of a series of a predictive facies and sequence stratigraphic models. This chapter considers how these models can be applied as tools with which to estimate the likely

distribution of high-quality in porosity and permeability zones in analogous subsurface reservoirs within similar subsurface settings.

Chapter Six: Conclusion

This chapter summarizes the main outcomes of the research and explains how the principal findings could be applied in future work. Specifically it explains how the proposed models developed as part of this research could be used in the exploration, appraisal and development of mixed fluvial and aeolian reservoir successions. Finally, consideration is given to potential future work and to possible future enhancements to the methods presented.

Table 1.1: Summary of Thesis Structure

| <i>Chapter</i> | <i>Title</i> | <i>Description</i> |
|----------------|--|--|
| 2 | Literature Review | Brief discussion on the sedimentological controls that operate within zones of aeolian-fluvial interaction; considers both fluvial-dominated and aeolian-dominated erg-margin systems. |
| 3 | Styles of Interaction between Fluvial and Aeolian Systems: Skelton Coast, Northern Namib Desert | A case-study of presently and recently active types of fluvial and aeolian system interaction through a detailed analysis of high-resolution satellite imagery data collected from Google Earth Pro software. |
| 4 | Preserved Sedimentary Record of Fluvial-Aeolian Interaction: Triassic Helsby Sandstone Formation, Cheshire Basin, UK | Outcrop analyses of identified lithofacies within the Helsby Sandstone Formation. The interpretation of lithofacies in a series of architectural panels allows the three-dimensional geometry of architectural elements to be determined. Facies distributions are shown laterally and vertically in response to a series of autogenic and allogenic controls. |
| 5 | Discussion | Consideration of the implications of fluvial and aeolian interaction for prediction of subsurface reservoir heterogeneity and flow behaviour using the interpreted data from chapters 3 and 4. |
| 6 | Conclusion and Future Work | A brief summary of the results of the research. considers how best to utilise the findings of this work to advance the state of the science in the future. |

Chapter 2

Literature Review

2.1 Aeolian-fluvial interaction at erg margins

In erg-margin areas, sedimentary processes associated with interaction between competing aeolian and fluvial systems play a significant role in shaping dryland environments. These types of interactions in desert systems do not occur independently but rather they are related dynamically via sediment flux (e.g. Kocurek, 1998; Bullard and Livingstone, 2002; Cain and Mounney 2011). The sediment flux is defined as the total volume or mass of sediment entering (influx, Q_i) and exiting an aeolian system (outflux, Q_o) (Kocurek and Havholm, 1993). The different sedimentary characteristics of the various environments present in desert systems each respond to interactions differently yet predictably. As such, changes between environments where aeolian processes dominate and those where fluvial processes dominate may be recorded in – and interpreted from – the accumulated and preserved sedimentary record (Fryberger, 1990a; Bullard and Livingstone 2002; Cain and Mounney 2011).

Many detailed features of aeolian-fluvial system interactions have been documented previously from both modern environments (e.g., Langford, 1989; Teller et al., 1990; Stanistreet & Stollhofen, 2002; Krapf et.al 2003, Svendsen et al., 2003; Cohen, 2010) and from the ancient geologic record (e.g., Langford and Chan, 1988; Ward, 1988; Herries, 1993; Trewin, 1993; Mounney et al, 1998, 1999; Sweet, 1999; Mounney and Thompson, 2002; Veiga et al., 2002; Mounney & Jagger, 2004; Cain & Mounney, 2009; Jordan & Mounney, 2010) (Fig. 2.1).

Types of aeolian-fluvial interaction have been classified by Langford (1989) who provided a detailed description of the relationship between fluvial and aeolian systems from Great Sand dunes Natural Monument, Colorado and Mojave river wash, California. In that study, six types of fluvial-aeolian interaction were identified: (1) streams dammed by aeolian topography; (2) water-flooded interdunes alongside channels and behind aeolian dams; (3) water-eroded dunes adjacent to channel margins and flooded interdune; (4) sediment of aqueous origin deposited in interdune areas; (5) fluvially derived groundwater that rises to flood interdunes; and (6) wind-eroded fluvial sediment that is re-deposited via aeolian processes. A recent study by Al-

Examples of Fluvial-Aeolian Interactions in Dune-field Margins

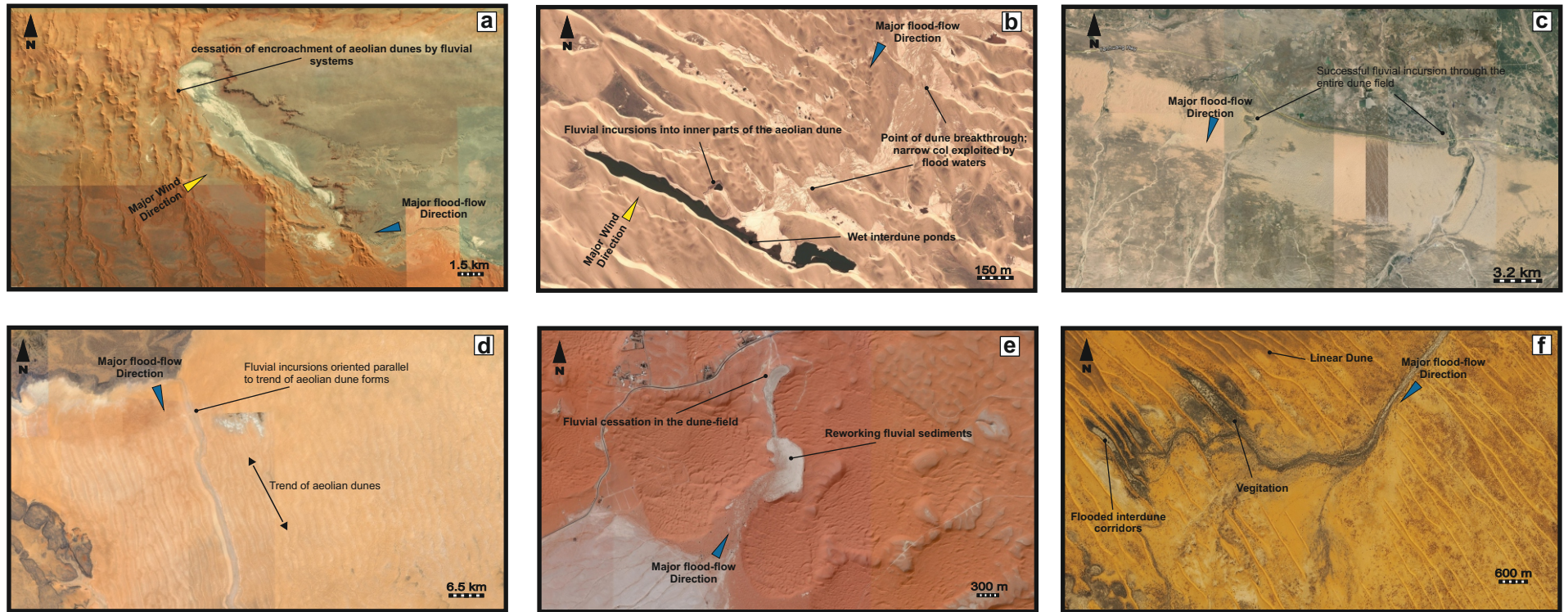


Figure 2.1: Examples of different styles of fluvial-aeolian interactions in erg margins: **(a)** cessation of encroachment of aeolian dune fields by fluvial systems. Tsondab Vlei River, Namib Desert (23 56 19.96 S 15 23 31.10 E). **(b)** Fluvial incursions oriented perpendicular to the trend of aeolian dune. Note ponded interdune as a result of cessation of the fluvial river. Hunkab River, Skeleton Coast Erg (19 49 2.58 S 13 3 25.69 E). **(c)** Successful breakthrough-rivers Mu Us Desert, China (40 15 15 N 09 46 35 E). **(d)** Fluvial incursions oriented parallel to trend of aeolian dune forms (Al-Masrahy & Mountney, 2015). Southern El Djouf Desert, Mauritania (18 04 17 N 11 11 09 W). **(e)** Termination of fluvial channel networks in aeolian dune fields (Al-Masrahy & Mountney, 2015). An Nafud Desert, Saudi Arabia (24 22 58 N 46 14 14 E). **(f)** termination of fluvial channel networks in aeolian dune fields (Al-Masrahy & Mountney, 2015). Simpson Desert, Australia (24 10 29 S 135 15 53 E).

Masrahy & Mountney (2015) introduced ten recognised types of fluvial–aeolian interaction using satellite images from 60 worldwide deserts: (1) fluvial incursions aligned parallel to trends of linear chains of aeolian dune forms; (2) fluvial incursions oriented perpendicular to the trend of aeolian dunes; (3) bifurcation of fluvial flow between isolated aeolian dune forms; (4) through-going fluvial channel networks that cross entire aeolian dune fields; (5) flooding of dune fields due to regionally elevated water-table levels associated with fluvial floods; (6) fluvial incursions emanating from a single point source into dune fields; (7) incursions emanating from multiple sheet sources; (8) cessation of the encroachment of entire aeolian dune fields by fluvial systems; (9) termination of fluvial channel networks in aeolian dune fields; (10) long-lived versus short-lived modes of fluvial incursion.

Within this thesis, a variety of styles of modern fluvial-aeolian interaction are documented from the Skeleton Coast Erg, Namibia, and from an ancient succession, the Helsby Sandstone Formation, UK. However, prior to considering these case-study examples in detail, a review of pertinent literature is required to appreciate and consider the controlling mechanisms that influence aeolian and fluvial depositional regimes in dune-field margin settings. A brief discussion of the sedimentary response of both aeolian-dominated and fluvial-dominated erg margins during and in the aftermath of major fluvial floods is provided here. Specific focus is given to factors which govern both modern system development and accumulation and preservation into the preserved rock record.

2.2 Aeolian dominated erg-margin settings

2.2.1 Dry versus wet aeolian systems

Dry aeolian systems (dry ergs) occur where the water table or its capillary fringe lies at a level beneath the accumulation surface (Kocurek and Havholm, 1993). As such, the water table does not influence sedimentation, and most of the loose sand particles above the accumulation surface are available for transportation. In dry ergs, aeolian process of deposition, bypass and erosion occur as a result of local aerodynamic conditions. Aeolian bedforms accumulate when sand deposited on the dune lee slope is not entirely eroded by the passage of the stoss slope and by the intervening interdune areas during migration (Kocurek and Havholm, 1993). Dry interdunes within such systems typically contain only thin (< 1 m) accumulations of sediments and are mostly preserved in the rock record as thin accumulations (< 1 m-thick sets) of wind-ripple strata between thicker

cross-bedded sets of aeolian dune origin. Aeolian dunes in dry ergs tend to construct (grow) to a size where they fill and eliminate most of the interdune flat before they commence climbing one another as they migrate (Hummel and Kocurek, 1984). Therefore, dry aeolian successions are dominated by cross-bedded dune strata and are associated with only thin accumulations of dry-interdune strata (Mountney 2006a) (Fig. 2.2).

By contrast, wet aeolian systems (wet ergs) are those where the water table or its capillary fringe are at or close to the accumulation surface. Within these systems, deposition, bypass and erosion processes along the accumulation surface are controlled by both aerodynamic configuration and also the moisture content of the substrate (Kocurek and Havholm, 1993). A net overall rise in the relative water-table level is an important mechanism that enables both dune and interdune accumulation in wet aeolian systems (Hummel and Kocurek, 1984; Pulvertaft, 1985). As a consequence, accumulations of wet aeolian systems consist of cross-bedded dune strata separated by damp or wet interdune units that themselves may be relatively thick (Fig. 2.3). Bypass or erosion processes within wet erg systems occur during a static or falling water table, respectively (Kocurek and Havholm, 1993). Sediments indicative of accumulation on a damp or wet substrate within interdune areas are more resistant to erosion during wind storm events because their moisture content means that the accumulated sediment possesses cohesive properties, such as moisture-induced grain-to-grain bonding (Kocurek and Havholm, 1993; Mountney, 2006a). Sedimentary features associated with interdune units allows recognition of a wet aeolian succession. Common sedimentary structures include wave-ripple forms and strata, wavy laminae, contorted bedding, adhesion structures, bioturbation, desiccation cracks and current-ripple laminae (Kocurek, 1981a, Ahlbrandt and Fryberger, 1981 and Hunter 1981; Collinson et al, 2006; Mountney, 2006a, 2012).

Accumulation of wet aeolian systems is most commonly determined by the ratio between the rate of aeolian bedform migration and the coeval rate of relative water-table rise. This gives rise to “climbing” dune and interdune strata (Kocurek, 1981b). Although the angle of climb of bedforms may be classified as subcritical, critical or supercritical (see Collinson et al., 2006), large-scale bedforms in erg systems almost always climb at very low, subcritical angles that are typically just a few tenths of a degree (Hunter 1977; Mountney and Thompson, 2002; Mountney, 2012) (Fig. 2.2).

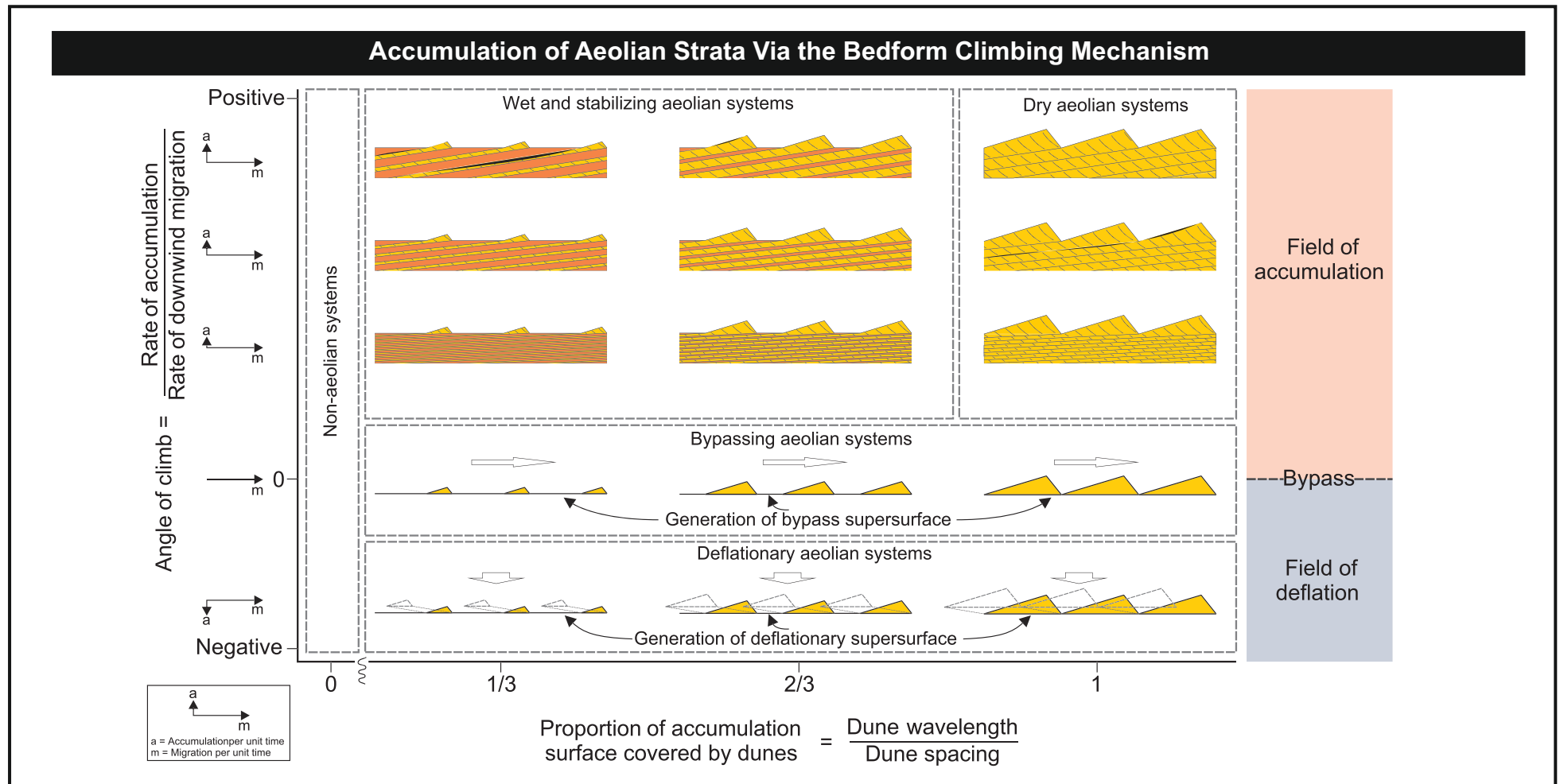


Figure 2.2: Spectrum of preserved dune and interdune architectures resulting from temporally and spatially invariable (i.e. static) aeolian system behaviour. The angle of bedform climb defines fields of accumulation and deflation, with bypass occurring when the angle of climb is zero. Accumulating dry aeolian systems require 100% dune cover. Accumulating wet or stabilizing systems have less than 100% dune cover. The angle of climb is determined by the ratio of the vertical accumulation rate and bedform migration rate. After Mounthey (2012).

Accumulation of Aeolian Systems

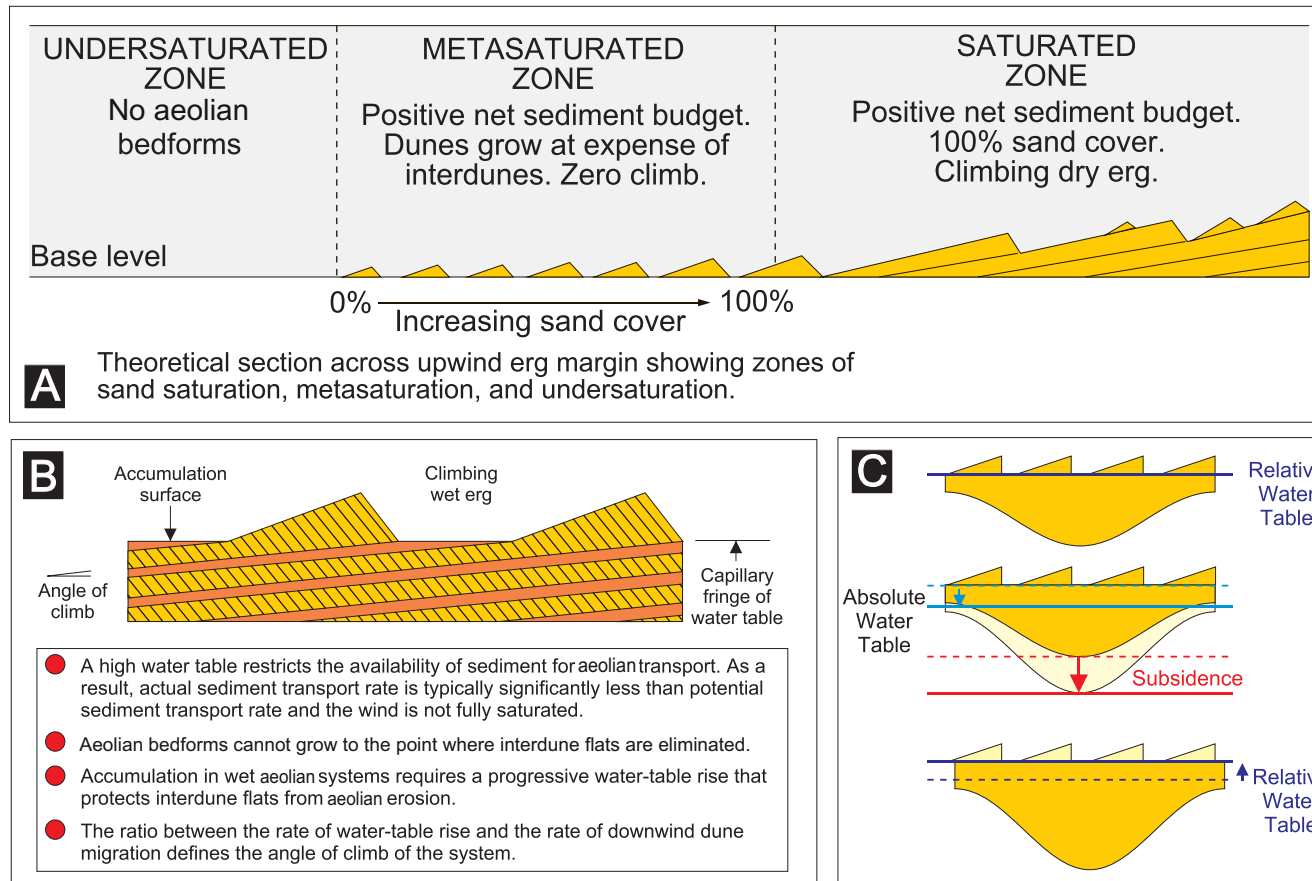


Figure 2.3: Models for accumulation in aeolian systems. A) In dry systems dune expansion to the point where interdunes flats are eliminated occurs prior to the onset of accumulation. Based on the discussion of Wilson (1971). B) In wet systems, both dunes and interdune flats accumulate in response to water-table rise. After Mounthey and Jagger (2004). C) Relative water table is determined by changes in absolute water table and subsidence. Modified after Mounthey (2006a).

2.2.2 Aeolian sediment state in erg margins

The construction of aeolian ergs is dependent on three main factors: sediment supply, sediment availability, and the transport capacity of the wind (Kocurek and Lancaster, 1999) (Fig. 2.4). The influence of climatic change over various time scales significantly influences sediment supply, sediment availability, and aeolian transport capacity of the wind, and these factors collectively govern the aeolian sediment state of a system (e.g., McKee et al., 1967; Herries, 1993; Howell & Mountney, 1997; Kocurek, 1999; Kocurek and Lancaster, 1999).

Erg-margin settings that are subject to fluvial flood activities are controlled by episodic water-table rise and fall that occurs coevally with on-going dune migration. Water-table level influences the supply and availability of sediment for aeolian transport (Mountney & Jagger, 2004). In dry aeolian systems, accumulation usually occurs when the sediment supply is available for wind transportation and this encourages the construction of aeolian bedforms to a size where intervening interdune flats are reduced to isolated depressions between dunes and downwind climbing then commences whereby aeolian dunes climb over one another to generate an accumulation (Kocurek, 1999).

By contrast, the availability of sediment for aeolian transport in wet aeolian systems is relatively limited and aeolian dunes in such systems tend not to be constructed to a size whereby adjacent interdune flats are minimised (Kocurek and Lancaster, 1999). Instead, erg accumulation occurs when on-going dune migration occurs synchronously with a component of rise in the level of the relative water table. As the accumulation passes beneath a relatively rising water table, so it is protected from potential aeolian deflation; accumulated sediments that lie beneath the water table in wet aeolian successions are no longer available for aeolian transport (Kocurek and Lancaster, 1999; Mountney, 2006a).

In general, the fluvial-aeolian interactions in erg-margin settings promote the distribution of fluvially derived sediment within the marginal parts of aeolian dune fields. During a major flood event that becomes distributed over a broad area of an aeolian dune-field margin, for example, the energy of the associated non-confined flows will be reduced. In such cases, erosion of aeolian bedforms by fluvial floods will be limited, though localised erosion of aeolian dunes may act generate an additional sediment supply to the fluvial system. Elsewhere within aeolian dune-field margin settings, fluvial flood waters may be confined within narrow interdune corridors between aeolian

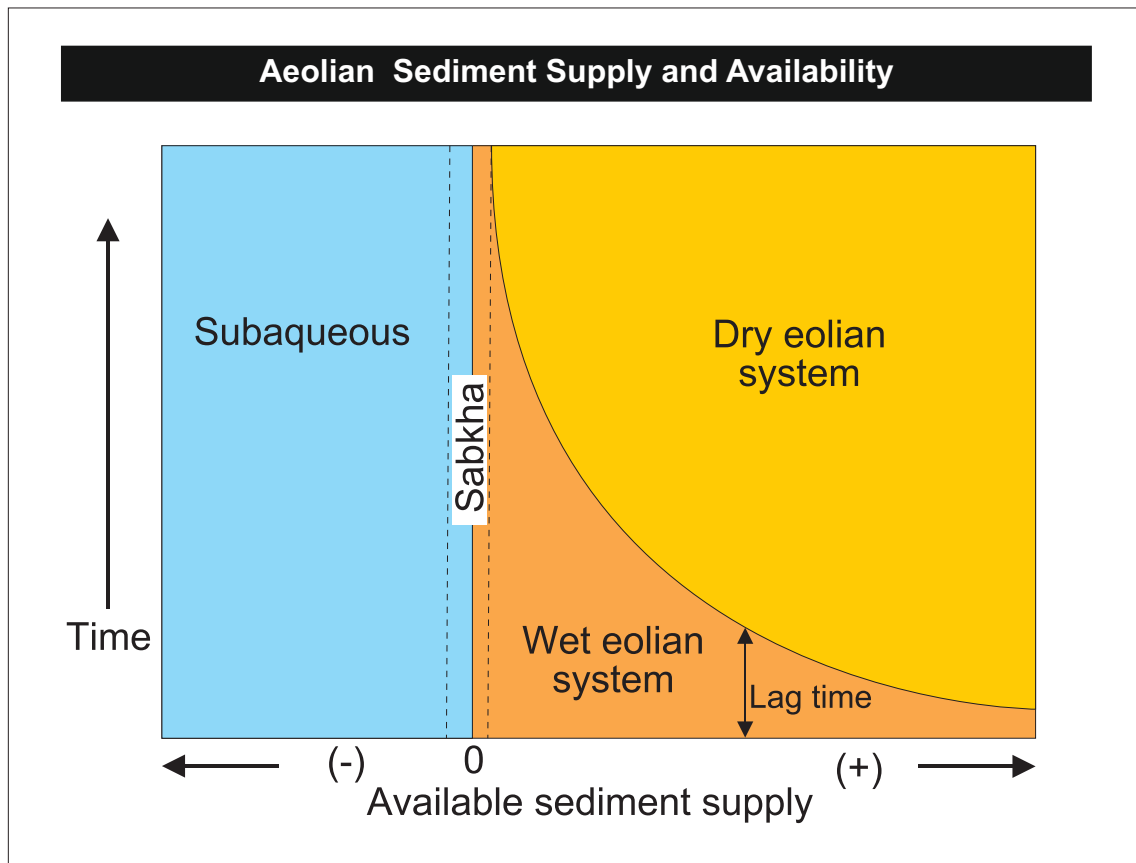


Figure 2.4: Fields for subaqueous environments, sabkhas, wet aeolian, and dry aeolian systems as a function of the available sediment supply over time. At any given time the available sediment supply is the percentage of the substrate covered by dry, loose sediment. The initial condition is one of a flat surface at the level of the capillary fringe of the water table. After Kocurek and Havholm (1993).

dunes; such flows may carve erosional channels into the floors of interdune corridors and may generate a supply of sediment for later aeolian dune construction (Al-Masrahy & Mountney, 2015).

2.2.3 Preservation of aeolian systems

Long-term preservation of aeolian strata occurs in cases where aeolian units are placed below a regional baseline of erosion, especially in water-saturated zones (Kocurek and Havholm, 1993). In general, subsidence, water-table rise, sea-level rise, surface stabilization, and exceptional circumstances are most prominent factors to promote the preservation within aeolian systems (Mountney, 2006a) (Fig.2.5). In dry erg systems, the preservation of aeolian dune and interdune successions can occur by subsidence such that the accumulation is progressively buried as additional accommodation is made available. Alternatively, an absolute rise in the water-table level can also enhance long-term preservation potential in dry aeolian systems where the level of the accumulation surface remains above the level of the water table such that the accumulation surface remains uninfluenced by the water table but the previously accumulated strata passes beneath this water table (Fig. 2.5.a, b).

In wet aeolian systems, the level of the water table plays a major role in defining accumulation space, such that preservation space and accumulation space are coincident (Kocurek and Havholm, 1993). A component of water-table rise can be either relative, where an accumulation of aeolian strata progressively subsides through a static water table, or absolute, where the water table rises in an absolute sense, for example as a result of shift to more humid climatic conditions (Fig. 2.5.c, d). Alternatively, aeolian strata can be inundated by sea-level rise that promotes preservation; aeolian successions may be transgressed and subsequently overlain by marine strata (Glennie and Buller, 1983; Jordan & Mountney, 2010, 2012; Wakefield & Mountney, 2013). The long-term preservation potential of aeolian strata may potentially be increased if the accumulation surface becomes stabilized by the development of armoured lags (e.g., serir) or by colonisation by vegetation, or by the precipitation of a surface crust of gypsum or calcrete (Bristow and Mountney, 2013). However, such stabilizing factors might not be permanent and once they are removed an accumulation may be subject to deflation (e.g., Loope, 1985).

Aeolian Accumulation and Preservation Space

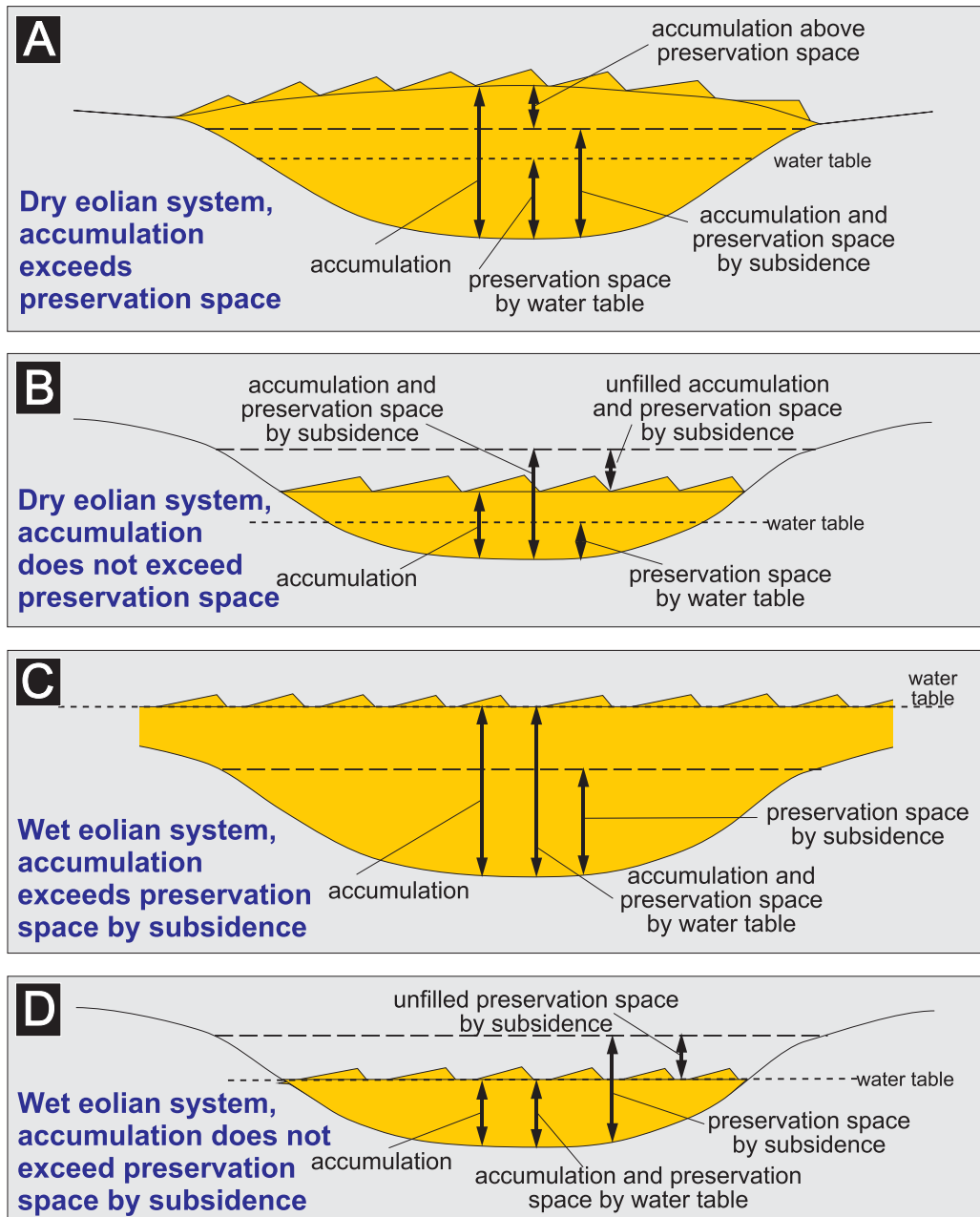


Figure 2.5: Components of accumulation and preservation space for dry and wet aeolian systems. A) Dry aeolian system: the accumulation has built above the preservation space because a positive net sediment budget exists. Long-term preservation potential of that part of the accumulation above the preservation space line is low. B) Dry aeolian system: the accumulation has not filled the available preservation space. C) Wet aeolian system: water-table rise has enabled the accumulation to build above that preservation space generated by subsidence. Deflation will occur if water table is fallen. D) Wet aeolian system: the water table is below the preservation space line and the preservation space remains unfilled. Modified after Kocurek and Havholm (1993), Mountney (2006a).

2.3 Fluvial dominated erg-margin settings

2.3.1 Flood characteristics in dryland rivers in erg margins

In dryland (desert) settings, four basic types of fluvial flood have been identified (Graf, 1988): flash flood events, single-peak flood events, multiple-peak flood events, and seasonal flood events. In general, flash flood events are mainly localised stream flows which are variable in their flood duration, some lasting for only a few minutes to a few hours as consequence of localised storms within only small drainage areas (Graf, 1988; Tooth, 2000b; North, 2003a). Flash floods are normally accompanied with high runoff coefficients and a dominance of Hortonian overland flow (Graf, 1988; Reid & Frostick, 1997; Tooth, 2000b). The rapid rise in the water level within channelised reaches generates hydrographs with near-vertical rising limbs (Graf, 1988). Single-peak flood events occur in basin-scale catchment areas as a result of tropical storms or frontal precipitation that may last from hours to days (Graf 1988). Multiple-peak flood events occur when multiple precipitation events occur or where several catchments feed a larger downstream catchment in the aftermath of a single storm but in a time-lagged manner. Multiple-peak flood events may be brief (< 1 day) or last for several days or weeks in dryland areas (Graf, 1988). Seasonal flood events are similar to the multiple-peak floods but they are highly influenced by monsoonal climatic effects. They occur on a relatively regular time interval with an annual repeat frequency (Graf, 1988). Flood magnitudes and frequencies in drylands are highly variable; high-magnitude but low-frequency events are more common than low-magnitude, high-frequency flood events (Graf, 1988; Tooth, 2000b). Dryland river systems are typically subject to significant downstream decreases in discharge and sediment transport due to transmission losses as a consequence of both infiltration and evaporation. Thus, such streams typically exhibit a downstream change in size, morphology and drainage pattern of their channels (Cain & Mountney, 2011). With increasing distance from channel source, the downstream changes along dryland rivers are characterised by variation of surface gradient, channel confinement, transmission losses, evapotranspiration and infiltration (Tooth, 2000a, b; Cain & Mountney, 2011). Fluvial streams commonly exhibit a transition from braided to single thread channels as river energy wanes (e.g. ephemeral rivers of Schumm, 1961).

2.3.2 Morphological features of dryland rivers in erg margins

Dryland fluvial systems exhibit a variety of channel morphologies that are generally wide and shallow as these types of system subject to episodic and potentially rapid widening and deepening during flood events. Braided plan-form patterns are the most common type in dryland fluvial systems, but single-thread, wide and shallow channels are also common in drylands (Reid & Frostick, 1997; Tooth, 2000b). Wide but shallow single-thread channels allow flood waters to spread out and remain shallow, thereby maintaining transport efficiency (Reid & Frostick, 1997).

In erg margins, channel morphologies can be identified in modern environments (Fig. 2.3). the distribution of breakthrough channels that pass into aeolian dune fields is determined by morphological dune type, the orientation of dunes relative to the direction of flood-waters, the shape of the interdune corridors along which flood waters pass in the dune-field margin settings, and the type and rate of aeolian dune and interdune migration relative to the frequency of fluvial flood events (Svendsen et al., 2003; Krapf et al., 2003 ; Al-Masrahy & Mountney, 2015). In contrast to the breakthrough channels, dune-dammed channels cannot penetrate far into dune fields due to the presence of large aeolian dune barriers or due to the “closed” (i.e. dune encapsulated) form of associated interdune depressions (Krapf et al., 2003; Al-Masrahy and Mountney, 2015) (Fig. 2.3.b).

2.3.3 Dryland river incursions into erg margins

In erg margins, flood are variable from completely channelised, to partly channelised, to largely non-channelised, to non-confined sheets (Olsen, 1987). Further, the extent of the penetration of these fluvial systems into dune fields is dependent on several factors: (1) morphological type and spacing of aeolian bedforms, which themselves control interdune width and shape; (2) the orientation of open interdune corridors relative to the angle of incidence of fluvial floods, (3) the rate of lateral migration of the dunes and their adjacent interdunes (Stanistreet & Stollhofen, 2002; Krapf et al., 2003; Al-Masrahy & Mountney, 2015). Langford (1989) described many commonly recognisable features of aeolian bedforms which control river flood pathways. Flood waters may cause extensive erosion of dunes and interdunes where channels cut through aeolian dune fields, and where fluvially associated groundwater floods interdune areas.

The distance that major breakthrough rivers penetrate into the aeolian erg margins is notable (many tens of km in several cases), especially in aeolian

dune fields where regularly spaced dune-trains of relatively straight-crested aeolian dunes are separated by wide interdune flats along which floods may pass (Al-Masrahy & Mountney, 2015). These open interdune flats can be flooded by high-magnitude events (Al-Masrahy & Mountney, 2015). As a result, extensive surfaces (i.e. flood surfaces of Langford & Chan, 1988) in aeolian deposits can develop. Such surfaces exert a major control on the mechanism of preservation of aeolian strata in the rock record (Langford & Chan, 1988, 1989; Cain & Mountney, 2011). The Uniab River of the Skeleton Coast Erg, Namibia, is characterised by open interdune corridors that occur adjacent to the river path that traverses through the aeolian dune field (Alkathery & Mountney, 2015; Chapter 3 of this thesis) (Fig. 2.3.b).

2.4 Implications for aeolian reservoir geology

From an applied perspective, understanding the spatial distribution of both low-porosity and low-permeability damp and wet interdune architectural elements within either aeolian-dominated or fluvial-dominated erg-margin successions is economically important because such preserved successions commonly act as stratigraphic heterogeneities that limit flow in both hydrocarbon reservoirs and in water aquifers, thereby compartmentalising subsurface reservoirs and aquifer bodies (e.g., Fryberger et al., 1990; North & Prosser, 1993; Mountney, 2006a). For successful aeolian reservoir characterisation of marginal aeolian dune-field settings, it is crucial to recognize and predict the arrangement of facies geometries by determining whether they were produced from events of regional or local significance and lateral extent (Herries, 1993).

In fluvial-dominated erg margins, fluvial sandstones can be found within aeolian deposits associated with flooded interdunes in response to climatic change that would affect the continuity of aeolian sandstones within the aeolian reservoir. The deposits of flooded interdune corridors, which are typically less laterally extensive, act as baffles to flow, particularly where they extend parallel to the strike of the aeolian bedforms. The downwind migration of flooded interdunes in front of advancing straight-crested dunes, will produce continuous sheet-like wet interdune deposits over local to regional areas (Langford & Chan, 1988; Herries, 1993; Mountney, 2012).

In aeolian-dominated erg margin settings, by contrast, short-lived flooded interdunes in between sinuous-crested dunes generate isolated lens-like interdune units of only localised extent. Further, the variety of aeolian

sedimentary structures preserved within dune toset elements tends to favour the development of relatively high-permeability zones, many dominated by aeolian grainflow strata in dune lee-slope deposits (Howell & Mountney, 2001). Tongue-shaped packages of inversely graded grainflow strata that form cross bedding within large dune sets typically record higher permeability than the surrounding packages of wind-ripple and gainfall strata (Howell & Mountney, 2001; Romain & Mountney, 2014). Grainflow-dominated dune architectural elements tend to act as preferential pathways for fluid flow. Although the presence of such elements may improve net reservoir quality, they may also bring associated problems in aeolian reservoir development and production, including early water breakthrough, for example.

The depositional model of Mountney and Thompson (2002) presents a useful example to summarise lateral relationships between adjacent fluvial, aeolian dune and interdune environments within an ancient erg-margin environment using outcrop data from the Helsby Sandstone Formation of the Cheshire Basin, UK (Fig. 2.6). It provides a test model for assessing reservoir heterogeneity and predicting the distribution of low-permeability interdune units that may act as baffles to fluid flow in the subsurface (Mountney & Thompson, 2002).

2.5 Conclusions

Understanding fluvial-aeolian interactions in modern aeolian dune-field margins is important in the development of predictive facies models for the recognition of the preserved expression of such interactions in ancient successions and hydrocarbon reservoirs. This chapter has provided a brief summary of the literature pertinent to the concepts and models developed later in this thesis that are applied to both aeolian-dominated and fluvial-dominated erg margins systems, and the associated processes that shape them. In both modern and ancient desert environments, the diversity of styles of fluvial-aeolian interactions results in complex geomorphological, sedimentological and preserved stratigraphic relationships.

Example of Depositional Model of Dune and Interdune Elements

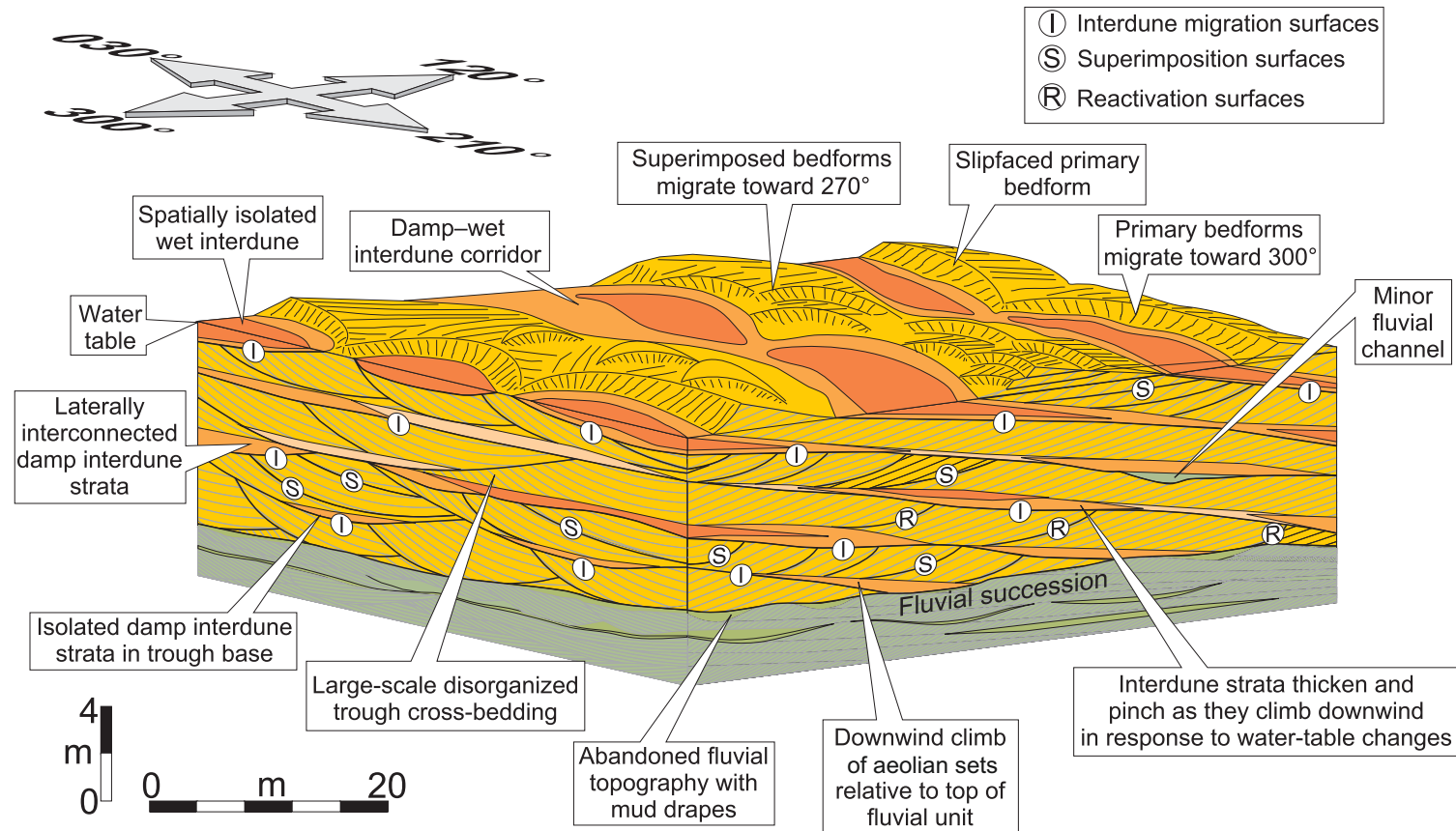


Figure 2.6: Example of depositional model from the Triassic Helsby Sandstone Formation, Cheshire Basin, UK. depicting downwind climbing of both dune and interdune elements. Interdunes exhibit downwind facies variability that reflects subtle changes in the level of the water table relative to the accumulation surface during accumulation. Lateral connectivity of the interdune elements is controlled by dune morphology, and both isolated interdune hollows (ponds) and interconnected, throughgoing corridors are recognized. Modified after Mounthey and Thompson (2002).

Chapter 3

Styles of interaction between fluvial and aeolian systems: Skeleton Coast, northern Namib Desert

3.1 Summary

The Skeleton Coast dune field (erg) of the Atlantic coastline of NW Namibia forms a ~300 km-long and 6 to 22 km-wide, north-northwest trending zone of active aeolian construction and accumulation that covers a 2000 km² region within which bedforms of various morphological types are present and attain heights up to 50 m. Fluvial systems of the Skeleton Coast are ephemeral and undertake marked changes in discharge in response to seasonal monsoonal rainfall events in their continental-interior catchments; they are characterized by networks of shallow braided channels with longitudinal sand bars.

Although the network of dunes forms a major obstacle to fluvial flow, several ephemeral rivers drain south-southwestwards through the erg towards the Atlantic Ocean. Thus, a range of styles of fluvial-aeolian interaction arise and characterize the erg margin region. The presently active system interactions include: (i) the establishment of long-lived major through-going open interdune corridors along which long-established rivers pass; (ii) the damming of river courses by active aeolian dunes that form barriers to flow, thereby resulting in the ponding of flood waters and the development of large, slowly draining flood basins; (iii) the passage of flood waters as sheet flows into the outer margins of aeolian dune fields.

The fluvial incursions into the desert basin are controlled by precipitation-event frequency and magnitude, fluvial runoff distance from catchment to receiving basin, sediment yield, changes to regional water-table level in response to flood events, topography of the accumulation surface, and long-term climate change. These factors conspire to determine fluvial discharge to the receiving basin in the form of confined floods within channelized networks or as widespread unconfined sheet flows in dune-field margins.

Geomorphological relationships have been examined through analysis of high-resolution satellite imagery data from Google Earth Pro software. Quantitative data relating to the geometry and morphology of dunes and interdunes have been recorded to demonstrate systematic changes with

increasing proximity to the major river courses. Results from analysis of remotely sensed data from the Skeleton Coast erg, Namibia, serve as the basis for the construction of a series of quantitative facies models that can be used to assess likely subsurface prospectivity in mixed fluvial-aeolian reservoir successions in both aeolian-dominated inner erg-margin settings and fluvial-dominated outer erg-margin settings.

3.2 Introduction

A 2000 km² region of the Skeleton Coast of north-western Namibia is characterized by an aeolian dune field (erg) that acts as a major obstacle to several ephemeral or intermittent fluvial systems, which flow from east to west towards the coastline of the Atlantic Ocean (Fig. 3.1). Where the westward flowing fluvial systems meet the eastern margin of the erg they exhibit a range of types of interaction between competing fluvial and aeolian processes. The Skeleton Coast erg is composed principally of large isolated aeolian barchan dune forms and chains of transverse, compound crescentic and barchanoid forms, most of which are actively migrating northwards (Lancaster, 1982, 1989). Various types of interaction between the fluvial systems and aeolian dunes are evident, both at the eastern and western erg margins and in the erg centre.

Some of the fluvial systems are intermittently dammed by aeolian dunes and the fluvial courses commonly undertake changes of course as aeolian dunes migrate across their paths. Hitherto, relatively few published studies have documented styles of aeolian-fluvial interaction within near-coast aeolian dune fields, such as those of the Skeleton Coast, using a combination of field-acquired data and high-resolution, remotely sensed satellite and aerial photograph imagery data (Krapf et al 2003, Stanistreet & Stollhofen, 2002; Svendsen et al., 2003; Teller & Lancaster, 1990, Lancaster, 1982; Langford, 1989).

The overall aim of this study is to document styles of interaction between coevally active aeolian and fluvial sedimentary systems within a series of modern aeolian dune fields and to contrast the processes known to occur in such settings with preserved sedimentary deposits of analogous systems present in the ancient rock record. The study of the Skeleton Coast erg discussed herein represents one part of the larger research programme summarized in this thesis that considers the sedimentary impact of aeolian-fluvial interactions. This aim will be achieved through analysis of remotely sensed satellite imagery, which is used to document types of system

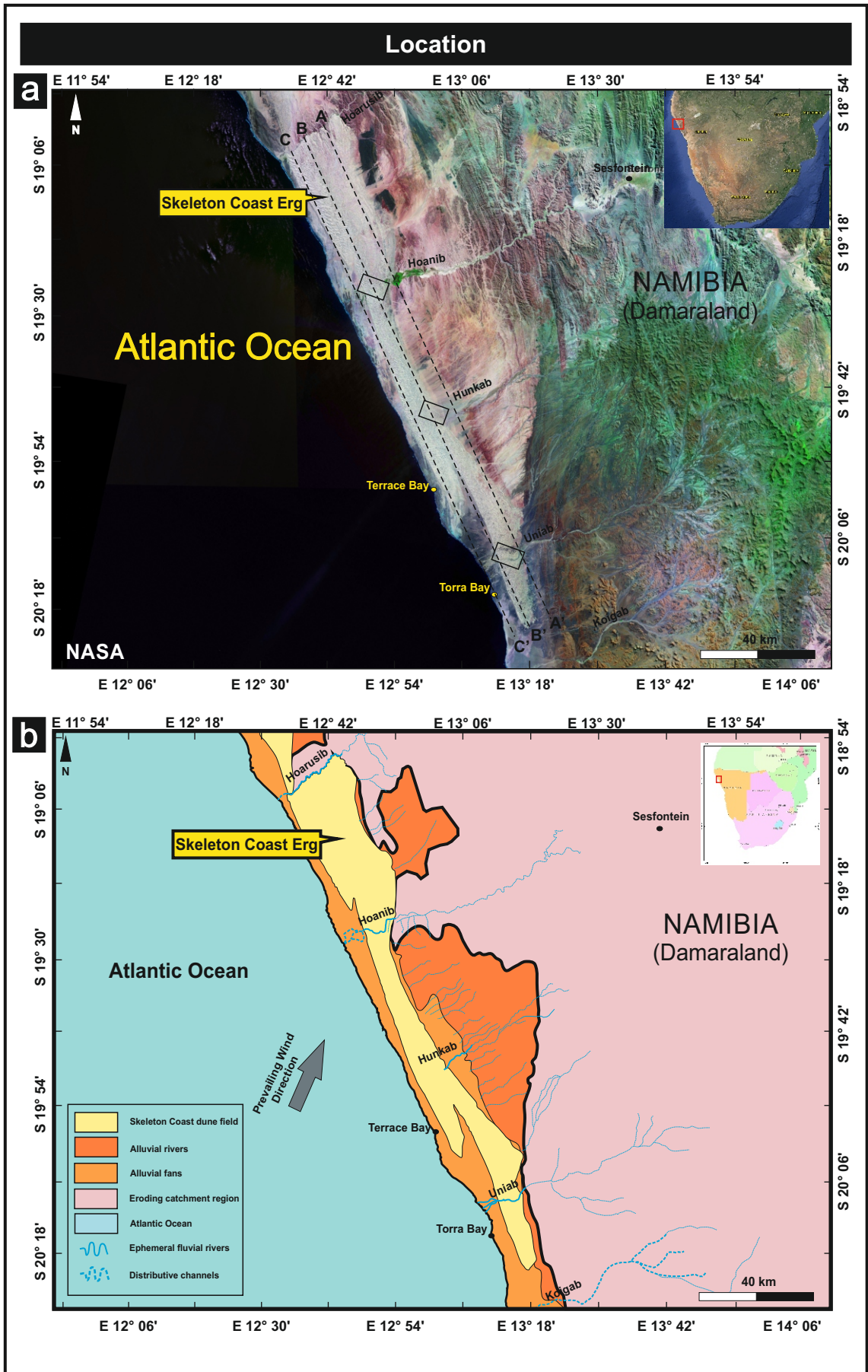


Figure 3.1: Location of the Skeleton Coast erg. (a) Satellite image displaying different intersecting ephemeral rivers that pass into the dune field towards the Atlantic Ocean. (b) Distribution of primary geomorphic landforms.

interaction in terms of formative processes, resultant landform generation, and mechanisms for accumulation of sedimentary successions that might ultimately become preserved in the stratigraphic record.

Specific objectives of this study are as follows: (1) to interpret the different types of the interaction between aeolian and fluvial systems through analysis of satellite imagery data; (2) to explain how morphological elements of aeolian and fluvial systems act to determine sediment distribution and sediment supply; (3) to demonstrate the intimate interdependent relationship between aeolian dunes, interdunes and ephemeral fluvial systems in arid and hyper-arid aeolian erg and erg-margin environments; (4) to produce conceptual facies models to show the expected form of sedimentary units in the preserved stratigraphic record.

This study is important because it provides a series of predictive models based on observed relationships that can be used to forecast likely preserved sedimentological and stratigraphic style in aeolian dune-field margins that are subject to repeated fluvial incursions.

3.3 Background

Fluvial-aeolian interaction is a descriptive term that is applied specifically to describe a range of geomorphic processes and resulting units of sediment accumulation that arise via the interaction of fluvial and aeolian systems, typically in the marginal areas of arid desert dune fields (ergs). Significant sediment can be transported via fluvial processes into aeolian dune fields as either bedload or suspended-load during flood events (fluvial sediment store). Such sediment is carried by flows that pass between active aeolian dunes along low-gradient and low-relief interdune corridors to be deposited within the outer margins, the inner margins or the central parts of aeolian dune fields. Thus, such fluvially derived sediments serve to generate a supply of sediment suitable for later aeolian reworking; as such they represent an aeolian sediment store that, following cessation of a flood event, may potentially be made available for aeolian reworking at a later time (Langford, 1989; Bullard & Livingstone 2002, Kocurek & Lancaster, 1999).

Various examples of fluvial-aeolian interaction have been documented in the published literature in recent decades and many studies have attempted to explain the processes of sediment accumulation associated with these interactions in modern systems (e.g., Langford, 1989; Bullard & Livingstone, 2002; Krapf et.al 2003, Stanistreet & Stollhofen, 2002; Svendsen et al.,

2003; Al Farraj & Harvey, 2004; Mountney & Russell, 2009). Such studies of modern systems are important in the development of techniques and criteria with which to recognize the preserved sedimentary record of such systems in the ancient sedimentary record (Langford & Chan, 1989, North & Prosser, 1993; Mountney & Thompson, 2002).

Although, numerous examples of fluvial-aeolian interaction are known from a wide range of modern geographic settings, relatively few studies have discussed the nature and form of such interaction in a systematic and detailed manner. Those studies that have considered this phenomenon have revealed and documented several types of interaction. Examples include Great Sand Dunes National Monument, Colorado (Langford, 1989), the Mojave River Wash, California (Langford, 1989), the Skeleton Coast Erg, northern Namibia (Krapf et al. 2003, Stanistreet & Stollhofen, 2002; Svendsen et al., 2003), the Kuiseb River, northern Namib Sand Sea (Teller & Lancaster, 1990), and the Cooper Creek Fan, Strzelecki Desert, Australia (Cohen, 2010).

In the ancient preserved sedimentary record, several examples of fluvial-aeolian interaction have been documented through analysis of outcropping successions. Selected examples are as follows: the Tertiary Tsondab Sandstone Formation, central Namib Desert (Ward, 1988); the Cretaceous Etjo Sandstone Formation, northwest Namibia (Mountney et al, 1998); the Triassic Helsby Sandstone Formation, UK (Mountney and Thompson, 2002); the late-Silurian Tumblagooda Sandstone, Western Australia (Trewin, 1993); the Permian Cedar Mesa Sandstone, SE Utah, USA (Langford and Chan, 1988; Mountney & Jagger, 2004); the Pennsylvanian to Permian lower Cutler beds, south-east Utah, USA (Jordan & Mountney, 2010, 2012); the Permian Organ Rock Formation, SE Utah, USA (Cain & Mountney, 2009); and the lower-Cretaceous Avile' Member of the Agrio Formation, Argentina (Veiga et al., 2002).

There remains a clear need to better understand the form of fluvial incursion into active aeolian dune fields and to develop a generalized set of facies models with which to account for the stratigraphic complexity associated with such types of interaction. Such models will have valuable predictive potential (Tooth, 2000; Belnap et al., 2011; Nanson and Tooth, 1999).

3.4 Skeleton Coast Erg and Geological Setting

The name “Skeleton Coast” was first coined by John Henry Marsh, author of an eponymous book that recounts the Dunedin Star shipwreck disaster in Namibia in November, 1942. Since the book was published in 1944, the erg system which is developed along this stretch of coastline has been officially identified on the maps as the “Skeleton Coast erg” (Wilson, 1973; Lancaster, 1982; Ward & Corbett, 1990).

The climate of the Skeleton Coast erg is hyper-arid with less than 50 mm average annual rainfall (Van Zinderen Bakker, 1984; Jacobson et. al, 1995). This coastal erg has developed because the regional climate is strongly influenced by the northward-flowing Benguela Current and its associated cold-water upwelling system offshore Namibia, which has resulted in anomalously low humidity and strong southerly, coast-parallel winds. The fluvial systems that pass into the main part of the erg system are fed by catchments of various sizes and which are characterized by a range of discharges (Van Zinderen Bakker, 1984; Jacobson et. al, 1995). The northern rivers (e.g. Hoanib), with relatively high discharges, are fed by relatively larger northern catchments compared to the southern catchments (e.g. Hunkab and Uniab) (Fig. 3.1).

The Skeleton Coast erg has been imaged by multiple generations of Landsat satellite imagery. It covers an area of ~2,000 km² of active aeolian construction and accumulation. This coastal erg is 160 km long, is located 1.5 to 10 km inland from the coastline, and trends north-northwest (Fig. 3.1). The dune-field is 6 km wide at the southernmost part but widens to 20 km at its northernmost limit. The erg has a southern limit east of Torra Bay, and widens northwards. In the northern part of the erg, the Hoarusib River acts to limit the further northward migration of aeolian dunes. Dunes in this region are up to 100 m high and the erg is 22 km wide at its northern limit (Svendsen et al., 2003).

Accumulated sediments of the Skeleton Coast are predominantly Late Pleistocene in age and overlie mostly sedimentary successions that are Cenozoic in age (Krapf et al., 2003, 2005). These successions themselves sit on the Lower Cretaceous Etendeka Group basement; a succession of thick volcanic rocks, present beneath the southern area of the erg (Krapf, 2003). The topography of the region partly defines the dune-field boundaries (Figs. 3.2): the local topography at the southernmost boundary of the erg is relatively high, where a 40 to 50 m-high basement high composed of

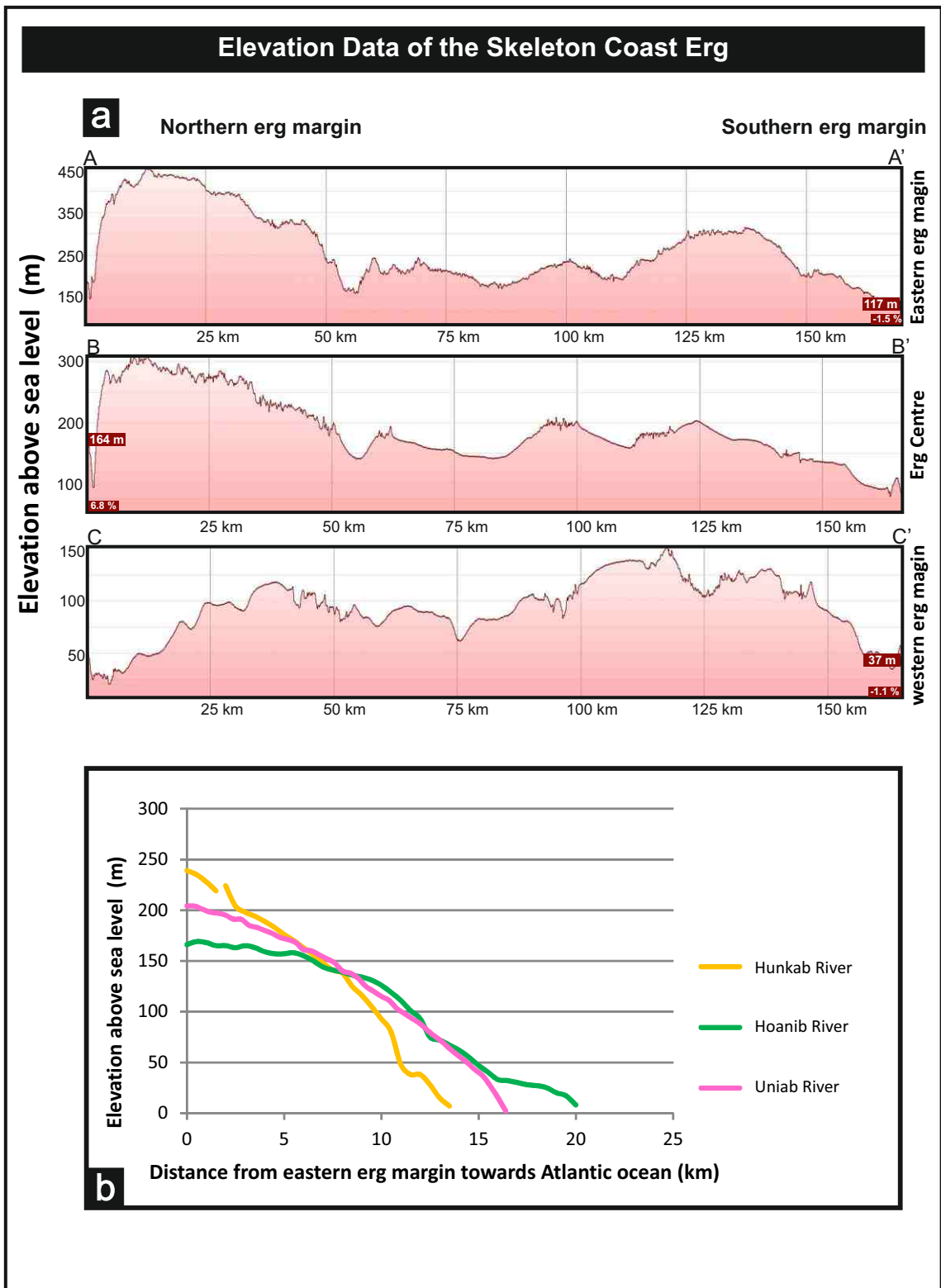


Figure 3.2: (a) Elevation data from the northern to the southern part of the erg along a transect through the eastern erg-margin (A-A'), central part (B-B'), and western erg-margin (C-C') of Skeleton Coast erg. Net shown on Figure 3.1 for the location of the cross-sections. **(b)** Elevation data of selected rivers above sea-level showing the slope from the eastern erg margin to the Atlantic Ocean.

Damara Schist outcrops to define the southern limit of the dune field (Lancaster, 1982; Krapf et. al, 2003; Livingstone & Warren, 1996). By contrast, the northern-most part of the dune field occupies the Hoarusib River valley and intermittent fluvial flow within this system prevents continued northward migration of the dunes.

Figure 3.2b depicts the change in river elevation above sea level based on data collated from the Google Earth Pro along the studied river courses, from the eastern erg-margin towards the Atlantic Ocean. The trends show slightly convex profiles, the form of which reflects the local topography: for each of the studied rivers, the gradient steepens as the river passes from a course flowing over basement rocks in the east, to a course flowing over Quaternary and Plio-Pleistocene sediments in the region of the erg and on the Atlantic coastal plain in the west.

In general, aeolian-dune morphology of the Skeleton Coast erg is controlled by prevailing SSW winds. Aeolian construction and accumulation has increased the level of the accumulation surface to a height 2 to 30 m above the coastal deflation surface in the downwind (i.e. northerly) part of the erg (Lancaster, 1982; Stanistreet & Stollhofen, 2002). Aeolian dunes of the Skeleton Coast erg are composed mainly of large isolated barchan forms and chains of transverse, compound crescentic and barchanoid forms, most of which are actively migrating northwards. Notably, the western margin is composed of a distinctive prominent 'dune wall' that comprises bedforms that are 20 to 80 m high, with large slip faces developed on their eastern flanks.

The dunes are formed by the coalescence of the western arms of crescentic or barchanoid dune ridges (Lancaster, 1982). The wall has acted as a trap for blown sand, mostly of medium and coarse sand grade that is itself was derived from the deflating coastal plain – a primary sediment source for the construction of the dune field. In many eastern parts of the erg, the dunes are predominantly simple barchanoid and transverse ridges that vary from 3 to 10 m in height (Lancaster 1982). Dune height and type gradually changes toward the northeast away from the 'dune wall' of the western erg margin: transverse ridges and barchanoid ridges characterize the central erg, and these gradually morph into individual isolated barchans forms at the eastern erg-margin. Although dunes are up to 80 m high at the western erg margin, they tend to become progressively smaller and towards the eastern erg margin (Lancaster, 1982). The sand supply is low at the eastern erg margin

because most the sand that is available for transport is trapped within the western and central parts of the dune-field that lie upwind (Lancaster, 1982).

Styles of fluvial-aeolian interaction within the Skeleton Coast erg have been the focus of previous field-based study (Lancaster, 1982; Stanistreet & Stollhofen, 2002; Svendsen et al., 2003; Krapf et. al, 2003). The earliest detailed field-based study that mentioned how fluvial-rivers intersect the aeolian dunes of the Skeleton Coast erg was that of Lancaster (1982), who described in detail the variety of dune-types and dune grain-size distributions within the Skeleton Coast region. He concluded that the mean grain size of sampled aeolian dune sands ranges between 1.7 to 2.2 phi across the erg with no consistent fining trend from south to north. Lancaster (1982) documented a trend of improved grain sorting regionally from southwest to northeast, and showed that the prevailing SSW wind direction is consistent with the direction of the sand supply and availability from different sources at the southern and western erg margin where each source provides new coarse and moderately sorted sands for aeolian construction (Lancaster, 1982).

Several major fluvial river courses intersect the eastern margin of the Skeleton Coast erg. Along these courses, major flood events lead to rapid rises to peak discharge that result in channel breaching and widespread flooding into adjacent interdune depressions at the eastern erg margin. Flooding events along these rivers inundate interdune areas adjacent to the main river-courses at points where the rivers intersect the dune field, leaving deposits of gravel, sand and silt, which are draped by clay-prone mud layers up to 15 cm in thickness and which represent waning-stage flood deposits. Pondered flood waters within interdunes gradually evaporate and infiltrate in the aftermath of each flood to leave deposits of thin beds of cohesive mud that tend to resist aeolian reworking and accumulate progressively over multiple floods. Discharge is rarely sufficiently high to result in dune breakthrough; instead, transient lake systems develop in the immediate aftermath of floods (e.g. Hoanib River).

Among the major rivers within the Skeleton Coast, the Hoanib River has previously been the subject of a detailed sedimentological case study of a style of fluvial-aeolian interaction (Stanistreet & Stollhofen, 2002). This study noted a range of associated interaction processes that were identified as being responsible for the deposition of thin mud-layers along adjacent interdunal-corridors in close proximity of the river's course at the eastern erg margin. Mud layers have accumulated in a large dune-bounded flood basin

(the Gui-Uin) that is episodically filled by high-discharge floods that emanate from a relatively large catchment. The schematic facies models that were developed as an outcome of the work by Stanistreet & Stollhofen (2002) summarize the following traits: (1) the ponding of initial flood waters to a point where the water level rises leading to initial dune break-through during recorded major floods at the eastern erg margin; (2) ponding of flood waters as a mechanism to allow for the settling from suspension of fine-grained mud fractions (clay and silt) resulting in sedimentation in flood ponds; (3) development of thin but laterally extensive mud beds that have a predictable geometry and architecture is governed by the form of the flood ponds that developed in the erg-margin setting. These traits have implications for better understanding the distribution of permeability baffles and barriers in subsurface reservoir successions that are exploited for oil and gas (cf. Mountney & Thompson, 2002; Ahlbrandt & Fryberger, 1981, Fryberger, 1990a) or which serve as water aquifers (Fryberger et. al, 1983, Lancaster & Teller, 1988) or repositories for the underground storage and sequestration of carbon dioxide (Mountney, 2012).

Detailed sedimentological observations of channel-shaped bodies of fine-grained strata have been related to fluvial break-through of the Uniab River into the Skeleton Coast erg (Svendsen et al., 2003). Sedimentological interpretations concluded that the channel-shaped depositional bodies present in the river course were formed via deposition from hyperconcentrated flows. Sedimentary structures such as massive and structureless beds, normally graded beds, basal pebble lags are recorded and shown to be interbedded with sand deposits of aeolian origin as a result of catastrophic dune-collapse into the active fluvial system. A generalized depositional model has been proposed to explain the sedimentological processes of dune damming, subsequent break-through and fluvial reworking of aeolian sediments (Svendsen et al., 2003).

In a study of five rivers (Koigab, Uniab, Hunkab, Hoanib and Hoarusib) that interact with the Skeleton Coast erg, Krapf et al. (2003) used small-scale NASA satellite imagery data to demonstrate styles of fluvial-aeolian interaction. This study discussed the main parameters that determine how mixed fluvial and aeolian systems interact, including the size of catchment areas that affect the fluvial discharge, the role of dune geomorphology within the erg, and the type of deposits as accumulate in the aftermath of flooding and dune damming of fluvial courses.

3.5 Methods

Geomorphological relationships have been examined through analysis of high-resolution satellite imagery data from Google Earth Pro software. The resolution of this imagery is sufficiently high to enable collection of detailed quantitative measurements relating to dune, interdune and fluvial forms, and to interpret their morphological significance. The satellite imagery data presented in this study has a stated spatial resolution of 15 m per pixel, though in reality the resolution is greater than this because individual 2-3 m diameter objects such as shrubs can readily be resolved. The collected images are characterized by 15 to 30 m-resolution multispectral Landsat data which have been pan-sharpened with panchromatic Landsat image processing software (cf. Al-Masrahy & Mountney, 2013). Each individual image has a resolution of 4800 x 2717 pixels. Multiple images have been tiled to provide large-format image mosaics suitable for detailed analysis.

Within the Skeleton Coast erg, three major rivers were selected for detailed analysis to investigate styles of fluvial-aeolian interaction. These rivers – from south to north, the Uniab, Hunkab and Hoanib (Fig. 3.1) – were chosen in this study for the following reasons: (1) they each record different styles of spatial change in the geomorphological elements of aeolian dune and interdunes in areas adjacent to the main fluvial courses; (2) they each demonstrate various styles of fluvial-aeolian interaction at the eastern margin of the erg where the fluvial systems pass into the dune field and where the fluvial-rivers impinge the dunes; (3) there are no previously published accounts of these rivers that utilize high-resolution satellite imagery data; and (4) the imagery available record examples of fluvial rivers in different states: both dry (e.g. Hoanib and Hunkab) and actively flowing (e.g. Uniab).

A variety of quantitative data relating to the geometry of 1552 dunes and 775 interdunes have been collected from the chosen-rivers. In total, 4656 geomorphological measurements of dunes (Fig. 3.3) have been recorded and these include the following: (1) mean bedform-spacing, which is defined as the distance from crest-to-crest (or toe-to-toe) between adjacent bedforms in an orientation parallel to the prevailing wind-direction (Breed and Grow, 1979); (2) bedform-wavelength including the minimum, average and maximum, which is the horizontal extent of a bedform in an orientation perpendicular to the trend of the bedform crestline (Rubin 1987); (3) plan-form wavelength and amplitude of along-crest sinuosity of dunes which together serve as a simple measure of bedform crestline sinuosity,

Definition of Parametres Used to Define Morphology

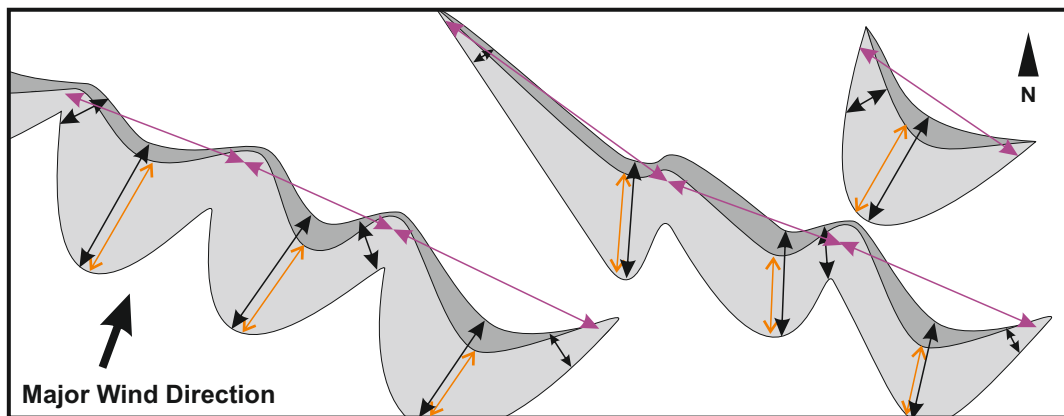
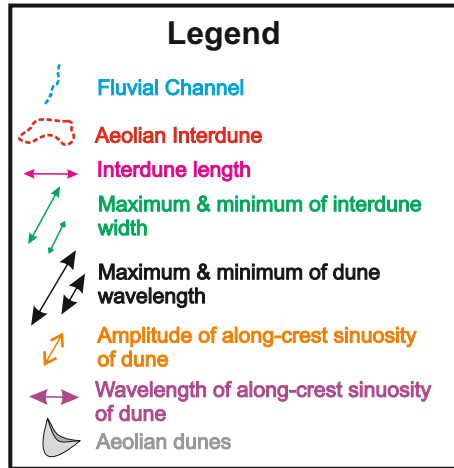
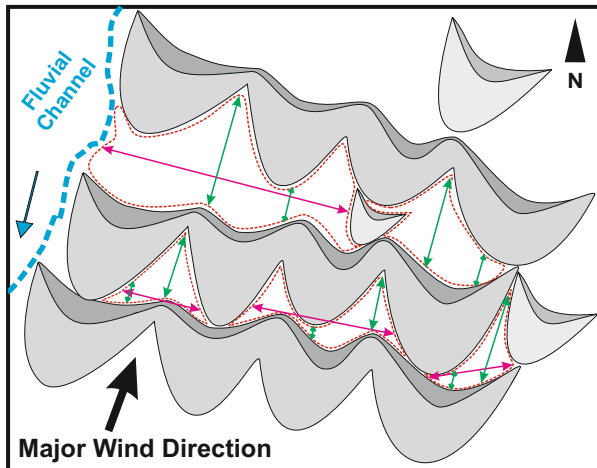
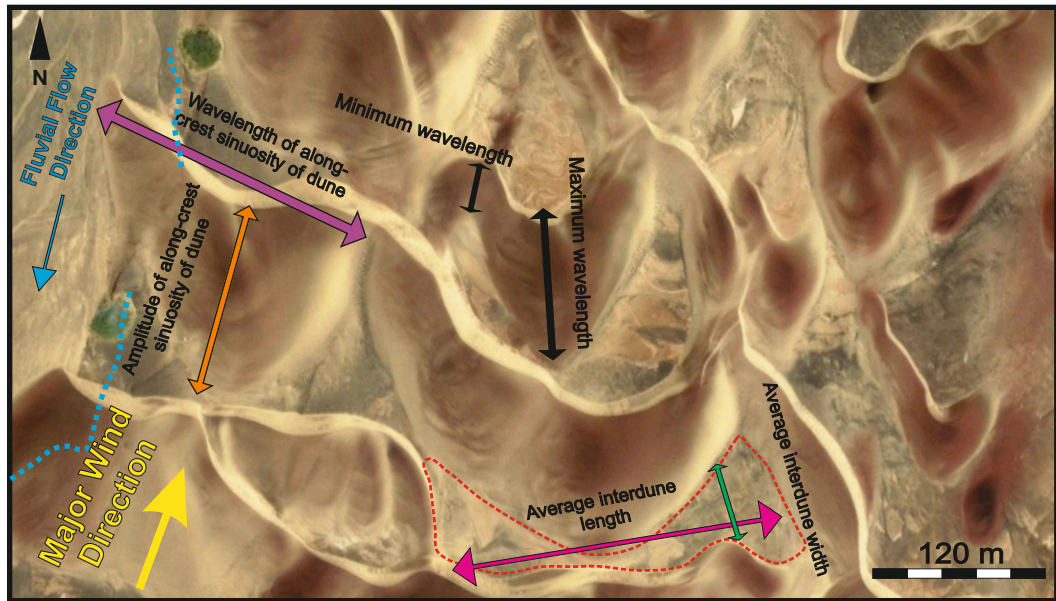


Figure 3.3: Example of dunes and interdunes within the catchment of the Uniab River showing the methodology, definitions and terminology used in this study for measurement of dune and interdune morphometry.

Additionally, 1550 geometrical measurements of interdunes have been recorded (Fig. 3.3), including those occupied episodically by the river courses. Measurements of interdune morphology include the following: (1) interdune length, which is the distance of a single interdune corridor that extends in an orientation parallel to the trend of the crestlines of the dunes that isolate the interdunes; (2) interdune width including the minimum, average and maximum, which is a measured line in an orientation perpendicular to interdune length; (3) interdune orientation, which describes the trend of an elongate interdune axis.

3.6 Styles of fluvial-aeolian interactions in the studied areas

3.6.1 Hoanib River (19 25 0.91 S 12 54 14.20 E)

The Hoanib River is the second largest ephemeral river, that interacts with the Skeleton Coast erg at latitude S 19°24' (Krapf et al., 2003). The 270 km-long river is fed by a 17,200 km² catchment area, centred close to the settlement of Sesfontein. The river, as depicted in satellite imagery presented here (Fig. 3.4), is ephemeral; however, the boundaries of the main course are clearly imaged. Several types of interaction are identified (Fig. 3.5), including dune damming of the main river course, fluvial breakthrough of damming aeolian dunes (Fig. 3.5a), dune migration over the dry river course to partly or completely fill the channel with aeolian sand (Fig. 3.5b and 3.5c), and the development of interdune flood-basin ponds (Fig. 3.5e and 3.5f).

Analysis of precipitation trends by Jacobson et al. (1995) in the Hoanib catchment reveals that the majority of the Hoanib river catchment area affected by annual precipitation between 100 and 300 mm. Major floods along the river that have been recorded historically have crossed the erg successfully to reach the Atlantic Ocean only three times in recent decades: 1995, 1997 and 2000 (Stanistreet and Stollhofen, 2002). During these past catastrophic-flood events, the water level within the flood basin reached the dune low-points (cols) enabling passage of flood waters into the interior of the dune field. The large flood basin associated with the Hoanib River repeatedly fills when high-discharge flood events occur, thereby forming a vegetated lake behind the dune barrier for several days or weeks in the aftermath of each major flood. Small distributary channels are developed within this large vegetated-basin at the eastern erg margin when relatively low-discharge flood events occur (Fig. 3.5a & 3.5d). These channels may break-through the dune barriers to form smaller secondary flood basins

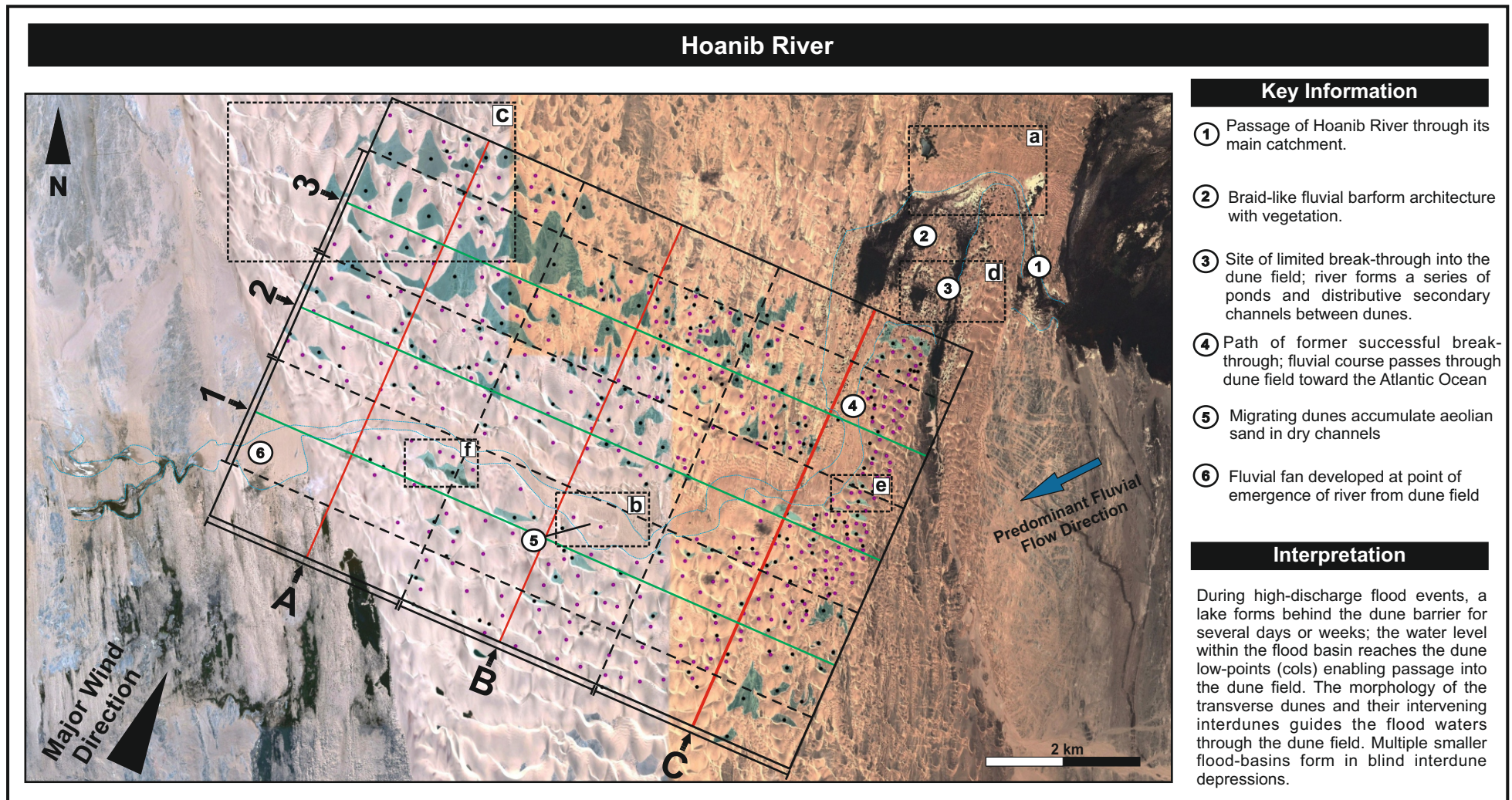


Figure 3.4: Study area of Hoanib River. The black spots are points from where data were collected and the interdunes are highlighted in blue (false colour). Red and green lines are transects representing trends of aeolian dune and interdune geomorphology along the transects that are parallel (A, B and C) and perpendicular (1, 2, and 3) to the prevailing wind direction. See Figure 3.5 for details of a - f.

Examples of Fluvial-Aeolian Interaction in the Hoanib River Catchment

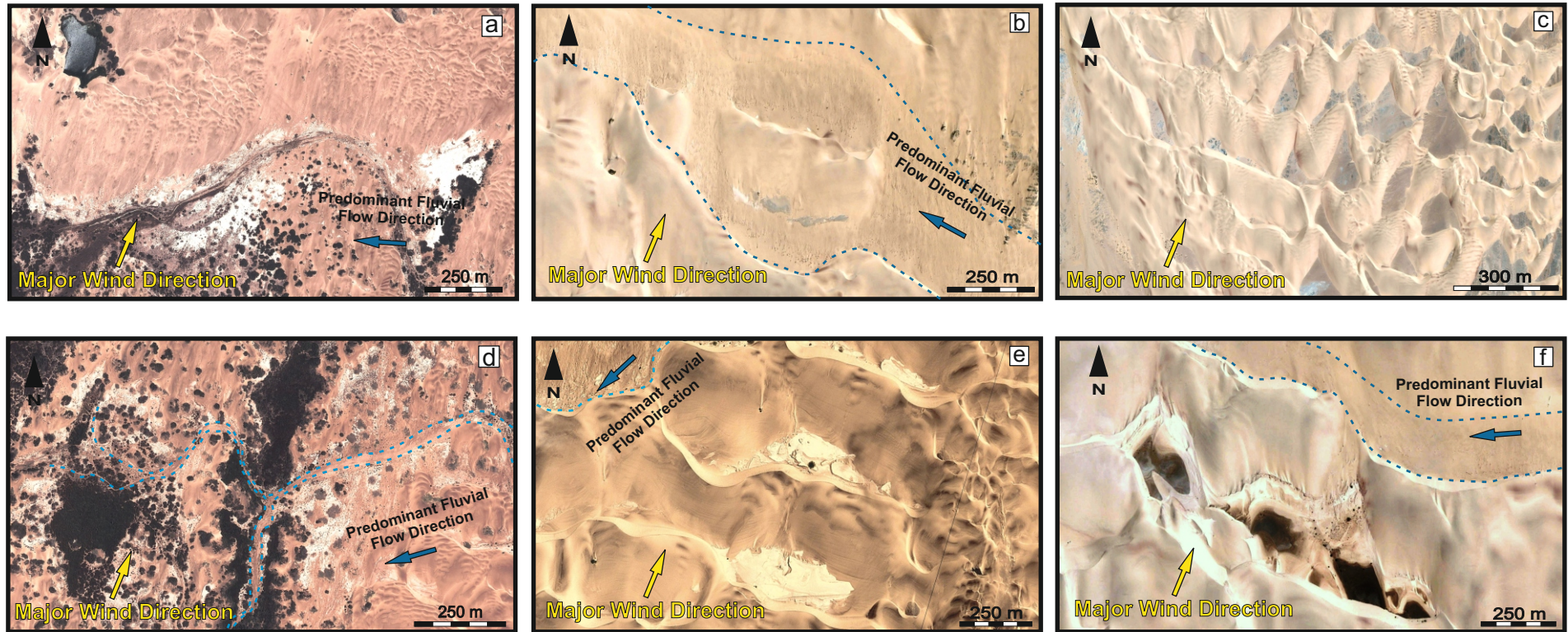


Figure 3.5: Examples of different types of fluvial and aeolian interaction within the Hoanib River: **(a)** migrating aeolian sand covering small distributary channels within flood-basin ponds at the eastern erg margin; note isolated wet interdune as result of a recent flood. **(b)** Aeolian dunes migrating over dry river channel. **(c)** Large dunes at the western erg margin. **(d)** Highly vegetated small distributary channels within the flood basin at the eastern erg-margin. **(e)** Dry southern interdunes filled with fine deposits near the eastern erg margin. **(f)** Wet isolated interdunes at the western erg margin. See Figure 3.4 for locations.

within the dune-field interior. Abundant vegetation (principally tamarisk but also other dryland bush species) have colonized the flood-basin floors (Stanistreet and Stollhofen, 2002).

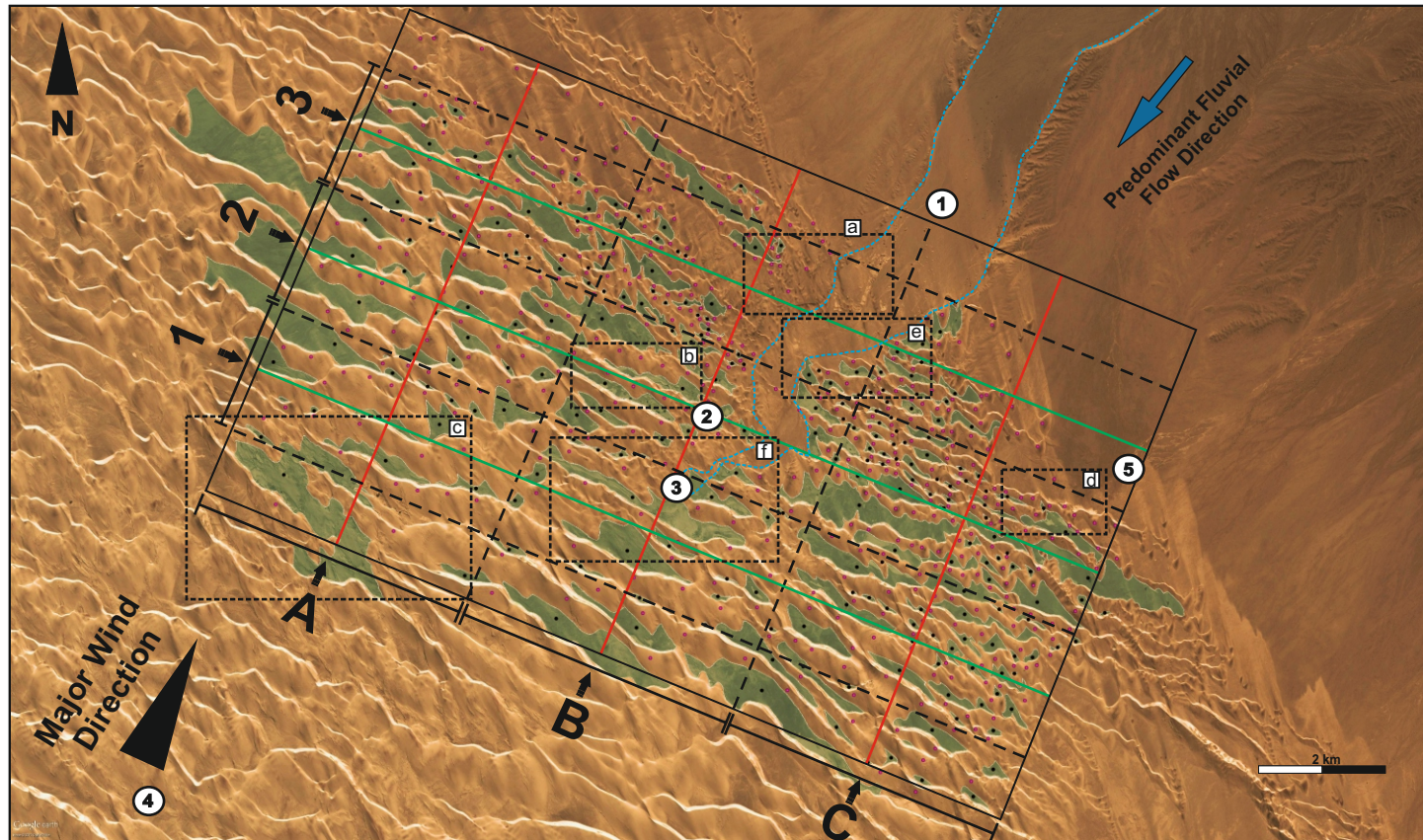
The morphology of the dunes where the river intersects the dune field at its eastern margin comprise a series of 20 to 50 m-high transverse and barchanoid dunes (Lancaster, 1982); these block the west southwest-flowing river course at the eastern erg margin to form an extensive 15 x 4 km flood basin (Gui-Uin). Vegetation in the Hoanib River acts as an obstacle that tends to anchor aeolian dunes, thereby encouraging their stabilization; additionally, this vegetation also partially blocks the river channel and may locally reduce water velocity during flood events (Fig. 3.5d). Increased surface roughness associated with vegetation colonization at the eastern erg margin has encouraged aeolian sand-sheet construction and accumulation. The influence of the vegetation within the flood basin also may stabilize the aeolian dunes sand and increase the possibility of development of sand sheets rather than dunes (cf. Kocurek & Nielson, 1986).

3.6.2 Hunkab River (19 48 25.80 S 13 4 9.44 E)

The 90-km-long Hunkab River is the shortest river considered in this study and is fed by a 700 km² catchment receiving mean annual rainfall below 100 mm (Jacobson et al., 1995). Several types of interaction between aeolian and fluvial processes and landforms have been identified (Fig. 3.7). The termination of a clearly defined ephemeral river (Fig. 3.7a) has broken through the Skeleton Coast erg only twice in recent decades: April 1995 and March 2000 (Jacobson et al., 1995).

The morphology of the dunes in the eastern erg margin (Fig. 3.7b), where the river is completely dammed (Fig. 3.7f), comprises 30 to 50 m-high compound crescentic dune bedforms (Lancaster, 1982; Krapf et al., 2003). The crestlines of the damming dunes are oriented north-northeast, which is perpendicular to the river flow direction (Fig. 3.7d). These dunes morph into transverse forms that are up to 50 m high towards the west (Krapf et al., 2003). The lack of along-crest sinuosity of the dunes means that the bounding interdune flat areas tend to be open, elongate, relatively wide and parallel to the trend of the dunes (Fig. 3.7b). This allows flood waters to readily penetrate along interdune corridors on both sides of the main river course, and thereby allow the accumulation of mud layers (Fig. 3.7c & Fig. 3.7f).

Hunkab River



Key Information

- ① Passage of Hunkab River through its main catchment.
- ② Braid-like fluvial barform architecture with vegetation.
- ③ Site of limited break-through into the dune field; river forms a series of ponds and distributary channels between dunes.
- ④ Path of former successful break-through: fluvial course passes through dune field toward the Atlantic Ocean.
- ⑤ Course of former floods obscured by aeolian dunes that have migrated over former channels; no evident course for fluvial systems to cross dune field.

Interpretation

Since 1995, the Hunkab River has been dammed by the dunes at the leading edge of the 15-km-wide dune field. No significant flood-basin lake has developed to the east of the dune field. This detailed image reveals no long-lived fluvial corridor through the dune field. The river channel is re-occupied rapidly by aeolian bed-forms in the aftermath of floods in response to aeolian sand migration following flood break-through events.

Figure 3.6: Study area of Hunkab River. The black spots are points from where data were collected and the interdunes are highlighted in blue (false colour). Red and green lines are transects representing trends of aeolian dune and interdune geomorphology along the transects that are parallel (A, B and C) and perpendicular (1, 2, and 3) to the prevailing wind direction. See Figures 3.7 for details of a-f.

Examples of Fluvial-Aeolian Interaction in the Hunkab River Catchment

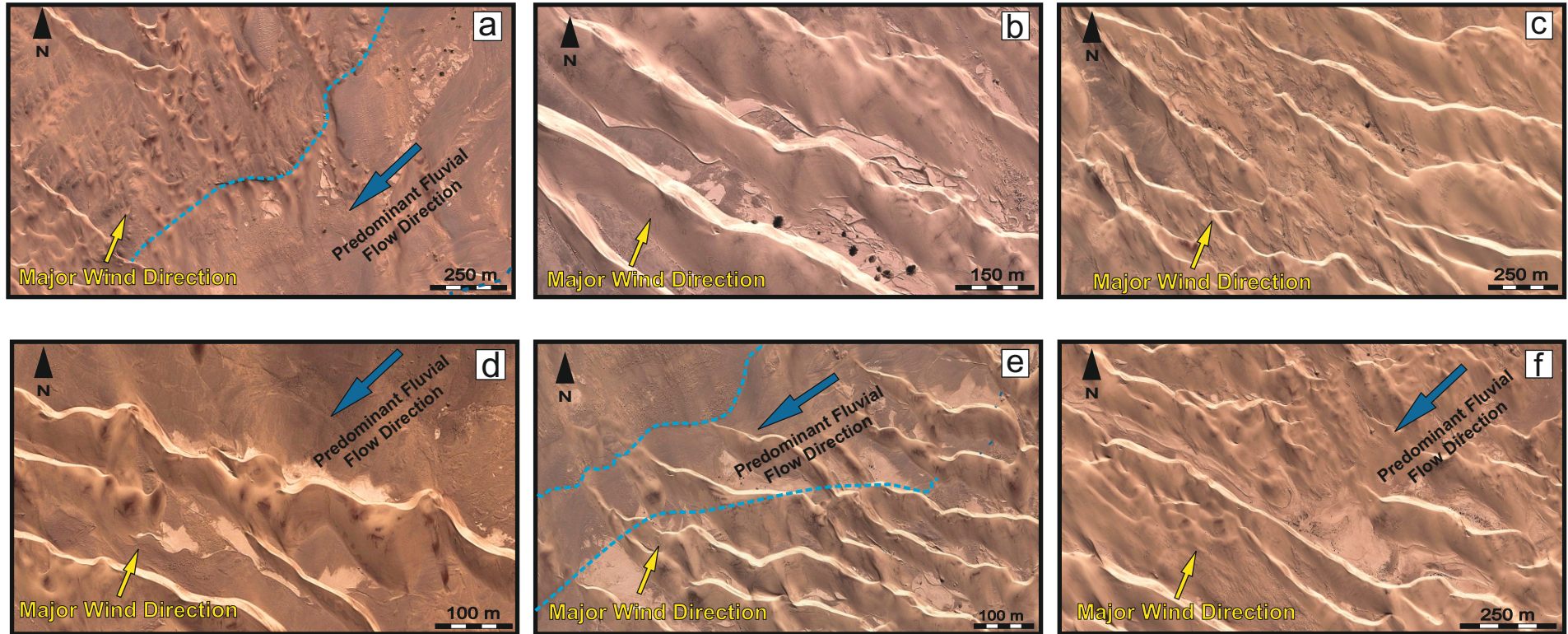


Figure 3.7: Examples of different types of fluvial-aeolian interaction within the Hunkab River: **(a)** Remnant of mud layers within the river incursion at the eastern erg margin. **(b)** Migrating dunes over former mud-rich interdunes. **(c)** Mixed aeolian and fluvial sediments at the river terminus. **(d)** Dammed fluvial sediments in front of damming dune walls. **(e)** Southern dune toes intersect the dry river; note the fine sediment in the ponded interdunes. **(f)** A dune wall completely blocks the river.

Mud layers are well developed on both sides of the river course, indicating that the elevation of the interdune depressions is very similar to that of the river channel bank. Interdune corridors at the southern side of the river are better connected than those at the northern side, which are relatively shorter and apparently less frequently occupied (Fig. 3.7e).

From the examined satellite images, numerous small flooded ponds are developed within interdune corridors between the dunes and these are characterized by deposits of fine sediments that are clearly visible in the images (Figs. 3.7a & f). The fine sediments have accumulated adjacent to the dune lee toe where ponded water successfully broke through the dune wall to enter into an open interdune corridor behind, depositing mud layers over the flat interdune floor.

In the aftermath of flood events, dunes migrated over these resistant mud layers preserving mud-prone elements whose preserved shape mimics that of the original interdune (Fig. 3.7b & 3.7c). In places, aeolian sand has become trapped on the damp pond surfaces via adhesion, thereby leading to the accumulation of a mixture of poorly sorted sediments (cf. Kocurek and Fielder, 1982).

3.6.3 Uniab River (20 8 53.86 S 13 17 50.56 E)

The 110 km-long Uniab River is located between latitude S 20° 09' and S 20° 11' and crosses the entire erg approximately 20 km to the north of the southern (upwind) erg boundary (Fig. 3.8). Satellite imagery records a recently active shallow braided fluvial channel which flows to the south-southwest and intersects a 7-km-wide erg along a well-established break-through corridor. The width of this break-through corridor ranges from 800 m close to the western erg margin to 2700 m close to the eastern erg margin. This active river drains a 4500 km² catchment area, of which only 2.3% receives mean annual rainfall of > 100 mm (Jacobson et al., 1995). Satellite images for the Uniab river course reveal different interaction types from at the eastern and western dune-field margins (Fig. 3.9). The river path has apparently remained stable during dry episodes between floods. Successive flood events have occurred on numerous occasions in recent decades. A major flood episode took place in early April 1995: the Uniab River broke-through the dune field which had dammed its path towards the sea for 13 years since 1982 (Jacobson et al., 1995). Subsequent flood events occurred in March 1997 and in April 2000 and these acted to flush aeolian sand from the river course, thereby preventing dune migration and accumulation within the channel from acting as an effective healing process. The aerial

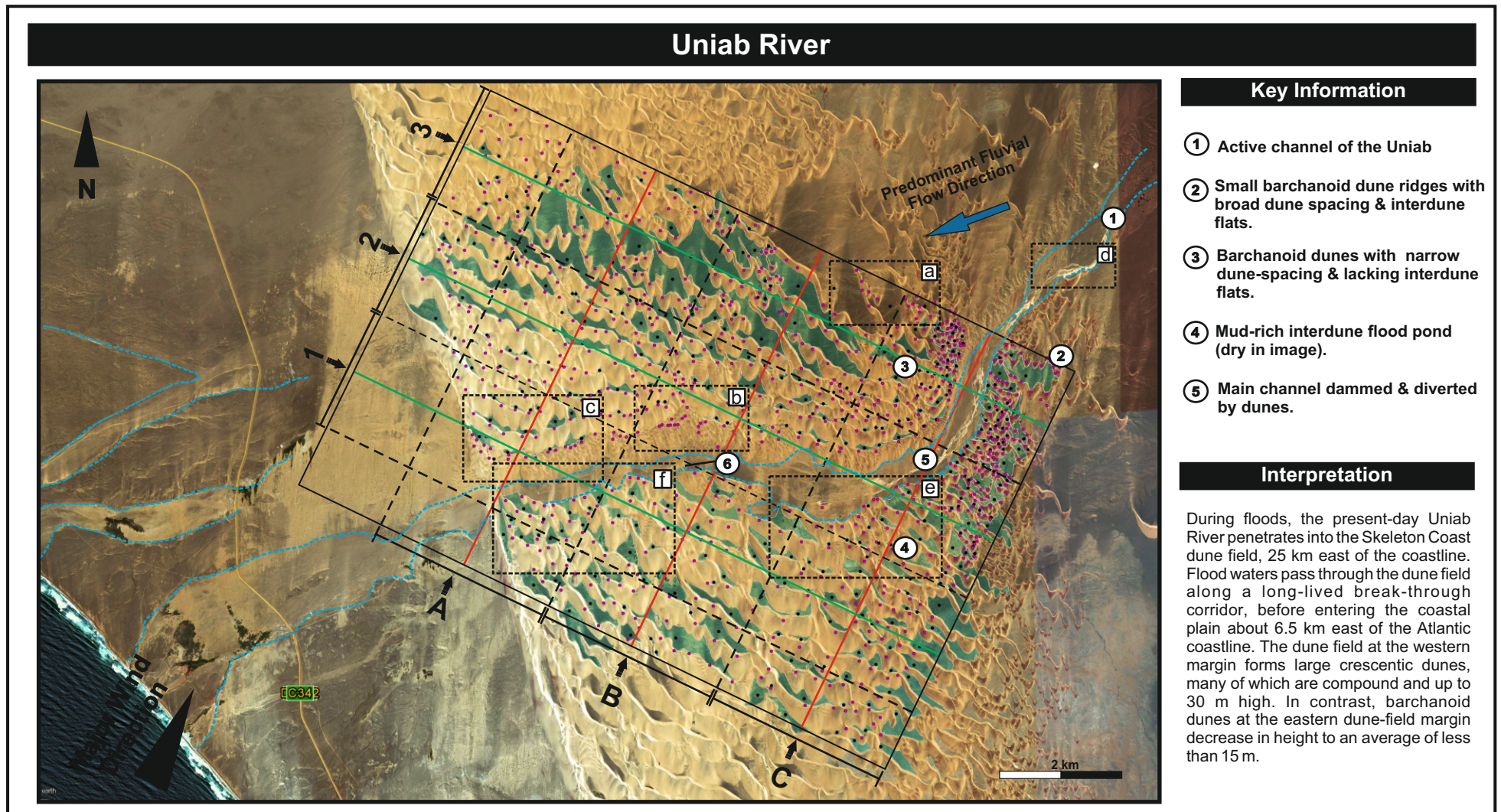


Figure 3.8: Study area of Uniab River. The black spots are points from where data were collected and the interdunes are highlighted in blue (false colour). Red and green lines are transects representing trends of aeolian dune and interdune geomorphology along the transects that are parallel (A, B and C) and perpendicular (1,2, and 3) to the prevailing wind direction. See Figure 3.9 for details of a - f.

Examples of Fluvial-Aeolian Interaction in the Uniab River Catchment

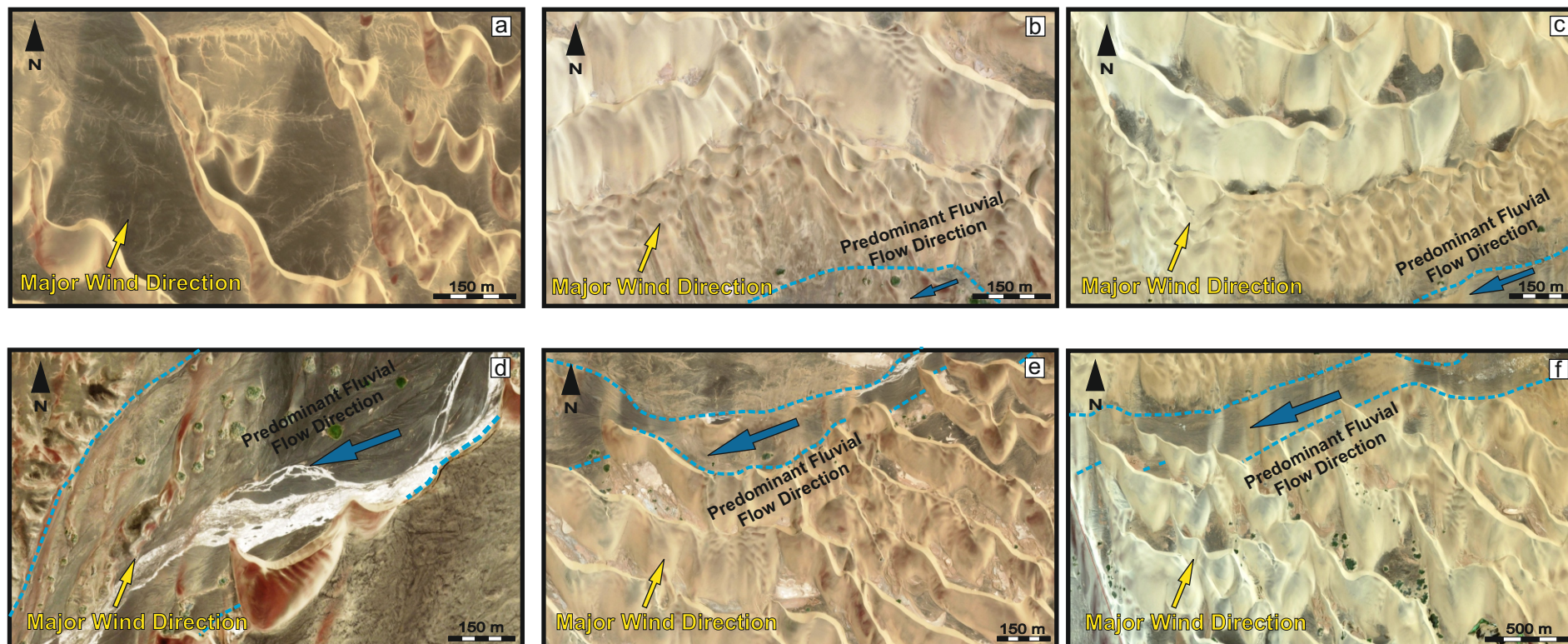


Figure 3.9: Examples of different types of interactions within the Uniab River: **(a)** Active oblique migrating dunes over former distributary channels at the eastern erg margin. **(b)** Aeolian sand reworked by fluvial process to the north of the river. **(c)** Advancing aeolian sediments towards NNE limiting the lateral expansion of the river. **(d)** Active barchans migrating over river. **(e)** Southern dune toes intersect the river bed in the central erg. **(f)** The southern dune toes almost blocking the river bed at the western erg margin.

photographs reveal several well-imaged examples of types of fluvial-aeolian interaction within the vicinity of this channel.

Where aeolian dunes adjacent to the fluvial course in the south, north-northeast trending dunes form 15 to 30 m-high barchanoid ridges, which systematically decrease in size from the western to the eastern erg margin, where the dunes take the form of scattered and spatially separated barchans. The largest crescentic dunes occur along the western and central part of the erg (Lancaster, 1982). Images record on-going dune migration over a recently active east-to-west channel which is itself oriented perpendicular to the prevailing wind direction (Fig. 3.9d and 3.5f).

Migrating dunes on the southern side of the river channel – some with horns – are separated by intervening damp flat areas that represent interdune corridors (Fig 3.9e). The southern interdune corridors along the eastern erg margin are of variable width (60 to 1040 m) where small barchans dunes merge together to become chains of larger barchanoid dune ridges. Towards the west (Fig 3.9c), these interdune corridors become more confined and have greater unbroken lengths (44 to 954 m): here they were apparently flooded in a manner whereby mud-laden flood waters ponded, leading to the gradual accumulation of mud layers (Fig 3.9e).

To the north of the river, by contrast, a distinctive belt of reworked sediment along the northern margin of the recently active channel acts as a positive topographic feature preventing flood waters from gaining access to adjoining interdune corridors (Fig. 3.9b and 3.9c); as such flood-basin ponds have not developed to the same extent here and associated mud layers have not accumulated in this region. Further, the elevation of the erg accumulation surface to the north of the river is 6 to 18 m above the level of the river itself (Fig. 3.8), thereby reducing the opportunity for fluvial incursion beyond the confines of the main channel. The dominant orientation of dune brink lines within this belt is parallel to the east-west oriented channel. This orientation swings round to become close to perpendicular further towards the north due to oblique migration of these dunes (Fig 3.9a). As a result, the interdune areas directly adjacent to the northern banks of the river are narrow, more isolated non-flooded depressions. However, they become wider and more open corridors between the oblique dunes further north (Fig. 3.9c).

Several distributary channels that are buried partially by aeolian sand are clearly visible in the studied satellite images and these document the paths of recent flood events. The small distributary channels, which are each up to 30 m wide, developed in the aftermath of earlier flood events. (Fig. 3.9a).

3.7 Results of geometrical relationships of dunes and interdunes in studied areas

The results of collated geometrical data within the selected rivers show the geomorphological relationship between adjacent dunes and the interdunes and their relationship to the adjacent fluvial systems. Graphs of collated data demonstrate relations between the morphological arrangement of aeolian dunes and interdunes for the present time from six cross sections within the vicinity of each of the three studied river courses. Graphed results of the studied areas show results that record important types of interaction and quantified spatial relationships associated with each of these river systems. For each river, three transects (A, B and C) – aligned parallel to each other and spaced 2.5 km for Hoanib and Uniab, 1.5 km for Hunkab apart – commence at a fixed line (arbitrary position chosen) in the central part of the erg and pass north-northeast-ward in a direction parallel to the prevailing SSW wind direction crossing the river system of interest (Figs. 3.4, 3.6, and 3.8). Additionally, for each river, three transects (1, 2, and 3) – again aligned parallel to each other and spaced 1.5 km for Hoanib and Uniab, 0.7 km for Hunkab apart – commence at a fixed line in the western part of the erg (arbitrary position chosen), and pass in a direction that is perpendicular to the prevailing SSW wind toward the eastern erg margin (Fig. 3.4, 3.6, and 3.8). Each cross section is a transect for which a range of data relating to dune and interdune morphology have been collected.

The results of the examined quantified measurements of dunes and interdunes within each studied river region include comparison between mean, maximum and minimum dune wavelength, wavelength of along-crest sinuosity of dunes, amplitude of along-crest sinuosity of dunes. Results of analysis of geometric measurements of interdunes (shown in blue false colour in all images – Figs. 3.4, 3.6 and 3.8) reveal relations between interdune width and length and relation to adjoining dunes and to the position of the nearby river. See the methods section for details.

3.7.1 Results from Hoanib River

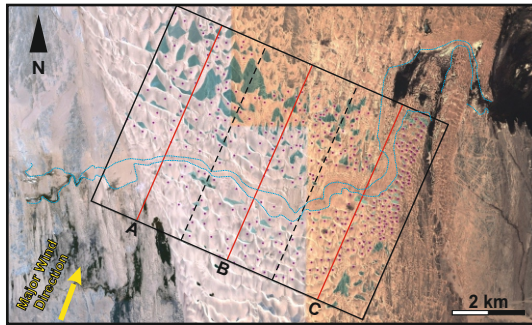
3.7.1.1 Description

Graphed results from observations in Hoanib River demonstrate various trends of aeolian dune geomorphology along transects that are both parallel (A, B and C) and perpendicular (1, 2, and 3) to the prevailing SSW wind direction (Fig. 3.10).

Transects A, B and C

Geometrical Results of Aeolian Dunes (Hoanib River)

Cross sections parallel to the prevailing wind direction NNE



Cross sections perpendicular to the prevailing wind direction NNE

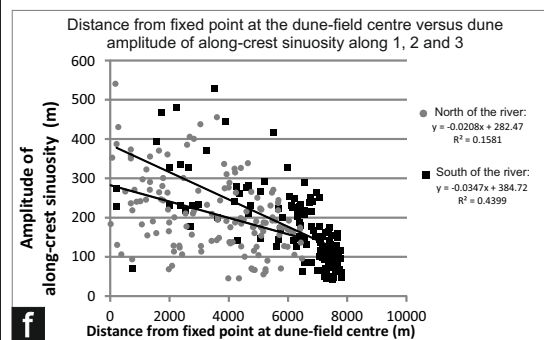
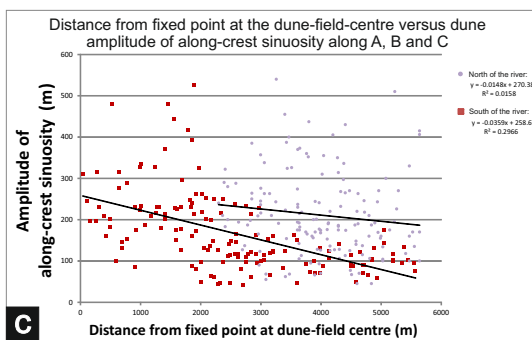
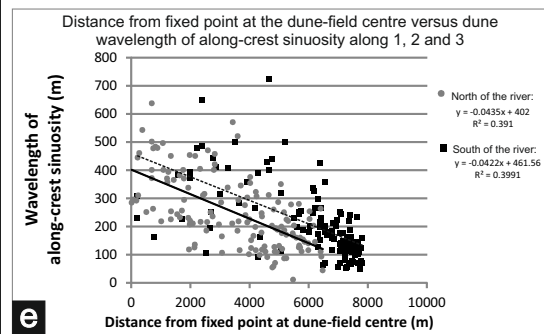
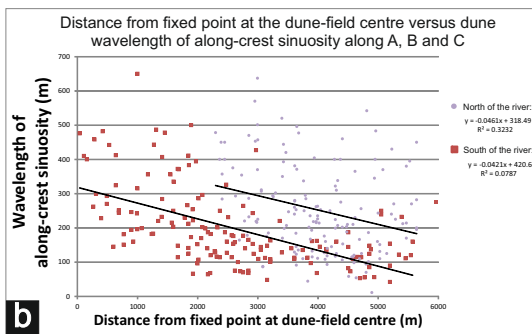
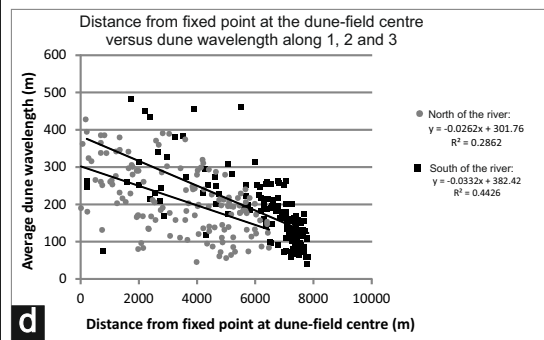
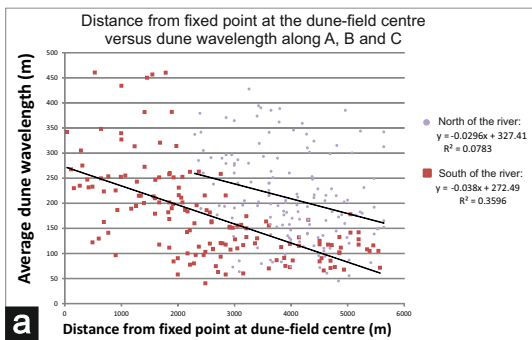
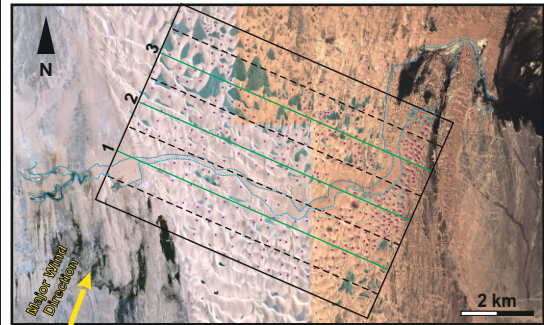


Figure 3.10: Examples of Hoanib River data summarizing spatial changes in morphology of dunes along cross sections parallel (A, B, and C) and perpendicular to the prevailing wind direction (1, 2 and 3): (a) Average dune wavelength along cross sections A, B and C towards the NNE. (b) Wavelength along-crest sinuosity along sections A, B and C towards the NNE. (c) Amplitude along-crest sinuosity along sections A, B and C towards the NNE. (d) Dune wavelength along cross sections 1, 2 and 3 towards the SSE. (e) Wavelength along-crest sinuosity along sections 1, 2 and 3 towards the SSE. (f) Amplitude along-crest sinuosity along sections 1, 2 and 3 towards the SSE.

Along transects A, B and C, which are aligned parallel to the prevailing wind direction, graphed results exhibit trends that record spatial variation in the arrangement of dune bedform patterns as a gradational transition occurs from combined barchanoid to simple dune bedforms. The dune bedforms decrease in size associated with an increase in interdune size from the dune-field centre. The dune bedforms show gradual decrease in size from the areas south of the Hoanib river toward the areas to the north of the river (Fig. 3.10, 3.11 and 3.12).

To the south (upwind) of the Hoanib River the following trends regarding spatial changes in dune and interdune morphology are noted in passing from SSW to NNE from the dune-field centre towards the southern river bank: (1) mean dune bedform wavelength (Fig. 3.10a) decreases from 460 to 70 m; (2) mean wavelength of the along-crest sinuosity of the dunes (Fig 3.10b) decreases from 650 m to less than 40 m, though the trend is relatively weak; (3) mean amplitude of the along-crest sinuosity of the dunes (Fig. 3.10c) decreases 500 m to less than 60 m; (4) mean interdune length (Fig. 3.11) decreases from 500 m to less than 10 m; (5) mean interdune width (Fig 3.12) decreases from 200 m to less than 10 m, though the trend is weak.

To the north (downwind) of the Hoanib River the following trends regarding spatial changes in dune and interdune morphology are noted passing from SSW to NNE from the northern river bank: (1) mean dune bedform wavelength (Fig. 3.10a) decreases from 420 to 50 m; (2) mean wavelength of the along-crest sinuosity of the dunes (Fig 3.10b) decreases from 600 m to less than 100 m; (3) mean amplitude of the along-crest sinuosity of the dunes (Fig. 3.10c) decreases 540 m to 45 m, though the trend is weak; (4) mean interdune length (Fig. 3.11) increases from 35 to 1380 m, though the trend is weak; (5) mean interdune width (Fig 3.12) increases from 11 to 700 m, though the trend is weak.

Transects 1, 2 and 3

Along transects 1, 2 and 3, which are aligned perpendicular to the prevailing wind direction, graphed results exhibit trends that record spatial variation in the arrangement of dune bedform patterns as a gradational transition occurs from the western to the eastern dune-field margin where the shape and extent of interdunes is determined by the morphology and spacing of surrounding aeolian dune forms (Fig 3.10d, e, f and Fig 3.13).

To the south of the Hoanib River, the following trends regarding spatial changes in dune and interdune morphology are noted passing from WNW to

Geometrical Results of Aeolian Interdune Lengths Along Cross Sections A, B and C in (Hoanib River)

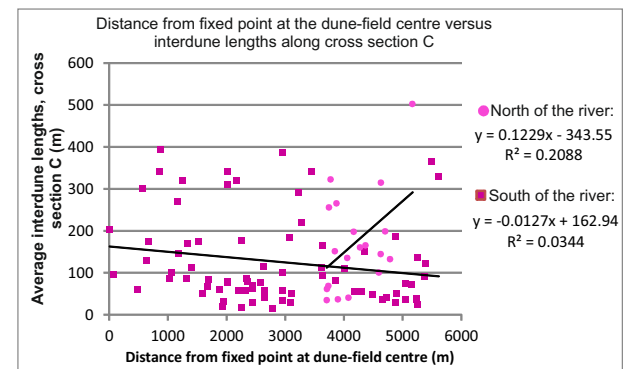
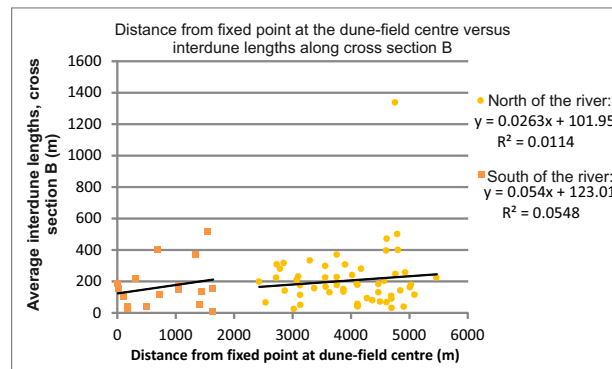
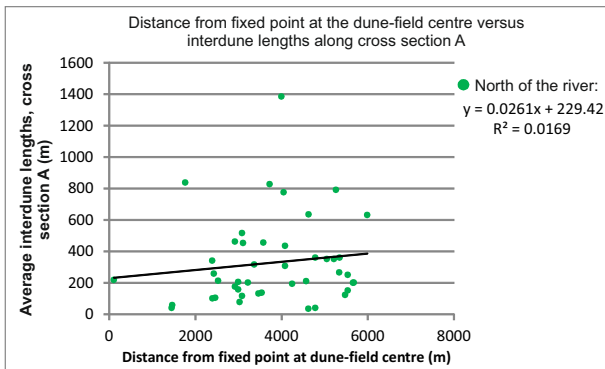
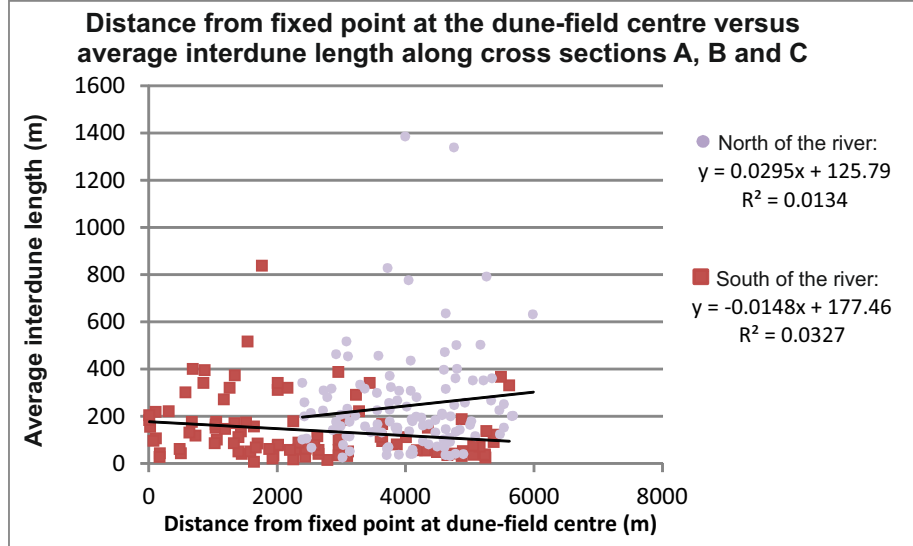
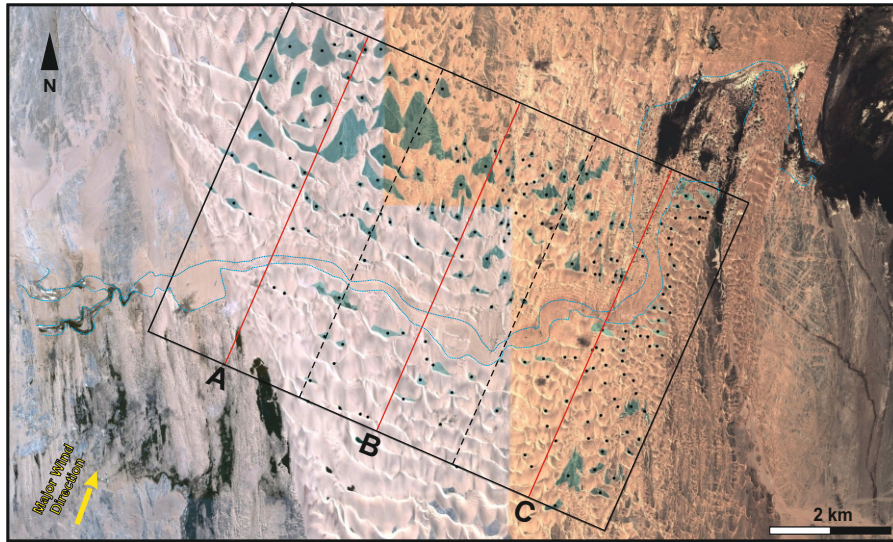


Figure 3.11: Examples of Hoanib River data summarizing spatial changes in aeolian interdune lengths along cross sections A, B, and C parallel to the prevailing wind direction towards the NNE.

Geometrical Results of Aeolian Interdune Widths Along Cross Sections A, B and C in (Hoanib River)

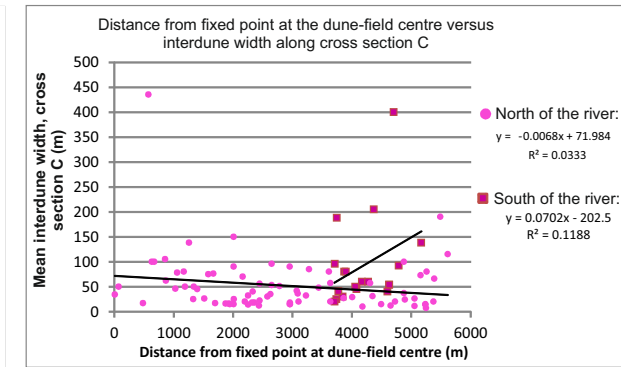
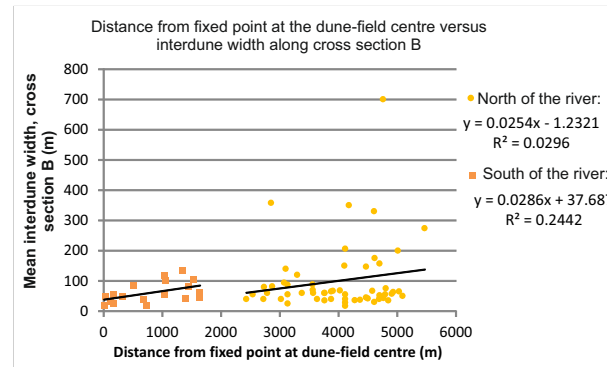
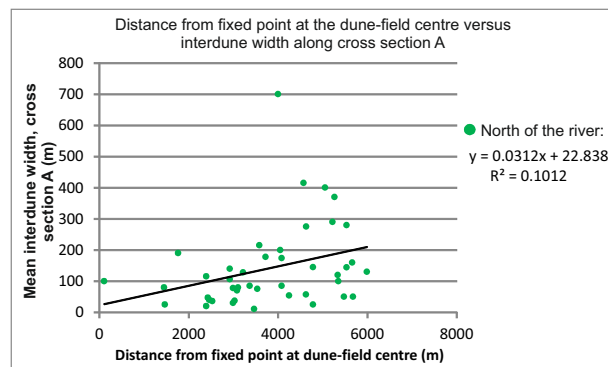
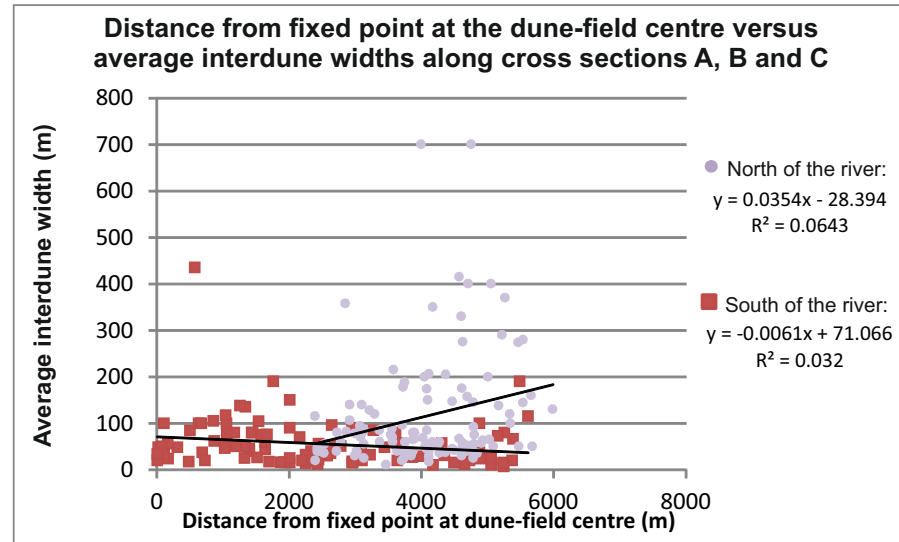
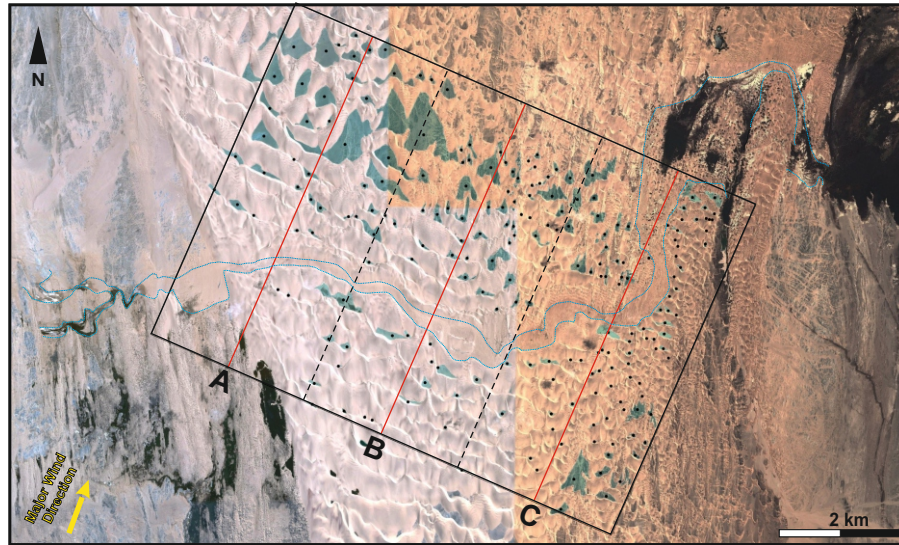


Figure 3.12: Examples of Hoanib River data summarizing spatial changes in aeolian interdune widths along cross sections A, B, and C parallel to the prevailing wind direction towards the NNE.

Geometrical Results of Aeolian Interdune Lengths and Widths Along Cross Sections 1, 2 and 3 in (Hoanib River)

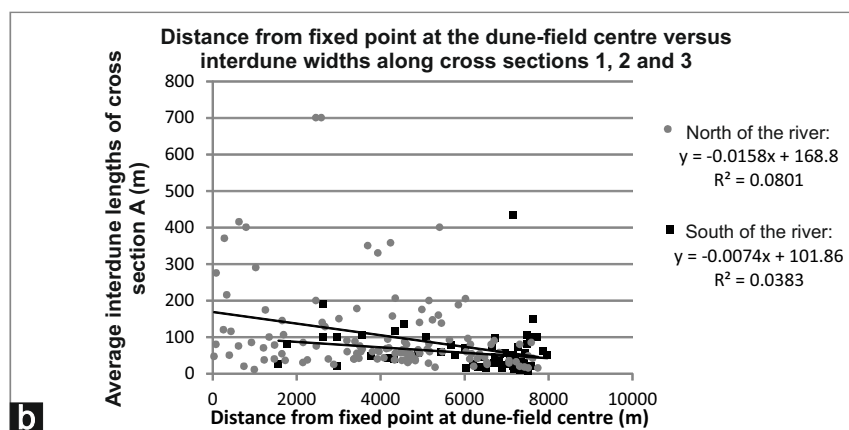
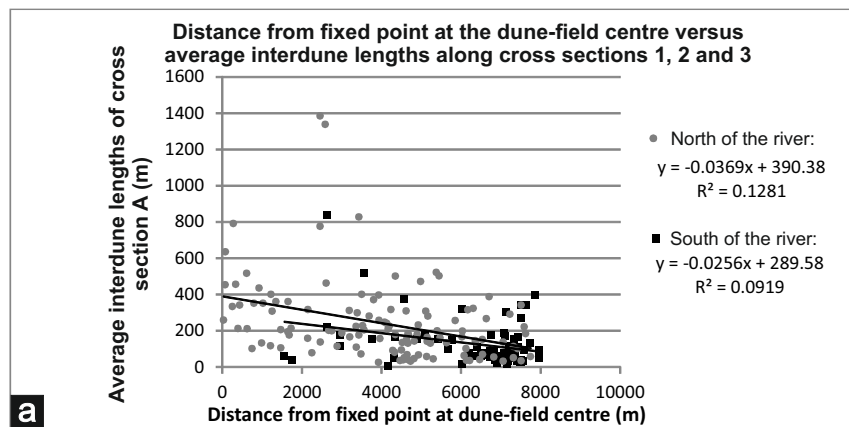
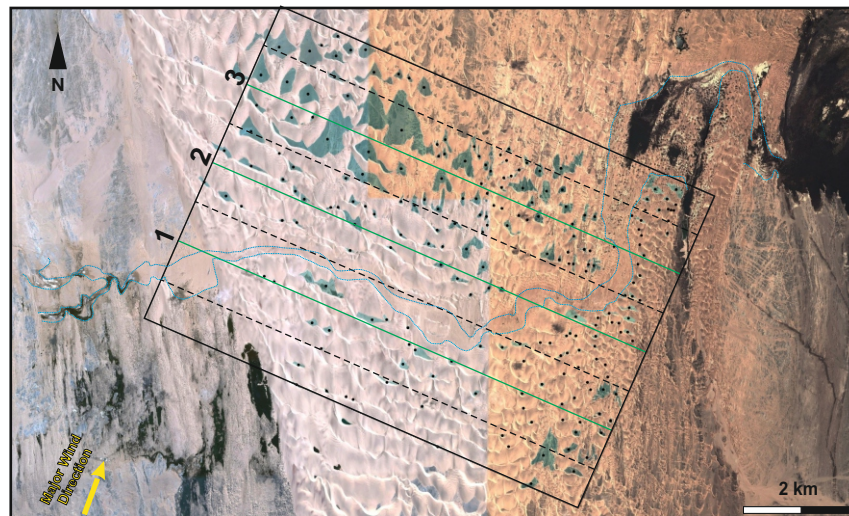


Figure 3.13: Examples of Hoanib River data summarizing spatial changes in aeolian interdune lengths and widths along cross sections 1, 2, and 3 perpendicular to the prevailing wind direction towards the SSE.

ESE: (1) mean dune bedform wavelength (Fig. 3.10d) decreases from 480 to 40 m; (2) mean wavelength of the along-crest sinuosity of the dunes (Fig 3.10e) decreases from 700 m to less than 50 m; (3) mean amplitude of the along-crest sinuosity of the dunes (Fig. 3.10f) decreases 500 m to less than 60 m; (4) mean interdune length (Fig. 3.13a) decreases from 385 m to less than 10 m, though the trend is weak; (5) mean interdune width (Fig 3.13b) decreases from 435 m to less than 10 m, though the trend is weak.

To the north of the Hoanib River, the following trends regarding spatial changes in dune and interdune morphology are noted passing from WNW to ESE: (1) mean dune bedform wavelength (Fig. 3.10d) decreases from 600 to less than 40 m; (2) mean wavelength of the along-crest sinuosity of the dunes (Fig 3.10e) decreases from 650 m to less than 40 m; (3) mean amplitude of the along-crest sinuosity of the dunes (Fig. 3.10f) decreases 500 m to less than 45 m, though the trend is weak; (4) mean interdune length (Fig. 3.13a) decreases from 1400 m to less than 20 m, though the trend is weak; (5) mean interdune width (Fig 3.13b) decreases from 700 m to less than 15 m, though the trend is weak.

3.7.1.2 Interpretation

The depicted trends demonstrate how the interdune morphology of Hoanib area is determined by spatial changes in the form and spatial distribution and orientation of dunes in the vicinity of the river.

Dunes to the north (downwind) of the river are relatively smaller than those to the south (upwind). In general, dune crestlines that are adjacent to the river in the dune-field centre and close to the western erg margin are aligned parallel to the direction of river flow, thereby preventing fluvial incursion into interdune areas during major flood events.

Further, the interdunes to the north of the river (Fig 3.10) are influenced by the degree of sinuosity of the planform morphologies of the dunes to the north of the river, especially at the eastern erg margin where dune forms become relatively more sinuous than the relatively simple transverse and barchanoid dunes at the western erg margin.

The increase in interdune size and connectivity in the areas to the north of the river and the corresponding decrease in dune size towards NNE results from an overall decrease in either the rate of generation of a sand supply for aeolian dune construction close to the eastern dune-field margin, the limited availability of that supply for aeolian dune construction, or a downwind

decrease in the sediment transport capacity of the wind (Kocurek and Lancaster, 1999).

Because the water table and its capillary fringe in the vicinity of the Hoanib River are determined by repeated floods, especially in the eastern part dune field, the damp nature of the accumulation surface of the interdunes in these areas tends to be maintained for protracted for much of the time. As such, the sediment on these interdune floors tends not to be available for aeolian transport and this therefore limits the potential for dune construction. Interdune flats are protected from deflationary processes due to the cohesive nature of the damp surface sediments (*sensu* Kocurek and Havholm, 1993). Flooded interdune ponds are more extensively developed to the south of the river channel, indicating gradient-induced preferential avulsion in that direction (Stanistreet & Stollhofen, 2002).

3.7.2 Results from Hunkab River

3.7.2.1 Description

As for the Hoanib River, results of analysis of dune and interdune bedform geomorphology in the vicinity of the Hunkab River represent different spatial changes along transects A, B and C (parallel to wind) and 1, 2, and 3 (perpendicular to the prevailing SSW wind direction) (Fig. 3.14, 3.15, 3.16 and 3.17).

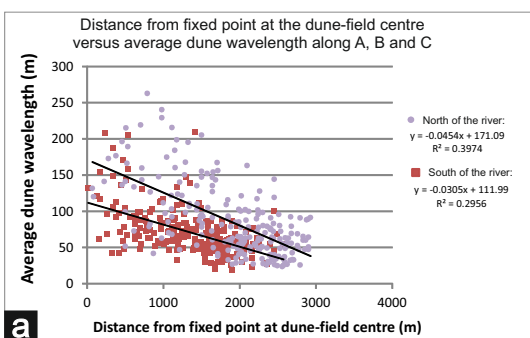
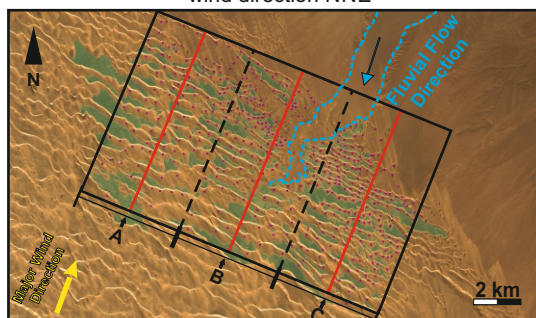
Transects A, B and C

To the south (upwind) of the Hunkab River the following trends regarding spatial changes in dune and interdune morphology are noted passing from SSW to NNE from the dune-field centre towards the southern river bank: (1) mean dune bedform wavelength (Fig. 3.14a) decreases from 207 to less than 30 m; (2) mean wavelength of the along-crest sinuosity of the dunes (Fig 3.14b) decreases from 450 m to less than 30 m, though the trend is relatively weak; (3) mean amplitude of the along-crest sinuosity of the dunes (Fig. 3.14c) decreases 290 m to less than 20 m; (4) mean interdune length (Fig. 3.15) decreases from 1060 m to less than 10 m; (5) mean interdune width (Fig 3.16) decreases from 200 m to less than 7 m, though the trend is weak.

To the north (downwind) of the Hunkab River the following trends regarding spatial changes in dune and interdune morphology are noted passing from SSW to NNE from the northern river bank: (1) mean dune bedform wavelength (Fig. 3.14a) decreases from 260 to 25 m; (2) mean wavelength of the along-crest sinuosity of the dunes (Fig 3.14b) decreases from 400 m

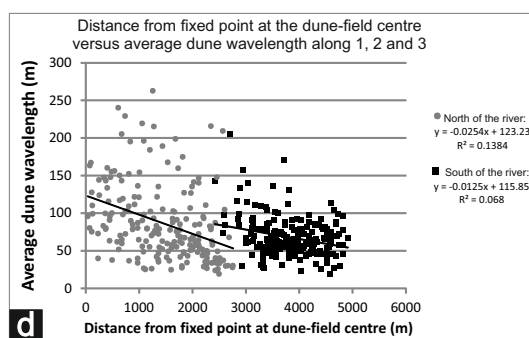
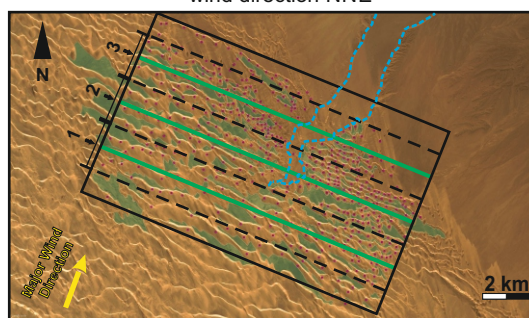
Geometrical Results of Aeolian Dunes in (Hunkab River)

Cross sections parallel to the prevailing wind direction NNE

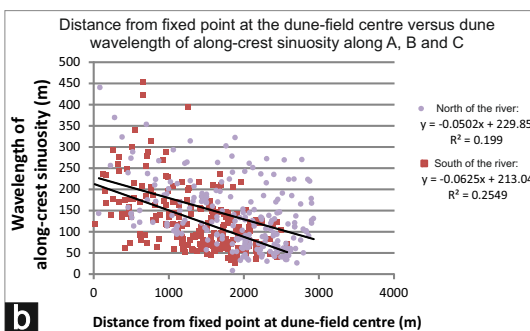


a

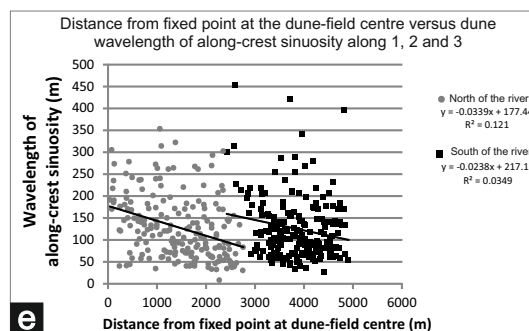
Cross sections perpendicular to the prevailing wind direction NNE



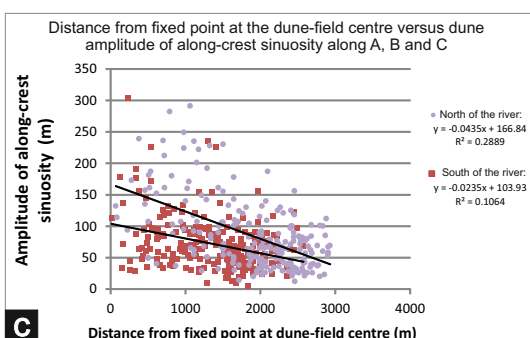
d



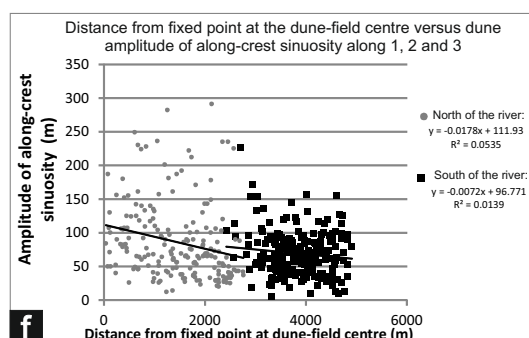
b



e



c



f

Figure 3.14: Examples of Hunkab River data summarizing spatial changes in aeolian morphology of dunes along cross sections parallel (A, B, and C) and perpendicular to the prevailing wind direction (1, 2 and 3): **(a)** Average dune wavelength along cross sections A, B and C towards the NNE. **(b)** Wavelength along-crest sinuosity along sections A, B and C towards the NNE. **(c)** Amplitude along-crest sinuosity along sections A, B and C towards the NNE. **(d)** Average dune wavelength along cross sections 1, 2 and 3 towards the SSE. **(e)** Wavelength along-crest sinuosity along sections 1, 2 and 3 towards the SSE. **(f)** Amplitude along-crest sinuosity along sections 1, 2 and 3 towards the SSE.

Geometrical Results of Aeolian Interdune Length Along Cross Sections A, B and C in (Hunkab River)

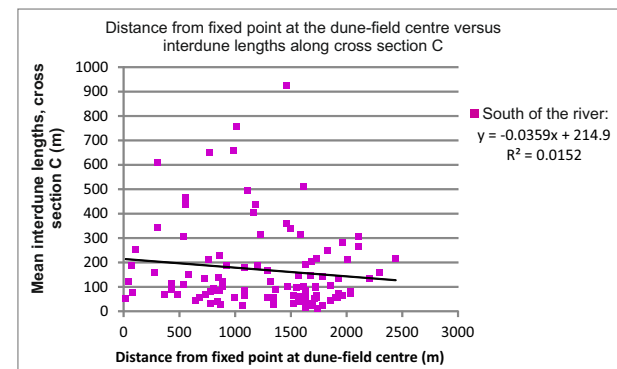
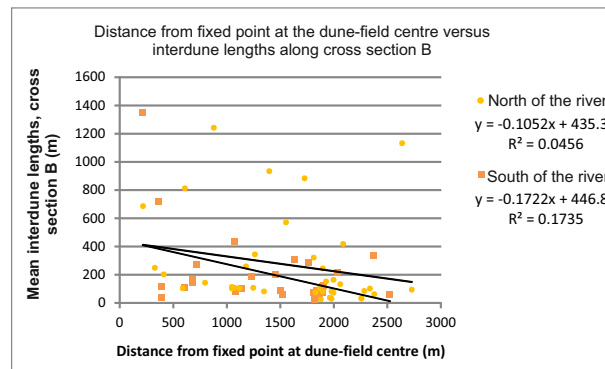
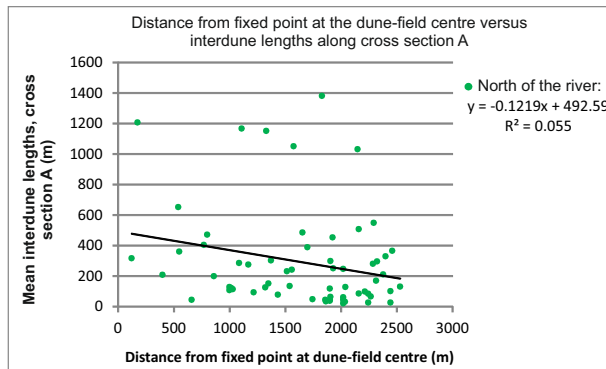
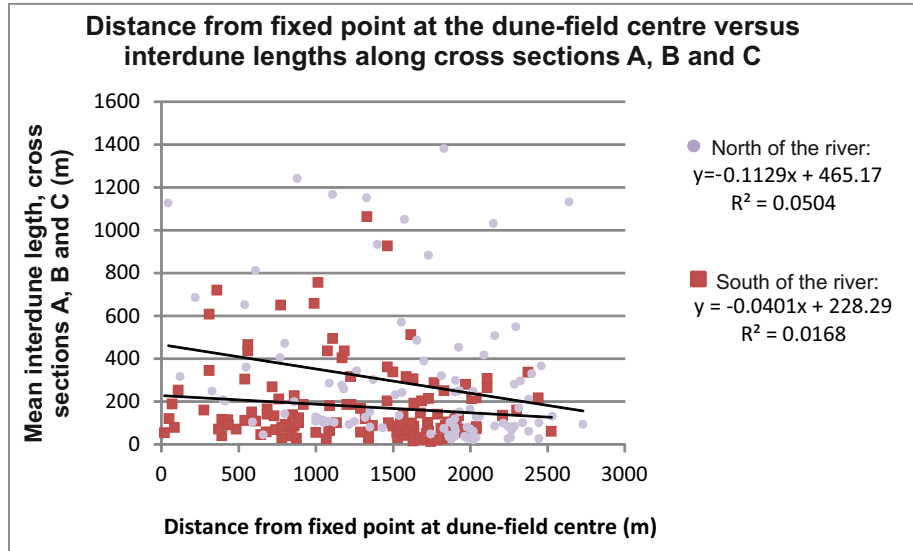
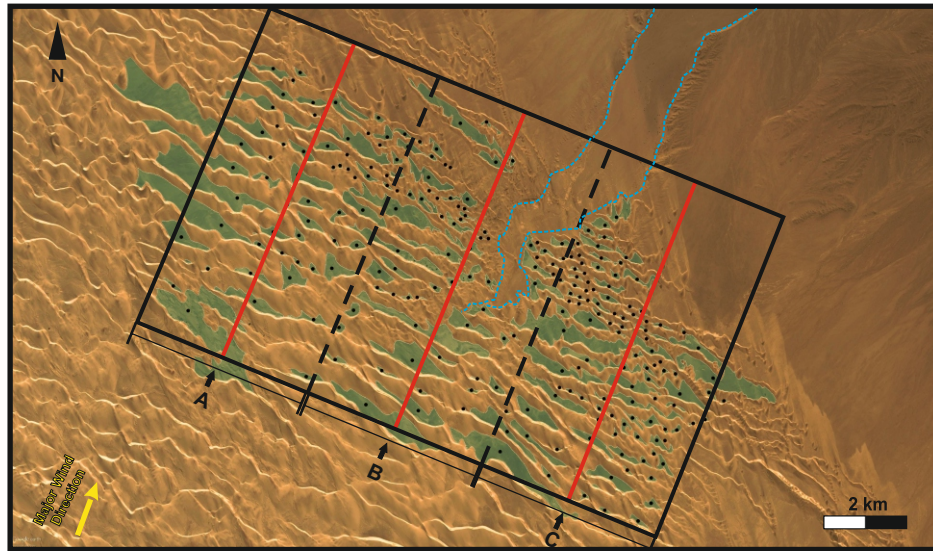


Figure 3.15: Examples of Hunkab River data summarizing spatial changes in aeolian interdune lengths along cross sections A, B, and C parallel to the prevailing wind direction towards the NNE.

Geometrical Results of Aeolian Interdune Width Along Cross Sections A, B and C in (Hunkab River)

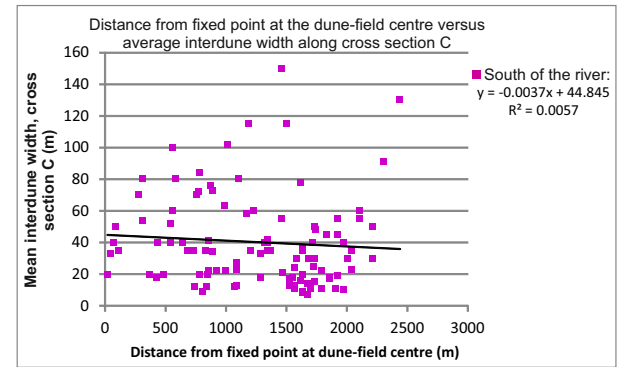
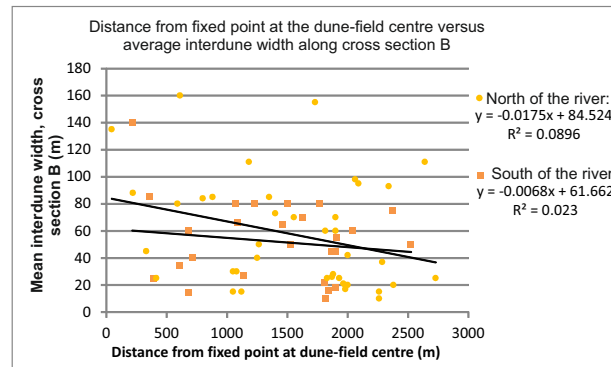
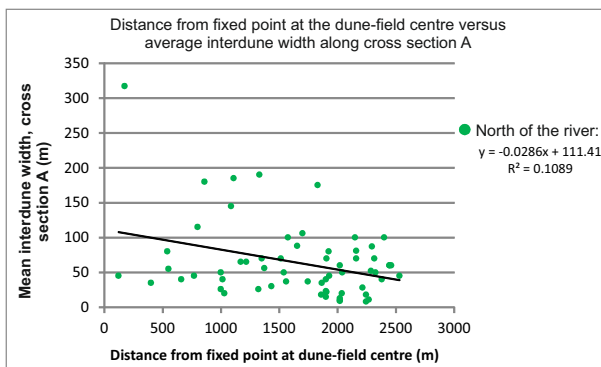
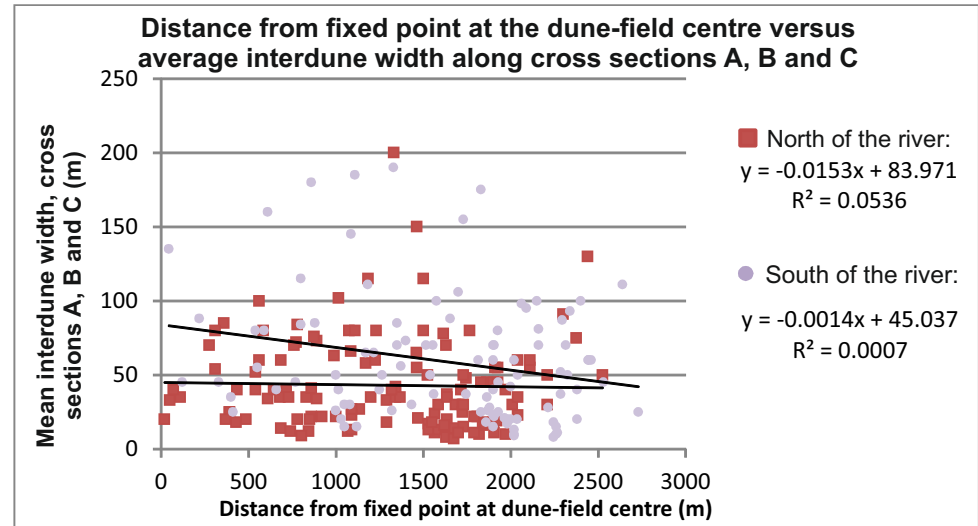
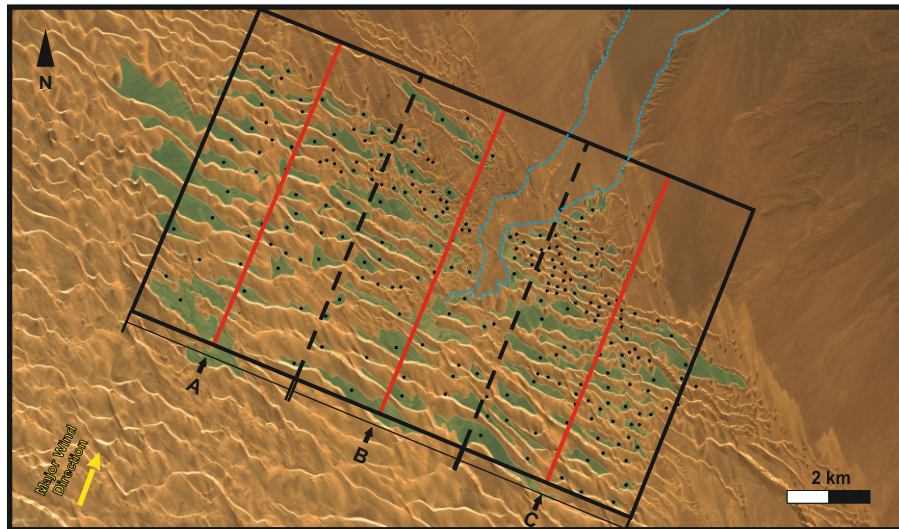


Figure 3.16: Examples of Hunkab River data summarizing spatial changes in aeolian interdune widths present time along cross sections A, B, and C parallel to the prevailing wind direction towards the NNE.

Geometrical Results of Aeolian Interdune Lengths and Widths Along Cross Sections 1, 2 and 3 in (Hunkab River)

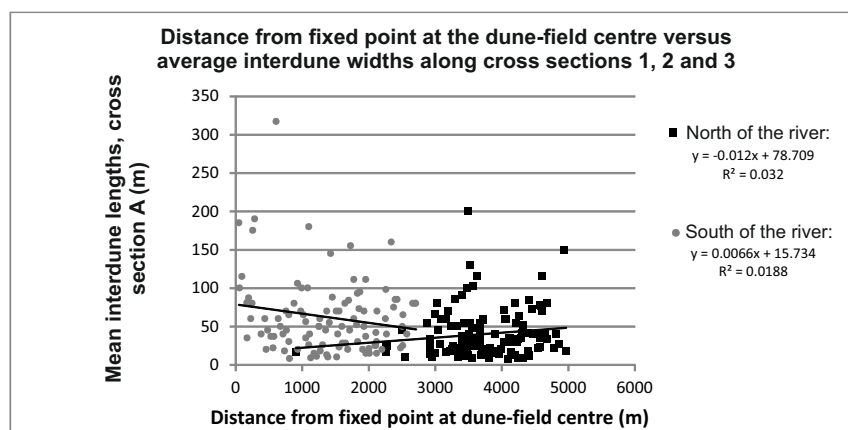
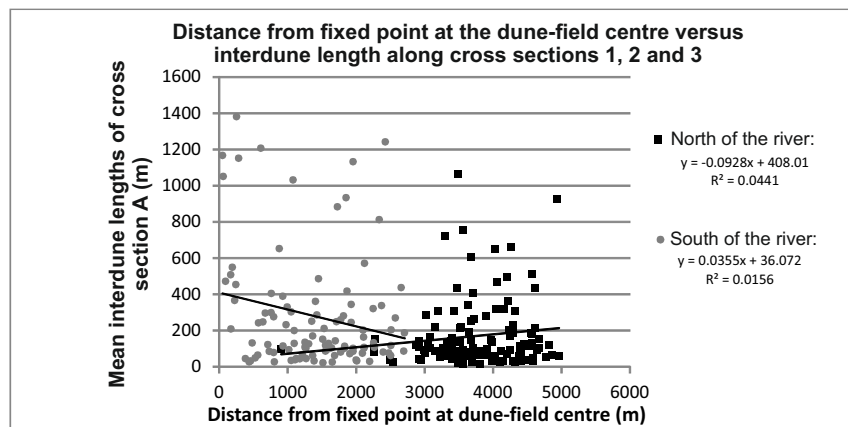
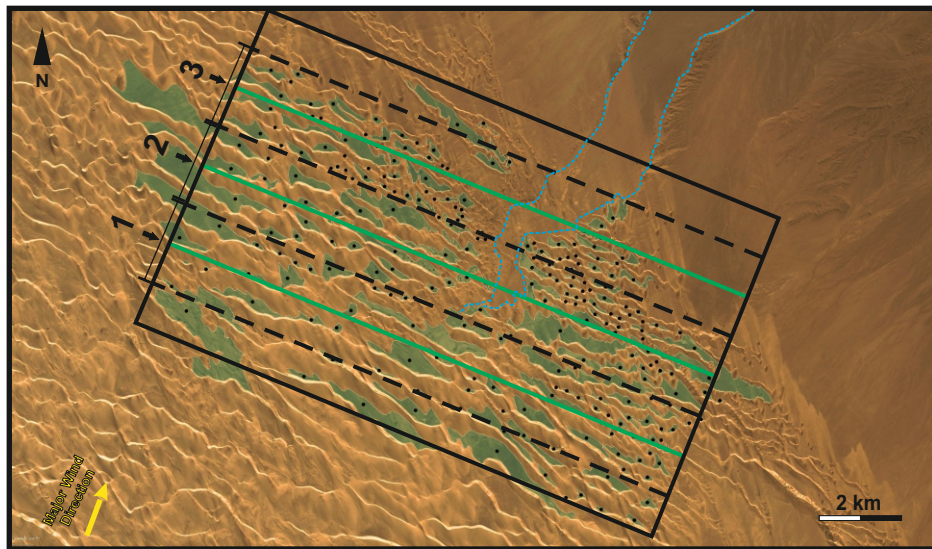


Figure 3.17: Examples of Hunkab River data summarizing spatial changes in aeolian interdune lengths and widths along cross sections 1, 2 and 3 perpendicular to the prevailing wind direction.

to less than 10 m, though the trend is relatively weak; (3) mean amplitude of the along-crest sinuosity of the dunes (Fig. 3.14c) decreases 300 m to less than 10 m; (4) mean interdune length (Fig. 3.15) is scattered and data ranges between 500 m and less than 10 m with no relationship; (5) mean interdune width (Fig 3.16) is scattered and data varies between 1400 m and less than 25 m with no relationship.

Transects 1, 2 and 3

Along transects 1, 2 and 3, which are aligned perpendicular to the prevailing wind direction, graphed results exhibit trends that record spatial variation in the arrangement of dune bedform patterns from the western to the eastern dune-field margin towards the ESE. The data describing morphological parameters for both dune bedforms and interdunes show either no or only weak trends (Fig. 3.14 and Fig. 3.17).

To the north of the Hunkab River, the following trends regarding spatial changes in dune and interdune morphology are noted passing from WNW to ESE: (1) mean dune bedform wavelength (Fig. 3.14d) decreases from 260 to less than 20 m; (2) mean wavelength of the along-crest sinuosity of the dunes (Fig 3.14e) decreases from 350 m to less than 10 m; (3) mean amplitude of the along-crest sinuosity of the dunes (Fig. 3.14f) slightly decreases from 350 m to less than 10 m, though the trend is weak; (4) mean interdune length (Fig. 3.17) decreases from 1400 m to less than 25 m with weak trend; (5) mean interdune width (Fig 3.17) decreases from 300 m to less than 10 m, though the trend is weak.

To the south of the Hunkab River, the following trends regarding spatial changes in dune and interdune morphology are noted passing from WNW to ESE: (1) mean dune bedform wavelength (Fig. 3.14d) decreases from 480 to 40 m; (2) mean wavelength of the along-crest sinuosity of the dunes (Fig 3.14e) decreases from 700 m to less than 50 m; (3) mean amplitude of the along-crest sinuosity of the dunes (Fig. 3.14f) slightly decreases from 500 m to less than 60 m, though the trend is weak; (4) mean interdune length (Fig. 3.17) increases from 1060 m to less than 15 m with a weak trend; (5) mean interdune width (Fig 3.17) increases from 200 m to less than 10 m, though the trend is weak.

3.7.2.2 Interpretation

Data relating to the morphology of aeolian dune bedforms in the vicinity of the Hunkab River exhibit similar trends to that of the dunes in the vicinity of the Hoanib River. However, the crestlines of elongate ridges in the eastern

dune field in the Hunkab area are aligned close to perpendicular to the main flow direction of the short break-through river. Furthermore, the aeolian dunes at the eastern margin are smaller than those in the areas of the Hoanib and Uniab rivers. Additionally, the nature of the flooding events in the Hunkab River is different to those of the Hoanib and Uniab rivers: the Hunkab River floods less frequently due its small catchment area (Krapf et al., 2003).

The aeolian bedforms to the north of the Hunkab River (Fig. 3.14) are dominantly larger dunes than to those to the south of the river. This may be because the dunes to the north have been constructed using a sand supply sourced from fluviially transported and reworked aeolian sediments or fluvial sediments provided by the river. By contrast, small dunes to the south of the river are densely distributed than those to the north. Additionally, the dunes to the south of the river are associated with interdunes that are frequently flooded more than the interdunes to the north of the river. Flooded interdunes to the south of the river are likely influenced by a water-table level above the interdune floor that limits aeolian activity for much of the time. Furthermore, the small dune bedforms and flooded interdunes to the south of the river likely act as sediment source for the northern region of the river during dry seasons.

The lengths and widths of aeolian interdunes are variable to the south of the Hunkab River (Figs 3.15, 3.16 and 3.17); open and wide interdunes are distributed and dispersed in the vicinity of the river with no systematic arrangement. Some of these corridors become isolated during dry seasons because small aeolian dunes migrate across the interdune flats (cf. Maia et al., 2005). By contrast, although interdunes to the north of the river are long, wide and open, access to these open corridors is blocked by a thin belt of reworked aeolian or fluvial sand transported towards NNE by wind storms (Fig. 3.6). Thus, the degree of interconnection of dry interdune corridors to the north of the river is influenced by the wide spaced dune bedforms than those in the south of the river.

3.7.3 Results from Uniab River

3.7.3.1 Description

Floods along the Uniab River break through the entire dune field (Fig. 3.8). Graphs of the morphometric data display predictable and important spatial changes and trends for aeolian dunes and interdunes in the vicinity of the river course (Figs 3.18 – 3.23). The presented data relating to dune and

interdune bedform geomorphology in the vicinity of the Uniab River along transects A, B and C (parallel to wind) and 1, 2, and 3 (perpendicular to the prevailing SSW wind direction) display both weak to strong trends.

Transects A, B and C

To the south (upwind) of the Uniab River the following trends regarding spatial changes in dune and interdune morphology are noted passing from SSW to NNE from the dune-field centre towards the southern river bank: (1) mean dune bedform wavelength (Fig. 3.18) decreases from 400 to less than 15 m; (2) mean wavelength of the along-crest sinuosity of the dunes (Fig 3.19) decreases from 845 m to less than 20 m; (3) mean amplitude of the along-crest sinuosity of the dunes (Fig. 3.20a) decreases 450 m to less than 10 m; (4) mean interdune length (Fig. 3.21) decreases from 1700 m to less than 40 m; (5) mean interdune width (Fig 3.22) decreases from 400 m to less than 10 m, though the trend is weak.

To the north (downwind) of the Hunkab River the following trends regarding spatial changes in dune and interdune morphology are noted passing from SSW to NNE from the northern river bank: (1) mean dune bedform wavelength (Fig. 3.18) decreases from 300 to less than 20 m; (2) mean wavelength of the along-crest sinuosity of the dunes (Fig 3.19) slightly decreases from 570 m to less than 20 m, though the trend is weak; (3) mean amplitude of the along-crest sinuosity of the dunes (Fig. 3.20a) decreases from 400 m to less than 15 m; (4) mean interdune length (Fig. 3.21) increases from 40 m to 1860 m with weak relationship; (5) mean interdune width (Fig 3.22) increases from 10 m to 600 m, though the trend is weak.

Transects 1, 2 and 3

Along transects 1, 2 and 3, which are aligned perpendicular to the prevailing wind direction, graphed results exhibit trends that indicative of non-systematic spatial variation in morphology from the western dune-field margin in an ESE direction towards the eastern dune-field margin. Both aeolian dunes and interdunes show weak or no trend in terms of spatial variability of form (Fig. 3.20 and Fig. 3.23).

To the north of the Uniab River, the following trends regarding spatial changes in dune and interdune morphology are noted passing from WNW to ESE: (1) mean dune bedform wavelength (Fig. 3.20b) decreases from 300 to less than 20 m; (2) mean wavelength of the along-crest sinuosity of the dunes (Fig 3.20c) decreases from 350 m to less than 10 m; (3) mean amplitude of the along-crest sinuosity of the dunes (Fig. 3.20d) slightly

Geometrical Results of Average Dune Wavelength in (Uniab River)

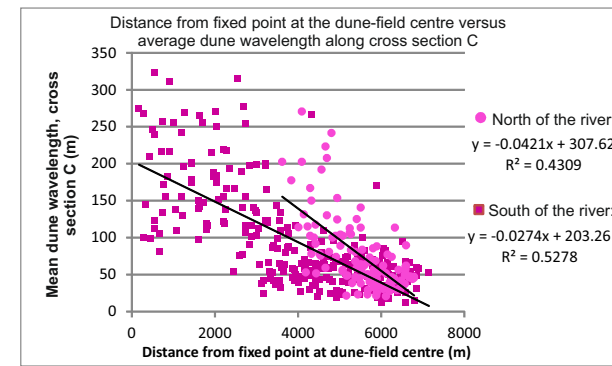
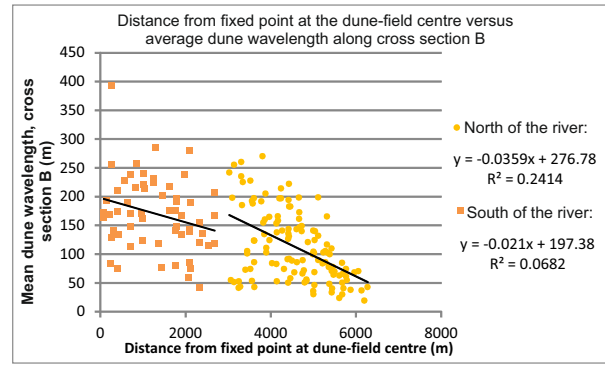
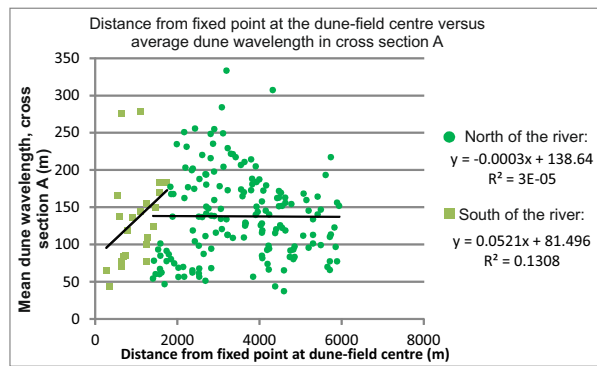
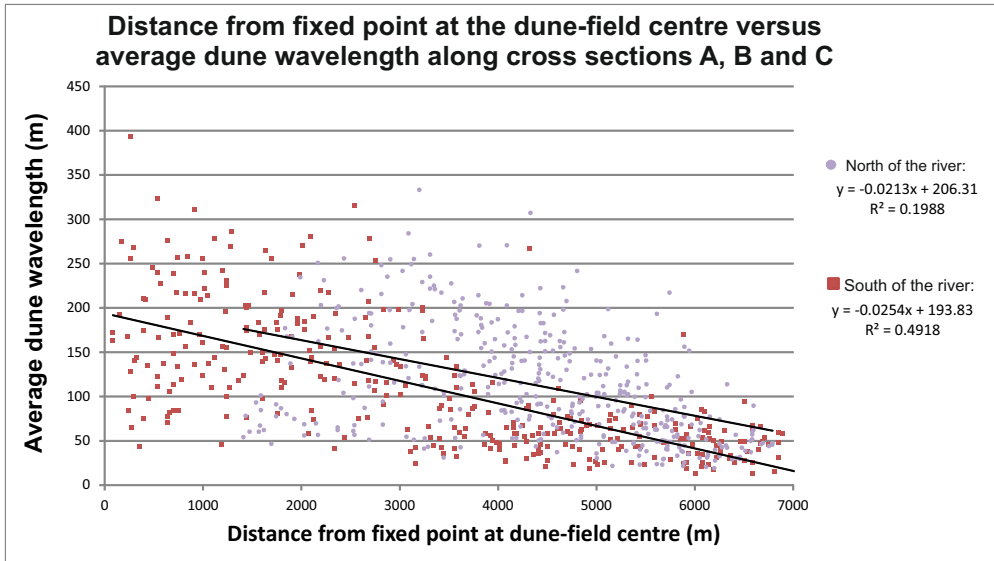
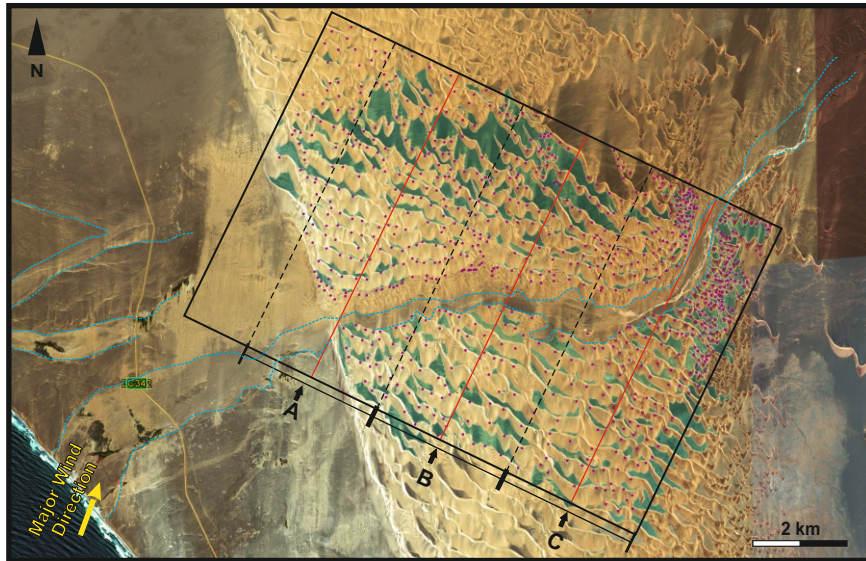


Figure 3.18: Examples of Uniab River data summarizing spatial changes in dune wavelength along cross sections A, B, and C parallel to the prevailing wind direction towards the NNE.

Geometrical Results of Wavelength of along-Crest Sinuosity of Dune in (Uniab River)

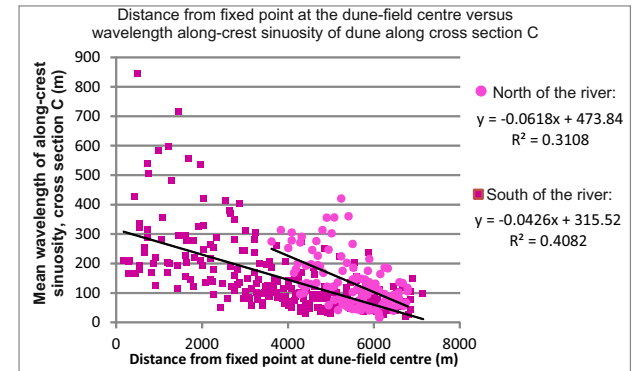
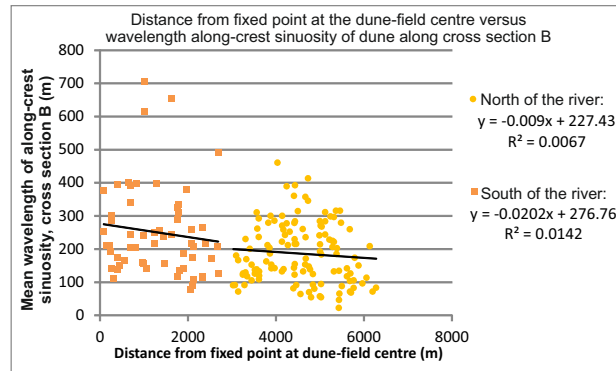
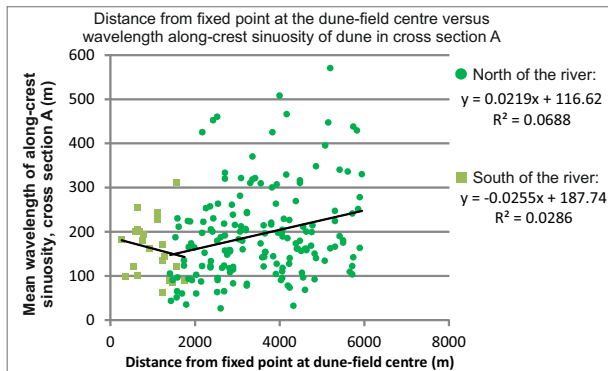
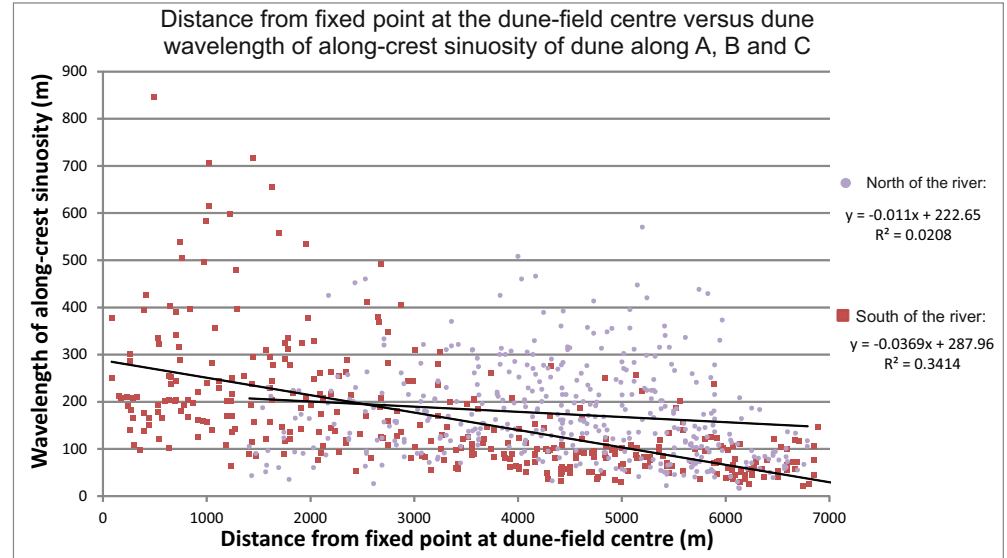
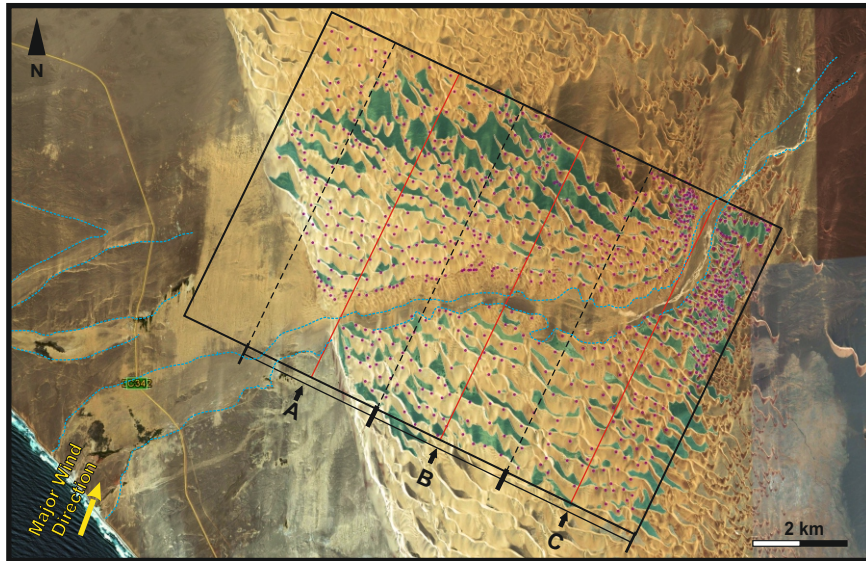
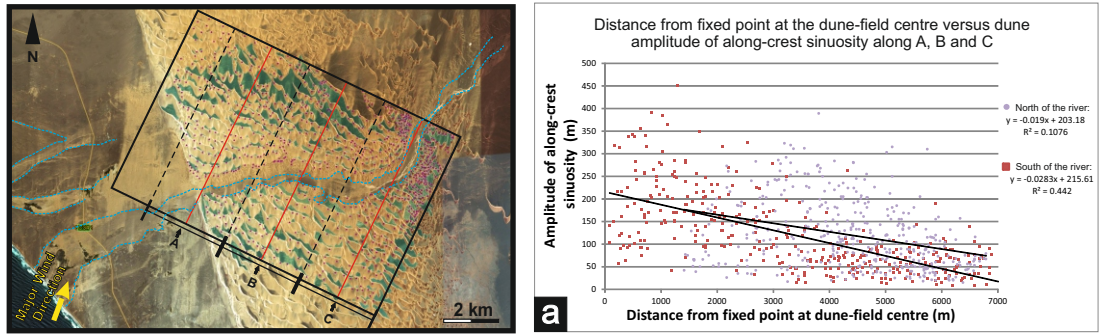


Figure 3.19: Examples of Uniab River data summarizing spatial changes in the wavelength along-crest sinuosity along cross sections A, B, and C parallel to the prevailing wind direction towards the NNE.

Geometrical Results of Aeolian Dunes in (Uniab River)

Geometrical results amplitude of along-crest sinuosity of dune along cross sections of A, B and C



Geometrical results of aeolian dunes along cross sections of 1, 2 and 3

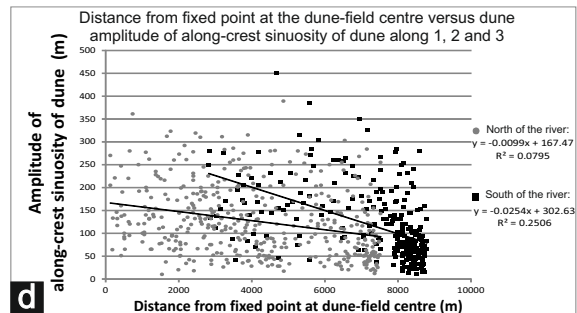
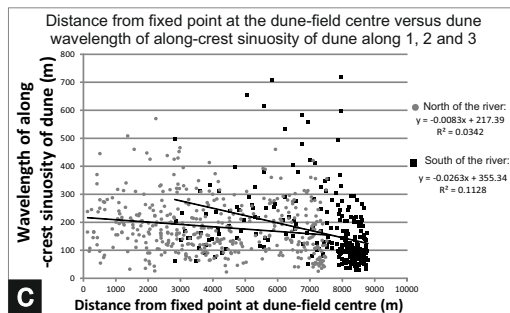
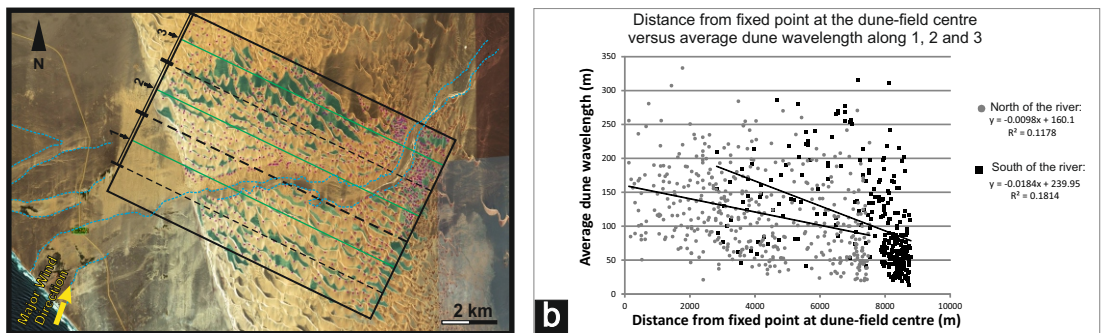


Figure 3.20: Examples of Uniab River data summarizing spatial changes in aeolian dunes along cross sections parallel (A, B, and C) and perpendicular to the prevailing wind direction (1, 2 and 3): **(a)** Amplitude along-crest sinuosity along sections A, B and C towards the NNE. **(b)** Average dune wavelength along cross sections 1, 2 and 3 towards the SSE. **(c)** Wavelength along-crest sinuosity along sections 1, 2 and 3 towards the SSE. **(d)** Amplitude along-crest sinuosity along sections 1, 2 and 3 towards the SSE.

Geometrical Results of Aeolian Interdune Lengths Along Cross Sections A, B and C in (Uniab River)

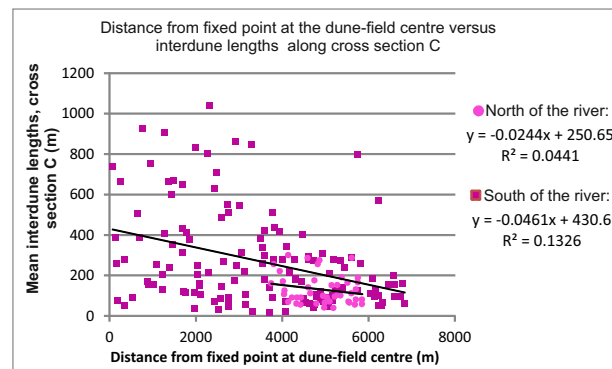
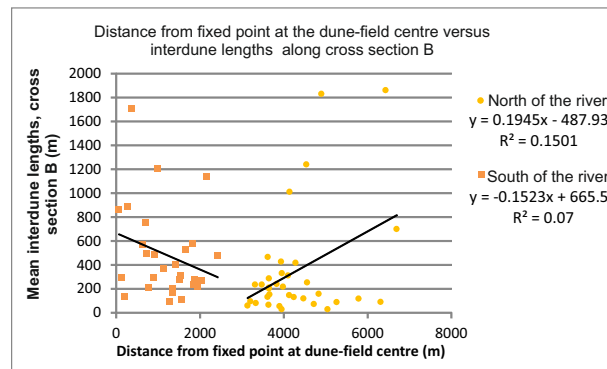
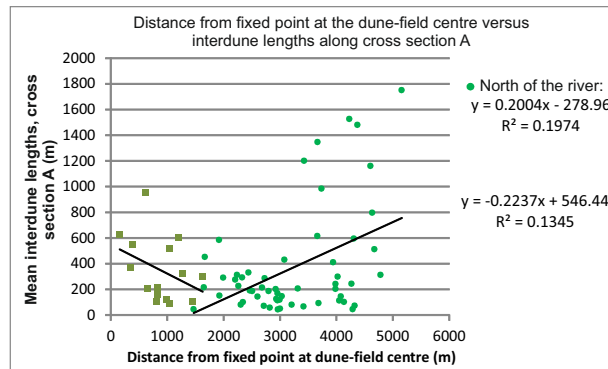
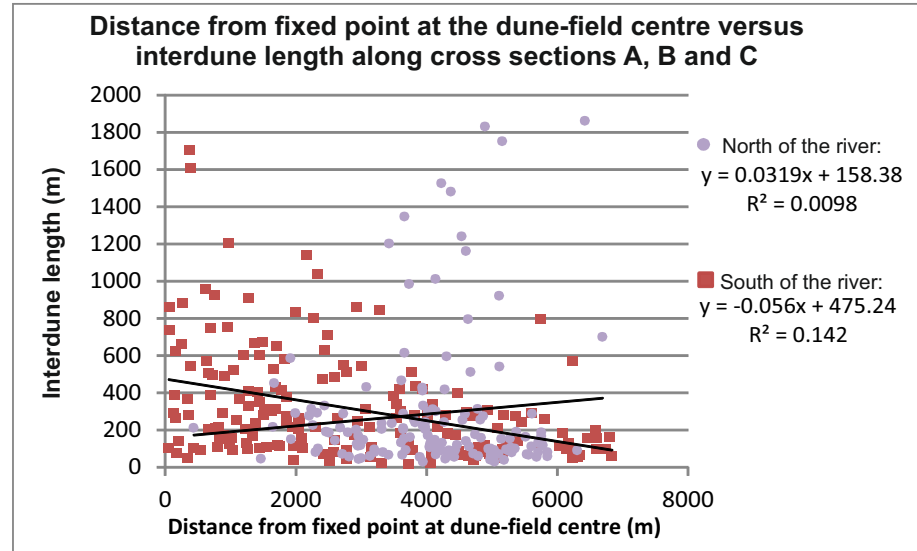
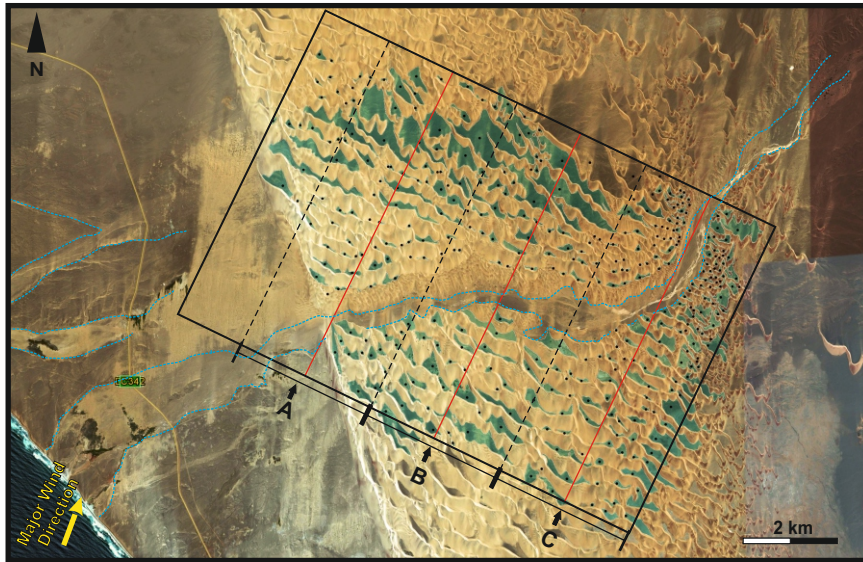


Figure 3.21: Examples of Uniab River data summarizing spatial changes in aeolian interdune lengths along cross sections A, B, and C parallel to the prevailing wind direction towards the NNE.

Geometrical Results of Aeolian Interdune Widths Along Cross Sections A, B and C in (Uniab River)

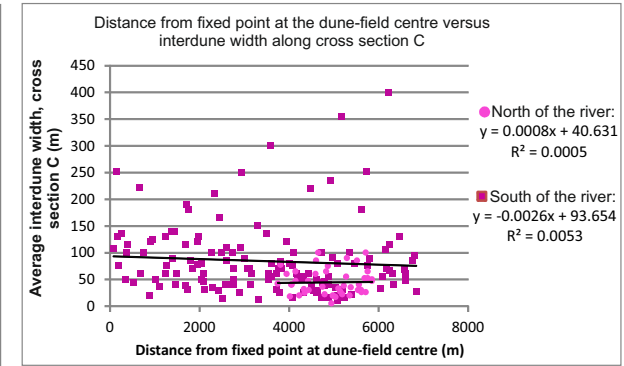
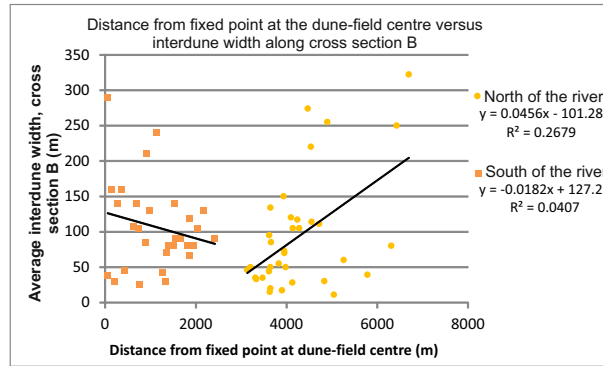
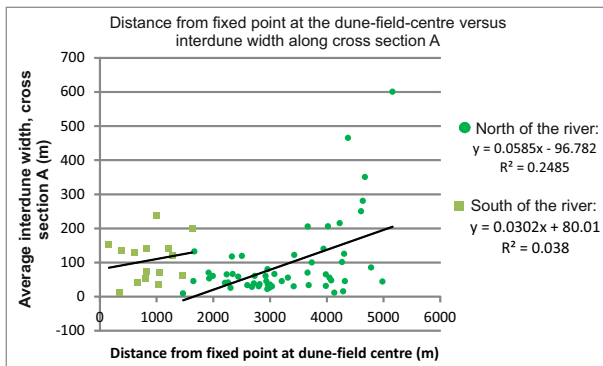
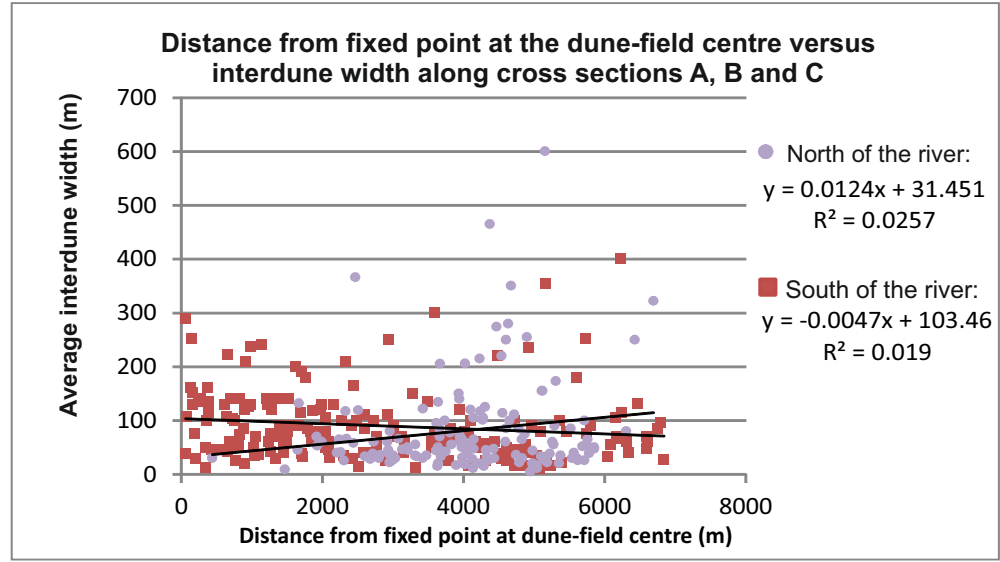
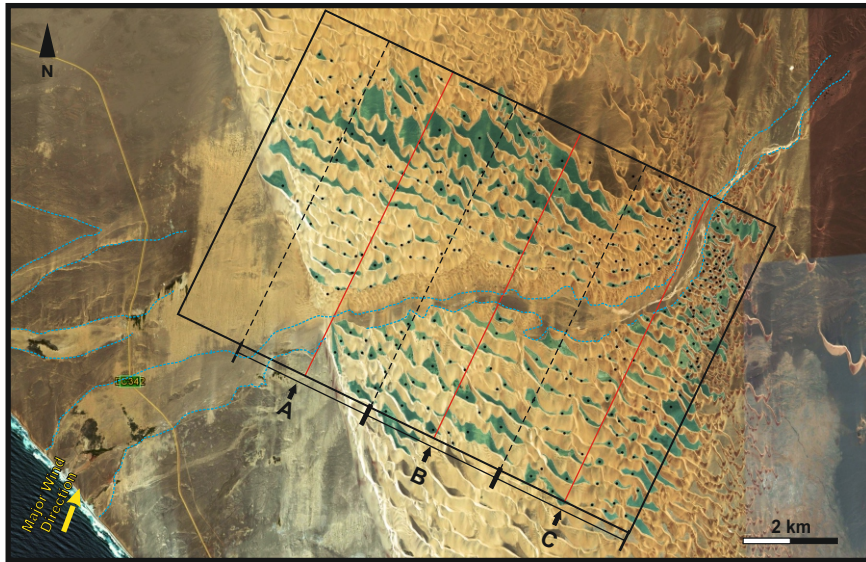


Figure 3.22: Examples of Uniab River data summarizing spatial changes in aeolian interdune widths along cross sections A, B, and C parallel to the prevailing wind direction towards the NNE.

Geometrical Results of Aeolian Interdune Lengths and Widths Along Cross Sections 1, 2 and 3 in (Uniab River)

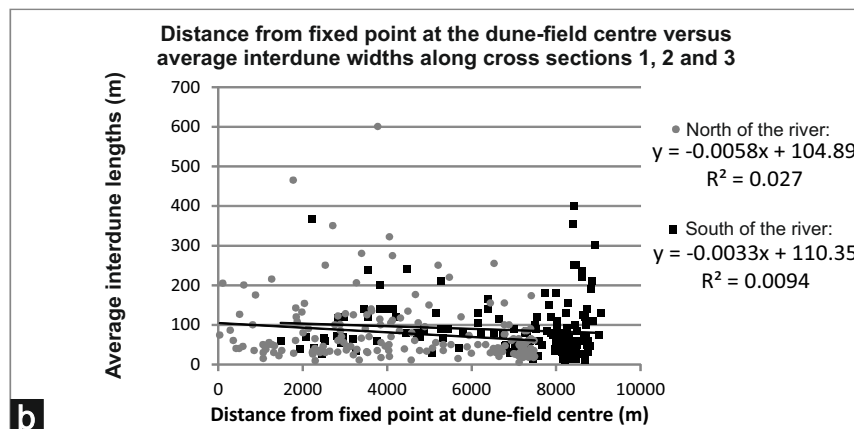
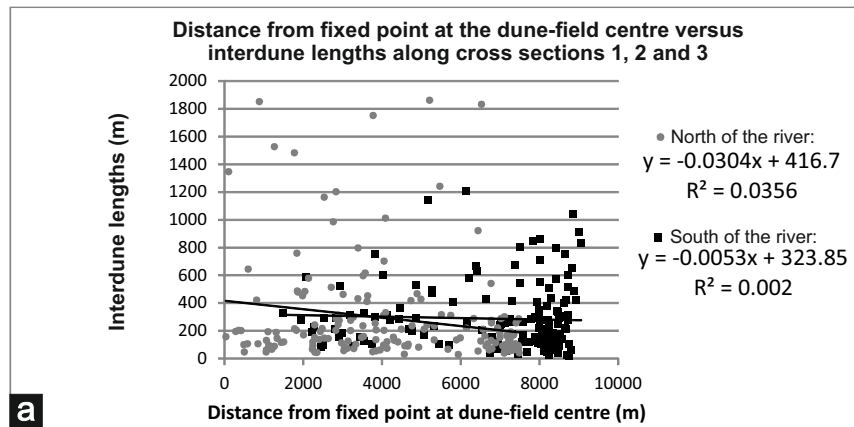
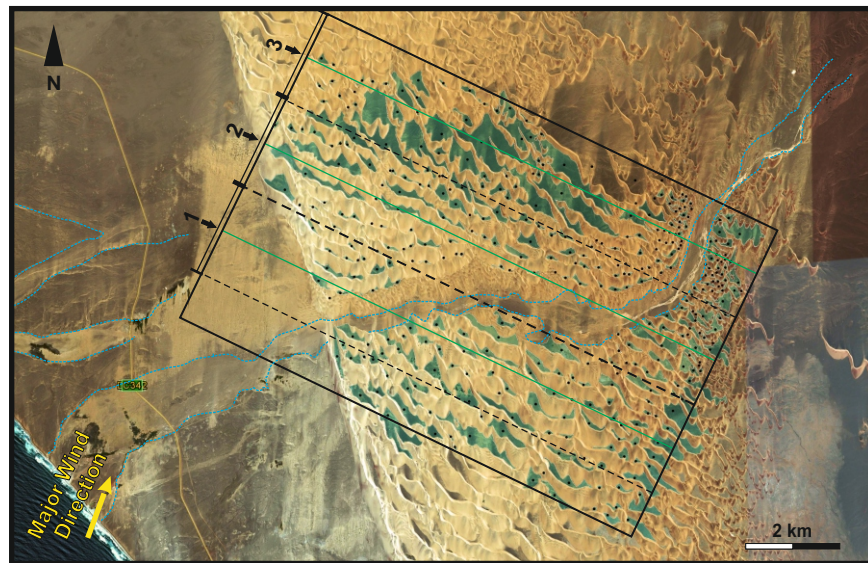


Figure 3.23: Examples of Uniab River data summarizing spatial changes in aeolian interdune lengths and widths along cross sections 1, 2, and 3 perpendicular to the prevailing wind direction towards the SSE.

decreases from 400 m to less than 15 m, though the trend is weak; (4) mean interdune length (Fig. 3.23a) is scattered and data ranges between 1800 m to less than 30 m with no relationship; (5) mean interdune width (Fig 3.23b) is scattered and data varies between 600 m to less than 10 m with no relationship.

To the south of the Uniab River, the following trends regarding spatial changes in dune and interdune morphology are noted passing from WNW to ESE: (1) mean dune bedform wavelength (Fig. 3.20b) decreases from 300 to less than 15 m; (2) mean wavelength of the along-crest sinuosity of the dunes (Fig 3.20c) decreases from 700 m to less than 30 m with a weak trend; (3) mean amplitude of the along-crest sinuosity of the dunes (Fig. 3.20d) slightly decreases from 450 m to less than 15 m, though the trend is weak; (4) mean interdune length (Fig. 3.23a) is scattered and data ranges between 1200 m to less than 15 m with no relationship; (5) mean interdune width (Fig 3.23b) is scattered and data varies between 400 m to less than 12 m with no relationship.

3.7.3.2 Interpretation

Results of analysis of spatial changes in dune and interdune morphology in the vicinity of the Uniab River demonstrate different spatial changes to the south and north of the river. The decreasing trends of dune bedforms in the figure (3.18, 3.19 and 3.20a) imply that a decrease in the sediment supply in a direction towards NNE. Dunes adjacent next to the Uniab river are smaller due to frequent floods into the adjoining interdune corridors (i.e. interdune expansions). The wet interdune corridors in this vicinity will reduce the availability of sediment for aeolian system construction.

The morphology of the interdune corridors to the south of the river, which are oriented parallel to the dune crestlines, is largely controlled by the spacing and the shape (morphology) of adjoining dunes (Lancaster, 1988). The spatial arrangements of narrow interdunes between barchanoid dunes to the south of the river allows flood waters to access open interdune corridors such that the interdunes become wider towards the eastern dune-field margin where simple barchans dunes are the dominant type. The orientation of the dune ridges oriented which is 35° - 40° from to the main river fairway (Figs 3.21 and 3.22) will limit the accessibility of flood waters into the adjoining interdune corridors. However, the flood waters are able to enter into adjoining interdunes in the area to the south of the river more easily at the eastern erg margin because the interdunes here are open and oriented close to perpendicular to the river course.

The area to the south of the Uniab River acts as a site of sediment storage for later dune construction in the area to the north of the river (downwind) such that sediments of reworked aeolian and fluvial origin can be transported to the NNE during dry seasons. The accumulation of aeolian sediments in the northern part of the river (Fig. 3.9c) prevent the lateral expansion of the river.

The observed large size of interdunes to the north of the river (Fig. 3.8) imply that either a high water table or flushed aeolian bedforms by previous major floods (cf. Krapf et al., 2003).

3.8 Discussion

The observed types of fluvial-aeolian interactions in this study, which are variable in their styles, allow the development of a discussion that considers the controlling factors that govern such types of interactions and the construction of a series of depositional models that can be used to predict the expected stratigraphic architecture of the preserved record of mixed fluvial and aeolian depositional systems (Fig. 3.24).

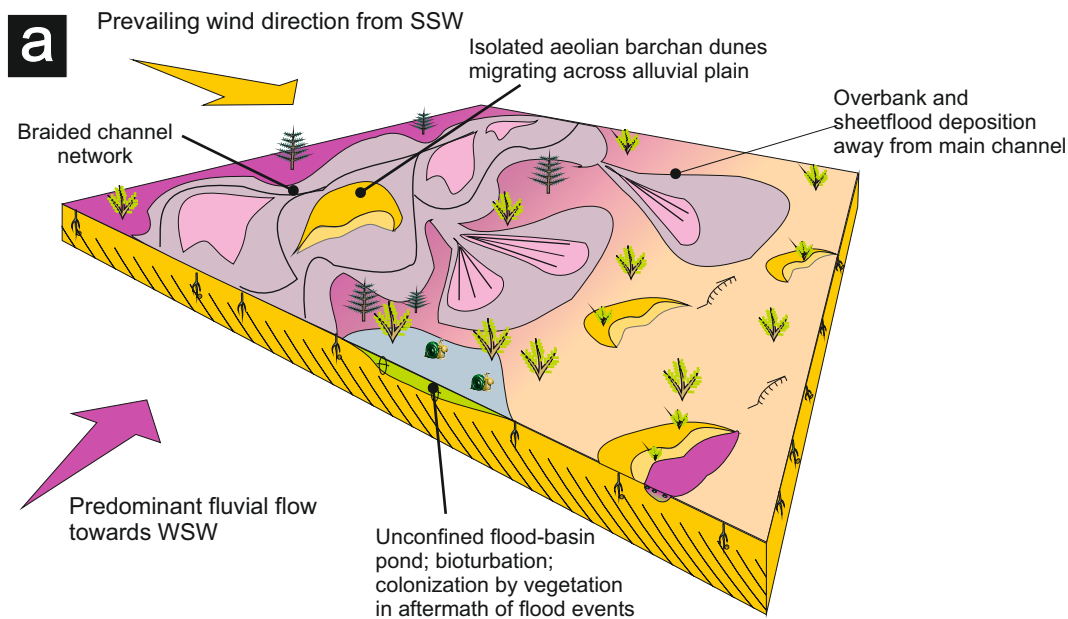
3.8.1 Dryland Rivers

Dryland rivers occur in regions influenced by arid or semi-arid climatic regimes. Dryland rivers are characterised by net surface water deficits and make up to 50% of the present global land surface (Thomas, 1997; Tooth, 2000b; Nanson et al., 2002). Drylands are characterised by low ratios between precipitation and potential evapotranspiration, by ephemeral and intermittent river discharge and by sparse, unevenly distributed and often temporally variable vegetation cover (Bullard, 1997; Tooth 2000b). Dryland rivers act as land-shaping agents over time (Frostick & Reid, 1987; Reid & Frostick, 1997). Observations from the ephemeral rivers of the Skeleton Coast of Namibia studied here indicate that fluvial processes are significant agents of both erosion and deposition.

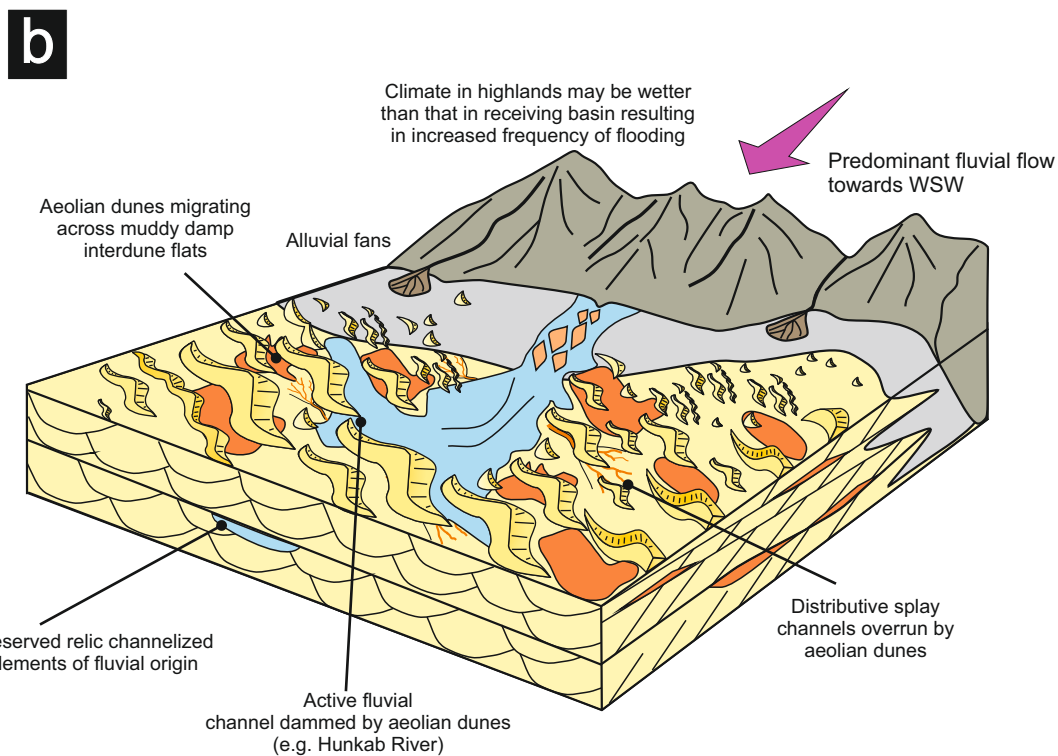
Shifts in climate act as a primary control on the relative dominance of fluvial over aeolian processes in mixed depositional desert settings (Clemmensen et al., 1994); this influences the location and nature of aeolian dune-field boundaries (Fig. 3.24a and 3.24b), which may expand and contract over time (e.g., Herries, 1993; Yang and Ding, 2013; Al-Masrahy & Mounthey, 2015).

High-energy flood events with high rates of fluvial discharge into aeolian dune-field margins increase the potential for erosion of existing aeolian

Summary Facies Models of the Skeleton Coast Erg



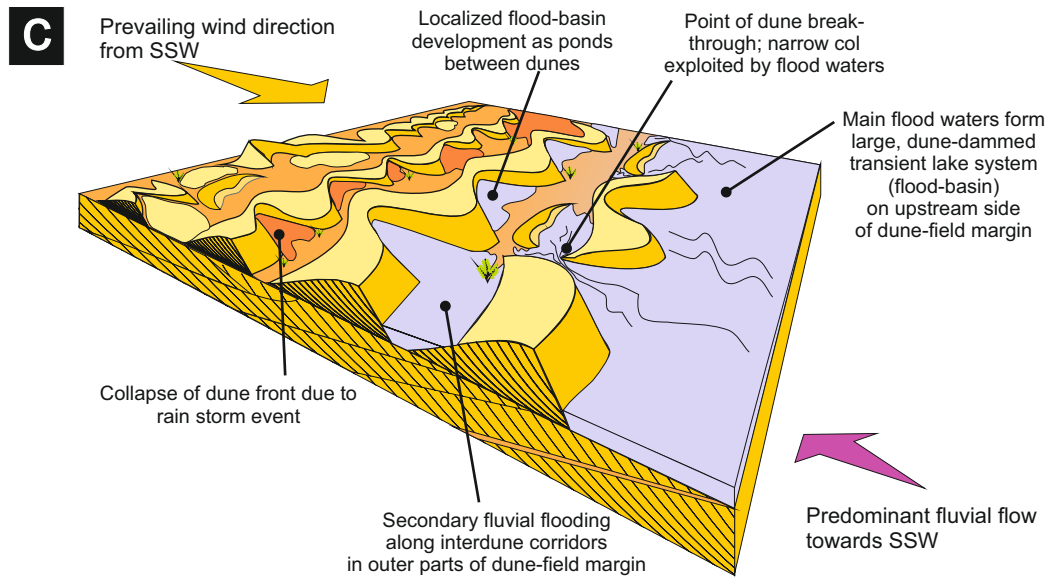
Major fluvial incursion over a low-relief, deflationary supersurface in an outer dune-field-margin setting. The level of the accumulation surface is typically determined by the capillary fringe of the ground water table. Modified in part from Stanistreet and Stollhofen (2002) and Jagger (2003). No particular scale implied.



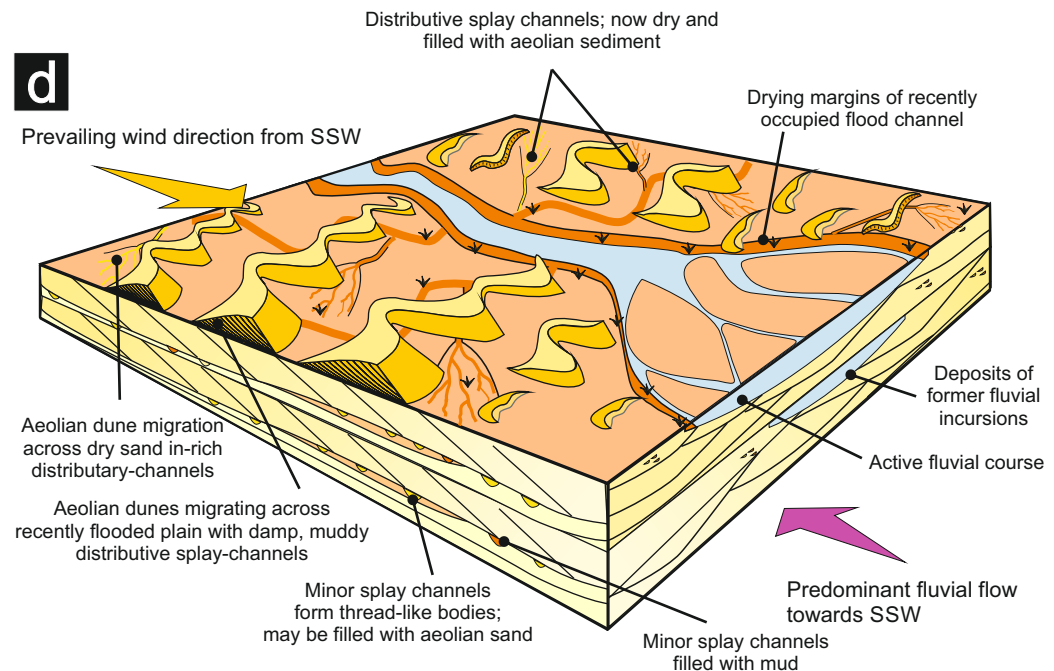
Major fluvial incursion into the marginal part of an aeolian dune field for which flood waters emerge from a mountain-front catchment. Aeolian dune bedforms act as damming ridges and migrate over former damp-interdunes due to climate change from wet to dry conditions in the aftermath of a major flood event. No particular scale implied.

Figure 3.24: Summary facies models showing the different types of interactions from the studied rivers. No particular scale implied.

Summary Facies Models of the Skeleton Coast Erg



Major fluvial flood prevented from entering dune field by continuous dune front with crestline oriented perpendicular to preferred direction of fluvial flow. Modified in part from Stanistreet & Stollhofen (2002) and Jagger (2003). No particular scale implied.



Aeolian dune bedforms migrating across a dry and damp accumulation surface in the aftermath of a major flood event; distributive splay channels become preserved; many are filled with aeolian sand deposits. No particular scale implied.

Figure 3.24: Continued.

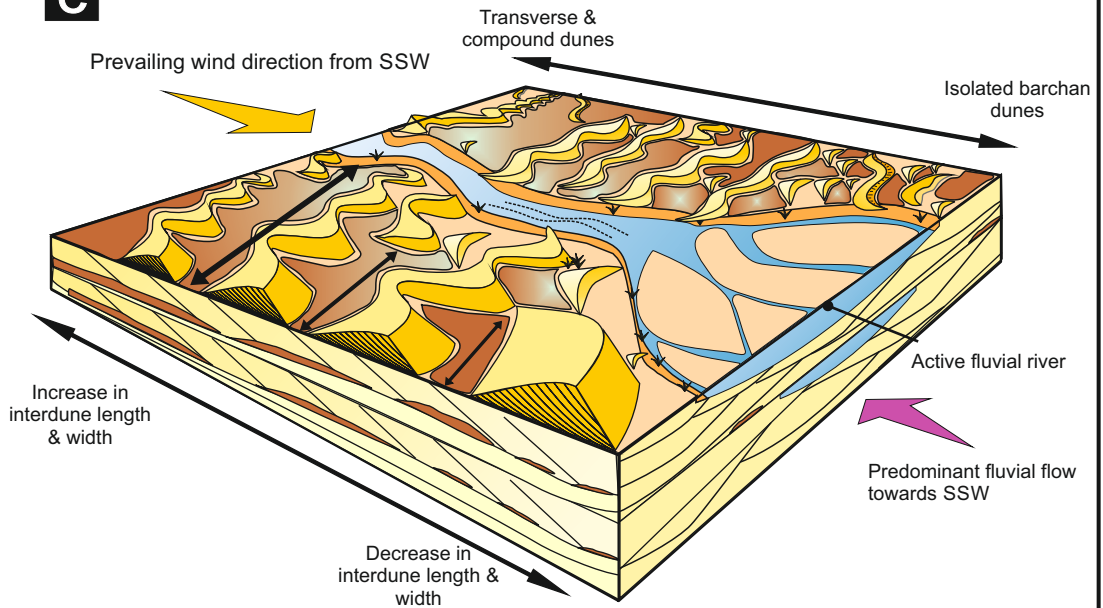
deposits (Fig. 3.24d). Under such conditions, erosion, transport and re-deposition of fluviually reworked aeolian sediments may occur where fluvial floods transport aeolian sediments downstream to areas where floods discharge wanes or terminates (Fig. 3.24c). This will influence the supply and availability of sediment for aeolian construction in dune-field settings. In areas where fluvial floods are frequent at dune-field margins, associated water-table rise may lead to the establishment of damp or wet interdune flats and the availability of aeolian sediment for dune construction will decrease. Additionally, dune migration may be curtailed by the presence of such damp interdunes, resulting in the stabilisation of both interdune corridors and adjoining dunes (Fig. 3.24c). A high water table may promote the establishment of surface stabilizing agents such as vegetation or chemical surface crusts (e.g. calcrete or gypcrete). Substrate erosion in these flooded interdune corridors may be limited by the presence of such stabilizing agents and deposition of mud layers from suspension load may occur in the aftermath of repeated flood events the penetrate into interdune corridors (Stanistreet & Stollhofen, 2002). Alternatively, damp interdune areas that lack mud layers may also develop where water table or its capillary fringe is close or at the interdune floor limiting the availability of sediment for aeolian construction (Kocurek, 1999).

The size of fluvial catchment areas is controlled by regional physiography. Rivers with larger catchments (e.g. Hoanib) tend to have a greater propensity for dune break-through into dune-field margins during major flood events. These catchments tend to generate a larger sediment supply, which may later be potentially suitable for aeolian construction (Kocurek, 1999). During relatively humid phases, the sediment availability in flooded aeolian systems is controlled by the water-table level relative to the depositional surface such that a relatively high water table will limit sediment availability enabling interdune expansion at the expense of aeolian dunes (e.g. interdunes in the south of the rivers Hunkab and Uniab) (cf. Stanistreet & Stollhofen, 2002; Mountney & Jagger, 2004). By contrast, sediment availability will be increased if the water table is relatively low, thereby promoting aeolian dune growth, possibly to a size where neighbouring interdunes become small and isolated elliptical depressions between large and semi-connected dunes (Wilson, 1971, 1973; Mountney & Jagger, 2004).

3.8.2 The Relationship between Aeolian Dune Morphology and Fluvial system

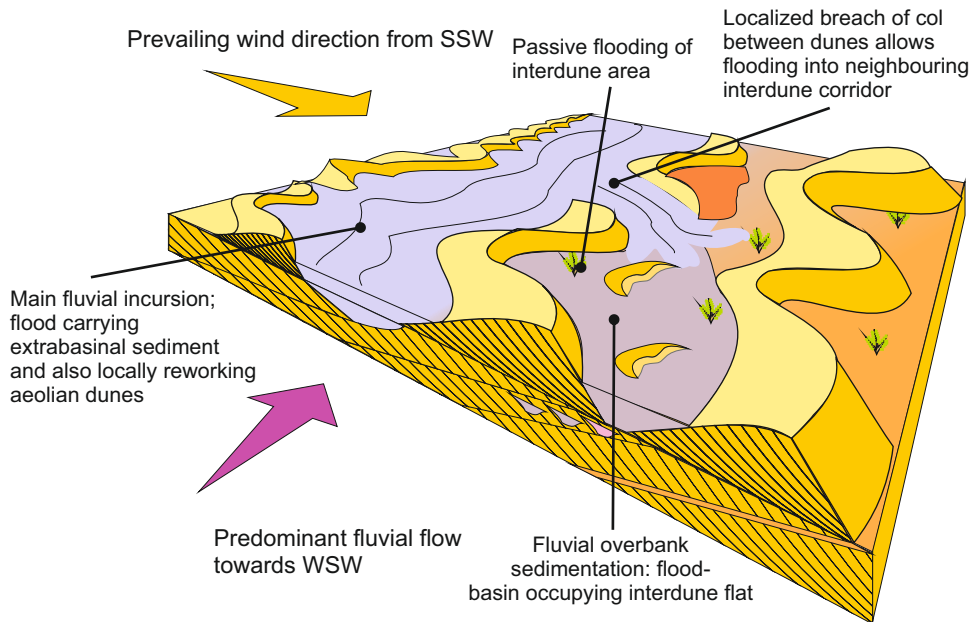
Summary Facies Models of the Skeleton Coast Erg

e



Major fluvial incursion into an aeolian dune field for which interdune corridors increase in length and width with increasing proximity to the upwind dune-field margin. Aeolian dune types vary from isolated barchan dunes to transverse dunes. No particular scale implied.

f



Major fluvial incursion into the central part of a dune field for which open interdune corridors act as major conduits for stream systems. Aeolian dune migration and construction may be on-going between flood events. Modified in part from Stanistreet and Stollhofen (2002) and Jagger (2003). No particular scale implied.

Figure 3.24: Continued.

Aeolian dune morphology within the studied region controls the shape of the adjoining interdunes. Fluvial systems are either able to penetrate along open interdune corridors (Fig. 3.24d and 3.24f) or are dammed by dunes that surround closed interdunes (Fig. 3.24c). The nature of interdune flooding observed in this study is determined by morphological dune type, the orientation of dune bedforms relative to the main fluvial flow direction, the geometric nature of interdune corridors which are defined in terms of their width and length, and the rate of aeolian dune and interdune migration relative to the flood frequency, as well as by the nature of the flood events themselves (cf. Langford & Chan, 1989; Al-Masrahy & Mountney, 2015).

Sinuuous-crested aeolian dune forms that commonly undertake high rates of bedform migration tend to be associated with the development of irregular and spatially isolated (i.e. closed) interdunes (Fig. 3.5c and 3.9c). By contrast, linear and straight-crested (or only slightly sinuous-crested) transverse or oblique dunes that commonly undertake low rates of dune migration tend to be associated with the development of elongate (i.e. open) interdune corridors (Fig. 3.24e). The form of the developed interdunes (open versus closed) exerts a primary control on potential fluvial penetration distance.

Aeolian dunes have greater capability to dam ephemeral rivers in cases where fluvial flow is oriented perpendicular to the trend of aeolian dune crestlines (Fig. 3.24c), whereas fluvial systems have a greater potential to penetrate along interdune corridors where aeolian dunes have crestlines that are aligned parallel to the fluvial flow direction (e.g. the north region of Uniab River, Fig. 3.24f).

High-intensity rainfall events that drain catchments with high gradients tend to favour catastrophic floods that are more effective at breaking through dune barriers, even where relatively large dune wall are established. For example, in the case of flood waters typically initially build up behind relatively large dunes for few days forming a large flood basin until the water level within this basin overtops the dune barrier at one or a series of low-points or cols (e.g., Krapf et al., 2003; Stanistreet and Stollhofen, 2002).

At the eastern dune-field margin, the river undertakes a marked change in form at the point where it intersects a significant number of stabilised aeolian dunes. Distributary channels (Fig. 3.5a and 3.5d) that wind their way between the vegetated dunes tend to develop in this vicinity. The presence of increased vegetation cover on dunes at the eastern dune-field margin of

Hoanib acts to limit the availability of aeolian sand that may be considered susceptible to re-working and transportation by floods.

3.8.3 Long-term preservation potential

Long-term preservation of accumulated deposits arising as a consequence of fluvial-aeolian interactions that have occurred for protracted episodes typically requires an episode of subsidence whereby the base-level increases in a relative sense over time (Kocurek & Havholm, 1993).

In wet aeolian systems, the preservation commonly occur where active aeolian dunes migrate alongside fluvially flooded interdune corridors such that frequent fluvial flooding tend to result in the temporary cessation of aeolian dune migration. In this case, water table may locally expand to cover large areas of flooded interdune corridors in front of advancing aeolian dunes (Simpson & Loope, 1985; Langford and Chan, 1988; Mountney, 2012). In erg-margins, the migration of aeolian dune bedforms associated with the repeated flooding of interdunes corridors favours the generation of bypass supersurfaces (so-called flood surfaces of Langford and Chan, 1988).

Intervening relationship between the upper surfaces of damp interdune areas and the adjoining aeolian-dune toesets can be developed in erg margins when dune encroachment over interdunes occur. The lateral continuation of damp or wet interdune areas is controlled by long-term fluctuations in water-table level over time in response to intermittent episodes of heightened ephemeral flash-flood events and changes in aeolian dune migration rate (Kocurek, 1981b, 1999; Pulvertaft, 1985; Mountney & Thompson, 2002). The sedimentary expression in this relationship is common in the ancient record as the onlap of fluvial or damp or wet interdune strata onto the lowermost flanks of the aeolian dunes (Pulvertaft, 1985; Mountney & Thompson, 2002; Cain & Mountney, 2009, 2011).

Where the dune-field margin is affected fluvial floods that penetrate into the erg margin with a higher frequency (e.g. Hoanib and Uniab), longer-term changes in water-table level may occur. The water table will respond to regional climate change such that the water-table rises during humid phases and falls during arid phases. The water table or its near-surface capillary fringe play an important role in promoting long-term preservation (Mountney, 2012; cf. Kocurek & Havholm, 1993).

The long-term preservation of aeolian accumulations requires the aeolian deposits to be placed below some regional baseline where erosion does not take place. This can be achieved by either, subsidence through a static water table or an absolute rise in water table level (Kocurek and Havholm, 1993; Bristow & Mountney, 2013; Rodríguez-López et al., 2014).

3.9 Conclusions

The following conclusions have arisen from this study.

1. The analysis of the remotely sensed images reveals the history of the Hoanib, Hunkab and Uniab rivers at locations where they interact with the aeolian erg, thereby yielding a valuable record of interaction between competing fluvial and aeolian processes.
2. The nature of aeolian dune-field boundaries, which may expand and contract over time is influenced by climatic shifts that act as a primary control on the relative dominance of fluvial over aeolian processes in mixed depositional desert settings.
3. High-energy flood events with high rates of fluvial discharge into aeolian dune-field margins increase the potential for erosion of existing aeolian deposits.
4. Under such conditions, erosion, transport and re-deposition of fluvially reworked aeolian sediments may occur where fluvial floods transport aeolian sediments downstream to areas where floods discharge wanes or terminates (Fig. 3.24c).
5. The supply and availability of sediment for aeolian construction in dune-field settings are determined by the frequency of erosion, transport and re-deposition of fluvially reworked aeolian sediments that may occur where fluvial floods transport aeolian sediments downstream to areas where floods discharge wanes or terminates.
6. The presence of increased vegetation cover on dunes at the eastern dune-field margin (e.g. Hoanib River) acts to limit the availability of aeolian sand that may be considered susceptible to re-working and transportation by floods.
7. Aeolian dune morphology within the studied region controls the shape of the adjoining interdunes. Fluvial systems are either able to penetrate along open interdune corridors (Fig. 3.24d and 3.24f) or are dammed by dunes that surround closed interdunes (Fig. 3.24c).
8. The nature of interdune flooding observed in this study is determined by morphological dune type, the orientation of dune bedforms relative to the

main fluvial flow direction, the geometric nature of interdune corridors which are defined in terms of their width and length, and the rate of aeolian dune and interdune migration relative to the flood frequency, as well as by the nature of the flood events themselves.

9. This study is important because it serves as the basis for a series of predictive models underpinned by observed relationships that can be useful in predicting aeolian high-quality reservoirs. The outcome of this work is the construction of a series of qualitative facies models that can be used to assess likely subsurface reservoir prospectively in mixed fluvial-aeolian successions that range from aeolian-dominated inner erg-margin settings to fluvial dominated outer erg-margin settings.

Chapter 4

Preserved sedimentary record of fluvial-aeolian interaction: Triassic Helsby Sandstone Formation, Cheshire Basin, UK

4.1 Summary

Outcrop analysis of the Helsby Sandstone Formation (Cheshire Basin) reveals the preserved stratigraphic expression of several types of ancient fluvial-aeolian interactions: (i) relationships indicative of systematic temporal change from an aeolian dune field characterized by small, isolated dry interdunes to one in which interdunes were large and interconnected such that they acted as conduits for fluvial flow whereby fluvial channels were able to penetrate into dune-field centre settings; (ii) evidence for fluvial reworking of aeolian dune deposits by erosive flows that resulted in temporary cessation in dune migration in the immediate aftermath of flood events. Observations from analysis of outcrop data have enabled lithofacies interpretations to be made and these have been combined with other published and unpublished data collected previously from the region of study to produce a series of predictive sedimentological models of the inferred palaeoenvironment and its preserved stratigraphic expression. The range of facies types present in deposits of the Helsby Sandstone Formation is indicative of a record of the accumulation of aeolian and fluvial sediment systems under the influence of an arid to semi-arid climate. In this chapter, the detailed interpretation of this ancient succession allows a more general discussion of the main forcing factors responsible for determining processes and facies distributions (both laterally and vertically) within both aeolian and fluvial systems more generally. This study serves to further develop our understanding of processes that operate in mixed fluvial and aeolian systems in arid and semi-arid environmental settings through the development of a series of sophisticated predictive facies models that can additionally be applied as tools for estimating the likely distribution of high-quality in porosity and permeability zones in analogous subsurface reservoirs within this basin or other similar basins.

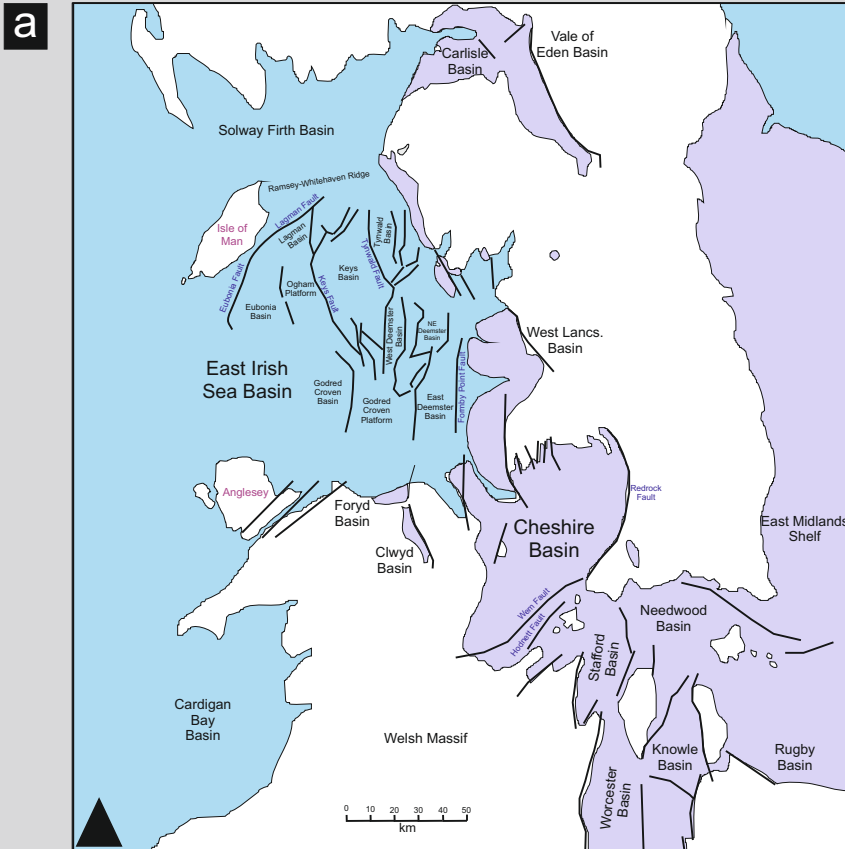
4.2 Introduction

Fluvial-dominated and fluvial-influenced aeolian dune-field (erg) margins are characterized by combinations of water-controlled and water-influenced

deposits that develop in response to a varied range of aeolian and fluvial transport processes. The nature of these controlling processes and the distribution of resultant preserved facies varies both 1) spatially from the erg centre to its margin, as a result of changes in system dominance, and 2) temporally, as a result of the influence of both intrinsic autogenic factors, such as switching of river courses, and extrinsic allongenic factors such as climate change where fluctuations between relatively more arid and relatively more humid conditions govern depositional conditions and the resultant form of the stratigraphic accumulations over longer episodes. The recognition of the preserved deposits of such systems in ancient outcropping successions requires detailed observation and analysis of lithofacies associations and architectural-element distributions to demonstrate evidence for the following: 1) widely distributed fluvial-incision into previously accumulated aeolian deposits at a specific stratigraphic level, indicating that the interaction between the two systems was temporally distinct; 2) abundant lateral facies-change from damp or wet aeolian-interdune facies, to dry aeolian dune and dry interdune facies, in a style indicative of aeolian dune migration that was contemporaneous with on-going changes in water-table level; 3) long-lived establishment and development of damp and fluvially flooded interdune corridors and their preservation as extended sheet-like bodies as a result of water-table fluctuations associated with frequent and repeated fluvial floods along interdune corridors.

The aim of this study is to examine a series of well-exposed successions of mixed aeolian and fluvial origin in the lower Thrustaston Member and middle Delamere Member of the Helsby Sandstone Formation of the Cheshire Basin (Fig. 4.1 and 4.2) to investigate the intimate relationship between a water-table controlled aeolian succession and a dryland fluvial succession, accumulation and preservation of which was influenced by fluvial flood events in a marginal aeolian dune field (erg) setting. Specific objectives of this study are as follows: 1) to describe and interpret the sedimentary facies of both ancient fluvial and aeolian deposits present in an ancient preserved erg-margin succession; 2) to discuss the nature of the interaction between aeolian, fluvial and floodplain sub-environments within the study area; 3) to integrate field-based observations made as part of this study with observations from earlier discussions and publications by others to further develop an understanding of the distribution of occurrence and lateral extent of lithofacies and architectural elements – and their spatial and temporal variations – in a fluvial-influenced erg-margin depositional system (cf. Mounthey & Thompson, 2002; Herries, 1993; Kocurek, 1981a; Langford &

Distribution of major Permo-Triassic basins in Central and North West England and the East Irish Sea



Stratigraphic subdivision of the Sherwood Sandstone Group (SSG) in the Cheshire Basin

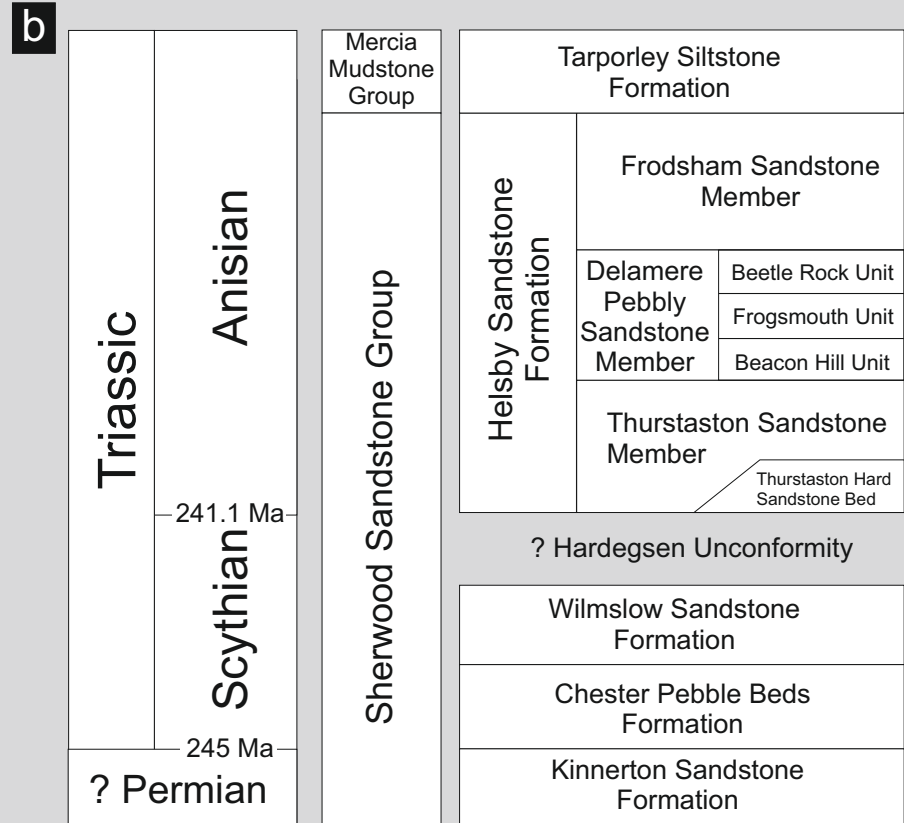


Figure 4.1: a. Central, NW England and the East Irish Sea basins (after Coward, 1995). b. Permo-Triassic stratigraphy of the northern Cheshire basin (after Harland *et al.*, 1990; Mountney & Thompson, 2002).

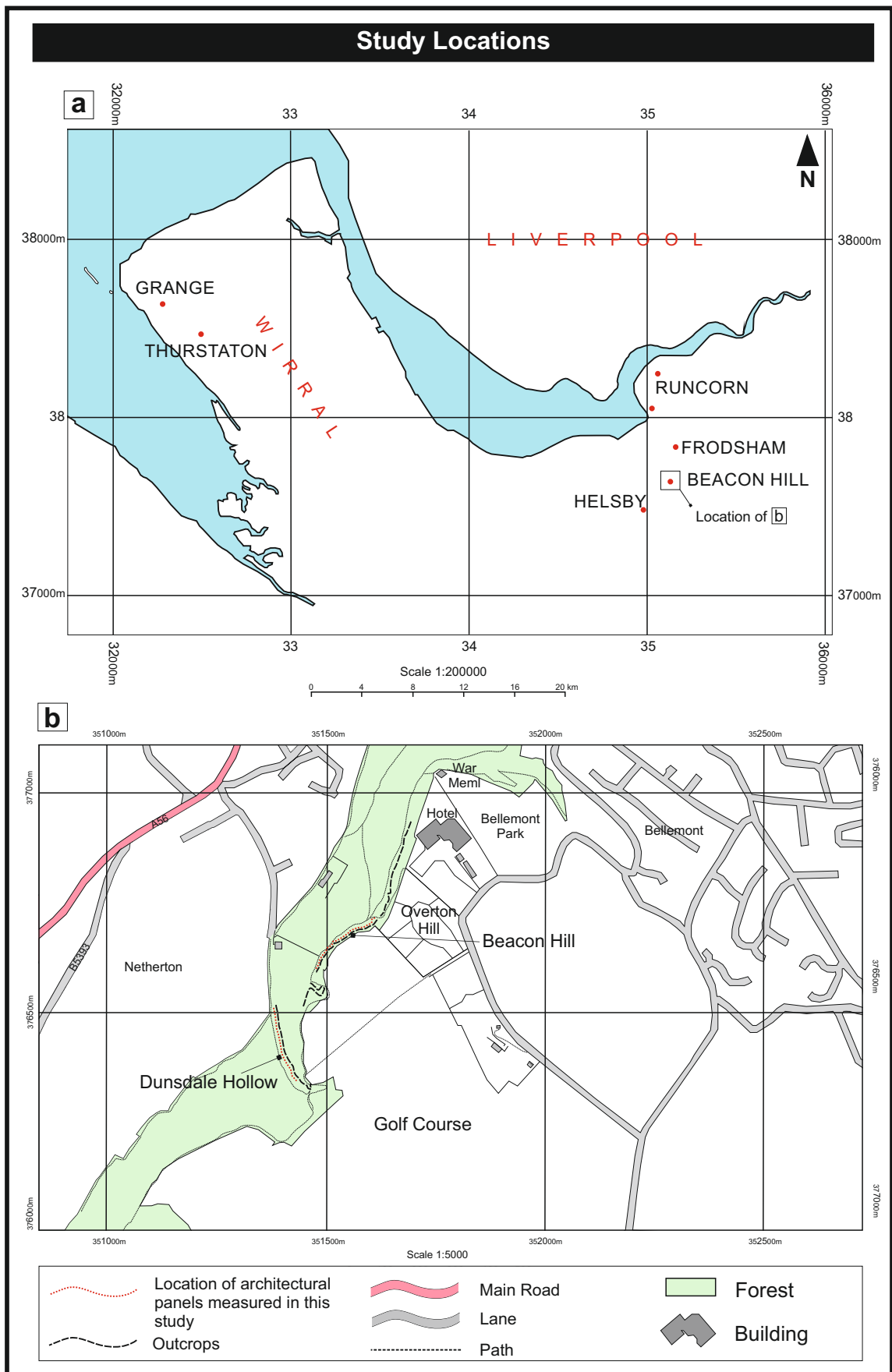


Figure 4.2: a. Location of studied Helsby Sandstone Formation outcrops within the Cheshire Basin in NW of England, UK (red points); b. Locations of detailed architectural panels from Beacon Hill and Dunsdale Hollow, Frodsham Town are shown in figures 4.19 & 4.20.

Chan, 1988; Mountney & Russell, 2006; Ahlbrandt et al., 1980; Rodriguez-Lopez et al., 2010).

A specific outcome of this work has been to demonstrate how careful field examination can be used to document detailed stratigraphic architectural relationships that can be used as the basis for the development of high-resolution, three-dimensional facies models that account for the style and mechanism of preservation of fluvial and aeolian deposits. A further outcome is the development of a discussion that considers the principal factors that influence sedimentation in erg-margin settings, including an examination of the role of water-table level, sediment supply, and sediment availability. Such models have valuable predictive potential (cf. Howell & Mountney, 2001; Al-Masrahy and Mountney, 2015; Tooth, 2000a; Belnap et al., 2011; Nanson and Tooth, 1999).

4.3 Background

Previous work has documented the range of sedimentary processes that are known to operate as a function of both allogenic and autogenic controls in erg-margin settings dominated by aeolian and fluvial sedimentary systems. Notable examples of fluvial-aeolian erg-margin interaction in modern systems include those of Great Sand Dunes National Monument, Colorado (Langford, 1989), the Mojave River Wash, California (Langford, 1989), the Skeleton Coast Erg, northern Namibia (Krapf et al. 2003, Stanistreet & Stollhofen, 2002; Svendsen et al., 2003), the Kuiseb River, northern Namib Sand Sea, (Teller & Lancaster, 1990), and the Cooper Creek Fan, Strzelecki Desert, Australia (Cohen, 2010). Notable examples from the ancient record include the Tertiary Tsondab Sandstone Formation, central Namib Desert (Ward, 1988), the Cretaceous Etjo Sandstone Formation, northwest Namibia (Mountney et al, 1998, 1999), the lower-Cretaceous Avilé Member of the Agrio Formation, Argentina (Veiga et al., 2002), the Jurassic Kayenta-Navajo transition, northeastern Arizona, USA (Herries, 1993), the Triassic Helsby Sandstone Formation, UK (Mountney and Thompson, 2002), the Permian Cedar Mesa Sandstone, SE Utah, USA (Langford and Chan, 1988; Mountney & Jagger, 2004), the Permian upper Rotliegend Group UK southern North Sea (Sweet, 1999), the Permian Organ Rock Formation, SE Utah, USA (Cain & Mountney, 2009), the Pennsylvanian to Permian lower Cutler beds, south-east Utah, USA (Jordan & Mountney, 2010, 2012), and the late-Silurian Tumblagooda Sandstone, Western Australia (Trewin, 1993).

The preservation of erg-margin-deposits in transitional regions that pass laterally from a zone where low-sinuosity braided fluvial systems (many subject to ephemeral flow) dominate to a zone where aeolian sandsheet, dune and interdune systems dominate results from a complex set of interacting processes. In wetter (i.e., more humid) climates, preservation commonly takes place where active aeolian dunes migrate alongside fluvial channels that occupy interdune corridors between these dunes. Episodes of fluvial flooding tend to result in the temporary cessation of aeolian dune climb; flood waters may locally expand to cover large areas of flooded interdune corridors in front of advancing aeolian dunes (Simpson & Loope, 1985; Langford and Chan, 1988; Mounney, 2012). In erg-margin areas, the on-going migration of aeolian dunes without climbing (i.e. zero angle of climb), coupled with the repeated flooding of interdunes, favours the generation of bypass supersurfaces (so-called flood surfaces of Langford and Chan, 1988) (Fig. 4.3).

By contrast, in actively migrating aeolian dune systems that are climbing such that long-term accumulation is taking place, the nature of the accumulation is controlled by the ratio between the rate of lateral migration of aeolian dunes and their intervening interdunes and the rate of vertical accretion of aeolian and related deposits (Rubin, 1987; Kocurek, 1999; Mounney, 2012). For so-called wet aeolian systems (*sensu* Kocurek and Havholm, 1993) that are climbing, the rate of vertical accumulation is typically determined by the rate of rise of a relative water table (e.g. Mounney and Thompson, 2002) (Fig. 4.3).

This work describes and interprets in detail aeolian damp-interdune set architectures that transition gradually into overlying aeolian-dune toesets. These observations are used to test the hypothesis that active aeolian migration occurred synchronously with water-table rise in aeolian-interdune corridors, which were repeatedly flooded by fluvial flood waters.

4.4 Geologic Setting of Cheshire Basin

The Cheshire Basin is one of a series of Permo-Triassic basins present in NW England (Fig. 4.1); it is faulted-bounded rift half-graben, the eastern margin of which has a NE-SW trend. The basin was initiated at the end of the Variscan (Late Carboniferous) period (Chadwick, 1997). The Cheshire Basin represents one part of a complex N-S trending major rift system, which extends for more than 400 km, from the English Channel Basin in the

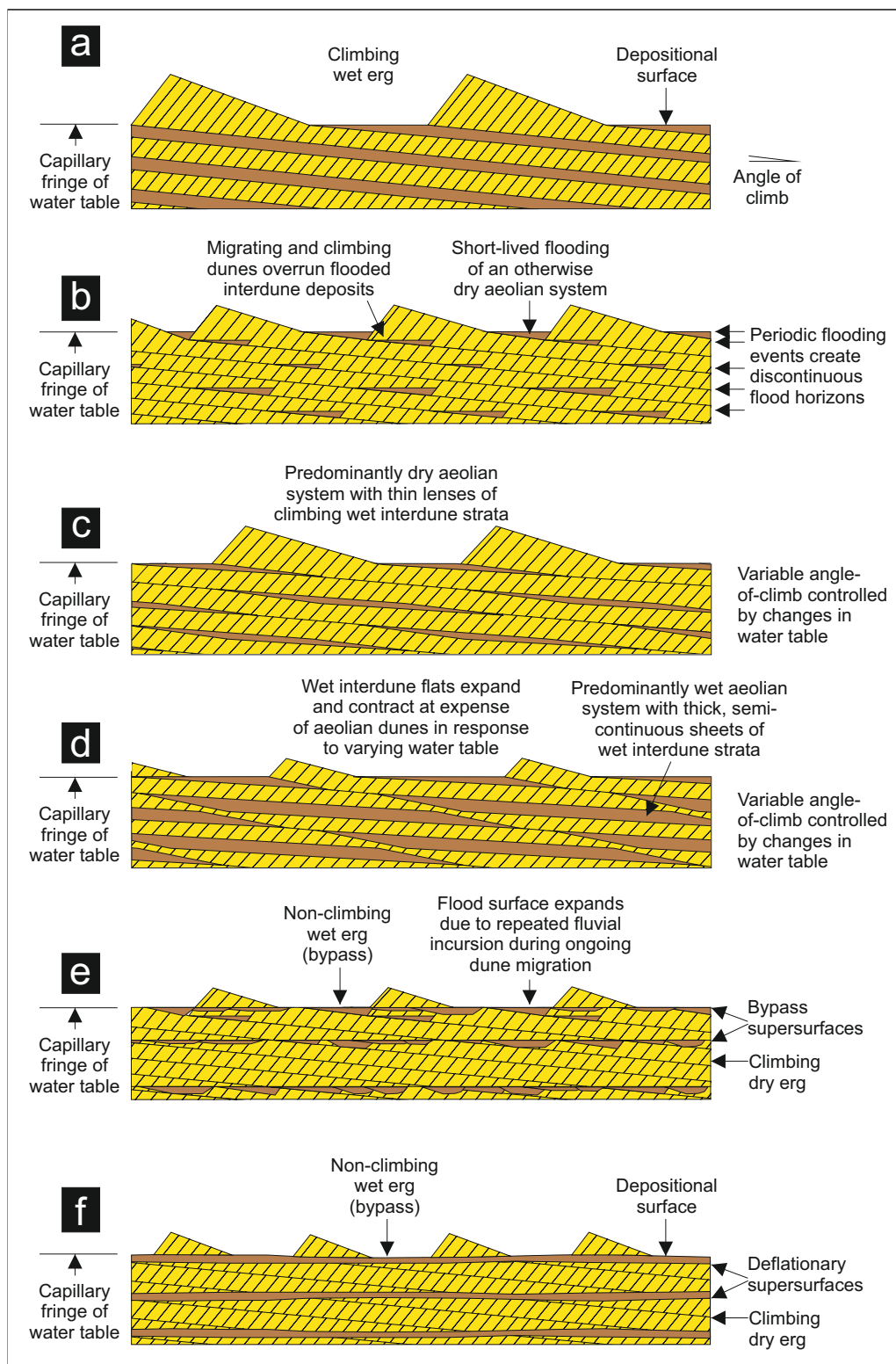


Figure 4.3: Different types of interdune geometries determined by changes in rates of : 1) water-table rise, 2) the rate of aeolian-dune migration and 3) the net aeolian sediment budget. (a) Entrada Sandstone, Kocurek (1981a).; (b) Navajo Sandstone (Herries, 1993); (c) and (d) Helsby Sandstone Formation (Mountney & Thompson, 2002); (e) Cedar Mesa Sandstone (Langford & Chan, 1988, 1989); (f) White Sands (Simpson & Loope, 1985; Loope & Simpson, 1992). After Mountney & Thompson, 2002.

south, to the East Irish Sea, Lancashire, Cumbria and Vale of Eden, Carlisle and Solway Firth basins in the north (Chadwick, 1997).

In Early to Mid-Triassic times, the NW England region occupied an arid climatic region in the northern part of the Pangaea supercontinent, Laurasia, where the climate was arid to semi-arid up to 15° N of the palaeoequator (Thompson, 1970; Warrington & Cook, 1992; Benton et al., 1994). Within the fill of the Cheshire Basin, the Sherwood Sandstones Group (SSG; Fig. 4.1) – which is named after type sections in the Sherwood Forest region of Nottinghamshire – represents a lower, predominantly arenaceous part of the Triassic basin fill (Anisian). This is overlain conformably by the Merica Mudstone Group (MMG) that comprises an upper, predominantly argillaceous red-bed sequence (Thompson, 1970 a, b). The lowermost formation in the MMG, the Tarporley Siltstone Formation, represents accumulation in an intertidal mudflat setting; these mudflats were transgressed in response to a relative sea-level rise (Ireland et al., 1978; Warrington & Cook, 1992). Overlying formations in the MMG are characterised by red and green calcareous mudstones, for example of the Blue Anchor Formation.

Based on seismic, well-log and core data, the total thickness of the preserved Permo-Triassic (SSG and MMG) and Lower Jurassic sediments within the depocentre of the Cheshire Basin is more than 4000 m (Chadwick, 1997). The SSG is characterized by red, yellow and brown sandstones of continental origin that accumulated mainly in a range of fluvial and aeolian environments (Warrington et al., 1980; Thompson, 1970a, b).

In the upper part of the SSG, the HSF in the Cheshire Basin rests unconformably on the Wilmslow Sandstone Formation (WSF; Warrington et al., 1980; Evans et al., 1993). Within the SSG, the HSF, known previously as the Keuper Sandstone, is the thickest rock-unit in the group in the Cheshire Basin, attaining a thickness of 230 m in the Prees Borehole (SJ 5572 3447) and 205 m in the Knutsford Borehole (SJ 7026 7785) (Evans et al., 1993). In these boreholes and in the surrounding region, the HSF is divided into three members: the basal Thrustaston Soft Sandstone Member, the middle Delamere Pebbly Sandstone Member, and the upper Frodsham Soft Sandstone Member (Warrington et al., 1980; Thompson, 1970b; Mountney & Thompson 2002; cf. BGS map, 2007). Numerous good-quality outcrops of the various members of the HSF are present in the northwest part of the Cheshire Basin, though numerous faults are also present and these tend to make lateral stratigraphic correlation problematic across the region.

Based on analysis of some of these outcrops in the region around Helsby, Frodsham, Runcorn and the Wirral, several detailed previous studies have sought to document the preserved stratigraphic architecture of each of these members, to thereby establish lateral and vertical trends in lithofacies and architectural-element distribution in these rock units (e.g., Strahan, 1882; Thompson, 1969, 1970 a & b; Øxnevad, 1991; Mountney & Thompson 2002; Howard et al., 2007). However, hitherto, no detailed account has been published of the details of the stratigraphic architecture of Beacon Hill and Dunsdale Hollow on Overton Hill near the town of Frodsham (Fig. 4.2). These localities are the focus of primary data collection for this study.

4.5 Methods and data

Outcrop data have been collected from a series of outcrops near to the towns of Frodsham, Runcorn and Helsby, northwest England: Beacon Hill, Dunsdale Hollow, Runcorn Expressway road-cut, and Helsby Hill and Quarry (Fig. 4.2). In this study, 14 lithofacies and 11 facies associations have been identified within both aeolian and fluvial strata. Three vertical graphic logs (Fig. 4.4) and a series of 11 two-dimensional stratigraphic panels that each represent stratigraphic sections that are 49 to 157 m long and 5 to 13 m high have been used to depict lithofacies and architectural-element relationships from outcrops at Beacon Hill, Dunsdale Hollow and Runcorn Expressway road-cut. Sedimentary features depicted on each of these panels demonstrate variability (both laterally and vertically) of facies distributions and architectural elements of both fluvial and aeolian strata. The panels highlight the geometric relationships between aeolian dune and interdune facies, and the styles of interaction between competing fluvial and aeolian environments. Two-hundred-and-fourteen foreset dip-azimuth readings have been measured from fluvial and aeolian cross-bedded lithofacies to record palaeo-bedform migration trends and to infer local palaeocurrent trends for both the fluvial and aeolian palaeoenvironments (Fig. 4.4).

4.6 Sedimentary Facies Analysis

Lithofacies described from exposed parts of the HSF within Cheshire Basin have been documented previously by several workers (e.g. Thompson, 1969; 1970a & b; Mountney and Thompson, 2002). However, no detailed studies of sedimentary architectural relationships have been published previously relating to the outcrops discussed herein. Within this study, 14

Summary Sedimentary Log of Beacon Hill

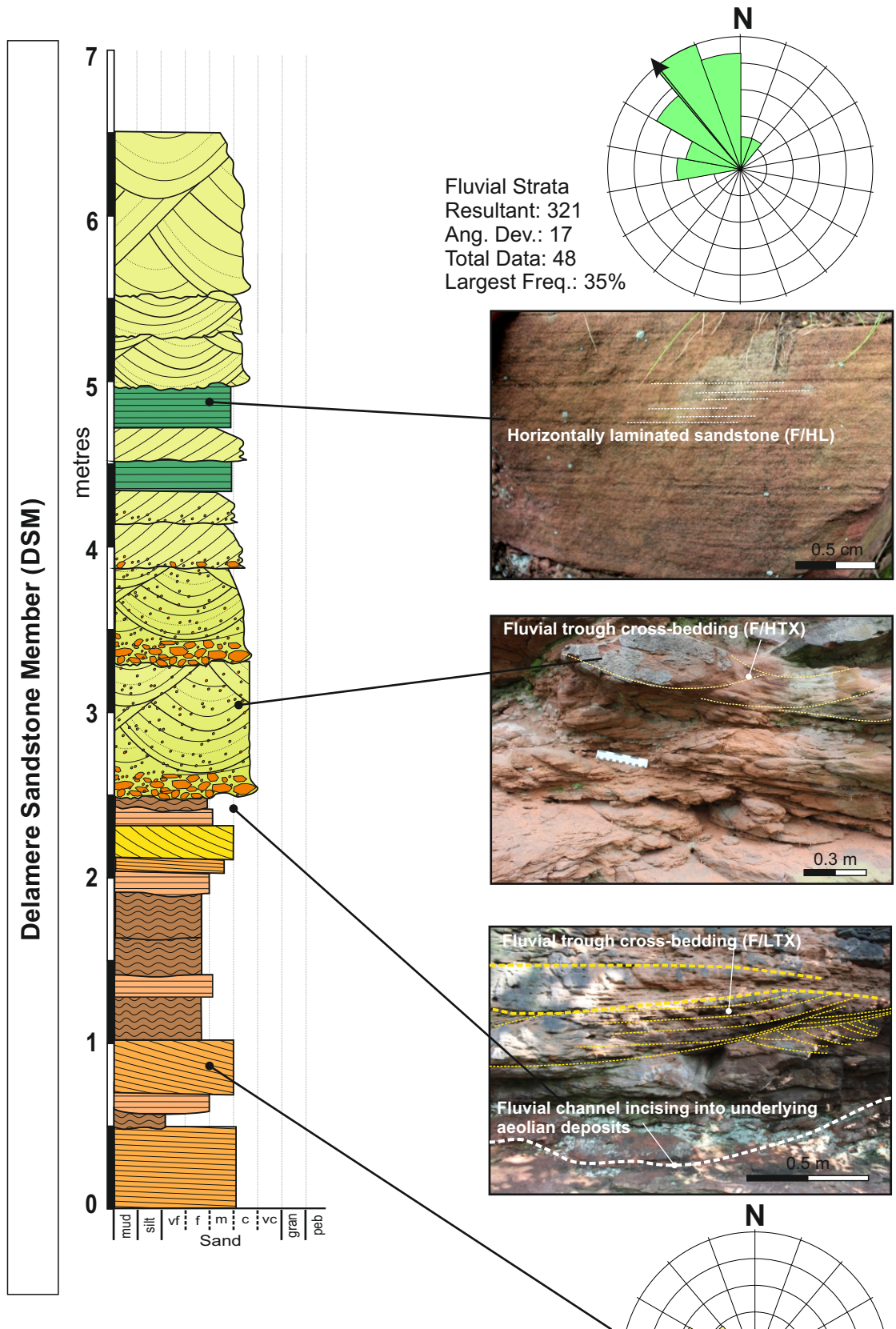


Figure 4.4a: Sedimentary log representing both aeolian and fluvial successions of the Delamere Sandstone Member (DSM) in Beacon Hill, Frodsham town. The log location is shown in figures 4.2.

Aeolian Strata
 Resultant: 252
 Ang. Dev.: 22
 Total Data: 95
 Largest Freq.: 28%

Summary Sedimentary Log of Dunsdale Hollow

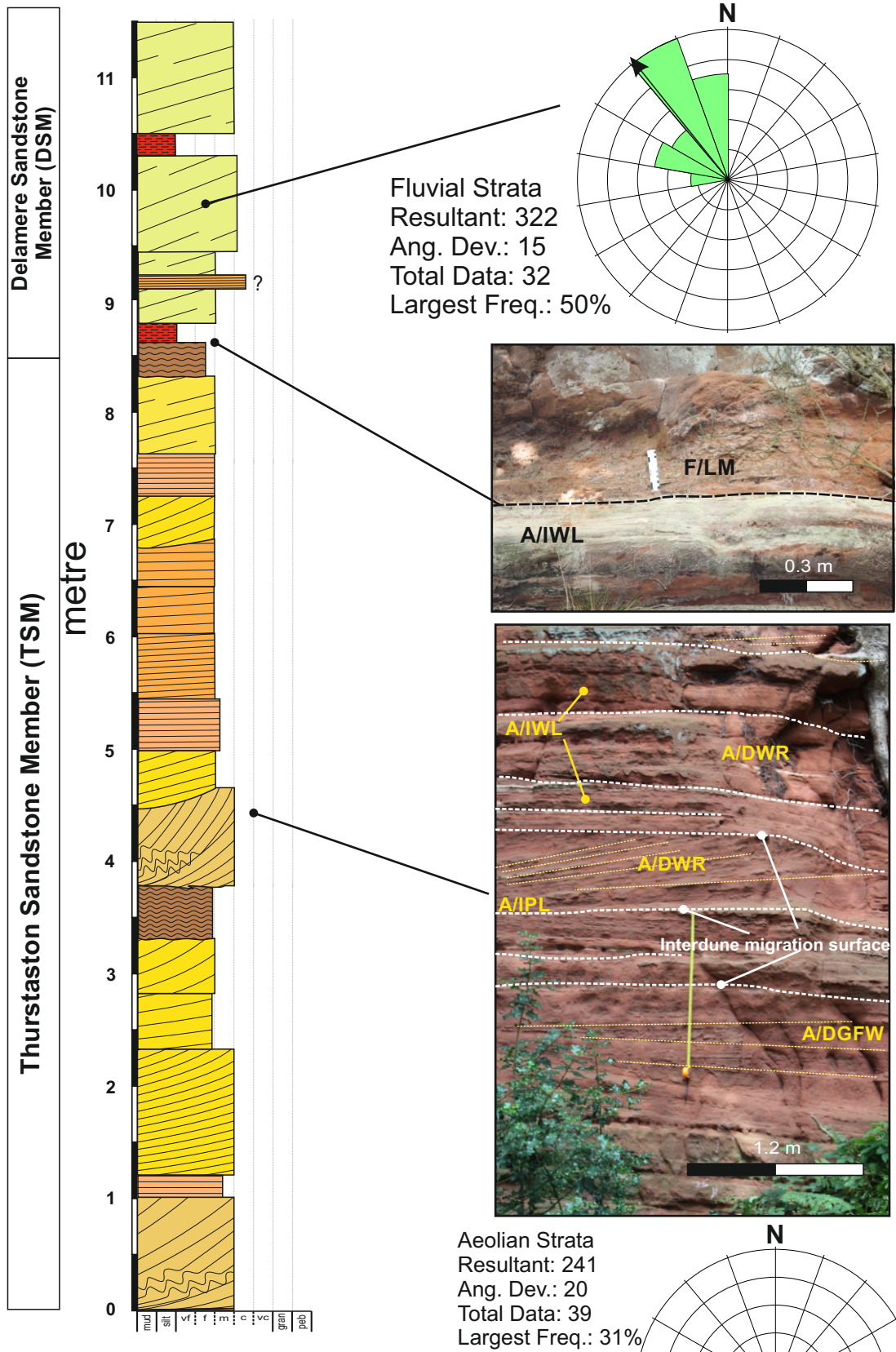
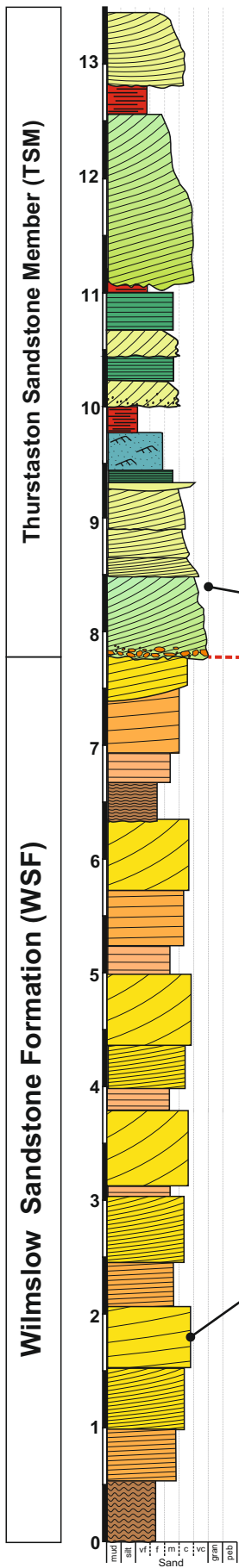
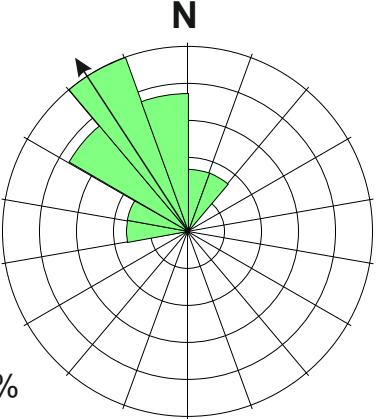


Figure 4.4b: Sedimentary log representing aeolian successions of the Thurstaston Sandstone Member (TSM) underlying the fluvial successions of Delamere Member (DSM) in Dunsdale Hollow, near Frodsham, Cheshire. The log location is shown in figures 4.2.

Summary Sedimentary Log of Runcorn Expressway Road-cut (A557)



Palaeocurrent data of TSM in Runcorn Expressway



Fluvial Strata
 Resultant: 327
 Ang. Dev.: 17
 Total Data: 23
 Largest Freq.: 39%



? Hardegsen Unconformity

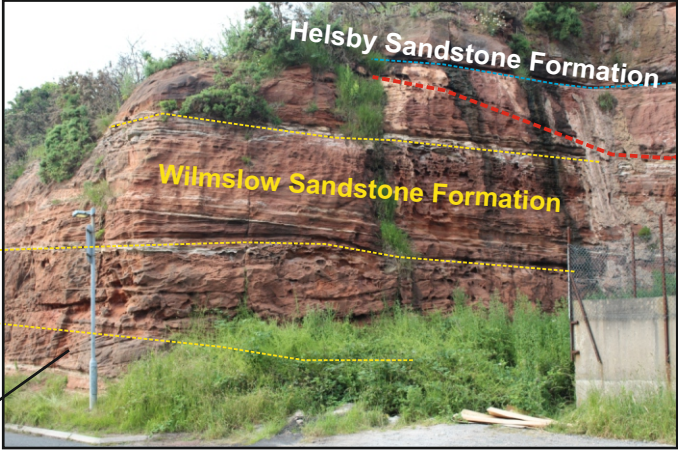








Figure 4.4C: the sedimentary log representing the fluvial successions of Thurstaston Member (TSM) of Helsby Sandstone Formation (HSF) overlying aeolian successions of the Wilmslow Sandstone Formation (WSF) at the Runcorn Expressway cutting (A557), Runcorn town. The log location is shown in figures 4.2.










Legend

Lithofacies Key

Aeolian Facies

| | | |
|---|--------|---|
|  | A/DGFW | Aeolian Dune: Grainflow Strata |
|  | A/DGFA | Aeolian Dune: Grainfall strata |
|  | A/DWR | Aeolian Dune: Wind-Ripple Strata |
|  | A/DDW | Aeolian Dune: Convoluted Laminated Strata |
|  | A/IPL | Aeolian Interdune: Planar Laminated Sandstone |
|  | A/IWL | Aeolian Interdune: Wavy / Crinkly Laminated Sandstone |

Fluvial Facies

| | | | | | |
|---|-------|--|---|------|--|
|  | F/EFM | Fluvial Extraformational Pebbly Sandstone |  | F/DC | Fluvial Deformed (De-watered) Sandstone |
|  | F/IFM | Fluvial Intraformational Pebbly Sandstone |  | F/LS | Fluvial Horizontally Laminated Sandy Siltstone |
|  | F/LTX | Fluvial Low-angle Trough Cross-bedded Sandstone |  | F/LM | Fluvial Horizontally Laminated Mudstone |
|  | F/HTX | Fluvial High-angle Trough Cross-bedded Sandstone | | | |
|  | F/HL | Fluvial Horizontally Laminated Sandstone | | | |
|  | F/CR | | | | |

Facies Association Key

Fluvial Associations

Channelized Fluvial Facies Association (CF)

Succession CF₁ Succession CF₃
 Succession CF₂ Succession CF₄

Non-confined Sheet-Like Fluvial Facies Association (NC)

Succession NC₁

Aeolian Associations

Aeolian Dune Facies Association (AD)

Succession AD₁ Succession AD₂

Aeolian Interdune Facies Association (ID)

Succession ID₁ Succession ID₂

Aeolian Sandsheet Facies Association (SH)

Succession SH₁ Succession SH₂

Architectural Element Key

Fluvial Elements

F₁ Multi-Storey, Multi-Lateral Amalgamated Channel-Fill Complex

F₁ Single-Storey, Multi-Lateral Amalgamated Channel-fill Complex

F₁ Fluvial Floodplain Element

Aeolian Elements

A₁ Aeolian Dune Element

A₂ Aeolian Interdune Element

A₃ Aeolian Sandsheet Element

Sherwood Sandstone Group (SSG)

HSF Helsby Sandstone Formation

FSM Frodsham Sandstone Member

DSM Delamere Sandstone Member

TSM Thurstaston Sandstone Member

WSF Wilmslow Sandstone Formation

Study Location Grid References (UK National Grid)

Beacon Hill **SJ 514763**









Dunsdale Hollow **SJ 514763**

Runcorn Expressway **SJ 505808**

Helsby Quarry **SJ 489748**

Grange Road-Cut **SJ 220869**

Key to Depositional Models

| | |
|---|-----------------------------------|
|  | Aeolian Dune |
|  | Aeolian Dune (toe set) |
|  | Aeolian Sandsheet / Dry Interdune |
|  | Aeolian Damp-Interdune |
|  | Fluvial-Mudstone |
|  | Abandoned Fluvial Channel |
|  | Fluvial-Channel Sandstone |
|  | Active Fluvial-Channel |




| | |
|---|---|
|  | Base of FSM / Top of DSM |
|  | Base of DSM / Top of TSM |
|  | Base of TSM / Hardegsen Unconformity / Top of WSF |

Figure 4.4: Continued.

distinct lithofacies are identified representing both fluvial and aeolian palaeoenvironments (Table 4.1).

4.6.1 Fluvial facies

In Table 1, eight discrete facies of fluvial origin are identified in the HSF based on primary observations from studied outcrops. These fluvial facies are described herein in detail. Commonly occurring combinations of these facies are assigned to 2 facies associations: channelized fluvial and non-confined sheet-like fluvial. Fluvial lithofacies comprise 35% of the studied succession.

4.6.1.1 Extraformational Pebbly Sandstones Facies (F/EFQ)

Description: This facies occurs as 0.1 to 0.25 m-thick sets of poorly-sorted, sub-angular to sub-rounded matrix- to clast-supported, quartz-dominated extraformational clasts that are up to 15 mm in diameter (Fig. 4.5). Although rare in this facies, intraformational clasts may also be present and these take the form of angular to sub-angular mud-clasts, up to 18 cm in diameter. Both extraformational and intraformational clasts “float” in a medium- to very coarse-grained sandstone matrix. Sets of this facies are mostly internally structureless, though rare cases in which internal sedimentary structures are present are noted, mostly in the form of poorly defined planar-tabular or trough cross-stratification. Sets of this facies commonly occur directly above prominent erosional bounding surfaces. This facies represents 3% of the total succession and 8.6% of the fluvial part of the succession.

Interpretation: The occurrence of this facies directly above erosional bounding surfaces, the coarse-grained nature of the sediments, the occurrence of pebble clasts that appear to float in a coarse sand matrix and the only rare presence of cross-stratification is indicative of rapid deposition from concentrated, sediment laden flows during major flood events (Jones & Rust, 1983; Collinson et al., 2006 p.153). This facies represents in-channel lags that accumulated preferentially in the relatively deeper erosional scour bases of fluvial channels, though could alternatively represent accumulation of the basal parts of fluvial mid-channel bars (Miall, 1996; Bridge, 2006 & 2008). The poorly developed cross-bedding in this facies, which varies from tabular to trough-shaped form, possibly indicates a changing water stage over a flood cycle (Harmes et al., 1963; Miall, 1996).

4.6.1.2 Intraformational Pebbly Sandstone Facies (F/IFM)

Description: This facies occurs as 0.1 to 0.4 m-thick sets of poorly-sorted, angular to sub- angular, matrix-supported, red-brown sandstone in which










| Colour Code | Code | Facies | Description | Interpretation |
|--|--------------|--|--|---|
|  | F/EFQ | Extraformational pebbly sandstone | Coarse sand to pebble grade, moderate- to poorly-sorted, sub-angular to sub-rounded and quartz-dominated clasts 0.3 - 3.5 cm diameter. This facies is 0.1 to 0.4 m thick sets. | Represents in-channel lags formed in deep channels as result of rapid deposition. |
|  | F/IFM | Intraformational pebbly sandstone | Coarse sand to cobble grade, moderate- to poorly sorted, angular to sub-rounded and mud-clasts dominated clasts 0.5 - 18 cm diameter. This facies occurs as 0.1 to 0.3 m thick sets. | Reworking of sediments that were eroded locally from either previous fluvial-floodplains or aeolian-interdunes during channel migration. |
|  | F/HTX | High-angle trough cross-bedded sandstone | Medium- to coarse grained matrix, moderate- to poorly sorted and sub-angular to sub-rounded. This facies dominated by tabular cross-bedding with tangential base. Set thickness is 0.3 to 1.5 m. | In-channel migration of pebbly sinuous-crested dunes (mid-channel) developed under strong flows. |
|  | F/LTX | Low-angle trough cross-bedded sandstone | Medium- to coarse grained matrix, moderate- to poorly sorted and sub-angular to sub-rounded. This facies consists of tabular-planar foresets with angular base. Set thickness is 0.3 to 1.5 m. | Macro-forms of mid-channel-bars or point bars with lack of pebbles. |
|  | F/HL | Horizontally laminated sandstone | Fine- to medium-grained, well- to moderate-sorted and rounded to well-rounded. This facies is defined by parallel and planar laminated sets that are 0.1 to 0.3 m thick. Laminae reaches up to 2 mm thick. | Deposited dominantly under upper-plane bed stage flows in confined-channels with medium-grained sandstones. |
|  | F/CR | Current-ripple laminated sandstone | Very fine to fine-grained, very well- to well-sorted and rounded to well-rounded. This facies is characterized by mud-draped of sub-critical climbing-ripple strata. A laminae range 5 to 40 mm thick. | Rapid deposition of abundant fine-grained suspended-load under sub-aqueous low-flow regime conditions in the waning stages of flood pulses during down stream migration of straight- and sinuous-crested ripples. |
|  | F/DC | Deformed sandstone (de-watered and convoluted sandstone) | Medium- to coarse-grained, well- to moderate-sorted and rounded to well-rounded. Convoluted strata include slump and water-escape structures. Sets range 0.1 to 0.6 m thick. | Rapid deposition in a sub-aqueous setting in which semi-lithified laminations are squeezed out causing disruption in the original structures. |
|   | F/LS F/LM | Laminated siltstone and mudstone | Sandy siltstone and silty mudstone, very well-sorted, rounded to well-rounded; faint laminations; Sets range 0.02 to <3 m thick. | Represents the deposition of suspension-load either within fluvial floodplain areas or with an abounded fluvial-channel where a waning flow was commonly present. |

Table 4.1: Table of observed lithofacies.







| Colour Code | Code | Facies | Description | Interpretation |
|---|--------|---------------------------------|--|--|
|  | A/DGFW | Aeolian grainflow strata | Fine- to coarse-grained, well to moderate-sorted, sub-rounded to well-rounded. This facies forms inclined tongues that are typically 0.05 to 0.23 m thick. | Represents gravity-driven sediments on aeolian dune lee slopes as a result of repeated avalanche processes when dune lee slopes exceed the angle of repose (32 - 34 degrees). |
|  | A/DGFA | A e o l i a n grainfall strata | Fine- to medium-grained, well-sorted, rounded to well-rounded. This facies typically forms 1 - 4 mm thick laminae separating grainflow sets. | Accumulation of saltating sediments on the upper part of the aeolian dune lee slopes during episodes of decreased wind velocity. |
|  | A/DWR | Wind-ripple laminated sandstone | Very fine- to medium-grained, well-sorted, rounded to well-rounded. Laminae are 1 - 3 mm thick (translatent strata). | Represent subcritically climbing wind-ripples formed in aeolian environments of either low-relief dune slopes or dry-interdunes or sandsheets as a result of tractional processes. |
|  | A/DDW | De-watered/convoluted sandstone | Fine- to coarse-grained, well to moderate-sorted, sub-rounded to well-rounded containing slumps and water-escapes structures. Sets are typically 0.1 to 2.8 m thick. | Represents slump degradation when semi-liquefied loosely packed sediments avalanche or slide down slope on the lee of damp or wet migrating dunes. |
|  | A/IWL | Wavy laminated sandstone | Very fine- to fine-grained, well-sorted, sub-rounded to well-rounded forms irregular and continuous laminations. Sets vary between 0.1 to 0.5 m thick. | Represents damp to water-lain deposits of damp and/or wet aeolian-interdunes that were developed under fluctuations in short-and/or long-term water-level rise. |
|  | A/IPL | Planar laminated sandstone | Very fine- to fine-grained, well-sorted, sub-rounded to well-rounded forms laminations that vary 0.1 to 0.3 m thick. | Represents the wind-ripple deposits of dry-interdune-flats in between migration aeolian-dunes. |

Table 4.1 cont.: Table of observed lithofacies.

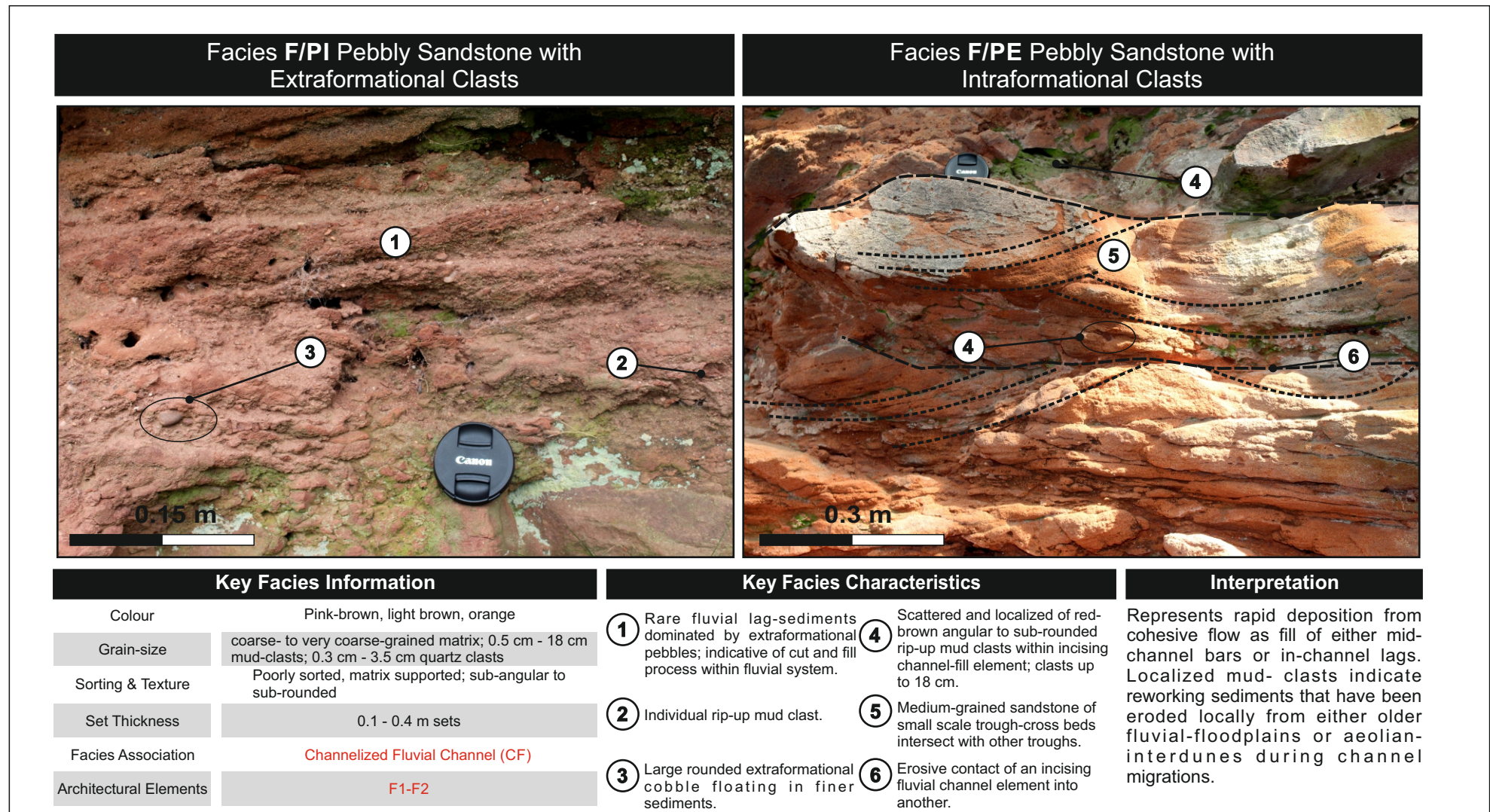


Figure 4.5: Pebbly sandstone lithofacies (F/PI & F/PE).

elongate-shaped, mud-dominated intraformational clasts are present (Fig. 4.5). Additionally, small to medium size pebble clasts of extraformational (mostly quartzite) origin commonly occur scattered within the medium- to very coarse-grained sandstone matrix. The red-brown mud clasts of intraformational origin attain lengths up to 18 cm, and are especially abundant in the lower parts of sets of this facies. Although in places scattered in their distribution, clasts in the sandy matrix are commonly normal graded. Sets of this facies most commonly occur directly above prominent erosional bounding surfaces. This facies occurs in common association with facies F/HTX, F/LTX and F/EMI (Table. 4.1). This facies represents 7% of the total succession and 20% of the fluvial part of the succession.

Interpretation: The mudstone clasts present in this facies primarily originated from dried mud-layers on top of stream beds that have been entrained and re-worked by a subsequent flood and transported as clasts within bed-load sediments (Reid & Frostick, 2011). Some of the clasts may have alternatively originated from fine sediments of interdune facies (A/ID) that were eroded and re-transported during major floods (Cain & Mountney, 2009; Svendsen et al., 2003). The scattered nature of clasts present in the sandy matrix of this facies may represent re-working of sediments as a result of either local floodplain erosion or bank collapse (Stear, 1985; Miall, 1996, p. 438; Cain & Mountney, 2009). The normal grading observed within this facies suggests deposition by waning flow regime during falling stage in energy of fluvial floods (Miall, 1996).

4.6.1.3 Trough Cross-Bedded Sandstone Facies (F/HTX and F/LTX)

Description: This facies occurs as 0.3 to 1 m-thick sets composed of poorly to moderately sorted, medium- to very coarse-grained, sub-rounded to rounded, cross-bedded sandstone. The foresets of both trough cross-beds (F/TX) planar cross-beds are delineated by coarse sand grains that form mm-thick lags. In some examples, both extraformational and intraformational mud clasts occur scattered throughout sets of this facies but are especially common directly above erosional lower contacts of sets and on the lower parts of cross-stratified foresets. In the Beacon Hill study area, the trough shapes of facies F/TX are readily recognized in sections oriented perpendicular to NNW palaeomigration direction. Internally within sets that are exposed in sections parallel to the generally NNW palaeoflow, both high- (F/HTX: 28° to 35°) and low-angle (F/LTX 10° to 18°) inclined planar cross-bedded foresets are prominent. Both planar foresets with an angular base

and curved foresets with an asymptotic and tangential base are observed (Fig. 4.6). Both types occur in sets that are typically 0.3 to 0.7 m thick and these occur stacked together to form cosets that are themselves up to 2.5 m thick. Internal set boundaries of both facies record first-order bounding surfaces with minor erosion (cf. Miall, 1996). Lag deposits of extra- and intraformational clasts are common at the base of some trough cross-bedded sets. Overall, this facies tends to occur most commonly directly above major channel scour surfaces. This facies represents 18% of the total succession and 51.4% of the fluvial part of the succession.

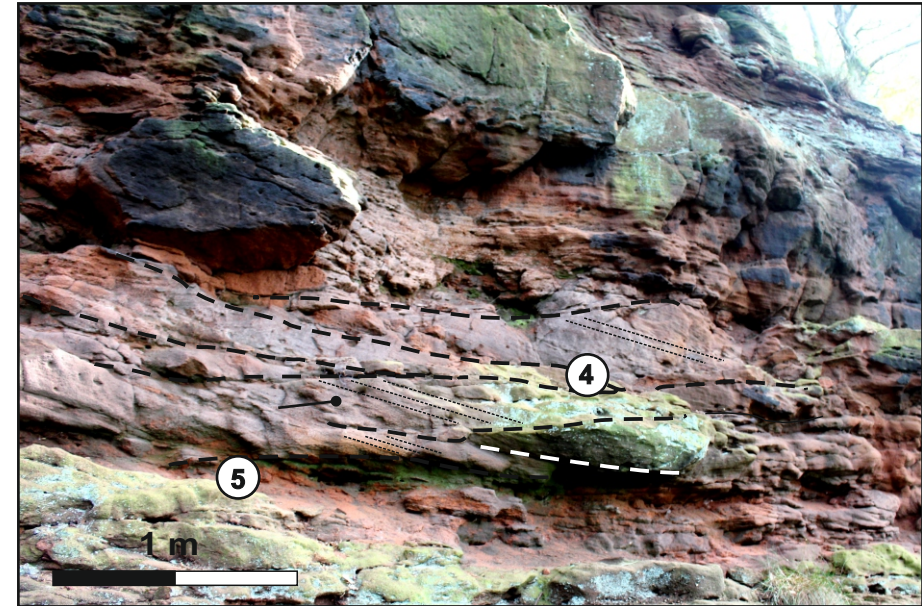
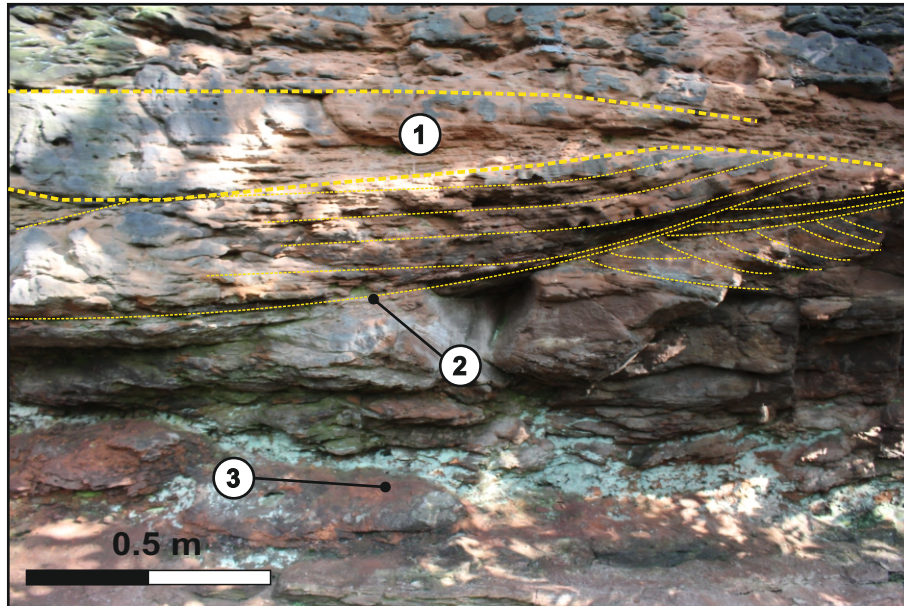
Interpretation: The trough cross-bedding that characterises this facies records the migration and accumulation of sinuous-crested transverse or slightly oblique fluvial dunes (cf. Rubin, 1987). The first-order surfaces that bound sets represent continuous sedimentation via trains of similar type of dunes, whereas the second-order bounding surfaces that bound cosets may indicate alteration in flow condition or direction with short-period time disruption (Miall, 1996). Such dunes were generated by relatively powerful flow where the shape of these dunes formed as a result of the eddying currents merging closely in scour pits at the front of the migrating dunes (Jopling, 1965; Collinson et al., 2006; Allen, 1982). With increasing in flow-strength, a backflow component is produced in front of the toesets of the lee face of the dunes associated with flow separation forming tangential foresets (asymptotic) within the dunes (e.g., Jopling, 1965, Collinson et al., 2006). The preservation of only toesets and foresets, and the absence of topsets demonstrates a subcritical angle of climb for which preservation of dune-stoss slopes was not possible (Rubin, 1987, Collinson et al., 2006).

4.6.1.4 Horizontally Laminated Sandstone Facies (F/HL)

Description: This facies occurs within the study areas as sets that are each 0.2 to 0.4 m thick. This facies is characterized by light-brown to red-brown, fine-grained, moderate- to well-sorted, sub-angular to rounded sandstone (Fig. 4.7). Individual laminations are 2 to 4 mm thick and parallel to flat-lying set bounding surfaces. Some well-developed examples of a laminae exhibit normal grading. Rarely, reworked quartz-pebbles and mudstone rip-up clasts that are < 1 cm in diameter are present in the sandy laminated deposits. This facies represents 3% of the total succession and 8.6% of the fluvial part of the succession.

Interpretation: Horizontally laminated sandstones in fluvial successions represent sediments that accumulated dominantly under conditions of upper-stage plane-bed flow (Bridge, 2003; Picard & High, 1973; Collinson et al.,

Facies F/HTX High-angle and F/LTX Low-angle Trough Cross-Bedded Sandstone



Key Facies Information

| | |
|------------------------|---|
| Colour | Light-brown, orange, pink-brown |
| Grain-size | medium- to coarse grained matrix; 0.5 cm - 18 cm clasts |
| Sorting & Texture | Poorly sorted, matrix supported; sub-rounded to well rounded sands; angular to sub-rounded clasts |
| Set Thickness | 0.3 - 1.7 m sets |
| Facies Association | Fluvial Channel (CF) |
| Architectural Elements | F1, F2, F3 |

Key Facies Characteristics

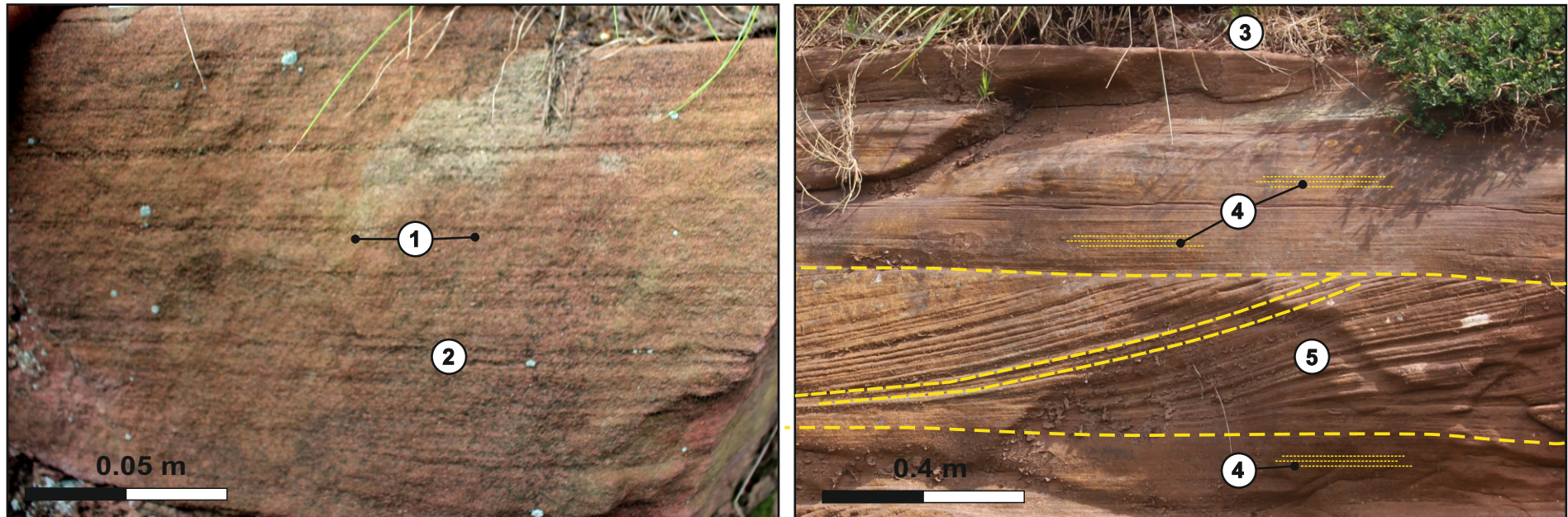
- ① Lenticular lenses of cross-bedded sandstone sets with base relief up to 1 m; contain rare intraformational mud-clasts.
- ② Erosional base of a channel element with high-proportion of mud-clasts.
- ③ Thin bed of mudstone in between fluvial and aeolian sandstone.
- ④ Erosional contacts of stacked, small- to medium-scale incised fluvial channel element largely filled with poorly sorted coarse sandstone and intraformational mud clasts.
- ⑤ Medium- to small-scale trough -cross beds up to 0.6 m thick with high proportion of localized mud-clasts.

Interpretation

Tangential foresets (asymptotic) in this facies indicates a flow-separation in front of the toesets of lee-face of in-channel dunes when flow-strength is increased. F/HTX and F/LTX represent mid-channel-dune migration of sinuous-barforms that developed under strong flows.

Figure 4.6 : Trough cross-bedded sandstone lithofacies (F/HTX & F/LTX).

Facies F/HL: Horizontally Laminated Sandstone



Key Facies Information

| | |
|------------------------|---|
| Colour | Light pink-brown to red-brown |
| Grain-size | 0.5 cm - 15 cm clasts; medium- to coarse grained matrix |
| Sorting & Texture | Poorly sorted, matrix supported; sub-angular to sub-rounded |
| Set Thickness | 0.3 - 2.5 m sets |
| Facies Association | Fluvial Channel (CF) |
| Architectural Elements | F1, F2, F3 |

Key Facies Characteristics

- ① Sub-rounded to sub-angular medium-grained sandstone; laminated, Beacon Hill.
- ② Fine-grained sandstone interlaminated with medium-grained sandstone showing subtle fining-up trend within this facies.
- ③ Laminated mudstone overlying horizontal lamination with basal sharp contact.
- ④ F/HL set thickness increases upward and passes into low-angle inclined fluvial dune foresets.
- ⑤ Medium-grained sandstone of fluvial dunes with sharp basal contact.

Interpretation

This facies represents deposits developed dominantly under upper-stage plane bed flows. Indicates suppressed turbulence within the flow; streaks of faster and slower moving water close to the bed. The fine-grained laminae may record deposition of the suspension load at the distal lee-side of bed waves. Common in late-stage channel filling.

Figure 4.7 : Horizontally laminated sandstone lithofacies (F/HL).

2006). Differences in turbulent in bed shear stress and sediment transport are primarily responsible for the development of such laminations (Bridge, 2003). The fine-grained components of each laminae may accumulate in response to deposition of the suspension load where the fine sediments settle on the underlying medium-grained laminae (Bridge, 2003; Reid & Frostick, 1997). The rare occurrence of reworked pebbles and rip-up clasts associated with this otherwise sand-dominated style of lamination indicates local erosion and re-deposition by rolling along in a sand traction carpet (Miall, 1996).

4.6.1.5 Current Ripple-Laminated Sandstone Facies (F/CR)

Description: This facies occurs as 0.15 to 0.3 m-thick sets of light-brown to brown, very fine- to fine-grained sandstone with accompanying minor mud-drapes that occur within sets as laminae that each have thickness of <1 mm and lateral extension of 0.5 m (Fig. 4.8). Ripple cross-lamination is characterized by cross lamination that is indicative of subcritical climb at angles of 5 to 10° relative to set basal bounding surfaces. The thickness of individual current-ripple laminations varies from 3 to 5 mm. This facies, which is well-exposed in the southern exposure of Runcorn Expressway Road, extends laterally as sets that are up to 1 m wide. This facies represents 1% of the total succession and 2.9% of the fluvial part of the succession.

Interpretation: This facies forms from the rapid deposition of abundant fine-grained suspended- and bed-load under sub-aqueous low-flow regime conditions in the waning stages of flood pulses during down-stream migration of straight- and sinuous-crested ripples (Stear, 1985; Collinson et al., 2006; Allen, 1968). The climbing ripple-laminated sets indicate high rates of vertical aggradation associated with on-going ripple migration driven by relatively weak unidirectional currents, likely in shallow water (Collinson et al. 2006; Stear, 1985). The subcritical angle of the climb indicates a balance between the rate of vertical aggradation and the rate of ripple migration such that the angle of the climb is less steep than the angle of the stoss slope of the ripples. This resulted in the scour of minor erosion surfaces as the troughs between successive ripples migrated, and thus no preservation of the stoss-side of the ripple forms (Collinson et al. 2006).

4.6.1.6 Deformed Sandstone (De-watered and Convolute Sandstone) Facies (F/DC)

Description: This rare facies occurs as 0.1 to 0.35 m-thick sets of red-brown, fine- to medium-grained sandstone that is soft-sediment deformed, resulting in the preservation of faint convolute-deformed foresets characterized by 5 cm-wide and 5 to 10 cm-high overturned folds in which both normal and reverse folding is evident, commonly with complex and variable fold-axes (Fig. 4.9). This facies represents 1% of the total succession and 2.9% of the fluvial part of the succession.

Interpretations: This facies serves as evidence for rapid deposition in a sub-aqueous setting in which semi-lithified laminations are squeezed out causing disruption in the original structures. Overturned folds of convolute laminations are most commonly formed when shear stresses increase in the upper parts of a sub-aqueous bedform and partially deform its primary (i.e. original) sedimentary structures (Collinson et al., 2006; Allen, 1982). The presence of this facies in fluvial successions suggests that the deformation occurred shortly after sediment accumulation (Selley, 1969; Nicholson, 1993). The structures of this facies may develop when fluids trapped within the pore framework of the deposited sands moved upwards to lower pressure areas, thereby causing narrow zones of fluidization of clean fine sands of pipe-like shape (Santos et al., 2012; Owen & Santos, 2014).

4.6.1.7 Laminated Siltstone and Mudstone Facies (F/LS and F/LM)

Description: This facies occurs as sets up to 2 m thick composed internally of laminated sandy-siltstone and mudstone facies (F/LS & F/LM). Good examples are evident in the Old Helsby Quarry section close to the town of Helsby (Fig. 4.10). Laminated sandy-siltstone deposits typically occur interbedded with massive, finer-grained mudstone (itself rarely laminated). The internal structures commonly include upward transitions from parallel lamination to massive beds that are gradational. Desiccation cracks, vertebrate footprints, and invertebrate burrows have been noted from this facies previously (cf. Mountney & Thompson, 2002; King & Thompson, 2000). This facies represents 2% of the total succession and 5.7% of the fluvial part of the succession.

Interpretation: This lithofacies records accumulation from suspension-load in quiet-water zones. When the turbulent current in the water calms, sediment held in suspension settles-down to the bed to form laminations. The thickness of these laminations is determined by the continuity of

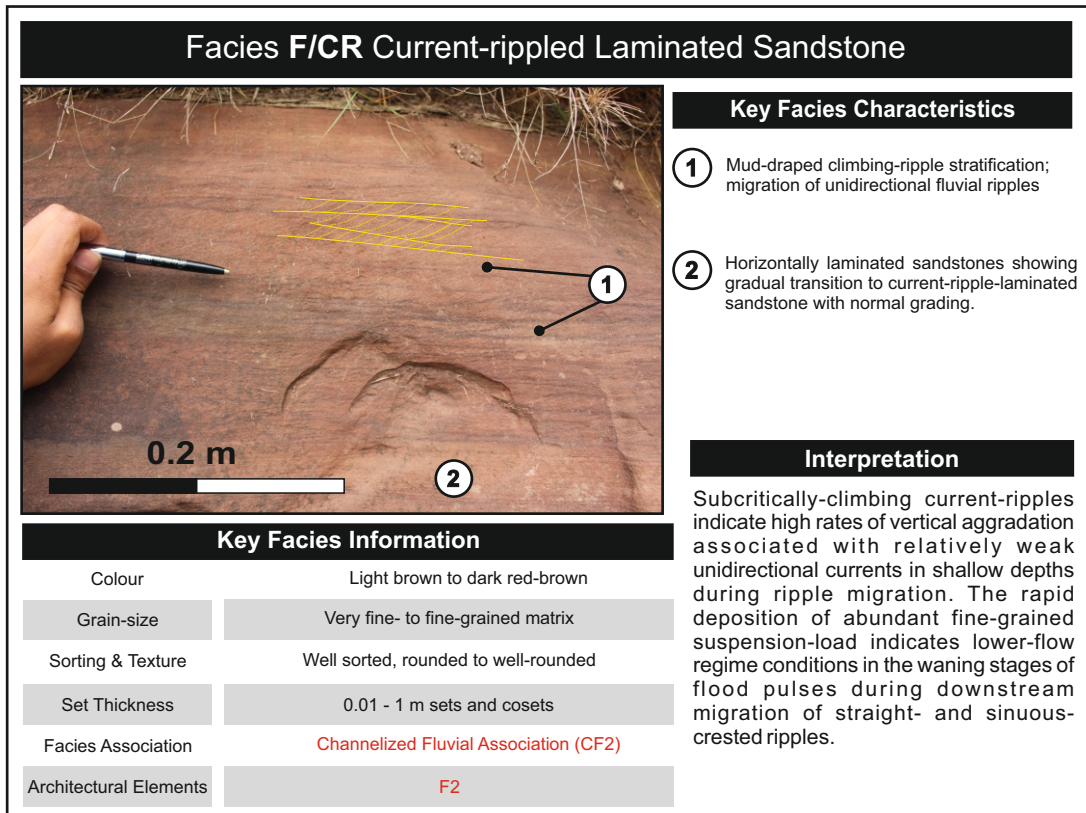


Figure 4.8 : Current-ripple laminated sandstone lithofacies (F/CR).

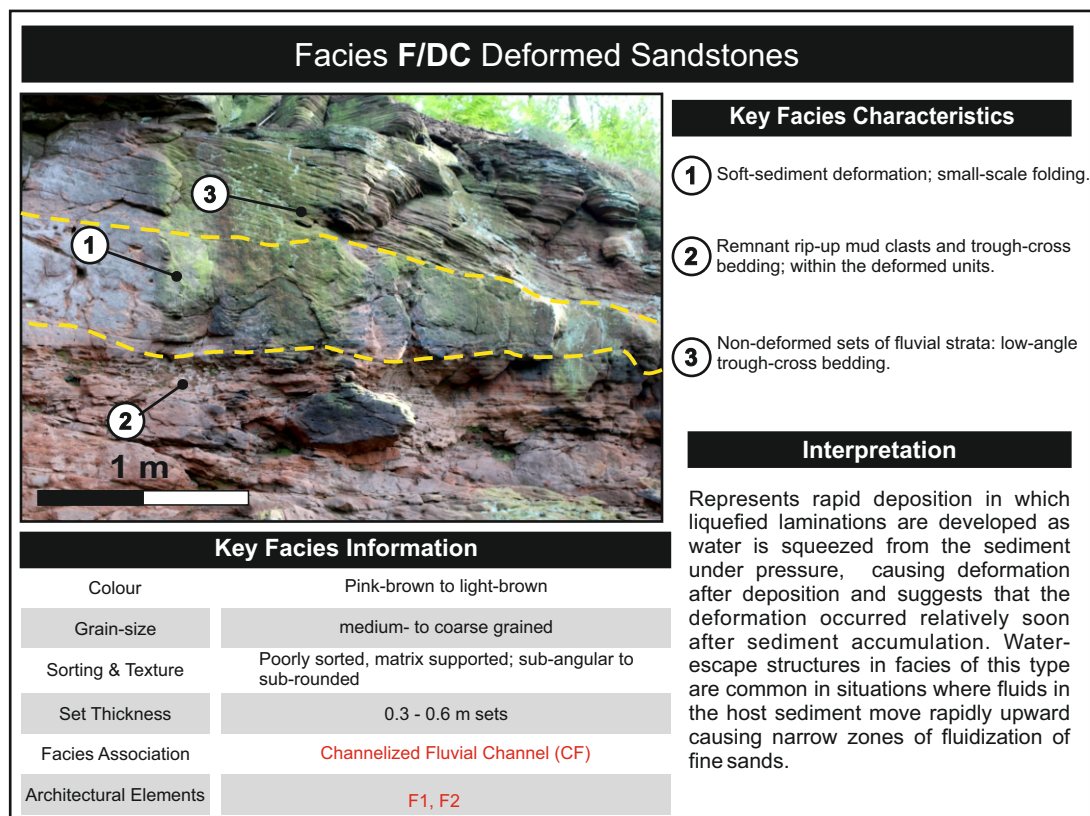
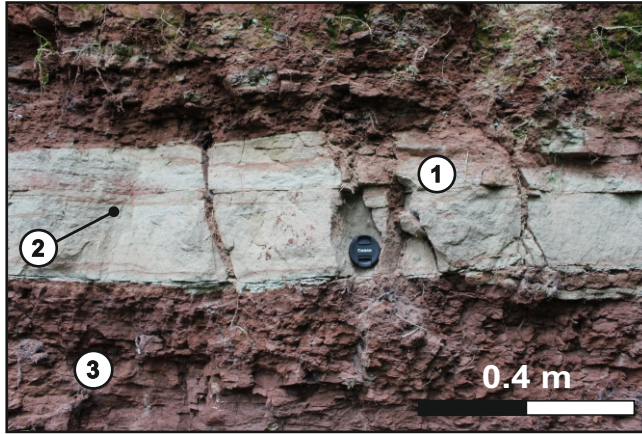


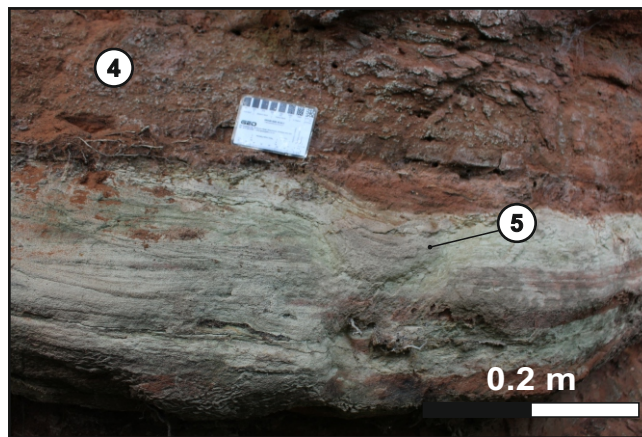
Figure 4.9: Deformed sandstone lithofacies (F/DC).

Facies F/LM Laminated Siltstone and Mudstone



Key Facies Characteristics

- ① Massive laminated siltstone interbedded between horizontally laminated mudstone.
- ② Faint horizontally-lamination within siltstone beds.
- ③ Thin, laterally extensive bed of red-brown silty mudstone.



- ④ Thin bed of red-brown silty mudstone of fluvial floodplain overlying sandy aeolian-interdune in Dunsdale Hollow (check the sedimentary log in Fig.4.4.b).
- ⑤ Deformed sets within sandy aeolian-interdune present animal vertebrate footprint marks that indicate moisture was at or above damp interdune surfaces.

Key Facies Information

| | |
|------------------------|--|
| Colour | Light-grey to dark red-brown |
| Grain-size | sandy siltstone, silty mudstone |
| Sorting & Texture | well sorted; rounded to well-rounded |
| Set Thickness | 0.02 - 1 m sets and cosets |
| Facies Association | Non-confined Fluvial Sheet-Like |
| Architectural Elements | F3 |

Interpretation

The presence of siltstone indicates deposition from sluggish flows associated with splays or widespread non-confined sheet-like flow, whereas mudstone beds indicate the deposition of suspension-load from standing ponds of water in fluvial floodplain areas where waning flow had ceased. Desiccation cracks, vertebrate footprint marks, and invertebrate burrows are rarely present.

Figure 4.10 : Laminated siltstone/mudstone lithofacies (F/LS & F/LM).

suspension-load supply and by weak density currents close to the bed (Collinson et al., 2006). This facies could have formed in a fluvial floodplain setting (Miall, 1996; Bridge, 2003), in an abandoned fluvial channel where flow waned to zero (Collinson et al., 2006), or in an aeolian interdune setting that was subject to fluvial incursion and ponding of flood waters (Mountney & Jaggar, 2004; Mountney & Thompson, 2002).

4.6.2 Aeolian Facies

Six aeolian lithofacies have been identified from the studied succession, though two of these (A/DGFW and A/DGFA) always occur intimately together and are therefore described together below. Discrete facies are identified based on recognition of basic types of aeolian sedimentary structures associated with aeolian dune, aeolian interdune, and aeolian sandsheet deposits (Table. 4.1). These aeolian facies are described herein in detail. Commonly occurring combinations of these facies are assigned to six facies associations: aeolian dune (AD₁ and AD₂), aeolian interdune (ID₁ and ID₂), and aeolian sandsheet (SH₁ and SH₂). Aeolian lithofacies comprise 65% of the studied succession.

4.6.2.1 Aeolian Grainflow Dominated (A/DGFW) and Grainfall Dominated (A/DGFA) Sandstone Facies

Description: This facies is characterized by light-brown to pinkish-brown, fine- to medium-grained, well to moderately-sorted, sub-rounded to well-rounded quartz arenite (Fig. 4.11) arranged into cross bedded sets. This facies occurs as cross-stratified sets that are 0.3 to 1 m thick and extend laterally up to 5 m. Sets are dominated by grainflow strata; individual grainflow deposits (30 to 60 mm thick) may be separated by 1 to 4 mm-thick grainfall laminations that are themselves composed of fine-grained sandstone. The individual grainflow units are characterized internally by massive or, rarely, inversely graded sandstone. In some cases, units comprising several grainflow deposits are grouped as packages in which each grainflow unit has a diffuse boundary. The inclination of the grainflow-dominated foresets that comprise the internal structure of these sets is 20° to 25°, within which grainflow tongues progressively pinch-out in a down-dip direction into low-angle inclined wind-rippled strata and then into horizontally laminated wind-ripple strata at the dune-set base (cf. Mountney, 2006b). Cosets of several sets of cross strata are up to 1.5 m thick. This facies represents 25% of the total succession and 38.5% of the aeolian part of the succession.

In orientations parallel to palaeo-migration direction (Fig. 4.4a & 4.19), the sets of grainflow and grainfall strata form compound, scalloped-shaped and tangential cross-bedded cosets with sub-horizontal sheets of wind-ripple strata in their lower parts, and sharp to irregular bounding surfaces at their base (cf. Mountney and Thompson, 2002; Rubin, 1987b; Crabaugh & Kocurek, 1993).

Interpretation: The grainflow component of this facies represents the deposits of gravity-driven sediments on the lee-slope of aeolian dunes that arose as a result of repeated avalanching processes when the dune lee-slope attained a critical angle of repose, typically 32° to 34° for dry sand, and thus failed, thereby generating a sand avalanche (cf. Hunter, 1977; Kocurek & Dott, 1981). In contrast, the grainfall component of this facies – which is typically difficult to distinguish (cf. Hunter, 1985) – where recognised, represents airfall deposition from a cloud of saltating grains that were carried out by strong winds over dune-brink, leading to accumulation on the upper part of aeolian dune lee slopes when the wind velocity momentarily reduced (Nickling et al., 2002).

4.6.2.2 Wind-Ripple Laminated Sandstone Facies (A/DWR)

Description: This facies is characterized by light-brown to pinkish-brown, very fine- to medium-grained quartz-arenite that is well-sorted and is composed of grains that are rounded to well-rounded. Sets of this facies are 0.15 to 0.3 m thick (Fig. 4.11). Laminations are 1 to 3 mm thick and are inclined at angles of 1 to 5° (cf. Hunter, 1977, 1981). These laminations usually pass in an up-dip direction into more steeply inclined grainflow facies (A/DGFW), whereas they commonly transition downdip into crinkly and/or planar laminations of facies (A/IPL). The boundaries of the individual laminations are typically difficult to make out due to the uniformity of the grain-size. However, laminations with weak inverse grading are observed in some places, though are rare. This facies represents 16% of the total succession and 24.6% of the aeolian part of the succession.

Interpretation: This facies represents wind ripple-lamination, which is a common and widespread deposit in aeolian environments in dry interdunes, sandsheets, and on low-relief lower dune slopes inclined at low angles. This facies records deposition via tractional processes (Hunter, 1977b, 1981; Fryberger & Schenek, 1988; Collinson, et al., 2006). In this case, the low-relief strata represent subcritically climbing wind-ripple deposits formed on the lower parts of gently inclined aeolian dune lee slopes; these are translational wind-ripple strata (*sensu* Hunter, 1977 & 1981).

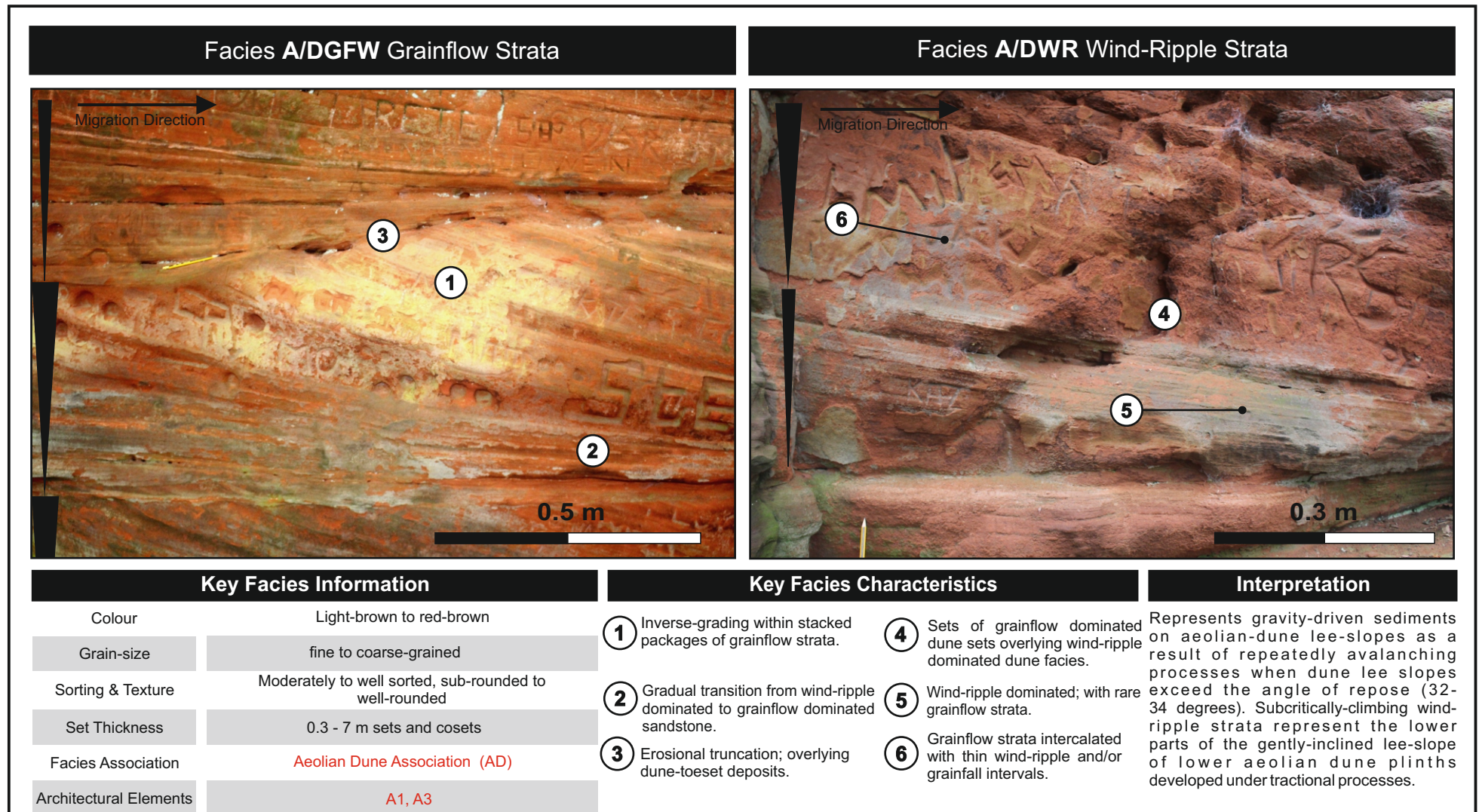


Figure 4.11: Grainflow dominated (A/DGFW) and wind-rippled-dominated (A/DWR) sandstone lithofacies.

4.6.2.3 De-watered/Convolute Sandstone Facies (A/DDW)

Description: This facies occurs as 0.1 to 1.5 m-thick sets that are characterized by light-brown to pinkish-yellow, fine- to coarse-grained, well to moderate-sorted, sub-rounded to well-rounded quartz arenite sandstone. The individual internal structures include small- to large-scale disturbed structures (2 – 13 cm in height) of previously deposited layers including convoluted and oversteepened strata that involve complex contorted instances of primary stratification (Fig. 4.12). The rotated layers form complex to gentle forms of synclines, anticlines and drag folds that reach up to 0.4 m and extend laterally up to 3 m. Internally, the boundaries of the primary layers are generally prominent and in places are massive sandstones within complex contorted zones. The layers are truncated towards the base and pass downward into non-deformed, distinctively bedded sandstones (cf. Doe & Dott, 1983). This facies represents 10% of the total succession and 15.4% of the aeolian part of the succession.

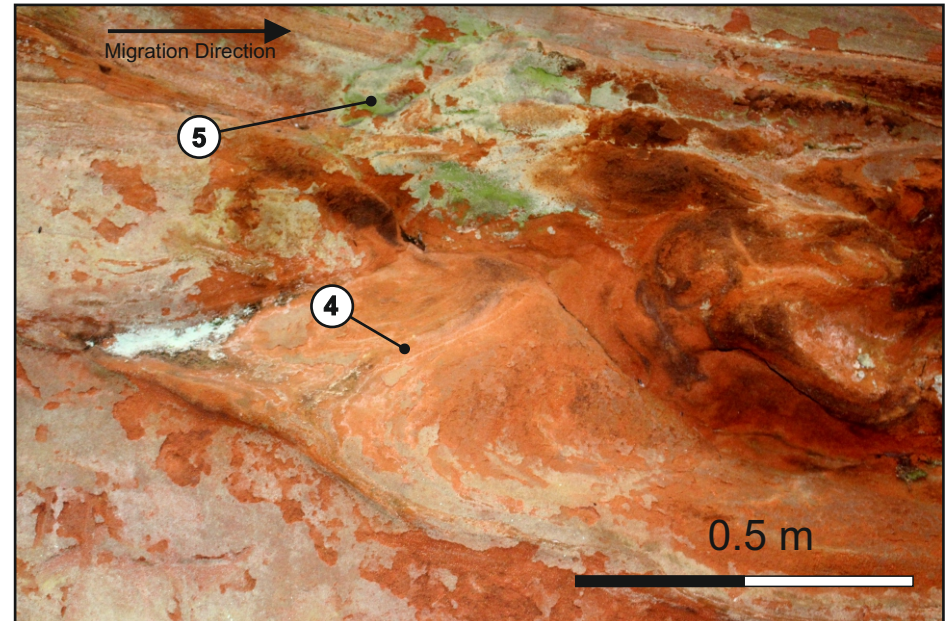
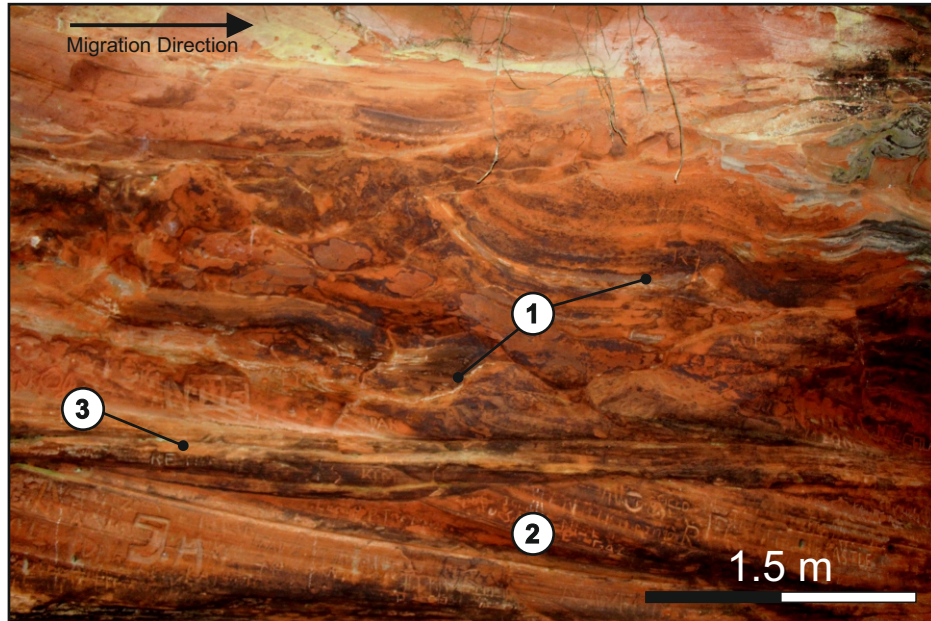
Interpretation: This lithofacies represents the accumulated product of slump degradation grainflows that arose when semi-liquefied, loosely packed sediments avalanched downslope as in-situ-water was squeezed out at the front of migrating dunes leading to the formation of complex of drag folds, flames, and over-steepened features (McKee et al., 1971; McKee & Bigarella, 1972; Doe & Dott, 1980; Owen, 1987),

4.6.2.4 Crinkly Laminated Sandstone Facies (A/IWL)

Description: This facies occurs as 0.1 to 0.3 m-thick sets of very fine- to fine-grained, well-sorted, sub-rounded to well-rounded sandstone (Fig. 4.13). Internally, 1 to 2 mm-thick plane-laminae are characterized by irregular crenulations. Stacked sets form cosets of this facies, which attain thicknesses up to 0.7 m and extend laterally for 3 to 16 m (e.g. panel H and I in Fig. 4.19). The boundaries of poorly developed wavy laminations are typically difficult to distinguish due to grain-size uniformity. The irregular waviness of laminations in this facies gradually grades vertically and horizontally into either wind-rippled laminated sandstones (A/DWR) or planar laminated sandstones (A/IPL). Convolute laminations are common in this facies but are rare in facies A/IPL. This facies represents 10% of the total succession and 15.4% of the aeolian part of the succession.

Interpretation: This facies represents water-lain or water-influenced deposits of damp interdunes or sandsheets that accumulated under conditions of short- or long-lived elevated or fluctuating water table brought

Facies A/DDW Convolved Bedded Sandstone



Key Facies Information

| | |
|------------------------|--|
| Colour | Light-brown to red-brown |
| Grain-size | fine- to medium-grained |
| Sorting & Texture | Moderately sorted, sub-rounded to well-rounded |
| Set Thickness | 0.1 - 2.8 m sets and cosets |
| Facies Association | Aeolian Dune Association (AD) |
| Architectural Elements | A1 |

Key Facies Characteristics

- ① Large-scale deformation (folds) within aeolian strata preserving the primary bedding; indicates ductile buckling of strata.
- ② Non-deformed strata of grainflow and wind-ripple origin.
- ③ Wavy lamination underlying the deformed strata indicative of wet interdune with rare deformation.
- ④ Remnant lamination still evident; demonstrates that the soft-sediment deformation was post-depositional.
- ⑤ Limited deformation in upper parts of sets.

Interpretation

This facies indicates rapid and catastrophic water escape from grainflow-dominated dune lee slopes as in-situ-water is squeezed out from the front of migrating dunes forming complex of drag-folds, flames, and over-steepened features.

Figure 4.12: Convolved bedded sandstone lithofacies (A/DDW).

by either fluvial floods (Langford, 1989; Stanistreet & Stollhofen, 2002) or rains (Hummel & Kocurek, 1984; Kocurek & Fielder, 1982). The poorly developed crinkly laminations can form on a damp surface in the aftermath of a falling water-table whereby aeolian sand adheres to the damp surface (Hunter, 1980b; Kocurek & Fielder, 1982). Alternatively, the wavy lamination may record deposition from weak, sub-aqueous currents that acted to modify previously generated aqueous current ripples (Kocurek, 1981). The convolute laminations in this facies are indenter marks, preserved animal vertebrate foot imprint marks; such structures indicate that moisture was at or above interdune surface (Fig. 4.10, key facies characteristics No.5).

4.6.2.5 Planar Laminated Sandstone Facies (A/IPL)

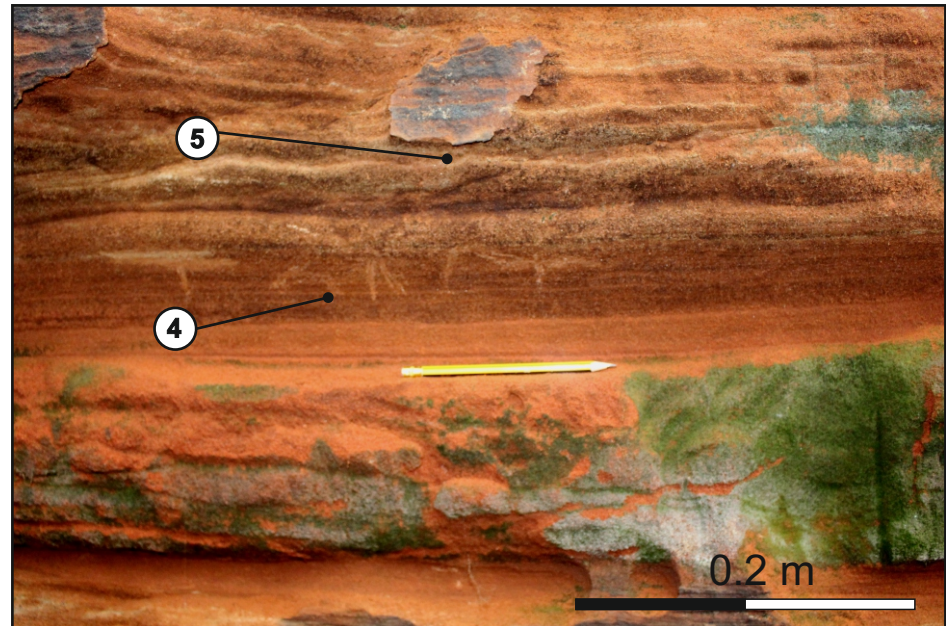
Description: This facies occurs as 0.1 to 0.3 m-thick sets of very fine- to fine-grained, well-sorted, sub-rounded to well-rounded sandstone characterized internally by parallel and horizontal-laminations. Silt grains (maximum 20%) are in some cases mixed with the dominantly sandy sediment. This lamination is characterized by 3 mm-thick pin-stripe laminae with weak inverse-grading and horizontal sharp bases. Boundaries of the individual laminations are typically difficult to distinguish due to the well-sorted nature of the sediment and a paucity of grain segregation and grain-size differentiation (Fig.14.3). The inclination of these laminations varies from 0° to 10° and examples of sets of this facies commonly exhibit a gradual lateral or vertical transition into-low angled-inclined wind-rippled-laminations (A/DWR). This facies represents 4% of the total succession and 6.2% of the aeolian part of the succession.

Interpretation: The planar laminations represent the deposits formed by tractional processes under a high wind velocity (Hunter, 1977b). The pinstripes are wind-ripple laminations that climbed at very low angles (Fryberger & Schenk, 1988, Kocurek, 1981). This facies therefore represents accumulation on a planar, near-horizontal or low-angle-inclined dry surface subject to aeolian processes (dry interdune flat or dry aeolian sandsheet). The structures indicate accumulation on an accumulation surface that was not influenced by the presence of moisture (Kocurek, 1981; Mountney, 2006a).

4.7 Facies Associations

Five facies associations are recognized within the fluvial and aeolian deposits of the Helsby Sandstone Formation encountered in this study:

Facies A/IWL Crinkly Laminated and Facies A/IPL Planar Laminated



Key Facies Information

| | |
|------------------------|---|
| Colour | Red-brown to dark red-brown |
| Grain-size | fine- to very fine-grained |
| Sorting & Texture | Well to very well sorted, sub-rounded to well-rounded |
| Set Thickness | 0.05 - 0.5 m sets and cosets |
| Facies Association | Aeolian Interdune Association (ID) |
| Architectural Elements | A1, A2 |

Key Facies Characteristics

- ① Sets of irregular and wavy lamination indicative of wet interdune.
- ② Wind-ripple strata toesets downlapping conformably onto wet interdune element.
- ③ Sharp contact separating damp interdune element and underlying aeolian dune strata.
- ④ Planar lamination suggesting that the interdune was characterized by a dry surface.
- ⑤ Gradual upward transition from planar to crinkly lamination, indicative of progressive wetting trend.

Interpretation

Planar lamination represents wind-ripple deposits of dry-interdune flats developed between migration aeolian-dunes. Sets of crinkly lamination indicate deposition under damp surface conditions of damp and/or wet aeolian interdunes that developed in response to fluctuating short- and long-term water-level variation.

Figure 4.13: Wavy laminated (A/IWL) and Planar Laminated (A/IPL) facies.

fluvial channelized, fluvial non-confined sheet, aeolian-dune, aeolian-interdune, and aeolian-sandsheet. Each of these facies associations is characterized by a predictable vertical succession of facies that itself records the expression of a distinctive sub-environment within the mixed fluvial and aeolian depositional system. The description of these facies associations and their typical vertical facies successions enables the record of sedimentary processes that gave rise to the deposition, accumulation, erosion, and preservation to be established.

Five commonly occurring fluvial facies successions are assigned to two fluvial facies associations: channelized fluvial (CF) and non-confined sheet-like fluvial (NC). Six commonly occurring aeolian facies successions are assigned to three aeolian associations: aeolian dune (AD), aeolian interdune (ID), and aeolian sandsheet (SH). Each succession and association reflects the nature of sedimentary processes of entrainment, transport, and deposition; they form the internal components of the architectural elements discussed in section 4.8.

4.7.1 Channelized Fluvial Facies Association (CF)

Facies successions CF1, CF2, CF3 and CF4 represent examples of common vertical successions arising from deposition within the confines of a fluvial-channel; they mainly comprise facies of coarser grained sediments and are characterized by erosional bounding surfaces that define the base of fluvial cycles (Fig. 4.14). The channelized fluvial facies association comprises 85% of the total fluvial sandstones encountered in this study.

Succession CF₁

This succession is composed of medium- to coarse-grained sandstone, and is dominated by facies F/LTX and F/HTX in which cosets (up to 5 cosets) are stacked vertically such that each forms a weakly fining-upward unit within an overall fining-upward succession or cycle (Fig. 4.14). Within a coset, individual sets vary in preserved thickness from 0.25 to 0.8 m and set thickness generally decreases toward the top of the coset. The base of each individual set is always dominated by extraformational pebble clasts (F/EFQ) and intraformational pebble clasts (F/IFQ) of variable size, in which their abundance decreases vertically upward in the set. The overall preserved thickness of each coset is 2 to 3 m. The irregular basal bounding surfaces of these successions cuts down abruptly and dips at angles of 10 to 20 degrees. The succession forms 70% of the total proportion of channelized fluvial association in this study.

CF - Channelised Fluvial Facies Association

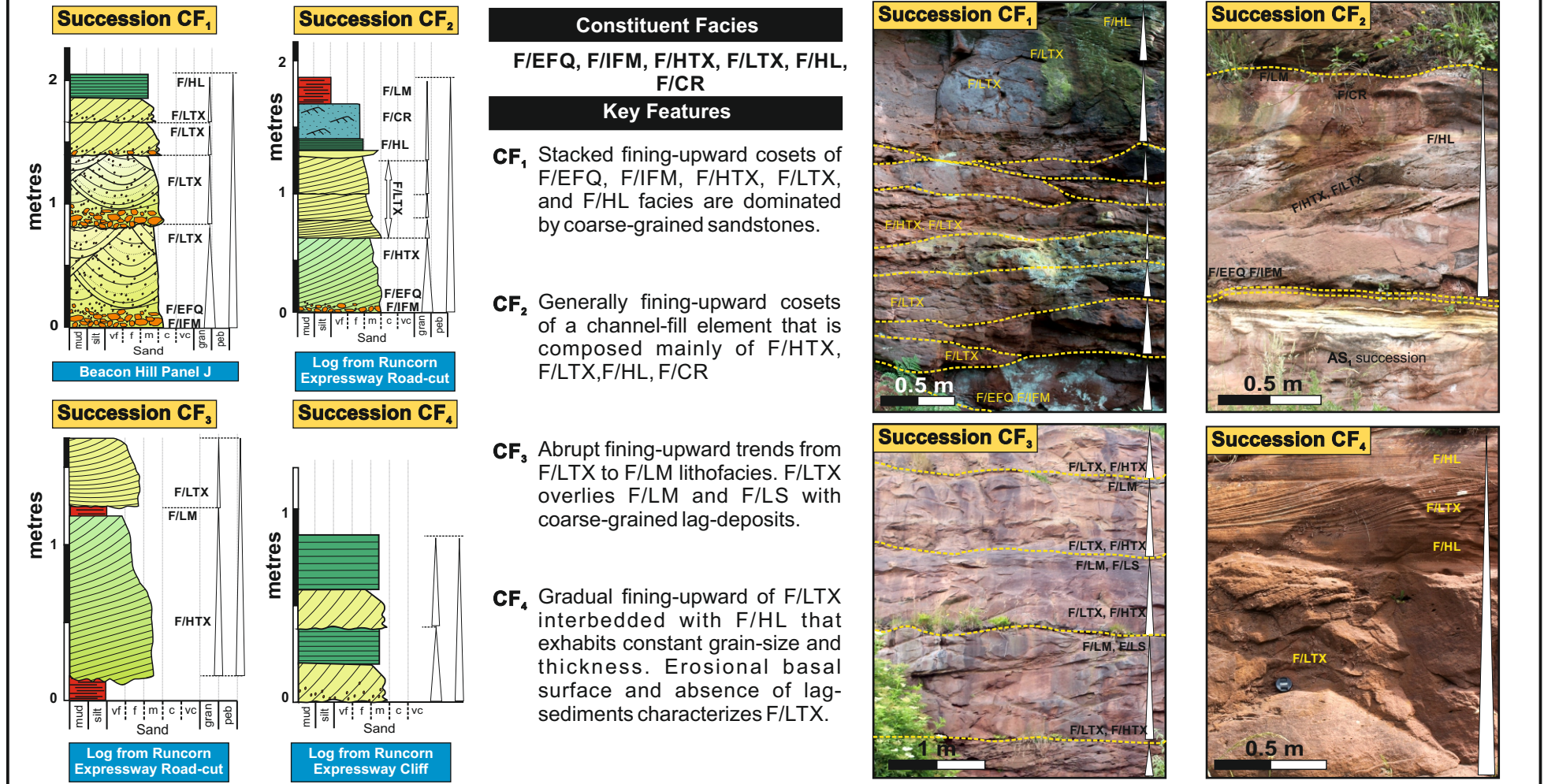


Figure 4.14: Channelized fluvial facies association (CF).

Succession CF₂

This succession represents a “classic” fining-upward channel-fill succession. It is characterized by both intra- and/or extraformational clasts at the base (F/EFQ and F/IFQ), high- and low-angle cross bedding (F/HTX and F/LTX), horizontal lamination (F/HL), current-ripple lamination (F/CR), and laminated mudstone (F/LM) at the top of the succession. Vertically stacked sets within this succession typically decrease upward from 0.7 to 0.1 m thick (Fig. 4.14). The preserved thickness of a typical complete succession is 2 to 3 m. Examples of this succession comprise 10% of the total proportion of the channelized fluvial association.

Succession CF₃

Although not common within the study region, examples of this succession are recorded at Runcorn Expressway within the Thurstaston Sandstone Member (TSM). This succession is characterized by an abrupt upward-fining trend in its upper part. Internally, the succession comprises arrangements of facies F/LTX and F/LM, which occur interbedded. Facies F/LTX passes upward abruptly into mudstone/siltstone facies (F/LM and F/LS) (Fig. 4.14). The preserved thickness of this succession is up to 2 m. Examples of this succession comprise 5% of the total proportion of the channelized fluvial association in this study.

Succession CF₄

This succession is characterized by a gradual fining-upward trend of facies F/LTX and F/HL in which individual sets occur interbedded in a repeating order. The base of the succession is defined by an irregular bounding surface inclined at angles of up to 5° to 10°. Sets directly overlying such erosional surfaces lack any coarse-grained pebble lag deposits. At the base of the succession, the grain-size of facies F/LTX exhibits a slight fining-upward trend of fine- to medium-grained sandstones, whereas the overlying sets of facies F/HL tend to be of a relatively constant grain-size of fine- to medium-grained sandstone with no trends throughout the succession. The preserved thickness of this succession is up to 1 m and forms 15% of the total proportion of the channelized fluvial association (Fig. 4.14).

4.7.2 Non-confined Sheet-Like Fluvial Facies Association (NC)

In contrast to the vertical facies successions that are representative of the channelized facies association, the non-confined, sheet-like fluvial association (NC) is characterized by a single succession (NC1), which represents fluvial floodplain deposits composed predominantly of relatively

fine-grained sediments with an absence of erosional features (Fig. 4.15). This association comprises 15% of the total proportion of fluvial deposits in this study.

Succession NC₁

This succession is primarily characterized by lithofacies F/LM and F/LS. The best exposed example of this succession is in the Helsby Quarry section (Fig. 4.15). In this succession, thin sets of F/LS interbedded with thicker sets of F/LM occur in a repeating order. The interbedded facies F/LM are of variable thickness, with individual beds varying from 0.2 to 0.6 m thick; by contrast, sets of facies F/LS are generally of constant thickness (~0.2 m). Examples of this succession typically have an overall preserved thickness of up to 2 m (Fig. 4.15). The succession forms 100% of the total proportion of non-confined sheet-like fluvial associations in this study.

4.7.3 Aeolian Dune Facies Association (AD)

Two distinct types of aeolian dune facies association (AD) reflect the preserved sedimentary expression of migrating aeolian dunes (Fig. 4.16): AD₁ and AD₂ are similar but are distinguished by minor differences in depositional style. The aeolian dune association comprises 70% of the total proportion of the aeolian deposits examined in this study.

Succession AD₁

This succession is characterized by different aeolian lithofacies including, from base to top, facies A/IPL, A/DWR and A/DGFW, which occur as sets that together comprise a compound coset, examples of which attain an overall thickness up to 2 m. Examples of this succession are most readily observed in sections parallel to aeolian transport. Individual compound cosets show an overall coarsening upward that is bounded by a low-angle-inclined (0° to 5°) interdune migration surface at the base. Internally within the coset, individual grainflow-dominated sets (A/DGFW) are commonly bounded by scalloped-shaped reactivation surfaces inclined at 10 to 25 degrees. Sets of this facies and the reactivation bounding surfaces commonly merge downdip with inclined wind-ripple strata (A/DWR). Stacked compound cosets are commonly arranged into vertically repeating cycles. These cyclic inverse-graded cosets are common within the Helsby Sandstone Formation in the northern cliff of Beacon Hill, where the outcrop attains an overall thickness of 3.5 m (Fig. 4.16). This succession represents 85% of the total proportion of the aeolian dune association.

Succession AD₂

This succession is similar to AD₁ in that coarsening-upward cosets are common; however it differs in that facies A/DDW is dominant in the upper parts, and facies A/DGFW is absent from this succession. The large-scale deformation (A/DDW) extends downslope from the top of the succession towards aeolian dune palaeomigration direction (ENE to WSW) for up to 5 m and merges with either inclined wind-rippled strata (A/DWR) or planar strata (A/IPL). Examples from the northern cliff of Beacon Hill show that the deformation is intense in places. Cosets within examples of this succession have an overall inverse grading and can be up to 2 m thick (Fig. 4.16). This succession represents 15% of the total proportion of the aeolian dune association.

4.7.4 Aeolian Interdune Facies Associations (ID)

The ID facies association represents the preserved sedimentary expression of accumulation in interdune settings, in which either aeolian or fluvial processes dominate. Two discrete interdune successions, ID₁ and ID₂, are recognised; both record predictable cyclic arrangements of interdune lithofacies in which A/IPL and A/IWL are dominantly interbedded and pass upward into overlying dune lithofacies (Fig. 4.17). This succession forms 20% of the total proportion of the aeolian deposits in this study.

Succession ID₁

This interdune succession comprises facies A/IWL and A/IPL that are composed overall of very fine- to fine-grained sandstone. Small-scale (~10 cm) coarsening upward trends occur within examples of this succession. Vertical trends tend to record an upward transition from facies A/IWL to A/IPL (Fig. 4.17). The overall preserved thickness of examples of this succession is up to 1 m and laterally examples are up to 5 m in extent. The succession occurs below aeolian dune successions in the Beacon Hill and the lateral transition between the interdune and dune the facies is here represented by a gradual upward transition from wavy laminations of damp interdune origin into wind-ripple lamination of the overlying aeolian dune toeset deposits, and up into grainflow lamination of the dune lee-slope deposits. This succession comprises 15% of the total proportion of the aeolian dune association and 25% of the total proportion of the aeolian interdune association.

Succession ID₂

Succession ID₂ is characterized by variable arrangements of aeolian lithofacies A/IWL, A/IPL, A/DWR and A/DGFW that occur as compound sets

NC - Non-confined Sheet-Like Fluvial Facies Association

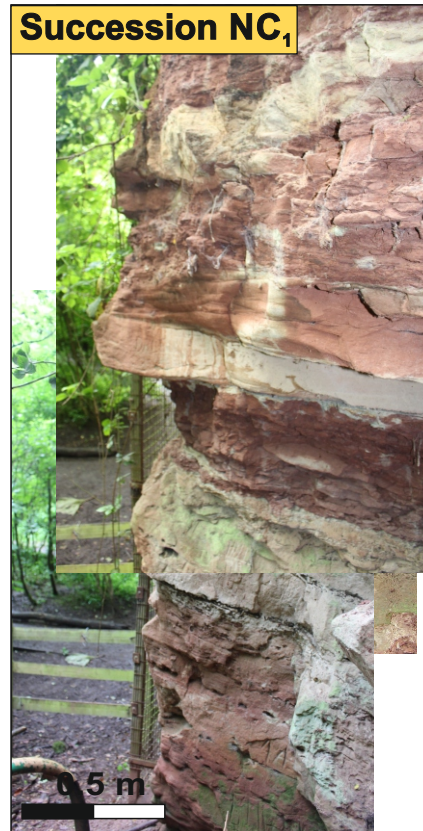
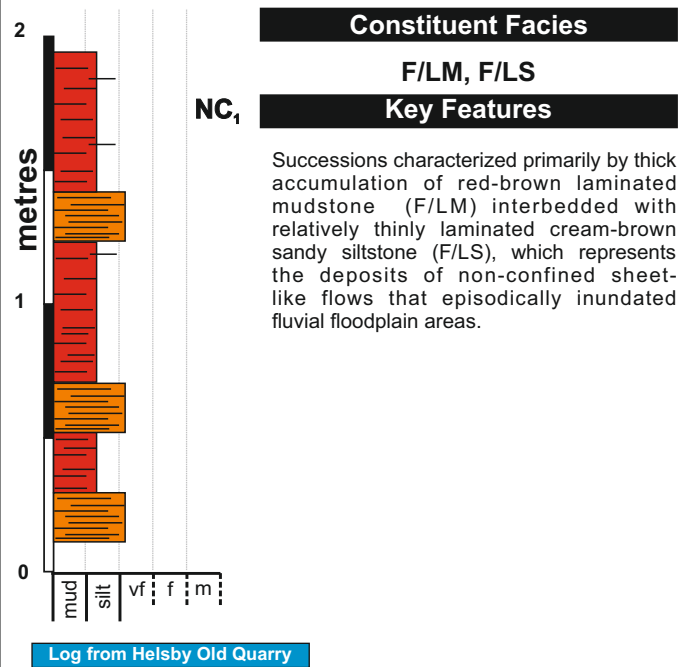


Figure 4.15: Non-confined fluvial sheet facies association (NC).

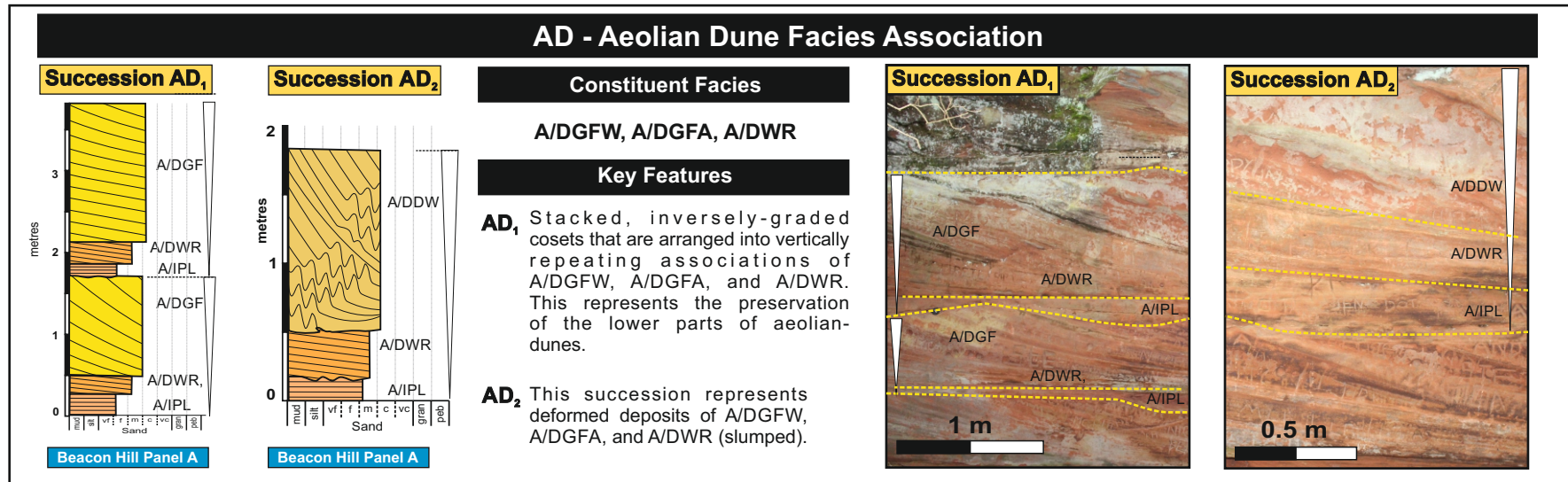


Figure 4.16: Aeolian dune facies association (AD).

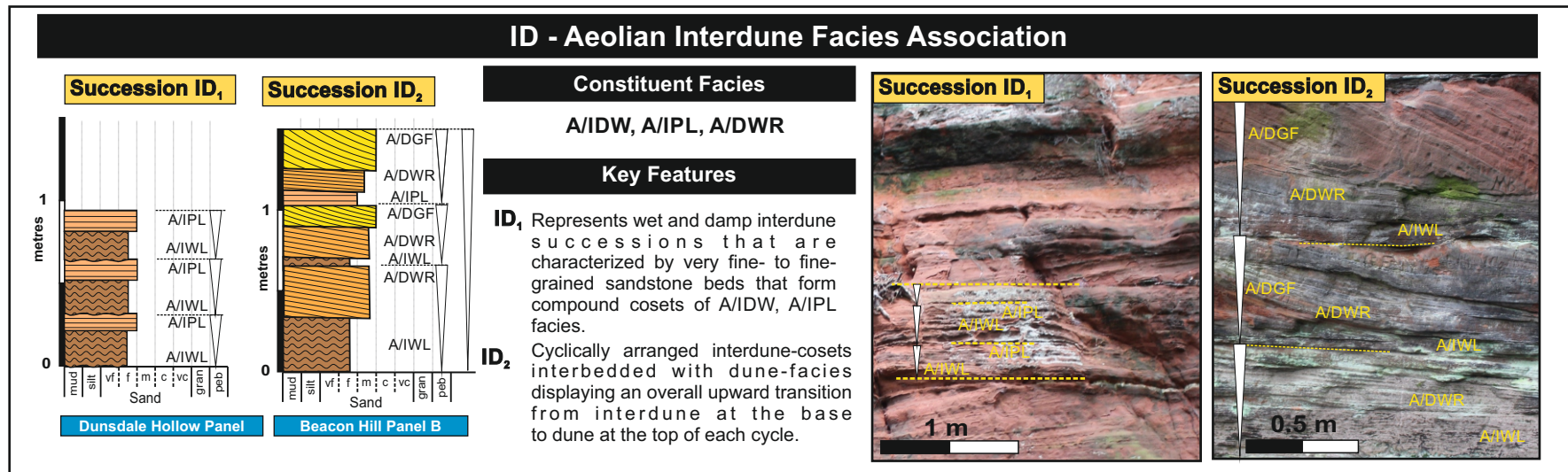


Figure 4.17: Aeolian interdune facies association (ID).

of variable thickness (3 to 25 cm). The stacked compound sets form cyclic cosets that exhibit an overall inverse grading and which reveal a gradual upward transition from interdune to dune lithofacies with an overall drying-up trend (Fig. 4.17). The overall preserved thickness of this succession is up to 1.5 m. This succession comprises 75% of the total proportion of the aeolian interdune association.

4.7.5 Aeolian Sandsheet Facies Associations (SH)

The aeolian sandsheet facies association (SH) represents the preserved sedimentary record of deposition within aeolian sandsheets – typical of inner and outer erg-margin settings – where lithofacies associations representative of large-scale aeolian-dunes are absent. Within this association, two styles of sandsheet are identified (SH₁ and SH₂) and these record vertical aeolian lithofacies transitions that reflect the variation of depositional processes within aeolian-sandsheet areas. The aeolian sandsheet association comprises 10% of the total proportion of the aeolian deposits in this study.

Succession SH₁

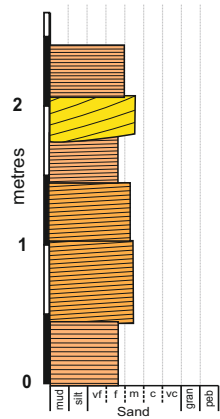
This succession is dominated by facies A/IPL and A/DWR, with an absence of facies A/DGFW and A/IWL. The succession is characterized by a slight coarsening-upward trend. The thickness of the sets that comprise examples of this succession is variable (0.1 to 0.5 m). Sets of facies A/DWR in this succession are typically thicker (0.2 to 0.5 m) than in the interdune successions (ID₂) in which they occur as 0.1 to 0.3 m-thick sets of A/IWL and A/IPL interbedded with A/DWR sets. The best-developed example of this succession is in the upper northeast part of the cliff of the Dunsdale Hollow locality, where the succession has an overall preserved thickness up to 2.5 m; however, here it is difficult to distinguish facies A/IPL and A/DWR due to the weathered nature of the exposure (Fig. 4.18). This succession forms 55% of the total proportion of the aeolian sandsheet association.

Succession SH₂

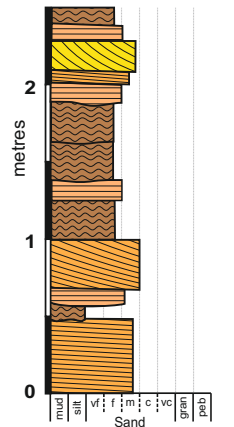
Succession SH₂ is composed primarily of facies A/IWL with only rare occurrence of facies A/DWR and A/IPL. The example in Fig. 4.13 (key facies No.1) depicts stacked cosets of thick crinkly strata of A/IWR arranged in sets, each of which is truncated by a horizontal bounding surface. The sets of crinkly laminated sandstone (A/IWL) are each 0.2 to 0.3 m thick. This succession occurs in study localities as stacked sets with an overall thickness up to 1 m in and lateral extension up to 5 m (Fig. 4.18). This

SH - Aeolian Sandsheet Facies Association

Succession SH₁



Succession SH₂



Constituent Facies

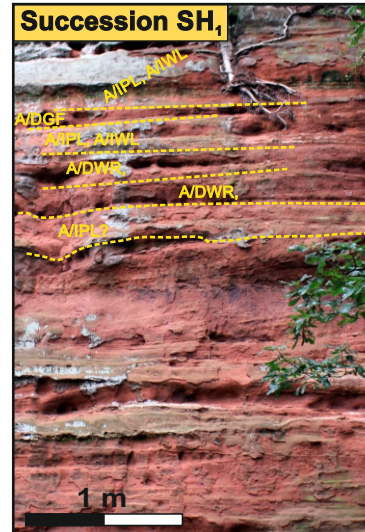
A/DWR, A/IWL, A/IPL,
A/DGFA

Key Features

SH₁ Succession consists primarily of stacked thick-cosets of wind-ripple strata of facies A/IPL and A/DWR with occasional A/DGF.

SH₂ Succession composed commonly of very fine- to fine-grained sandstone with rare medium- to coarse-grained sandstone. This represents sandsheet deposits in the outer-marginal settings of aeolian dune-fields.

Succession SH₁



Succession SH₂

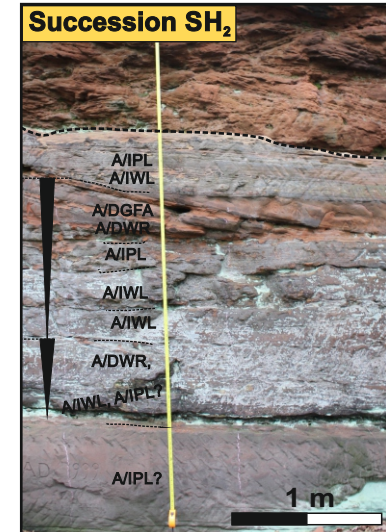


Figure 4.18: Aeolian sandsheet facies association (SH).

succession forms 45% of the total proportion of the aeolian sandsheet association.

4.7.6 Interpretation of Facies Associations

The channelized fluvial facies association (CF) described above records the preserved expression of deposits that accumulated under a variable set of processes that operate in confined fluvial channels. Successions CF1, CF2, CF3, and CF4 are similar and each records the in-channel migration of bedforms under the influence of episodic flow of high-energy-fluvial-discharge. Rapid-reworking of parts of the adjoining overbank region was common, thereby generating locally-derived rip-up mud clasts (Reid & Frostick, 2011, p. 324), that were carried and moved as bed load along the channel bases, before becoming preserved in the lower parts of fining-upward channel-fill depositional cycles (Bridge, 2006, p. 107). The predominance of coarse-grained sandstone and the fining-upward trends of these successions records relatively rapid deposition of sandstones via bedform migration during and in the immediate aftermath of fluvial-flood events (Stear, 1983). The finer-grained upper parts of the successions record a variety of forms of suspension-load deposition. Differences in the four recognized types of CF succession arise from the presence of interbedded mudstone layers in examples of CF2 and CF3; these suggest that suspension-load sediments were deposited in response to a decrease in fluvial flow velocity (Bridge et al., 2008, p. 147), whereby waning flow conditions arose in response to channel abandonment and channel avulsion processes (Slingerland & Smith, 2004; Mohrig et al., 2000; Bridge et al. 2008, p. 441; Miall, 1996, p. 177-178).

In contrast to CF successions, the non-confined facies association (NC₁), examples of which are well exposed in the Old Quarry section at Helsby Hill (Fig. 4.15), records the preserved expression of deposition from suspension in flooded fluvial overbank settings. Such settings were characterized by repeated drying out as demonstrated by the presence of desiccation cracks and associated rare vertebrate footprints and invertebrate burrows (cf. Mountney & Thompson, 2002; King & Thompson, 2000). These deposits are likely to be more prevalent in the preserved succession but are under-recognized because they tend not to weather to form clean outcrops.

The aeolian facies association (AD) in this study records the accumulation of downwind- and oblique-migrating aeolian dunes (Mountney & Thompson 2002; Mountney, 2006a). Episodic avalanches down the lee slopes of these dunes generated the grainflow-dominated packages of cross strata of facies

A/DGFW (Hunter, 1977, 1981; Kocurek & Dott, 1981). Reactivation surfaces are locally common, suggesting that dune lee slopes underwent episodic (or possibly periodic) reworking by winds that varied in their direction or intensity (Brookfield, 1977). By contrast, the aeolian interdune successions (ID) record deposition within hollows or flat-based corridors between migrating aeolian dunes (Mountney, 2006a); many such interdunes are here interpreted to have acted as corridors along which fluvial flood waters episodically passed (cf. Stanistreet & Stollhofen, 2002; Hummel & Kocurek, 1984). As such, these damp or wet interdunes represent corridors that facilitated fluvial incursion into aeolian dune fields (cf. Cain & Mountney, 2011; Al-Masrahy and Mountney, 2015).

Where an intimate association with overlying aeolian dune facies can be demonstrated, an interdune setting for deposits of crinkly lamination of damp interdunes can be established (Mountney, 2006a,b). The transition between the interdune and the dune facies is represented by a gradual upward transition from crinkly laminations of the damp interdune into wind-ripple lamination of the dune toeset and up into grainflow lamination of the dune lee slope (Crabaugh & Kocurek, 1993). Such relationships demonstrate that both the interdune and adjacent dune facies developed simultaneously (Mountney and Thompson, 2002; Pulvertaft, 1985) and accumulated via bedform climbing whereby the succeeding dune migrated over the preceding interdune.

Successions of aeolian sandsheets (SH) represent the accumulation of aeolian sediments that were deposited in flat-areas where sediment supply was restricted such that large-scale aeolian dunes were not constructed, or at least did not accumulate (Kocurek & Nielson, 1986). SH₁ successions are mainly dominated mainly by wind-ripple strata that reflect dry substrate conditions. Due to the dominance of lithofacies indicative of accumulation on a damp or wet substrate (A/IWL), succession SH₂, is interpreted to record accumulation under the influence of a damp or wet substrate, influenced by mixed of adhesion processes and non-confined distal fluvial flood processes (cf. "sheetflood" of Cowan, 1993).

4.8 Palaeocurrent Data

Two-hundred-and-fourteen palaeocurrent measurements (recorded as dip-azimuth of cross-bedding foresets) have been measured from the outcrops of the Beacon Hill and Dunsdale Hollow localities near Frodsham, from both fluvial and aeolian sandstones, to assist with architectural-element analysis.

In the major cliff section at Beacon Hill (Fig. 4.19), the outcrop of which is oriented 050° to 230° (Fig. 4.19), two rose-diagrams represent fluvial and aeolian palaeocurrent directions of the lowermost part of the overlying DSM (Fig. 4.4a). Most of the prominent sets of fluvial trough cross-strata exposed in the Beacon Hill outcrop (panel J, Fig. 4.19) are evident from sections oriented perpendicular to the palaeo-fluvial flow direction with a mean direction of 321° of palaeochannel direction ($n = 48$; standard deviation = 17) (Fig. 4.4a). The trends of the palaeocurrent readings of fluvial strata at Beacon Hill indicate that most of the fluvial-bedforms preserve a record of downstream migration in fluvial channels that passed towards the northeast. By contrast, the inclined aeolian strata (A/DGFW and A/DWR) are most clearly exposed in sections parallel to the palaeo-transport direction of the aeolian dunes with a mean direction of 252° in the Beacon Hill locality (Fig. 4.4a). In the major cliff section at Dunsdale Hollow (Fig. 4.20), which is oriented 160° to 340° , the palaeocurrent measurements are from the uppermost part of the aeolian Thurstaston Sandstone Member (TSM). The inclined foreset strata of the aeolian dunes in this section record net sand transport and dune migration towards WSW, with a mean foreset azimuth direction of 241° ($n = 39$; standard deviation = 20) (Fig. 4.4b).

4.9 Architectural Elements

Architectural elements are defined as constituents of a depositional system with characteristic facies associations that are interpretable as sub-environments and which can be described in terms of their three-dimensional form and geometry (Colombera et al, 2013). Architectural elements in this study are classified into two main groups: fluvial and aeolian. Recognised elements possess unique geometric properties and internal facies associations. Analysis of the geometrical arrangement of these elements has enabled the generation of high-resolution, three-dimensional, quantitative architectural models. Overall, six architectural elements are identified from examined outcrops of the Helsby Sandstone Formation (Fig. 4.21). Three are of fluvial origin: multi-storey, multi-lateral amalgamated channel-fill complex (F1); single-storey, multi-lateral amalgamated interbedded channel-fill (F2); and fluvial floodplain (F3). The other three are of aeolian origin: aeolian dune (A1); aeolian interdune (A2); aeolian sandsheet (A3).

Beacon Hill Architectural Panle

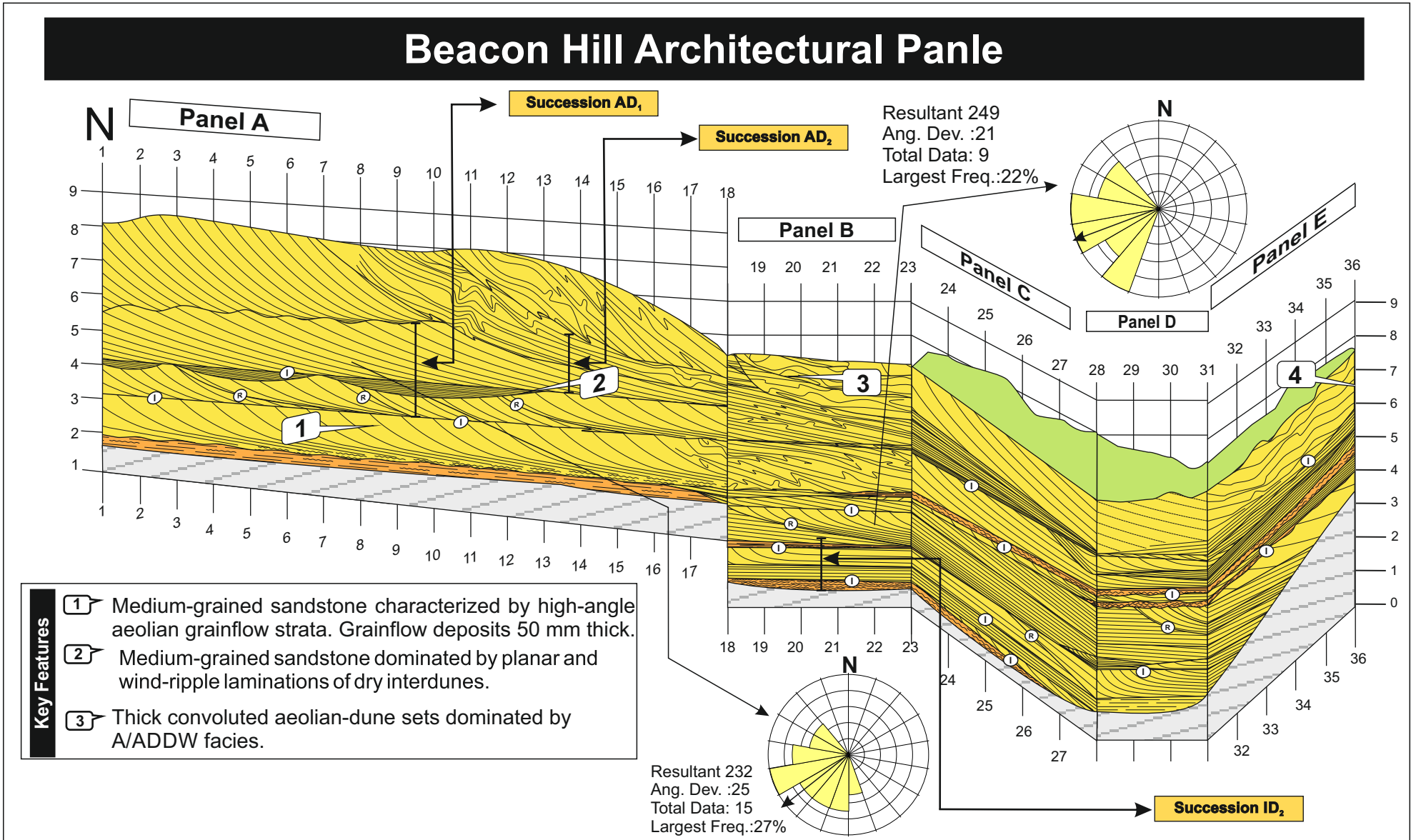
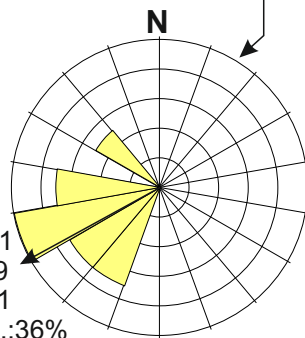
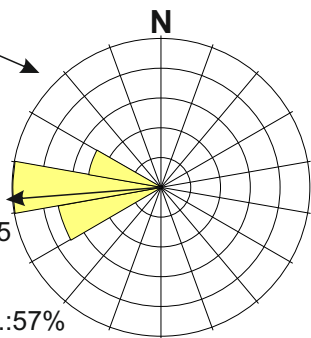
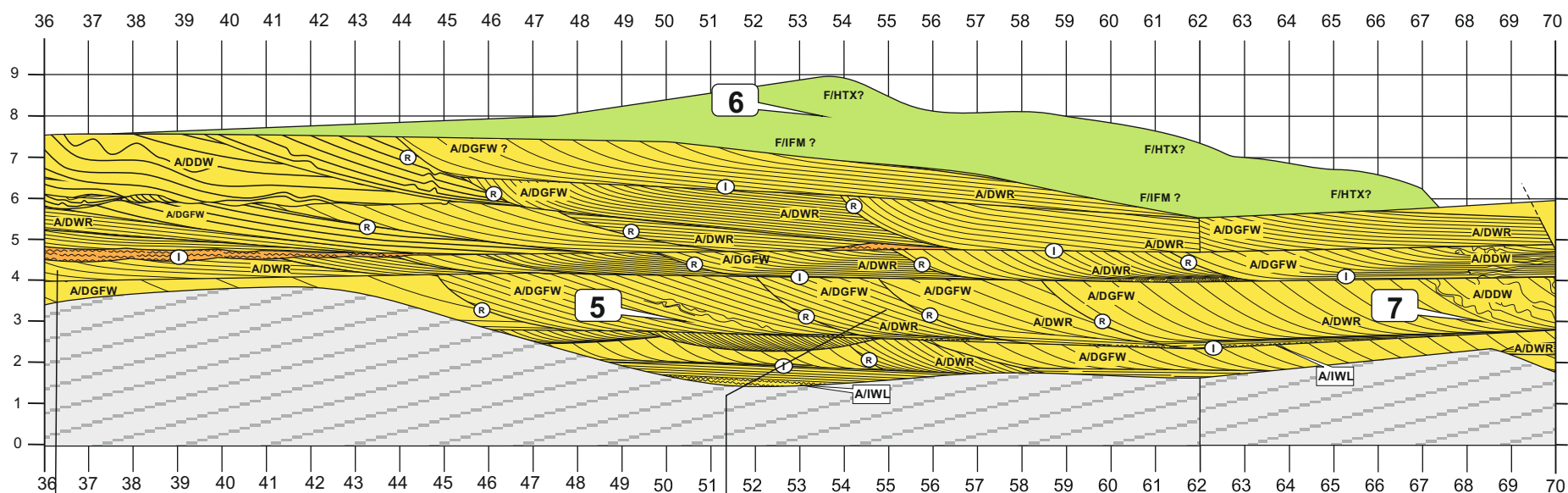


Figure 4.19: Composite diagram of architectural panels depicting the stratigraphic architecture of the Delamere Member as observed at Beacon Hill. Individual panel names are referred to throughout the text. See figure 4.2 and 4.29 for location.

Beacon Hill Architectural Panel

Panel F



- Key Features**
- 5 Medium-grained grainflow dominated sandstone with small-scale slump structures.
 - 6 Channelized fluvial successions (CF1, CF2) cut into aeolian successions (SH2, AD1, AD2).
 - 7 Medium- to coarse grained sandstone of thick convoluted aeolian-dune sets.

Figure 4.19: Continued.

Beacon Hill Architectural Panel

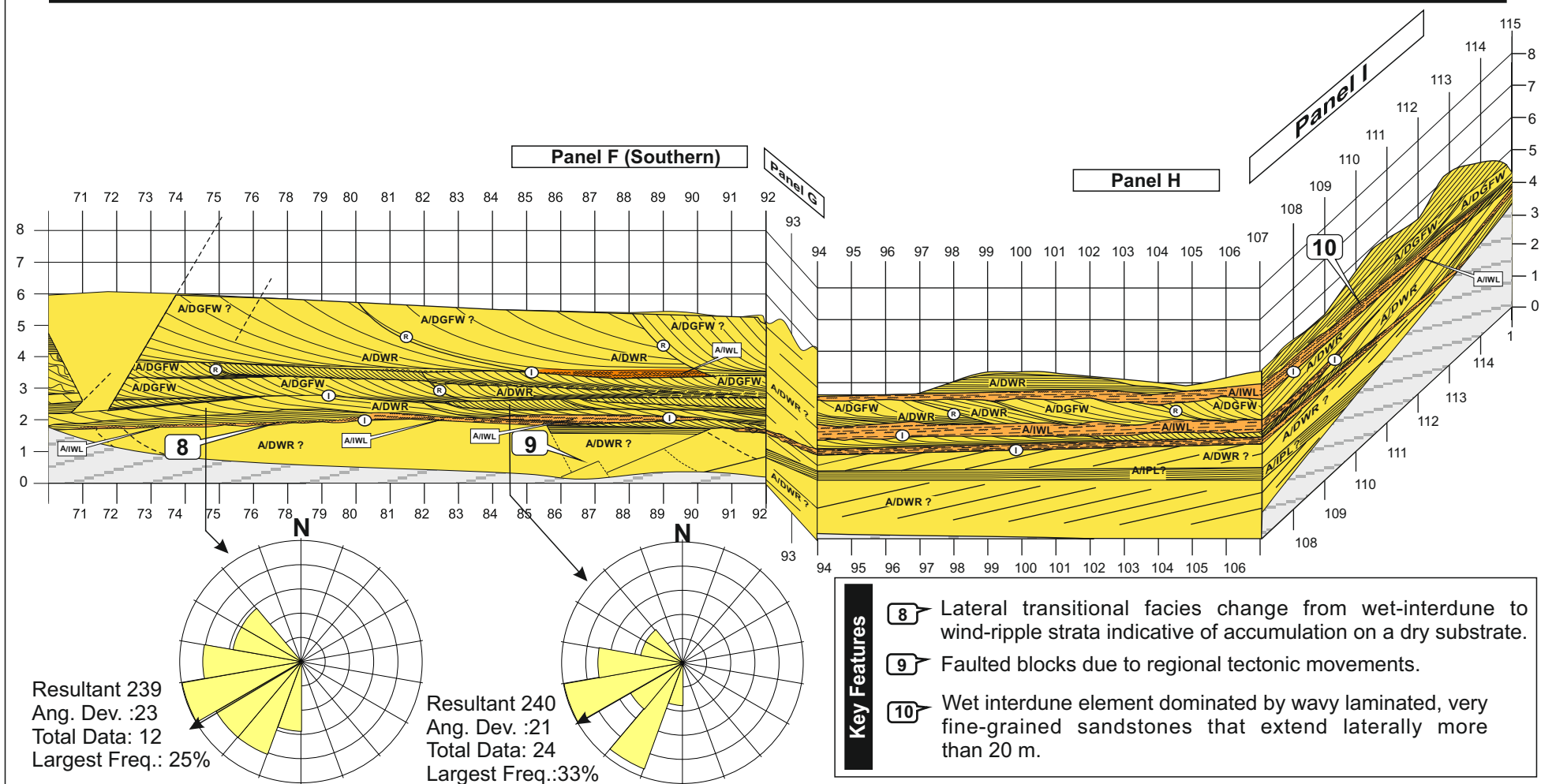


Figure 4.19: Continued.

Beacon Hill Architectural Panel

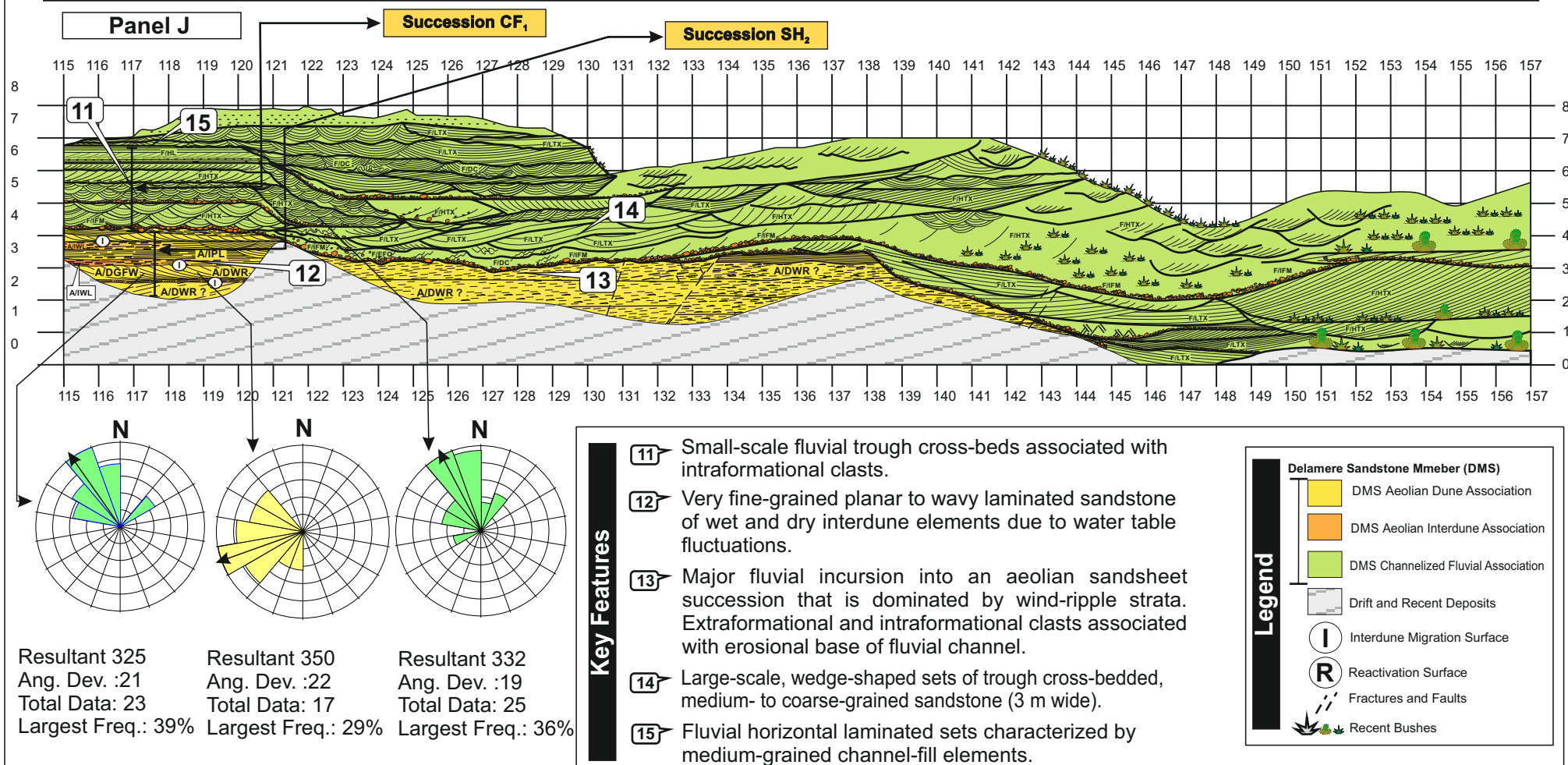


Figure 4.19: Continued.

Stratigraphic Architecture from Dunsdale Hollow, Frodsham, UK

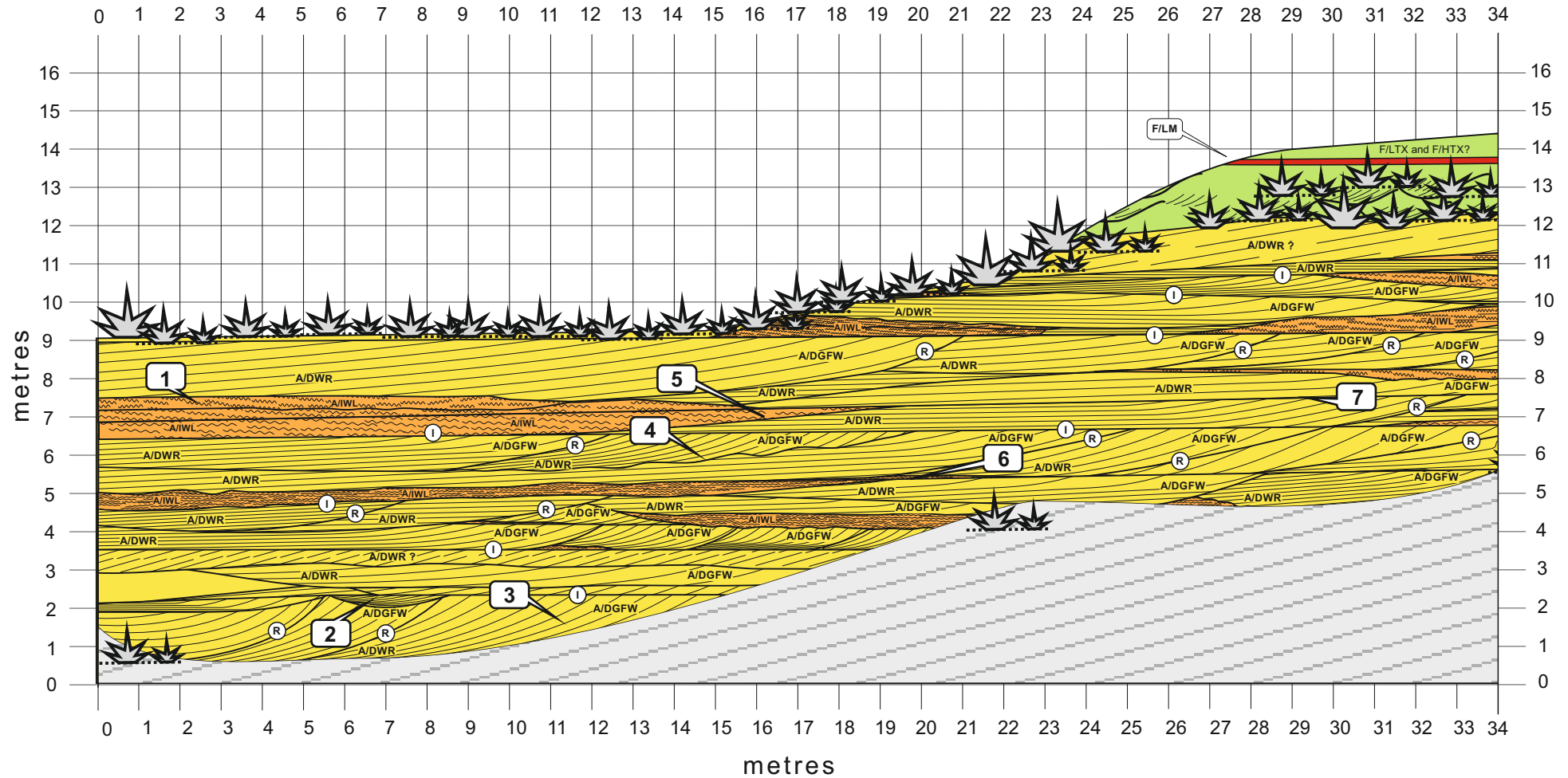
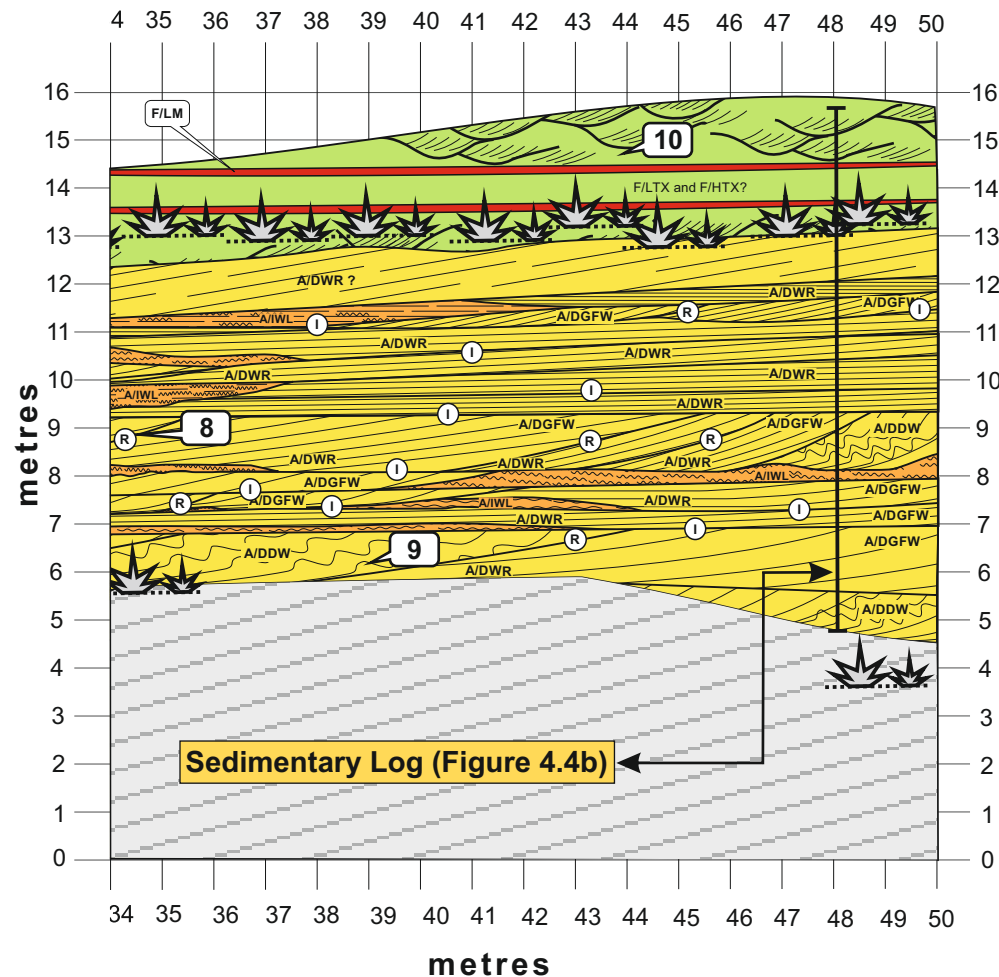


Figure 4.20: Architectural panel depicting the stratigraphic architecture of the aeolian strata of the Thurstaston Sandstone Member as observed in Dunsdale Hollow. See figure 4.2 and 4.29.

Stratigraphic Architecture from Dunsdale Hollow, Frodsham, UK

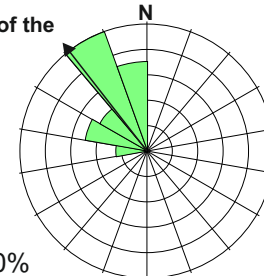


Features of Sedimentological Interest

- 1 Stacked beds of wet interdune origin dominated by wavy laminated, very fine-grained sandstones that extend laterally more than 13 m. Indicates cyclic deposition within interdune depressions.
- 2 Fine-grained sandstone dominated by planar and wind-ripple laminated sandstone; dry interdunes.
- 3 Aeolian grainflow strata characterized by high-angle stratified, medium-grained sandstone. Grainflow deposits are each up to 60 mm thick.
- 4 Erosive base of aeolian dune dominated by grainflow strata.
- 5 Lateral transitional facies change from wet interdune to wind-ripple strata indicative of dry interdune.
- 6 Lateral thinning wet-interdune facies element.
- 7 Stacked aeolian dune-strata separated by interdune surface.
- 8 Stacked aeolian dune-strata separated by interdune surface.
- 9 Deformed aeolian-dune sets (slumped) as a result of wetting of advancing-dune.
- 10 Fluvial succession cuts into older aeolian strata.

Palaeocurrent data of the fluvial strata of the Delamere Member (DSM).

Resultant: 322
Ang. Dev.: 15
Total Data: 32
Largest Freq.: 60%



Palaeocurrent data of the aeolian strata of the Thurston Member (TSM).

Resultant: 241
Ang. Dev.: 20
Total Data: 39
Largest Freq.: 31%

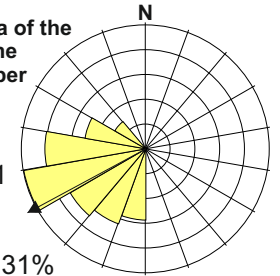


Figure 4.20: Continued.

Architectural elements and associated facies successions

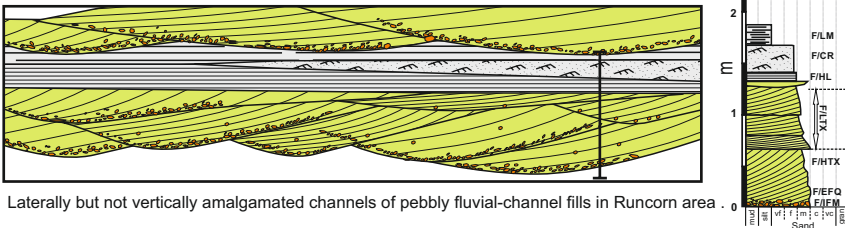
Fluvial channelized elements

F1 Multi-Storey, Multi-Lateral Amalgamated Channel-Fill Element



Vertically and laterally amalgamated of pebbly fluvial-channel fills typical of the Delamere Member in Beacon Hill.

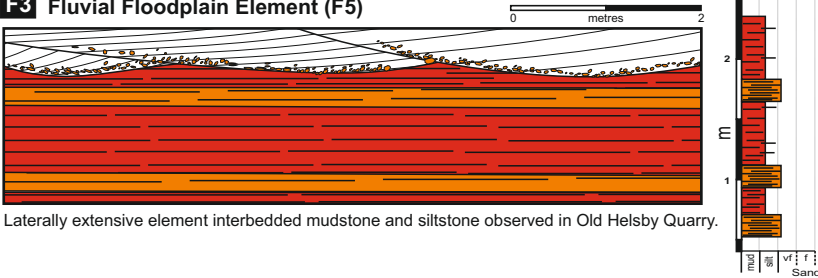
F2 Single-Storey, Multi-lateral Amalgamated Channel-Fill Element



Laterally but not vertically amalgamated channels of pebbly fluvial-channel fills in Runcorn area.

Fluvial non-confined element

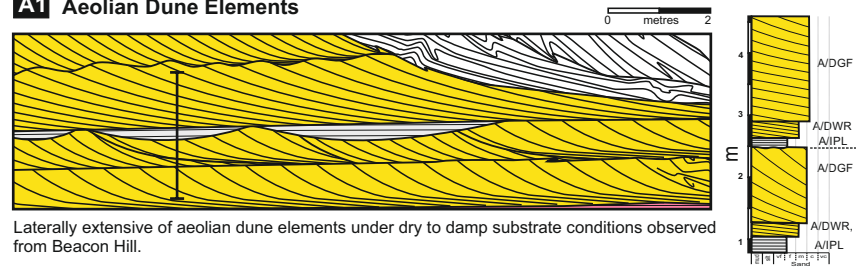
F3 Fluvial Floodplain Element (F5)



Laterally extensive element interbedded mudstone and siltstone observed in Old Helsby Quarry.

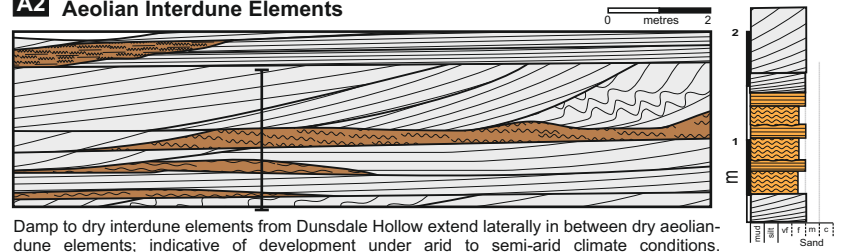
Aeolian dune and interdune elements

A1 Aeolian Dune Elements



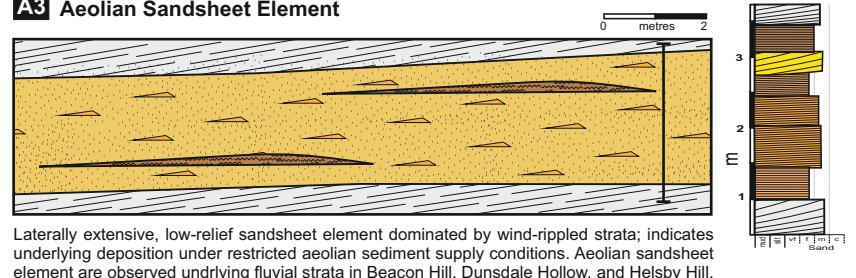
Laterally extensive of aeolian dune elements under dry to damp substrate conditions observed from Beacon Hill.

A2 Aeolian Interdune Elements



Damp to dry interdune elements from Dunsdale Hollow extend laterally in between dry aeolian dune elements; indicative of development under arid to semi-arid climate conditions. Damp-interdune elements are dominated by crinkly laminated sandstones (A/IWL), whereas dry-interdune elements are characterized by planar laminated sandstone (A/IPL). See table 4.1.

A3 Aeolian Sandsheet Element



Laterally extensive, low-relief sandsheet element dominated by wind-rippled strata; indicates underlying deposition under restricted aeolian sediment supply conditions. Aeolian sandsheet element are observed underlying fluvial strata in Beacon Hill, Dunsdale Hollow, and Helsby Hill.

Figure 4.21: Summary diagram of architectural elements observed from the outcrops of the Helsby Sandstone Formation.

4.9.1 Fluvial Architectural Elements

4.9.1.1 Multi-Storey, Multi-Lateral Amalgamated Channel-Fill Complex (F1)

Description: Examples of this element are well-exposed along Beacon Hill cliff where they can be traced for 400 m in lateral extent. However, only 42 m of this element along the Beacon Hill cliff is included in the detailed architectural panel (Fig. 4.19). This is because the cliff-face is mostly quarried and covered by recent bushes, making access problematic (panel J in Fig.19). Examples of this element comprise amalgamated channelized strata and are composed internally of stacked CF1 and CF4 successions. The lateral extent of each individual storey within this element ranges from 1 to 5 m (Fig. 4.22).

Sets of horizontally laminated, current-rippled, fine-grained sandstone (facies F/CR) are rarely present towards the top of this element. By contrast, intra- or extra-formational pebbly-lags are common and characterize the basal parts of this element (Fig. 4.22). Incision associated with low-angle-inclined, 3rd- to 4th-order bounding surfaces demonstrates the complex nature of the internal channel-fill between storeys within this element, whereas 5th-order bounding surfaces define the base of this element and demonstrate evidence of significant incision into underlying older elements (0.5 to 1.5 m relief associated with incision into underlying elements). The palaeocurrent readings within this element show consistent directions with a vector mean of 321° (n = 48; standard deviation = 17) (Figs. 4.4a, 22).

Interpretation: This element of laterally extensive amalgamated channel-fill complexes represents the deposits of aggrading braided-channel-belts. Repeated intense fluvial erosion processes were active during major flood events (cf. Flores & Pillmore, 1987). In outcrop, this element records the preservation of stacked sandy meso-forms as multiple storeys; these forms migrated predominantly downstream within the parent fluvial-channels (Miall, 1996). The erosional bounding surfaces that define each of the multiple storeys were generated in response to lateral channel migration and in-channel dune migration within a main channel-belt, or in response to channel-bar expansion (Picard & High, 1973; Bridge, 2003). As a result, repeated erosion events cut out the uppermost parts of underlying deposits and re-worked them in the channel thalweg to leave basal lags of mixed of pebbly- and rip-up-mud- clasts directly above the erosional-features (cf. Cain & Mountney 2009; Miall, 1996; Bridge, 2003; Gibling, 2006). The model presented here (Fig. 4.22) provides a detailed reconstruction of the

arrangement of observed packages of laterally and vertically amalgamated storeys in sections perpendicular and parallel to the predominant palaeoflow direction.

4.9.1.2 Multi-Lateral Amalgamated Interbedded Channel-Fill (F2)

Description: F2 elements comprise laterally amalgamated channel-fill storeys that each extend laterally of between 5 to 10 m and vertically up to 2 m thick (Fig. 4.23). A well-exposed example of this laterally extensive element can be traced for 50 m at Runcorn Expressway Road in orientations parallel to the regional palaeoflow (SE to NW). The representative vertical facies succession in this element (CF₂) records a single storey that includes a single coset of cross strata. Packages of storeys are developed laterally adjacent to each other to form this multilateral element. Internally, the gradational upward fining within the fill of this element is notable: cross-bedded sandstone, fine-grained, horizontally-laminated and current-rippled strata are topped by thin reddish-mudstone layers. This element is characterized by a sharp basal surface that exhibits up to 1 m of incision and relief. The basal erosional surfaces of F2 elements are directly overlain by, and intimately associated with intraformational mud-clasts. Basal erosional bounding surfaces exhibit lateral incisions into adjacent storeys. The palaeocurrent readings within this element show consistent directions with a vector mean of 327° (n = 23; standard deviation = 17) (Fig. 4.5c).

Interpretation: This element, with its fining-upward succession of cross-bedded sandstones and associated local intraformational clasts directly above the basal surfaces, records erosion and reworking processes of alluvial sediments in a braided fluvial system (Fig. 4.23). The development of the lateral-amalgamations of incising channels within this element indicates frequent avulsion processes of active channels at a single stratigraphic horizon (cf. Gibling, 2006; Banham & Mountney, 2014). The fining-upward trends in this element indicate the abandonment stage of single flood events as a suspension-load cap to the abandoned channels (Reid & Frostick, 1997; Bridge, 2006). This type of element occurs commonly in non-confined dryland areas where non-cohesive sandy overbanks of non-aggrading channels have high potential to be eroded during rapid channel migration or in the aftermath of repeated avulsions (cf. Cain & Mountney, 2009; Banham & Mountney, 2014).

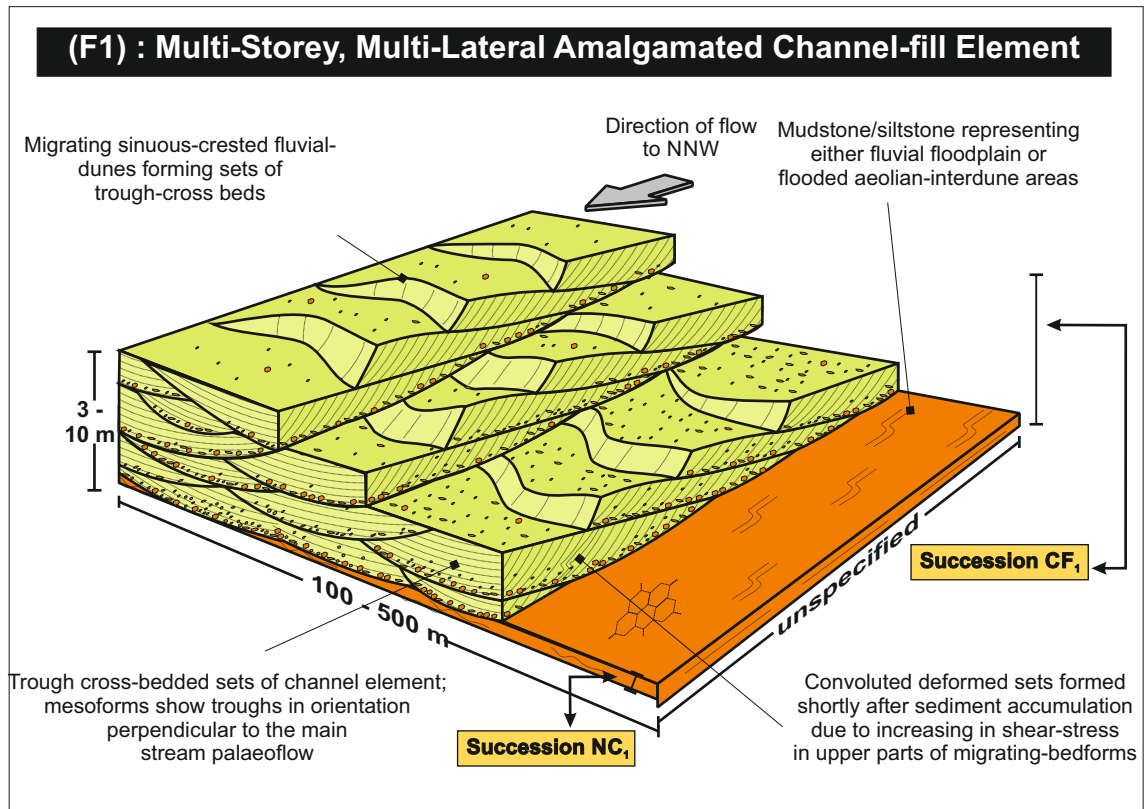


Figure 4.22: Multi-storey, multi-lateral amalgamated channel-fill element (F1).

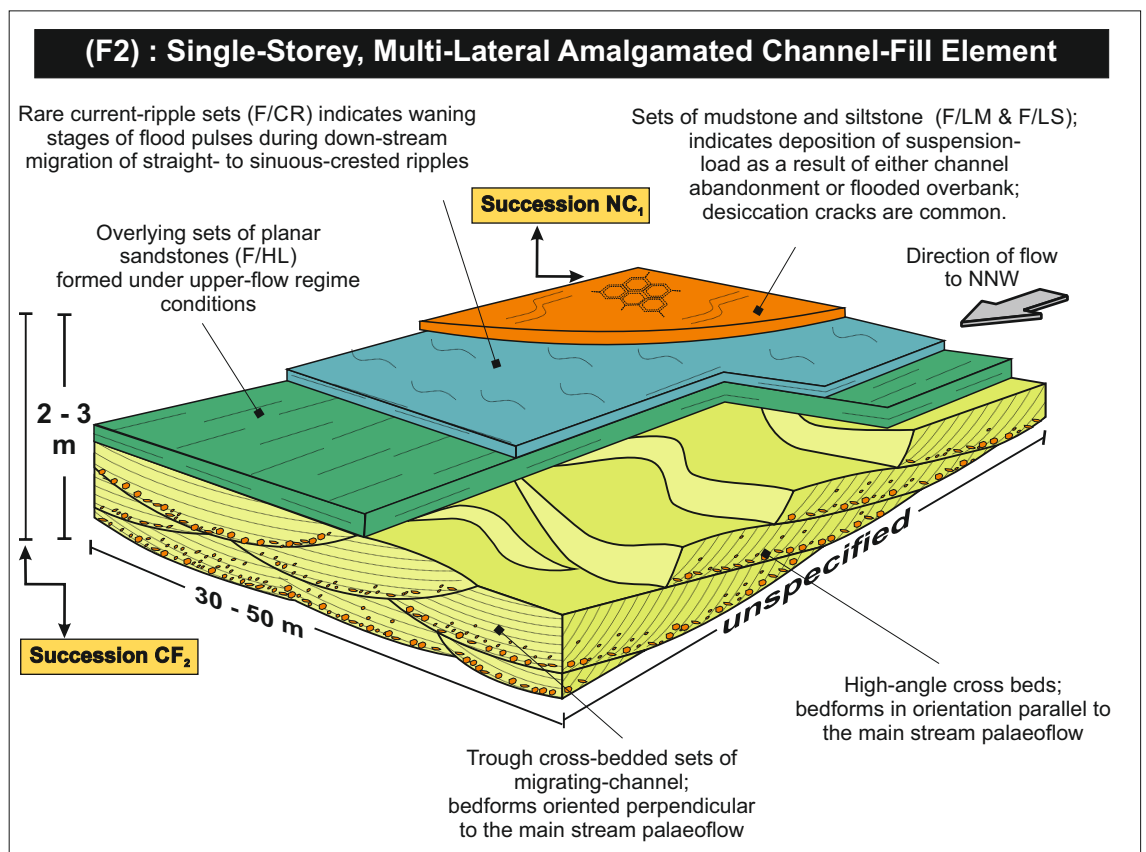


Figure 4.23: Single-storey, multi-lateral amalgamated channel-fill element (F2).

4.9.1.3 Fluvial floodplain Element (F3)

Description: Fluvial floodplain elements are composed of NC₁ successions. Examples are present in the Runcorn Expressway road-cut (Fig. 4.5c) and Beacon Hill (Fig.19) outcrops, though F1 and F2 channelized elements are dominant at these localities. Floodplain (F3) elements are especially well-developed and well-exposed exposed in the Helsby Hill Quarry section (Fig. 4.15); here they have a tabular geometry and are up to 2 m in thickness and extend for at least 15 m laterally (determination of lateral extent is limited due to lack of outcrop continuity in this locality and in the Beacon Hill area more generally). The lateral extent of floodplain elements in Runcorn Expressway section are up to 200 m (Fig. 4.24) and some examples will likely be significantly more extensive though lack of continuous outcrop precludes establishment of true length scales.

Interpretation: Floodplain deposits record sedimentation in non-channelized overbank settings in response to flood events that episodically passed water across the fluvial floodplain. Much of the associated sedimentation was via suspension settling from sluggish flow or from standing ponds (Bridge, 2003; Cain and Mountney, 2009; Banham and Mountney, 2014). Alternatively, mud-layers within this element may be the result of settling of suspended sediments from pools of standing water in abandoned channels in the aftermath of floods or channel avulsion events, or in isolated floodplain lakes (Bridge et al. 2008, p. 441; Miall, 1996, p. 177-178).

4.9.2 Aeolian Architectural Elements

4.9.2.1 Aeolian Dune Architectural Elements (A1)

Description: This element is common in the studied successions. It is composed internally of facies successions AD1 and AD2 (Fig. 4.16). In sections parallel to aeolian dune palaeomigration direction (ENE-WSW) (Fig. 4a,b and Fig. 4.19), the inclined strata of grainflow and wind-rippled dune sets downlap onto either of the following: 1) near-horizontal wind-rippled facies (A/IPL) and/or crinkly laminated facies (A/IWL) that form the upper parts of the fill of interdune elements; 2) sharp basal bounding surfaces. The scallop-shaped strata of sets of the dune elements are traceable toward a mean direction (~252°) for 5 to 10 m (Fig. 4.25). The basal surfaces that delimit examples of this element are inclined at <1° relative to palaeohorizontal indicators, such as lamination planes in red mudstone facies (F/LM) in the interdune elements. The maximum thickness of this element is 4 m, though most examples are <1 m. Examples of this element

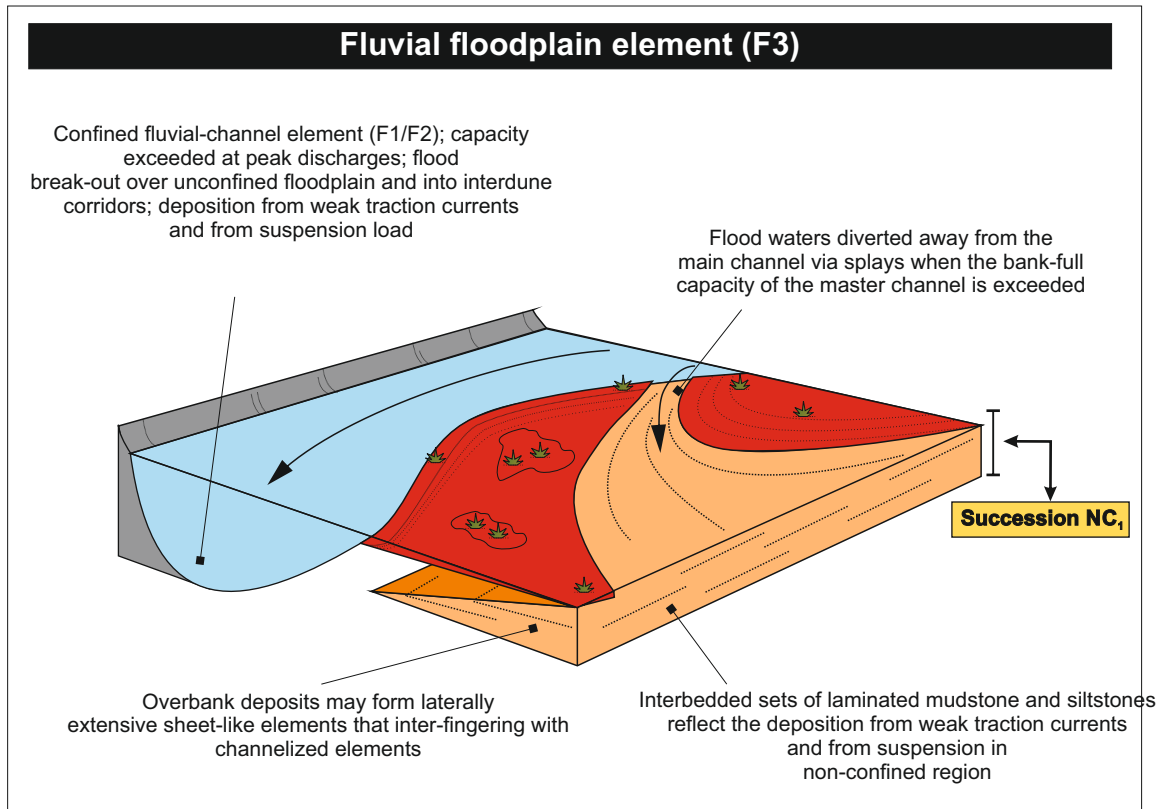


Figure 4.24: Fluvial floodplain element (F3).

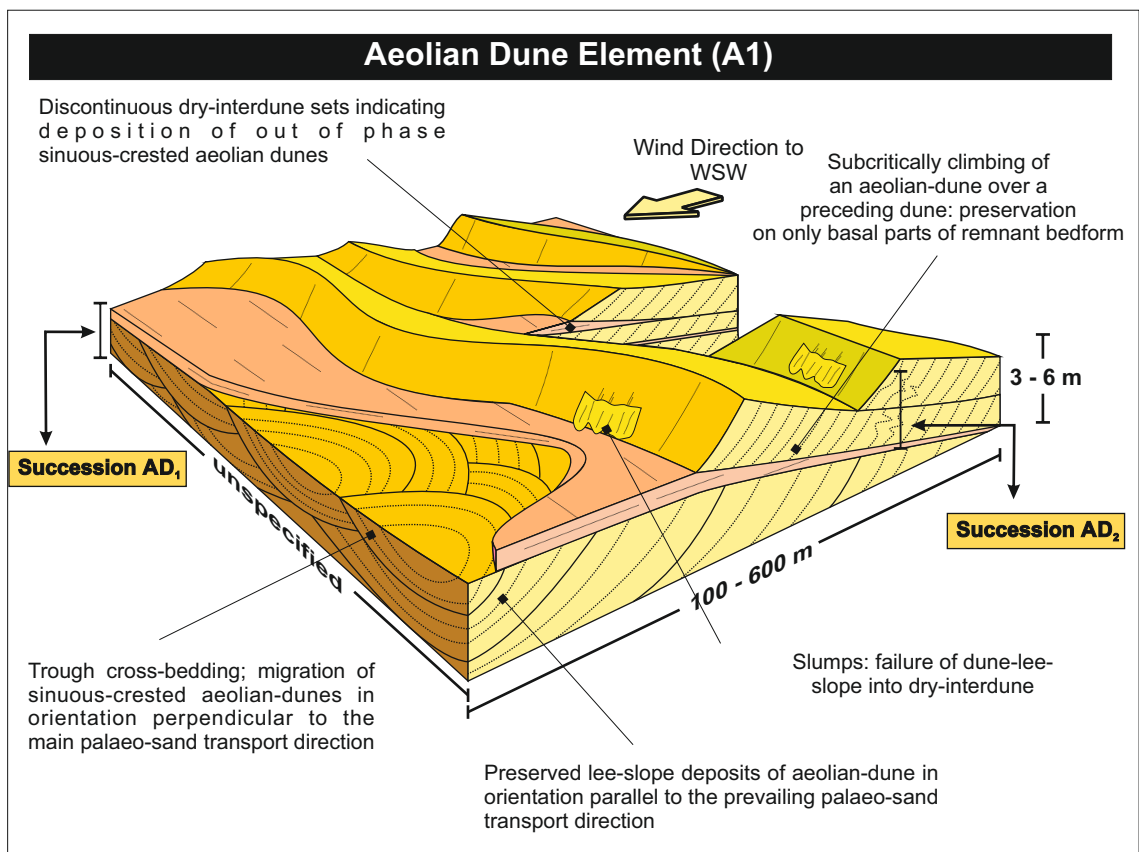


Figure 4.25: Aeolian dune element (A1).

can extend laterally for 4 to 8 m. In this element, the palaeocurrent readings have a vector mean of 252° from Beacon Hill ($n = 95$; standard deviation = 21) and a vector mean of 241° from Dunsdale Hollow ($n = 39$; standard deviation = 20) (Figs. 4.4b, 7). The compound scallops within sets of aeolian dune strata comprise tongue-shape structures that are intercalated downward with packages of underlying wind-ripple facies (A/DWR) and are systematically truncated toward south-westward by a scalloped-shape bounding surface. Internally, each compound set is composed of stacked sets of wind-ripple strata that are directly overlain by grainflow strata and cut out by a succeeding compound-set towards the southwest.

Interpretation: This element represents the preserved deposits of migrating aeolian dunes within water-table controlled dune fields such that only the lowermost parts of the relic dunes are preserved as sets (Fig. 4.25; Hunter, 1977; Kocurek, 1991). The observed scallop-shaped packages of grainflow and wind-rippled strata in the outcrop are representative of the upward transition from interdune, to dune plinth (toeset), to dune foreset sedimentation (Kocurek, 1981; Mountney & Thompson, 2002; Mountney, 2006a). These scalloped sets of cross-bedding accumulated via migration of aeolian-dunes coincidentally with dune-accumulation in damp interdunes under conditions of unsteady flow-velocity and/or flow-direction (Rubin, 1987b, Crabaugh & Kocurek, 1993). Slump degradation grainflow deposits are common within sets of this element and represent evidence of downslope-avalanching and collapse of loosely packed sediments under the influence of surface moisture (Mountney, 2006a). The bounding surfaces of cosets that directly underlie the wind-ripple strata are aeolian interdune migration surfaces (*sensu* Kocurek, 1996, cf. first-order bounding surfaces of Brookfield, 1977). The palaeocurrent measurements indicate palaeomigration towards the WSW, based on the azimuth of dune lee-slope foreset deposits (Figs. 4.4a, 19).

4.9.2.2 Aeolian Interdune Architectural Elements (A2)

Description: Observed interdune elements occur adjacent to closely associated aeolian-dune elements either as isolated lens-shaped or elongate and more laterally extensive bodies (Fig. 4.26). Interdune elements are characterized internally by successions ID₁ and ID₂. The geometric shapes of interdune elements vary as follows: 1) 0.1 to 0.3-m-thick isolated, lens-shaped interdune elements that are up to 5 m in lateral extent and composed of near-horizontally laminated wind-ripple laminations (A/IPL) with basal curved-like bounding surfaces (Fig. 4.19, panel A); 2) 0.5 to 0.7-m-

thick, elongate and laterally extensive interdune elements that are up to 16 m in lateral extent and are dominated by crinkly lamination (A/IWL) with flat basal bounding surfaces (e.g. panel H and I in Fig. 4.19). Both recognised types of the A2 element commonly exhibit interfingering at the uppermost boundary with overlying aeolian-dune elements (A1). Both varieties of the interdune element extend laterally for 3 to 15 m in sections parallel to dune palaeo-migration directions (ENE-WSW). In very rare cases within this study, a 0.1-m-thick-reddish-sheet-like-mud-layer that extends for 10 m with a sharp base has been seen in interdune elements in the Beacon Hill locality (panel I and J, Fig. 4.19).

Interpretation: The variations of lithofacies within this element indicate accumulation of interdune strata under a range of conditions. The presence of dominantly wind-rippled strata within some instances of this element indicates that deposits accumulated on a dry substrate within interdune hollows in-between actively migrating aeolian dunes (Kocurek, 1981; Mountney & Cain, 2009). In contrast, examples of this element composed internally of very-fine-grained-irregular-laminated of strata (A/IWL), which is common in this study, are indicative of damp interdune deposits that accumulated under the influence of a water-table capillary fringe that was in contact with interdune accumulation surface (Kocurek, 1981; Pulvertaft, 1985). The transition from damp-interdune upward to dry-interdune strata is indicative of fluctuations in the ground-water table, possibly influenced by ephemeral flash flooding in the interdune areas (Hummel & Kocurek, 1984; Langford, 1989) or by localized intense rainfall events. Alternatively, this transition might be a response to minor climatic changes or alteration in the balance between episodic subsidence and on-going accumulation (Kocurek & Havholm, 1993, Mountney & Thompson, 2002). The rare sheet-like mud layers represent ponds within open interdune corridors that were directly flooded by fluvial waters (cf. Stanistreet & Stollhofen, 2002). The lateral interfingering between aeolian dune A1 and interdune A2 elements indicates that both elements accumulated synchronously whereby aeolian dunes advanced over and interfingered with the fringes of dry and damp interdunes (Carbaugh & Kocurek, 1993; Pulvertaft, 1985; Mountney and Thompson, 2002).

4.9.2.3 Aeolian Sandsheet Architectural Elements (A3)

Description: Aeolian sandsheet elements are characterized internally by facies associations SH₁, SH₂ and ID₁. In the Beacon Hill and Dunsdale Hollow sections, examples of this element extend laterally for 20 m and

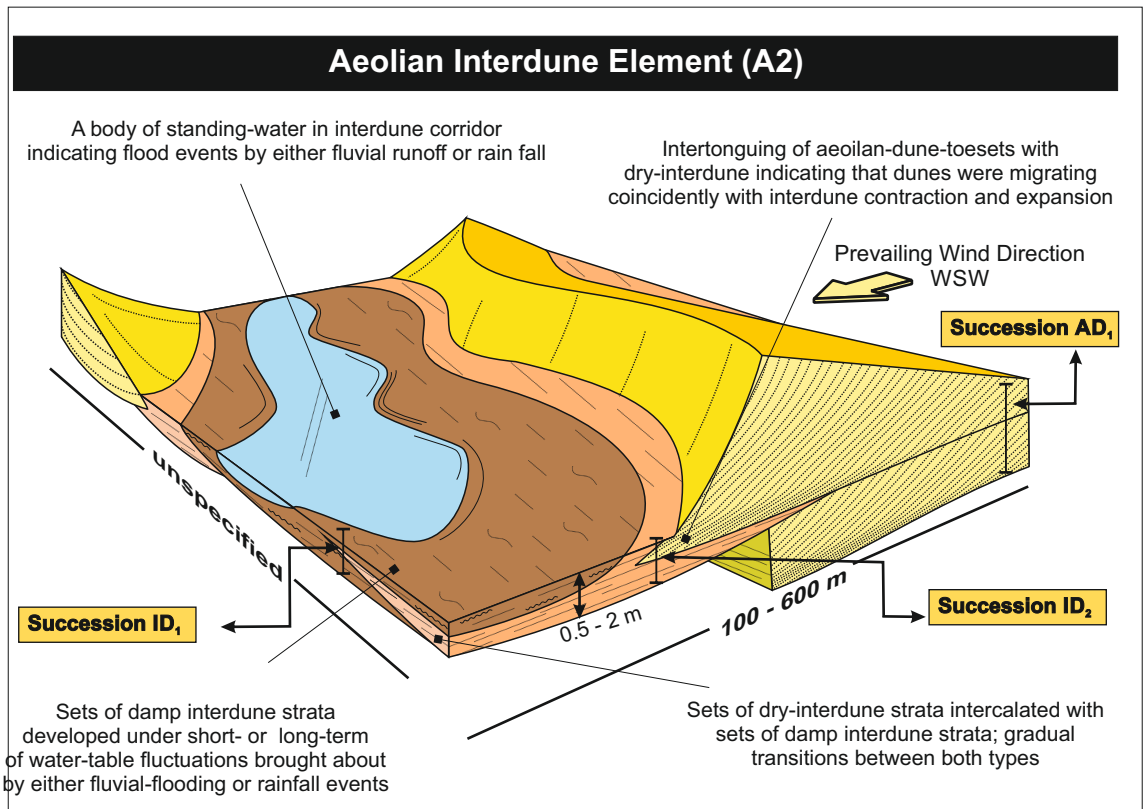


Figure 4.26: Aeolian interdune element (A2).

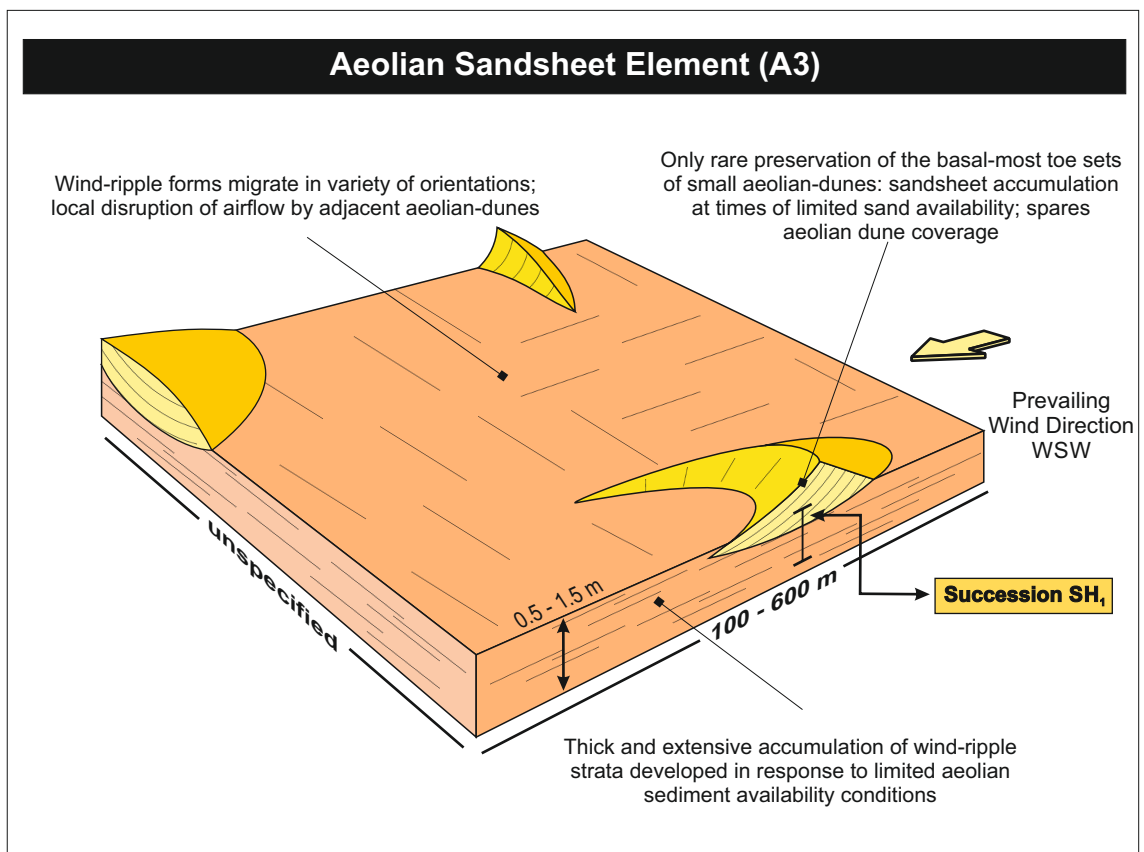


Figure 4.27: Aeolian sandsheet element (A3).

attain thicknesses up to 1.5 m (Fig. 4.27). Both vertical and lateral extents of aeolian sandsheet elements gradually become less where aeolian-dune elements are dominant in the stratigraphy and in such places, sandsheet elements cannot be differentiated from dry interdune elements (A2).

Interpretation: This element represents the preserved deposits of laterally extensive sandsheets that occupied the outer margins of aeolian dune fields, where slipfaced dunes were not generally present, probably due to insufficient supply or availability of sediment suitable for dune construction and accumulation (Kocurek and Nielson, 1986, Kocurek, 1999). The accumulation of low-relief sandsheet elements resulted from relic deposits of small, migrating aeolian dunes (Fryberger et. al, 1983). The interbedded aeolian dunes associated with examples of this element suggest that the dunes developed independently through slow rates of accumulation (Fryberger et al., 1983).

4.10 Discussion

The Permo-Triassic Sherwood Sandstone Group successions of the Cheshire Basin were influenced by many processes that are known to operate in mixed fluvial and aeolian settings developed under arid and semi-arid climatic regimes (Alkathery & Mountney, 2015). Collectively, these processes acted to determine the nature of the preserved sedimentary record. Observed sedimentary signatures in outcrops of the Helsby Sandstone Formation in the northern part of the Cheshire Basin were directly affected by the following factors: (1) the influence of water-table rise in mixed aeolian and fluvial systems, (2) fluvial flooding into the margins of aeolian dune fields, (3) aeolian morphological controls in determining how and where floods could extend in mixed fluvial-aeolian systems, (4) different mechanisms that acted to govern accumulation and preservation in mixed aeolian and fluvial systems.

4.10.1 The influence of water-table rise in aeolian deposition

Water-table rise is a requirement for aeolian accumulation and preservation in wet aeolian systems (Kocurek, 1999; Mountney, 2012). In wet aeolian systems, the water table lies at or close to the accumulation surface such that accumulation is influenced by the water table itself or its capillary fringe (Kocurek & Havholm, 1993). Within dune-field margins, the absolute level of the water table varies over short time scales in response to fluvial floods or seasonal rainfall events, or over longer time scales due to climatic change

(Kocurek & Fielder, 1982; Hummel & Kocurek, 1984; Langford, 1989; Stanistreet & Stollhofen, 2002). Additionally, the water table may vary in a relative sense due to subsidence whereby an aeolian accumulation may subside beneath a static water table (a so-called relative water table rise; Kocurek, 1999). In modern aeolian systems, where the paths of ephemeral flood-waters are controlled by aeolian topography, water-table levels are spatially variable from erg-margin to erg-centre regions and are controlled at least partly by the intensity of rainfall runoff and/or ephemeral flood events, and the open or closed nature of interdunes that serve as pathways for the floods (Al-Masrahy and Mounney, 2015). Collectively these factors determine the potential penetration distance of fluvial floods into aeolian dune-field margins.

A rise in water-table level has the greatest impact on the lower elevation parts of aeolian systems, such as elongate (i.e. spatially open) interdune corridors where short-lived ponds of standing water may develop in breached interdunes adjacent to the main path of a fluvial flood. This leads to so-called passive flooding. Additionally, ponds of water may accumulate in isolated (i.e. spatially closed) interdune depressions that are surrounded by aeolian dunes on all sides when local run-off of intense (typically seasonal) rainfall waters flow from local dune slopes to pond in isolated interdune depressions (Stanistreet & Stollhofen, 2002; Langford, 1989; Lancaster & Teller, 1988). Alternatively, an absolute rise of water-table level may also occur in flat-like areas of sand sheets which are also influenced by periodic ephemeral flood events either in hot arid and semiarid (Basilici et al., 2014; Kocurek & Nielson, 1986), or in cold arid and semiarid settings (cf. Mounney & Russell, 2004, 2009).

The rise of the water table to meet the accumulation surface or to within its capillary fringe of the accumulation surface for short periods of time results in the adhesion of saltating sand to the damp surface, thereby leading to the accumulation of adhesion deposits (Kocurek and Fielder, 1982; Olsen et al., 1989). If the water table falls, by contrast, the sediments on the accumulation surface will dry out and become prone to aeolian deflation. Localised and short-lived changes in the water-table level may change the availability of sediment for aeolian transport by either protecting it from deflation or liberating it for construction of adjacent aeolian dunes (Hummel & Kocurek, 1984).

In aeolian systems where on-going dune-migration and fluctuations of absolute water-table occur, water-table levels will vary both spatially and

temporally over short scales. This will result in episodic wetting and drying of interdune accumulation surfaces and this will be recorded as subtle yet observable vertical and lateral variations in preserved facies successions. Evidence for such changes are common in the HSF described herein. Interaction styles between aeolian dune and interdune elements in Beacon Hill and Dunsdale Hollow commonly show transitional lateral changes from wet or damp interdunes to dry interdunes, to dune-toeset deposits (Fig. 4.28). This demonstrates that episodic changes in water-table level were common and repeated and this was most likely driven by ephemeral flood events that were contemporaneous with on-going aeolian-dune migration and climb at a low but positive, and potentially variable angles (cf. Mountney & Thompson, 2002). The nature of the interdune units with variable geometries of damp/wet interdune elements in the study areas discussed herein was most obviously controlled by temporal changes in rates of aeolian-dune migration and rates of relative water-table rise and fall.

4.10.2 The characteristics of fluvial-floods in aeolian dune-fields

In arid and semi-arid environments, aeolian dune-field margins exhibit large variations in types of fluvial floods from channelized to non-channelized (Olsen, 1987; Al-Masrahy and Mountney, 2015). The nature and form of these floods are themselves influenced by flood magnitude, flood frequency, flood duration and factors such as rate of transmission losses as a result of flow across sandy substrates, and rates of loss due to evaporation. Additionally longer-term trends in flood behaviour may reflect either permanent or temporary climate changes. In settings where successive monsoonal floods last for a few months (e.g. multiple peak events of Graf, 1988, p.91), floods with high-magnitude and high-frequency characteristics can cause partial or complete erosion of previously accumulated aeolian dune successions, resulting in the reworking of older aeolian deposits and the transport of these eroded sediments further downstream as channel bedload sediments (Reid & Frostick, 1987). Consequently, high-magnitude episodic flood events promote rapid channel incision, followed by aggradation and avulsion in which new channel deposits are partially sourced from pre-existing channel-deposits producing multi-storey and multi-lateral channel elements with little opportunity for the in-situ preservation of mud-prone overbank elements (e.g. fluvial element F1 Fig. 4.22, Fig. 4.28a). In this type of situation, regional annual flood events tend to trigger a rapid rise in water-table level which temporarily increases the water table in both

Summary Stratigraphic Architecture of Beacon Hill, Frodsham

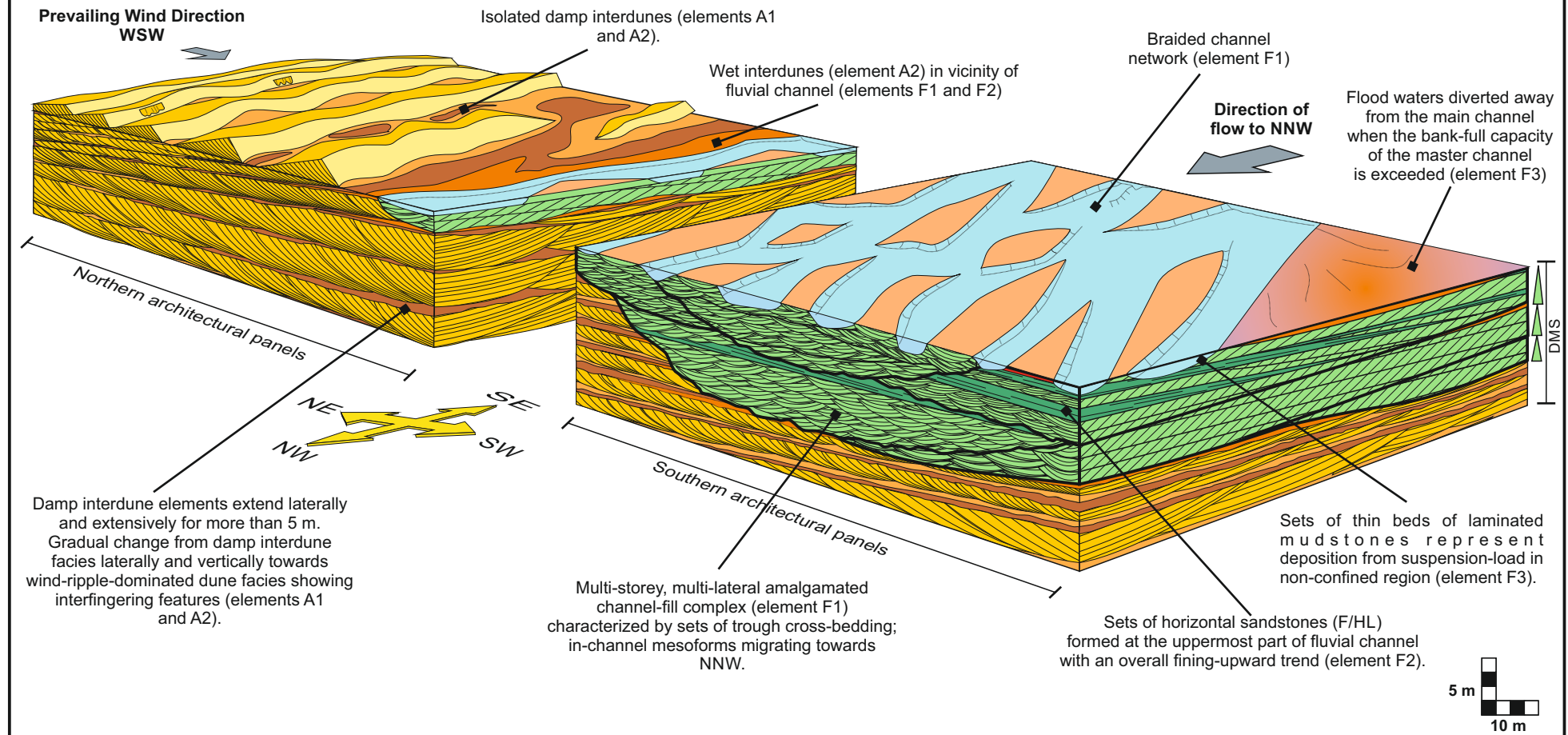


Figure 4.28a: Depositional model summarizing facies and architectural element arrangements of from Beacon Hill, Frodsham.

Summary Stratigraphic Architecture of Dunsdale Hollow, Frodsham U.K.

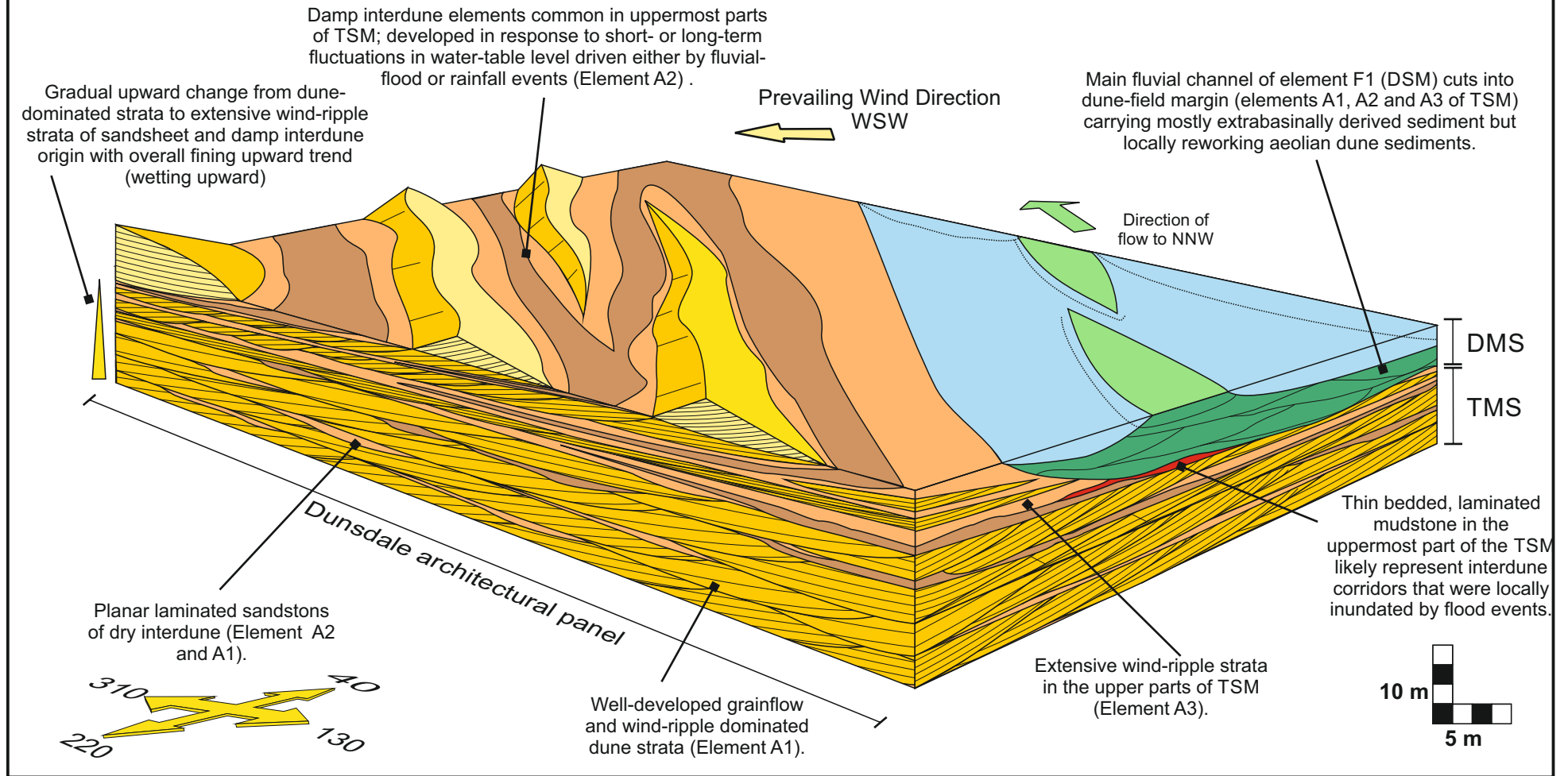


Figure 4.28b: Depositional model summarizing facies and architectural element Dunsdale Hollow, Frodsham.

isolated and open interdune depressions beyond the region subject to direct flooding. An outcome of this is the limitation of the availability of sand in interdune areas that is potentially suitable for aeolian transport and aeolian bedform construction. This is a so-called availability-limited system in the terminology of Kocurek and Lancaster (1999).

The aeolian damp interdunes facies within the HSF, for an instance, which transition gradually vertically to aeolian-dune toesets facies are dominant in parts of the stratigraphy closely associated to fluvial-incursion deposits. Such observed relationships may record water-table fluctuations in interdune areas adjacent to zones of fluvial flooding of dune-field margins (Mountney & Thompson, 2002).

Localised fluvial floods that pass directly into and along aeolian dune-field margins occur when tropical storms take place and may last from few hours to days (e.g. single-thread ephemeral channels of Reid & Frostick, 2011, p.310). These short-lived floods are normally high-magnitude and subject to rapid rates of downstream decrease in their discharge due to one or a combination of the following: 1) transmission losses as a result of floodwater infiltration or evaporation; 2) a sudden termination of floodwaters caused by damming by aeolian bedforms; or 3) distribution of floodwaters into multiple open interdune corridors (Tooth, 2000a; Stanistreet & Stollhofen, 2002; Al-Masrahy & Mountney, 2015).

4.10.3 Cyclic depositions of fluvial and aeolian sediments in deserts

At a regional scale, it has been suggested previously that cyclic accumulation of competing fluvial and aeolian systems in the Cheshire Basin could reflect a response to regional climate fluctuations from relatively more arid to more humid in response to orbitally forced Milankovitch-style cyclicity (Clemenssen et al., 1994; also cf. Howell & Mountney, 1997; Swezey et al., 1999; Chan, 1999; Bailey, 2001). Because evidence for large-scale interdigitation between accumulations of fluvial and aeolian origin is not demonstrated directly by the preserved successions examined in this study, the hypothesis that regional climatic oscillation acted as a main controlling effect is difficult to establish unequivocally but is worthy of further consideration (cf. Mountney et al., 1999). The observed cyclicity of cyclically repeated occurrences of deposits that represent ephemeral fluvial and damp and wet aeolian dune-interdune successions could represent the preserved expression of temporal and spatial variations of fluvial-aeolian system interactions as a palaeo-dune-field margin shifted back-and-forth across the

basin, thereby recording a window into the transit from fluvial-dominated to aeolian-dominated sub-environments for the episode of time represented by the stratigraphy (Langford & Chan, 1989; Mountney, 2006b; Mountney & Jagger, 2004; Trewin, 1993; Al-Masrahy & Mountney, 2015).

The cyclic arrangement of large-scale packages of aeolian strata and their intercalation with large-scale packages of fluvial strata can be interpreted in several ways. Most simply, they reflect the preserved expression of lateral shifts in the position of the erg margin over time. Such shifts would be governed by changes in both fluvial and aeolian sediment supply, sediment availability, transport rate, basin subsidence, groundwater level, and other factors (Kocurek & Lancaster, 1999; Nash, 2011; Mountney, 2012). The large-scale stratigraphic pattern can be interpreted to indicate repeated cycles of aeolian construction, accumulation and deflation processes and associated cycles of flash-flood progradation (encroachment) and retrogradation (retreat) of fluvial systems into and away from an aeolian dune-field margin (Mountney & Jagger, 2004; Mountney, 2006a, b; Clemmensen et al., 1989).

during relatively more humid phases, the generation of sediment supply through time within aeolian systems would be partly controlled by flood frequency and the amount of sand-sized sediment transported into the region (that would be suitable for later aeolian reworking) via ephemeral fluvial systems (Langford & Chan, 1989; Mountney et al., 1999; Bullard & Livingstone, 2002). The supply of sand-sized sediment would be availability limited at the time of accumulation (time lagged) but would become available for use in aeolian dune construction during the ensuing switch to a later drier climatic episode (i.e. a dry phase), during which the wind would be able to liberate and reuse sand made available on drying interdune surfaces that were responding to water-table fall (Kocurek & Havholm, 1994; Kocurek and Lancaster, 1999).

During episodes of minimum aridity, an increase in seasonal run-off activity into dune-field margin areas would be expected and this might be accompanied by a localised increase in the water table to a level at or close to accumulation surface. During relatively more humid climatic phases, deflation processes would have been relatively restricted because of the inability of the wind to entrain damp or wet sand that was adhered to the accumulation surface (e.g. due to increased cohesion association with wet grain bonding) (Crabaugh & Kocurek, 1993; Kocurek & Lancaster, 1999; Mountney, 2012). Thus, the amount and rate at which sediment would have

been made available for aeolian bedform construction would have been limited, possibly encouraging the construction and accumulation of aeolian sandsheets but not major aeolian dunes (Kocurek, 1999).

During humid climatic phases, multiple flash-flood events apparently dominated the outer dune-field margin areas and generated deep fluvial-incisions into underlying, older aeolian successions. On-going fluvial activity coincident with slow but on-going subsidence resulted in vertical aggradation and stacking of ephemeral channels to form F1 multi-storey channel elements (Miall, 1996; Stear, 1983; Langford & Chan, 1989; Gibling, 2006).

The high sand proportion within the bedload fill of these F1 elements likely reflects the reworking of older aeolian dune accumulations following channel incision. Some of these sands would likely have been carried further downstream by the fluvial system to later be distributed in interdune corridors, thereby generating a time-lagged supply for later aeolian construction. This is a major reason why mud-prone deposits are apparently scarce in the HSF (cf. Kocurek, 1981; Lancaster & Teller, 1988; Mountney & Thompson, 2002; Mountney & Jagger, 2004). Restricted sediment supply, resulted in the expansion of the interdune areas at the expense of aeolian-dune construction during humid climatic phases (Kocurek & Nielsen, 1986; Kocurek, 1999).

During more arid climatic phases, the water-table and its capillary fringe would fall below the accumulation surface such that dry, loose sand would become available for aeolian transportation given sufficient wind velocity (Kocurek & Lancaster, 1999). In this situation, damp interdune corridors would become less common and would tend to transform over space and time into isolated dry interdune hollows as a result of the construction (i.e. growth) of surrounding aeolian-dunes (Kocurek & Havholm, 1993).

4.10.4 Intertonguing of damp interdune and adjacent aeolian dune strata

In the studied localities, the spatial distribution of damp interdune elements is varied and in many cases this reflects the influence of water table, especially at the northern to the southern parts of the Beacon Hill cliff (Figs. 4.19 & 4.20). Sandy damp-interdune elements (A2, e.g. with adhesion structures of facies A/IWL, Fig. 4.13) in the southern part of Beacon Hill cliff (e.g. panels J, H, and I, Fig. 4.19) are abundant and are characterized by damp interdune deposits that occur interdigitated at a small scale (i.e. demonstrate intertonguing) with overlying aeolian dune-toesets of A1

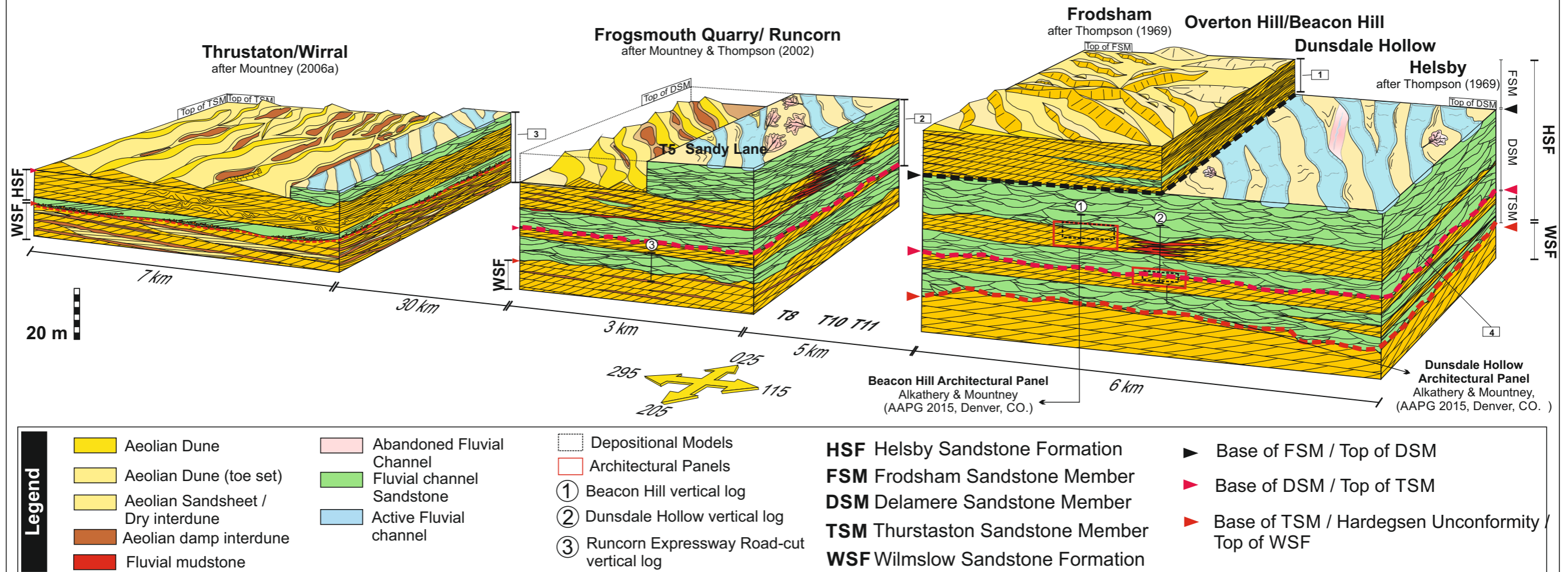
elements. Intertonguing relationships are laterally continuous for > 7 m towards in sections oriented parallel to palaeo-bedform transport (to the SSW). However, instances of similar relationships are more isolated (< 3 m) in the northern parts of the same cliff (e.g. panels A, B, and F, Fig. 4.19).

The intertonguing damp interdune and dune elements at the Beacon Hill locality reflect processes of coeval interdune and dune accumulation (see Gradzinski & Jerzykiewicz, 1974; Pulvertaft, 1985; Loope & Simon, 1992; Mounney & Thompson, 2002; Mounney, 2006b). Similar field observations were described by Mounney and Thompson (2002) who noted the close relationship between aeolian damp and wet interdunes and dunes of Delamere Sandstone Member in Frogsmouth Quarry at Runcorn (Fig. 4.29), and proposed summary models showing the relationships that indicates vertical accumulation of low-angle migrating damp-interdunes coincidentally with on-going active migration of aeolian-dunes during overall accumulation. An outcome of that study was the proposal of a method to calculate the small angle-of-climb based on small-scale observations of stratal relationships in outcrops of limited extent. In wet aeolian systems, the angle-of-climb of aeolian bedforms and laterally adjoining interdunes is known to fluctuate in response to changes in the ratio between the rate of water-table rise and the rate of aeolian-migration in wet systems (Mounney & Thompson, 2002).

The lateral downwind continuation of sandy damp-interdune elements, the geometry of which was dictated by the morphologic arrangement and spacing of adjacent dunes, is partly determined by long-term fluctuations in water-table level over time in response to intermittent episodes of heightened ephemeral flash-flood events or changes in regional basin subsidence rates, as well as changes in aeolian dune migration rate (Kocurek, 1981b, 1999; Pulvertaft, 1985; Mounney & Thompson, 2002). Isolated interdune elements (A2) of limited extent that occur surrounded by aeolian dune elements (A1) (e.g. panel A, Fig. 4.19) are indicative of the presence of spatially isolated (i.e. closed) interdune depressions that were surrounded by aeolian dunes and which therefore formed non-connected dry-interdune hollows between dunes (e.g. sinuous crested-bedforms, out-of-phase planform dune morphologies of Mounney, 2006a).

Such features are common in aeolian dune fields composed of relatively small, rapidly downwind migrating transverse or oblique aeolian dunes that construct (i.e. grow) to close off formally open interdune corridors by reducing them to isolated interdune depressions (Rubin, 1987). Such

Depositional Model of Helsby Sandstone Formation, Cheshire Basin, UK



| | | | | | |
|--------|-----------------------------------|--|---|---|---|
| Legend | Aeolian Dune | Abandoned Fluvial Channel | Depositional Models | HSF Helsby Sandstone Formation | Base of FSM / Top of DSM |
| | Aeolian Dune (toe set) | Fluvial channel Sandstone | Architectural Panels | FSM Frodsham Sandstone Member | Base of DSM / Top of TSM |
| | Aeolian Sandsheet / Dry interdune | Active Fluvial channel | ① Beacon Hill vertical log | DSM Delamere Sandstone Member | Base of TSM / Hardegsen Unconformity / Top of WSF |
| | Aeolian damp interdune | Fluvial mudstone | ② Dunsdale Hollow vertical log | TSM Thurstaston Sandstone Member | |
| | | ③ Runcorn Expressway Road-cut vertical log | WSF Wilmslow Sandstone Formation | | |

| | | | |
|---------------------------------------|---|---|---|
| Features of Sedimentological Interest | <p>① The Frodsham Sandstone Member (FSM) is an aeolian succession that represents the preserved deposits of large dome-shaped aeolian dunes (Thompson, 1969), deposits of which are well-exposed in a railway cutting near Frodsham town, UK. Aeolian sets are up to 20 m thick and preserve convex-up foresets that dip in a variety of orientations. The FSM records the development of an extensive aeolian dune field in the central Cheshire Basin.</p> | <p>③ The Thurstaston Sandstone Member (TSM) is exposed as a dominantly aeolian unit at its type section, a 300-m-long, 5-10 m high road-cutting at Thurstaston. This section exposes the uppermost part of the WSF and the aeolian TSM of the HSF. The two formations are separated by The Hardegsen Unconformity. The base of the TSM is characterized locally by a prominent fluvial channel element (Mounthey, 2006; Thompson, 1970a).</p> | <p>Architectural panels from the Helsby Sandstone Formation (HSF) document the detailed stratigraphy in orientations both parallel and perpendicular to aeolian transport direction, enabling a quantitative 3D reconstruction of genetically related aeolian dune, interdune and fluvial channel and non-channelized sheet-like elements. Sets of aeolian dune strata are composed of grainflow and translent wind-ripple strata and are divided by a hierarchy of bounding surfaces originating from oblique migration of superimposed dunes over slipfaceless, sinuous-crested parent bedforms, together with lee-slope reactivation under non-equilibrium flow conditions. Silty-mudstone and sandstone interdune units are characterized by wind-ripple-, wavy- and subaqueous wave-ripple-laminae, desiccation cracks, mud flakes, raindrop imprints, load casts, flutes, intraformational rip-up clasts, and vertebrate and invertebrate footprint impressions and trackways. These units result from accumulation on a substrate that varied from dry-, through damp- to wet-surface conditions. Interdune ponds were flooded either by fluvial incursions or rises in groundwater table and were periodically subject to gradual desiccation and re-flooding. Red silty-mudstone beds of subaqueous origin pass laterally into horizontally laminated wind-ripple beds indicating a progressive transition from wet-, through damp- to dry-surface conditions single interdune elements.</p> |
| | <p>② Mounthey and Thompson (2002) described a series of two-dimensional stratigraphic panels of the DSM of the HSF in old sandstone quarries (Runcorn Hill). This work recognized significant lithological changes between fluvial and aeolian sedimentation. The aeolian unit is characterized by small aeolian dune sets with a variety of intervening wet, damp and dry interdune units; fluvial units are composed of channelized elements, many with mudstone intraclasts and water escape structures.</p> | <p>④ The Helsby Hill area is the type section for the HSF and the the 3 members are all well exposed. Thompson (1970) and Mounthey & Thompson (2000) identified complex interactions between aeolian and fluvial units using extensive outcrop data from measured sections at Helsby Hill Main Cliff and Helsby Old Quarries.</p> | |

Figure 4.29: Depositional model of Helsby Sandstone Formation (HSF) across Cheshire Basin.

behaviour it typical of aeolian systems that are subject to a relatively rapid net fall of relative water-table level that stimulates the liberation of a time-lagged aeolian sand supply and promotes rapid dune construction. However, such small, isolated interdune depressions may be subject to localised short-lived flooding when the infiltration capacity of the sandy substrate is temporarily exceeded by short-lived but intense rainfall events (Mountney, 2006b).

The relationship between the upper surfaces of damp interdune elements and the overlying aeolian-dunes elements herein is varied, recording encroachment of dunes over interdunes, transitional change of intertonguing interdunes towards overlying dunes, and erosional scours of dunes into underlying interdunes as shown in (Fig. 4.11, 4.12, and 4.13) (cf. Pulvertaft, 1985; Mountney & Thompson, 2002).

The lateral extensive, very-low-angle-inclined surfaces (cf. Brookfield, 1977; Kocurek 1996) which are best seen in sections parallel to aeolian transport direction (WSW) along the Beacon Hill and Dunsdale Hollow localities, and which separate overlying interdune elements from underlying aeolian dune elements, are either interdune migration surfaces that separate trains of adjacent dunes and interdunes that are climbing at very low angles, or else they are bypass supersurfaces that record a zero angle-of-climb (Langford and Chan, 1988, 1989; Jagger & Mountney, 2004).

4.10.5 Preservation controls in aeolian accumulations

The reconstructed aeolian dune and interdune architecture for the studied part of the HSF (Figs. 4.25 & 4.26) likely records the episodic preservation of wet aeolian systems in response to slow but on-going basin subsidence that occurred synchronously with active dune and interdune migration. This mechanism enabled the slowly accumulating succession to pass progressively beneath the relative water table. The relative rate of change of water-table level experienced by the system represents both the change induced by slow but on-going subsidence, plus any short-lived changes associated with local climate changes or autogenically-driven fluvial incursions (e.g. monsoonal rainfall and ephemeral flash-flood events) (Crabaugh & Kocurek, 1993, 1998; Kocurek & Havholm, 1993; Kocurek, 1999). A key factor controlling the long-term accumulation and preservation of wet aeolian systems is temporal and spatial changes in the angle-of-climb that itself is influenced by changing rates between water-table rise and aeolian dune migration (Mountney & Thompson, 2002; Mountney, 2012).

Models for both climbing (e.g. Entrada Sandstone, Utah) and non-climbing (e.g. White Sands, New Mexico) wet aeolian systems have been documented based on studies of ancient and modern systems (Simpson and Loope, 1985; Crabaugh & Kocurek, 1993; Loope & Simpson, 1992, 1993) (Fig. 4.3). In climbing wet aeolian systems, aeolian dune accumulation is controlled by a steady increase in relative water table and a rate of aeolian sediment supply that is sufficient to maintain bedform migration; collectively these factors determine the angle-of-climb in such systems (Crabaugh & Kocurek, 1993). By contrast, the non-climbing systems are characterized by flat and extensive wet/damp interdune units as a result of water table rise associated with very slow vertical growth of thin dune-toesets due to low sediment supply and restricted accommodation availability (e.g. metasaturated zone of Simpson & Loope, 1992, 1993). In such circumstances, the accumulation of thin sets of migrating-dunes occurs only when accommodation space becomes available as subsidence or groundwater rise take place preserving the lower parts of migrating-dunes (Simpson & Loope, 1992, 1993). Most of the elapsed time is represented by bypass supersurfaces.

Determining whether the aeolian system represented by the studied succession accumulated via either climbing or non-climbing behaviour is problematic because of the limited lateral extent of the outcrops and the general difficulty in identifying surfaces that can be convincingly and reliably interpreted as indicators of palaeohorizontal such that potential angles of climb of just a few tenths of one degree can be measured. For these reasons, the calculation of angle-of-climb of system in this work that was proposed by Mountney and Thompson (2002) has not been attempted. However, the observed interaction between aeolian dune bedforms and damp or wet interdunes (Fig. 4.19 panel H, I and J) does indicate a water-controlled aeolian system or so-called wet aeolian system (cf. Kocurek & Havholm, 1993). Additionally, the lateral extent of sharp interdune surfaces between dune and interdune elements (Fig. 4.19 & 4.26) may indicate that the level of water table fluctuated, possibly as a result of longer-term (20 – 400 ka; Clemmensen et al., 1994) changes in the amount of precipitation or in climate (cf. Crabaugh and Kocurek, 1993). Additionally, the lateral extent of sharp interdune surfaces between dune and interdune elements (Fig. 4.19 & 4.26) may indicate that the level of water table fluctuated, possibly as a result of longer-term (20 – 400 ka; Clemmensen et al., 1994) changes in the amount of precipitation or in climate which together influence of the water table (cf. Crabaugh and Kocurek, 1993). The water table change within

interdunes can result a change in the vertical successions such that fall of water table will generate dry-interdunes and rising water table will form damp/wet interdunes (Kocurek & Havholm, 1993; Mountney & Thompson, 2002).

The water table fluctuations within interdunes can result a gradual change in the vertical successions from damp- to dry-interdunes associated with expansion and contraction of the interdune deposits at the expense of adjacent dunes. Furthermore, the observed amalgamated damp interdune deposits in the studied areas (Fig. 4.20) may indicate a non-climbing system whereby amalgamated damp interdunes developed as a result of relative or absolute water table (Fig. 4.3). The hypotheses of competing models of climbing versus non-climbing for the accumulation and preservation of damp or wet interdune strata which have been argued previously (Loope & Simpson, 1992, 1993; Kocurek & Crabaugh, 1993), are better viewed as end-members of a range of depositional styles (Mountney & Thompson, 2002). Thus, the system studied here could have accumulated via multiple minor episodes of climb, followed by minor deflation and/or bypass, followed by renewed climb. Via this mechanism, the system may have incrementally accrued at times when the circumstances fortuitously enabled accumulation. Technically, each episode of accumulation would represent an individual aeolian sequence, though discerning such sequences and their super-bounding surfaces from dune and interdune sets that arose via a “traditional” climbing mechanism is not possible for the outcrop data available.

4.11 Conclusions

The following conclusions have arisen from this study.

1. The Helsby Sandstone (Anisian) formations is well-suited to sedimentary lithofacies analysis and to the development of an architectural-element-based approach with which to develop a set of detailed depositional models for water-controlled erg-margin systems.
2. The cyclicity of fluvial and aeolian architectural elements within the Helsby Sandstone Formation indicates that accumulation occurred in response to episodic ephemeral river incursion into the margins of an aeolian dune field. This arose in response to one or a combination of the following mechanisms: 1) in response to large-scale autogenic factors such episodic regional-scale river avulsion; 2) in response to climate

change that caused repeated expansion and contraction of the margin of an aeolian dune field, possibly in response to Milankovitch-style cyclicity.

3. The widespread occurrence of fluvial multi-storey, multi-lateral amalgamated channel-fill elements (F1) suggests that the fluvial system undertook repeated avulsion and new channel development in response to a series of large-magnitude flood events that led to rapid channel incision, aggradation and avulsion.
4. The presence of lateral downwind continuation of intertonguing damp-interdunes with adjacent thin dune toesets in the Beacon Hill and Dunsdale Hollow aeolian successions indicates that the interdunes migrated coevally with on-going aeolian-dune migration.
5. The nature of the interdune surfaces that bound damp/wet interdune elements was controlled by the temporal changes in the ratio between the rate of aeolian-dune migration and rate of relative water-table rise.
6. In this study, the interpretation of the analysed aeolian strata demonstrates that the aeolian-dunes were water-table influenced and controlled.
7. The preserved form of aeolian dune and interdune architectural elements records the episodic preservation of a wet aeolian system, possibly resulting from an absolute rise of continental water-table in response to a long-lived temporal increase in climatic humidity due to environmental change during the episode of accumulation, though more likely due to the progressive subsidence of the accumulation beneath the absolute water table.

Chapter 5 Discussion

5.1 Introduction

This chapter discusses the implications of fluvial and aeolian interaction as a set of sedimentary processes that might potentially influence preserved hydrocarbon reservoir quality. This chapter additionally considers how the generic facies models developed as an outcome of this work might be implemented to better predict reservoir stratigraphic heterogeneity and flow behaviour. This chapter makes use of interpreted data from chapters 3 and 4; specifically, it summarises (1) the importance of utilising identified interaction styles from the Skeleton Coast erg as a modern analogue example of aeolian systems impacted by a variety of types of fluvial incursion, and (2) interpreted outcrop data from the Triassic Helsby Sandstone Formation, Cheshire Basin, England, as an outcropping analogue example of an ancient mixed aeolian-fluvial succession. The modern examples and detailed outcrop architectural panels of this ancient preserved succession can be used as a hydrocarbon reservoir analogue to better understand likely reservoir stratigraphic heterogeneity.

From the studied examples of the Skeleton Coast erg, the generic models presented as an outcome of the research findings in Chapter 3 allow for discussion herein of the role played by low-permeability, mud-prone fluvial flood elements within otherwise sand-prone aeolian successions. Such low-permeability elements can form effective baffles and barriers to flow in analogous reservoir successions (Stanistreet & Stollhofen, 2002). Facies models describing sedimentary architectural relationships in ancient outcrop analogue successions (Chapter 4) are useful because they describe the geometry and distribution of relatively low-permeability fluvial and damp/wet interdune elements within otherwise relatively high-permeability aeolian dune-dominated successions. Such models serve to provide summaries of three-dimensional stratigraphic relationships present at sub-seismic scale that cannot be determined directly from subsurface reservoir datasets provided by core and well-log data that are essentially one-dimensional in form.

5.2 The Skeleton Coast Erg, Namibia as a Modern Analogue Example

The types of fluvial-aeolian interactions identified in the Skeleton Coast erg (Chapter 3) offer a modern analogue for fluvial-dominated to aeolian-dominated erg-margin systems recognised in ancient preserved successions. In general, fluvially-flooded interdunes are characterised by relatively low porosity and permeability characteristics, meaning that the preserved architectural elements of such bodies potentially may act as baffles and barriers to fluid flow and, therefore, exert a significant impact on overall reservoir character (Stanistreet & Stollhofen, 2002). The geometry and degree of interconnectivity of preserved examples of such elements may differ significantly from the morphological form of the original sub-environments that existed at the time of sedimentation. However, models that predict the development and lateral growth of such elements, in response to gradual migration of the geomorphic forms as the sedimentation process is on-going, enable the likely facies composition, geometry and interconnectedness of the preserved form of the elements to be predicted (Mountney & Thompson, 2002; Mountney, 2012).

Where flood waters access interdunes during active flood events, variations in the morphology of the interdune flats dictates the distance to which flood waters can invade into aeolian dune fields; neighbouring interdunes may be locally interconnected to each other, thereby serving as open corridors that allow passage of flood waters; alternatively, neighbouring interdunes may be isolated from each other by dunes and only damp or wet interdune ponds may develop in response to local rainfall or water-table rise. In general terms, isolated and closed interdunes are more common in the aeolian-dominated erg-margin settings (e.g. Hoanib and Hunkab), whereas open and interconnected flooded interdune corridors are more common in fluvial-dominated erg-margin settings (e.g. Uniab and Hunkab). Fluvial incursion into interdune areas favours the accumulation of mud layers of variable thickness where flood waters pond (Langford & Chan, 1989; Stanistreet & Stollhofen, 2002). The mud layers of wet interdunes from modern examples can be identified and used as an analogue in the preserved record. However, the flooded interdune elements in ancient record commonly occur as laterally extensive and connected bodies that reflect their growth and spread over repeated flood events over time. The geometry of preserved flooded interdune elements is determined by the size of the original form, its longevity of existence, its rate of lateral migration and the repeat frequency

of each flood event that occupies these interdunes (e.g. the Helsby Sandstone Formation, Cheshire UK, Mountney & Thompson, 2002; the Permian Cedar Mesa Sandstone, SE Utah, USA, Mountney & Jagger, 2004).

In the ancient record, mudstone layers of wet interdune elements can occur as laterally extensive layers as a result of repeated fluvial incursion into large and long-lived interdune corridors (e.g. Hunkab, Skeleton Coast erg). However, mudstone layers cannot usually develop in isolated interdune hollows because mud-laden waters of floods cannot access these closed interdunes. Instead, groundwater level in these spatially isolated depressions may rise, leading to wet ponds that trap sand reworked locally from surrounding dune flanks or modest amounts of loess (Langford & Chan, 1989; Mountney & Jagger, 2004).

Where interdune corridors are present between non-climbing migrating dunes (e.g. Hunkab), deposition of fluvial sediments will occur in the aftermath of flash flood events in the outer erg-margin area. This may be coincident with a rise in water-table level such that wet interdune sedimentation may occur coevally within the aeolian-dominated inner erg margin where interdune depressions are closed in form (i.e. surrounded on all sides by dunes) (Langford & Chan, 1988, 1989). Repeated flooding of a dune field for which the bedforms are non-climbing but still actively migrating will favour the development of bypass flood surfaces (Langford and Chan, 1988, 1989; Mountney, 2012). Deposition associated with sediment reworking during short-lived but intense rainfall events in a generally dry system may occur where climbing migrating dunes occur surrounding isolated interdunes depressions that are typically dry but which may be episodically flooded by water-table rise and rare intense rainfall events (Herries, 1993). Where such interdunes are open to an outer erg margin, they may be subject to fluvial incursion and associated sedimentation.

Where fluvial activity is common and occurs repeatedly for protracted episodes within aeolian dune-field margins (e.g. the Uniab and Hunkab rivers of the Skeleton Coast erg), ridges of dunes in the erg margin region may remain active during periods of on-going fluvial flooding such that fluvial processes commonly progressively erode and rework the lowermost flanks of advancing dunes and also inundate interdune areas where associated mudstone layers have accumulated over time (Fig 5.1). In such cases, fluvial channels may occur contemporaneously with deflation surface generation (Fig. 5.1). Deflationary supersurfaces characterised by extensive and

The Evolution of Flooded Interdunes

1) Stage 1: Dry river channel crossing dune field. Floor of channel comprises both fine-grained sediment of fluvial origin from earlier flood and recent aeolian sediment blown onto channel.

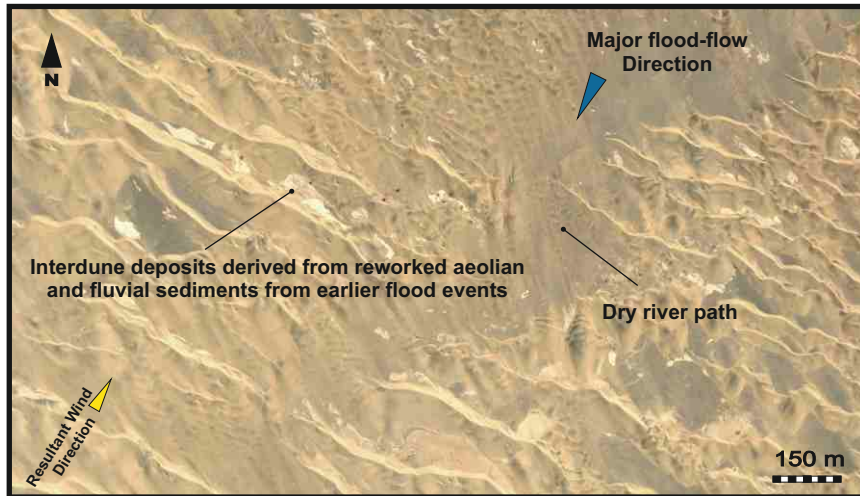


Image date:
October 2003

2) Stage 2: Sediment-laden flood waters carry reworked aeolian detritus dammed by dunes in erg margin; access to the adjacent interdunes. The fluvial flood is saturated with reworked aeolian sediments.

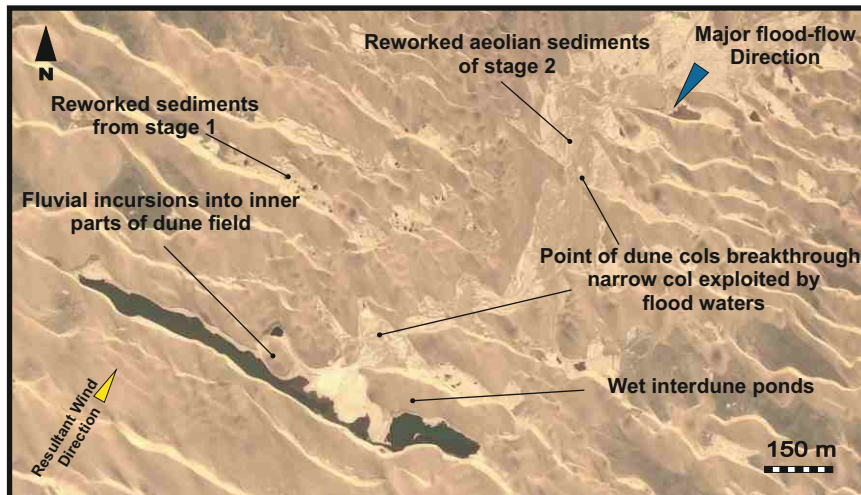


Image date:
March 2011

3) Stage 3: Deposition of suspended-load mud-dominated sediment in interdunes. Infiltration and evaporation of ponded flood water in the aftermath of a flood.

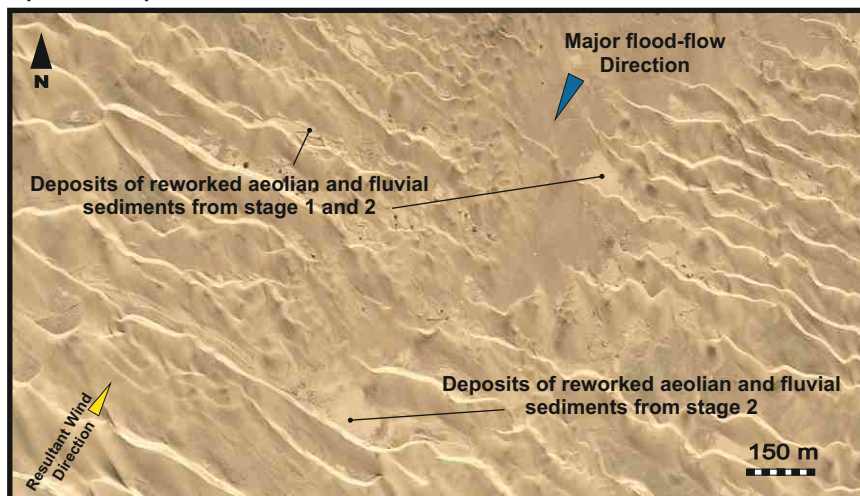


Image date:
May 2012

Figure 5.1: A time series of images of the same part of the main Hunkab River Channel, Skeleton Coast erg, Namibia showing example of series stages from different dated images of the same location in Hunkab River showing the evolution of flooded interdunes by fluvial incursions in Hunkab River and associated mud layers (e.g. low-permeability barrier).

laterally continuous erosion planes occur when the sediment budget is negative (Kocurek, 1991; Havholm et al., 1993; Mountney, 2006a). In aeolian erg-margin systems, deflation surfaces can be controlled by the rate of water-table fall over time such that the existing aeolian accumulation above water-table level will be subject to progressive and gradual deflation by the wind as it dries in response to a falling water table. Within the same area, the exposed mudstones of flooded interdunes are less likely to be deflated during water-table fall because they tend to form indurated and cohesive deposits. Instead, desiccation cracks may be developed and represent an important paleoenvironmental clue about the nature of the accumulation surface at the time of supersurfaces development and not necessarily about the nature of the accumulation surface at the time when the aeolian system was accumulating (Kocurek & Hunter, 1986; Stanistreet & Stollhofen, 2002; Mountney, 2006a). This can leave such mudstone units preserved at a level higher than the surrounding desert plain as “platforms”.

5.3 The Helsby Sandstone Formation, Cheshire Basin as a Hydrocarbon Reservoir Analogue

The Helsby Sandstone Formation (HSF) of the Cheshire Basin, England, is a direct onshore equivalent to the oil- and gas-bearing main reservoir unit in the East Irish Sea Basin (EISB), the Ormskirk Sandstone (Fig. 4.1). The HSF therefore serves as a valuable outcropping direct analogue for reservoir heterogeneity assessment of the EISB subsurface Ormskirk Sandstone succession.

The distribution of observed facies and their occurrence as associations that comprise elements with distinct geometrical properties (Chapter 4) allows the HSF to be applied as an analogue for mixed aeolian-fluvial reservoir systems from other basins. The various types of interdune elements present in the HSF (dry, damp and wet) may represent baffles and barriers to fluid flow that possess subtly different flow properties. Different interdune geometries in the HSF were considered previously by Mountney & Thompson (2002) and have been examined in more detail in this study. Ribbon-like interdune geometries occur as a result of infrequent fluvial flooding along interdune corridors; isolated lens- or lozenge-shaped interdune elements occur due to short-lived flooding between sinuous-crested dunes (see Chapter 2 for details; Fig.2.6).

The Triassic Sherwood Sandstone Group, East Irish Sea Basin (Morecambe Gas Field)

The Ormskirk Sandstone Formation (OSF) of the Triassic Sherwood Sandstone Group, East Irish Sea Basin (EISB) is considered in this section as a reservoir example. The OSF comprises both oil and gas reservoirs in the EISB, the largest of which is the Morecambe Gas Field, currently owned and operated by Centrica. The Sherwood Sandstone Group (SSG) of the EISB is interpreted as representing the accumulated deposits of a major braided fluvial system that evolved under the influence of a semi-arid climatic regime; minor aeolian dune fields and sandsheets developed in parts of the basin between the major braided fluvial networks (Colter & Barr 1975; Bushell 1986; Cowan, 1993). Aeolian and fluvial facies associations in the SSG comprise a range of lithofacies types, including mudstones that form thin layers apparently conformable with bedding, some such layers preserving characteristic features such as mud curls (Cowan, 1993). Six aeolian and fluvial facies of OSF were recognised by Cowan (1993) from cored wells. These are flat laminations of aeolian sandsheets, cross-bedding of aeolian dunes, large-scale planar-tabular cross-stratification of major fluvial channels, trough cross-lamination of ephemeral fluvial channels, wavy laminations of fluvial sheet-like deposits, and playa lake mudstones (Cowan, 1993).

Meadows & Beach (1993) argue that deposition of the sandflats (mainly aeolian dune, sandsheet and fluvial sheet-like deposits) was water-table controlled such that aeolian processes associated with relatively low water-table level were dominated by the accumulation of wind-rippled sandsheets and small aeolian dunes. Moreover, Meadows & Beach (1993) interpret the recorded fluvial channel sequences within the OSF as having a fill composed of locally fluvially reworked aeolian sediments.

Intervals of drying-upward and wetting-upward cycles within the sequence are interpreted to record gradual changes between high and low water-table levels; this indicates a temporally varying climate control on sandsheet development (Meadows & Beach, 1993). Identification of the aeolian facies is crucial since they tend to have markedly better reservoir characteristics compared to the surrounding sediments of the EISB (Cowan, 1993). The size and distribution of elements of aeolian facies and their occurrence between elements of fluvial origin gives rise to a highly heterogeneous reservoir, which has created problems during field production (Cowan, 1993). Thin aeolian intervals (< 10 m thick) form important primary reservoir zones in the EISB but are easy to misidentify during well testing (Cowan, 1993).

The volume of commercial hydrocarbon production that is derived from the aeolian parts of reservoirs in the OSF is disproportionately high, accounting for 50% to 70% of the flow into the well bore despite aeolian sandstone elements only comprising 5 to 10% of the total reservoir volume (Cowan, 1993).

The studied HSF outcrops considered in detail in this research are closely comparable to the main OSF reservoir in the Morecambe Gas Field. Both successions are characterised by aeolian and fluvial facies that accumulated in close proximity to each other in an intimate interdigitating manner. Both accumulated under the influence of a semi-arid environment (Cowan, 1993). Both record evidence of fluvial incursion into an aeolian dune-field margin. The fluvial and aeolian sandstone reservoirs in the EISB are of a very similar age and occurred laterally equivalent to those of the HSF reservoir in a similar basin setting. An important consideration in the Cheshire Basin is the associated top seal to the HSF, which consists of a thick sequence of interbedded sands and silty mudstones, the Tarporley Siltstone Formation and overlying mud-prone clastic units of the Mercia Mudstone Group. These argillaceous units are absent from the EISB where instead basal evaporites of the Mercia Mudstone Group act as top seal (Jackson et al., 1987; Mikkelsen & Floodpage, 1997).

Another well-documented outcropping example of a succession that records interaction between fluvial and aeolian environments in a dune-field margin setting is the Cedar Mesa Sandstone, SE Utah, USA, for which accumulation of aeolian dune and interdune elements was markedly influenced by the level of the water table and the position of a lateral dune-field margin into which major fluvial channelised systems passed (cf. Langford & Chan, 1988; Mountney & Jagger, 2004; Taggart et al. 2010). The cycles of wetting-upward and drying-upward recognised in the Triassic OSF of the Morecambe Gas Field (Meadows & Beach, 1993) are likely to have originated through similar processes to those of the Helsby Sandstone Formation (Thompson, 1970; Mountney & Thompson, 2002; Chapter 4, this study) and the Cedar Mesa Sandstone (Langford & Chan, 1988; Mountney & Jagger, 2004; Taggart et al. 2010).

5.4 Heterogeneity Complexity of Mixed Fluvial and Aeolian Reservoirs

Preserved mixed fluvial and aeolian successions are dominated by aeolian successions associated with a complex spatial variation in sedimentary

architecture that records a transition from a near exclusively dry aeolian erg-centre setting, to a water table-controlled aeolian-dominated inner dune-field margin setting, to a fluvial-dominated outer dune-field margin setting (cf. Mountney & Jagger, 2004).

In hydrocarbon reservoirs, dune and non-dune (fluvial) facies typically exhibit large porosity and permeability contrasts (Herries, 1993; Romain & Mountney, 2014). Fluvial and aeolian facies associations in the Rotliegend Group, southern North Sea, for an example, have wide range of porosity (Φ) and vertical permeability (k_h) within reservoir sand bodies (Sweet et al., 1996). Typical values vary between interpreted facies associations: aeolian dune association (Φ : 13 -15%, k_h : 0.62 – 3.2 mD); aeolian sandsheet association (Φ : 11 -13%, k_h : 0.57 – 0.62 mD); fluvial channel fill association (Φ : 11%, k_h : 0.033 – 0.045 mD).

Successful aeolian reservoir characterisation in dune-field margin settings requires a recognition of facies geometries (i.e. architectural elements). This requires an understanding of whether and how architectural elements are influenced by factors of long-lived and regional extent, or by local factors that are short-lived. Significant and widespread expansion of fluvial systems into erg margins and associated dune-field retreat may occur in response to the onset of wetter climatic conditions (e.g. Clemmensen et al., 1994). Alternatively local fluvial incursions into otherwise aeolian-dominated systems may occur as a result of localised interdunal flooding that is short-lived (e.g. Herries, 1993). In each of these cases, the geometries of preserved fluvial and aeolian elements would differ markedly but would be difficult to predict from one-dimensional core or well-log data alone in subsurface settings. The extent and continuity of fluvial elements in dune-field margin reservoir successions dictates the lateral and vertical continuity of aeolian sandstone elements within such reservoirs (Herries, 1993).

The three-dimensional model depicted in Figure 5.2 demonstrates the typical sedimentary architecture of aeolian dune-field margin successions based on understanding gained from case studies of the Skeleton Coast erg and the Helsby Sandstone Formation of the Cheshire Basin. The presented model exhibits the following traits:

- 1) Spatial variations of aeolian accumulation in dune-field margin settings that are controlled by episodes of water-table rise accompanied by on-going dune migration. Elevated water-table level will affect rates of sediment supply and availability for aeolian transport as interdunes become progressively flooded.

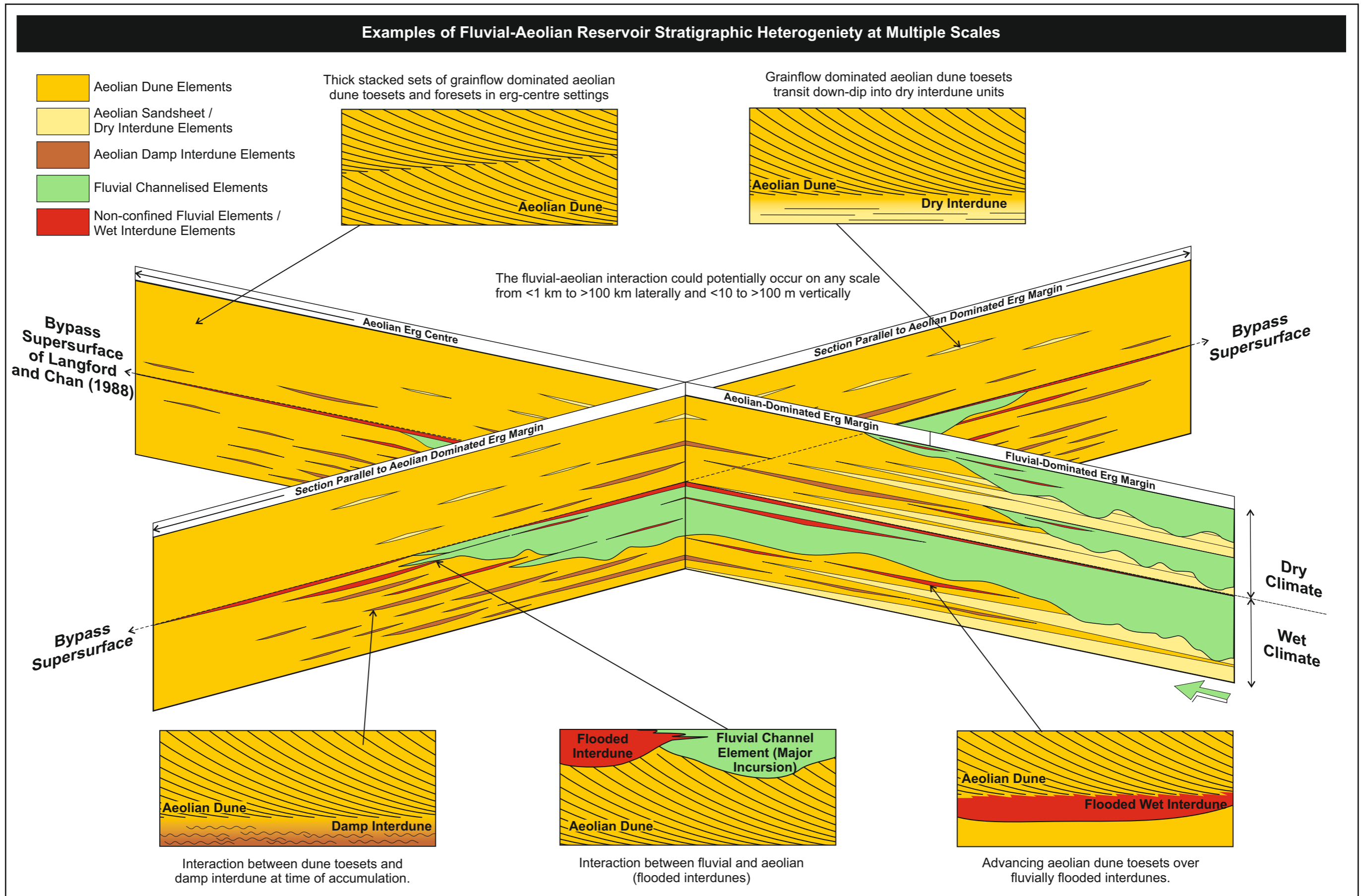


Figure 5.2: Three-dimensional depicting stratigraphic heterogeneity at multiple spatial scales in an aeolian erg margin setting subject to fluvial incursion. This model can be used to account for complexity in fluvial-aeolian reservoirs within erg-margin settings. The model depicts spatial variations of fluvial and aeolian elements in the transitional zone from erg margin to its centre. Examples of geometric and facies relationships between fluvial and aeolian elements at smaller scale are shown. No particular scale implied.

2) Variations in the geometry of dry, damp and wet (flooded) interdune elements that aligned parallel and perpendicular to fluvial flood corridors reflect the morphologic expression of the interdunes. Spatially isolated (closed) interdunes may remain dry if the water table remains below the accumulation surface; examples include interdune hollows in erg-centre settings. Locally interconnected interdunes may become damp and/or wet as flood waters pond in the aeolian-dominated dune-field margin. Fully interconnected (open) interdunes that are linked directly to major fluvial fairways will be prone to repeated flooding; such types are more common in fluvial-dominated dune-field margin settings.

3) Small-scale intercalations between aeolian dune and interdune facies are useful to demonstrate whether sedimentation in dry, damp and wet interdune settings took place coincidentally with on-going aeolian bedform migration.

The model depicted in figure (5.2) serves as a useful tool for assessing possible causes of stratigraphic heterogeneity in complex and problematic exploration targets in erg-margin successions where expansion of fluvially flooded interdunes in front of migrating dunes was responsible for generating the preserved architectural expression.

5.5 Summary

This chapter has applied the generic findings from the case studies of Chapter 3 (study of the modern Skeleton Coast erg) and Chapter 4 (study of the Triassic Helsby Sandstone Formation) to develop a set of predictive tools that can potentially be employed for characterisation of subsurface hydrocarbon reservoirs. The examples from the Skeleton Coast erg can be used as a potential modern analogue for comparable subsurface reservoirs interpreted as successions of arid or semi-arid origin in which fluvial systems penetrate into dune-field margins in a variety of ways, in some cases passing directly through the dune field during major flood events. The development of low-permeability barriers associate with preserved flooded interdune elements can be explained through analysis of historical satellite-imagery data. The preserved sedimentary succession in the Helsby Sandstone Formation can be used as an outcrop analogue for hydrocarbon reservoirs in subsurface settings.

The generic model presented here (Fig. 5.2) is useful in assessing the likely hydrocarbon reservoir potential of erg-margin successions; it documents the expected extent and connectivity of a range of types of low-permeability

baffle units associated with fluvial flooding into otherwise aeolian-dominated erg-margin systems. The model predicts the typical spatial variation of aeolian and fluvial elements from an erg margin to its centre. The distribution of these elements can assist in the identification and ranking of the most attractive exploration targets in subsurface examples for which other types of primary sedimentological data (e.g. core data) are lacking or sparse.

Chapter 6

Conclusions and Future Work

6.1 Introduction

The overall aim of this research to understand the complexity associated within fluvial and aeolian depositional systems in arid to semi-arid erg-margin settings, and to develop generic models for these systems that account for the influence of allocyclic versus autocyclic controlling factors on the accumulation and preservation of modern and ancient successions. A series of sedimentary models have been developed within this thesis in order to explain the evolution and sedimentary expressions of modern and ancient desert systems. This chapter presents a brief set of conclusions to summarise the primary learnings from case studies of the Skeleton Coast erg and the Helsby Sandstone Formation. In doing so, this chapter discusses some general issues related to the sedimentology and stratigraphy of arid to semi-arid desert successions. This chapter also highlights the potential future work that could be undertaken as a follow-on to this research to further develop our understanding of sedimentary processes associated with, and mechanisms of preservation of, erg-margin systems.

6.1.1 The Skeleton Coast Erg, Namibia

1. The analysis of the remotely sensed images reveals the history of the Hoanib, Hunkab and Uniab rivers at locations where they interact with the aeolian erg, thereby yielding a valuable record of interaction between competing fluvial and aeolian processes.
2. The nature of aeolian dune-field boundaries, which may expand and contract over time is influenced by climatic shifts that act as a primary control on the relative dominance of fluvial over aeolian processes in mixed depositional desert settings.
3. High-energy flood events with high rates of fluvial discharge into aeolian dune-field margins increase the potential for erosion of existing aeolian deposits.
4. Under such conditions, erosion, transport and re-deposition of fluvially reworked aeolian sediments may occur where fluvial floods transport

aeolian sediments downstream to areas where floods discharge wanes or terminates (Fig. 3.24c).

5. The supply and availability of sediment for aeolian construction in dune-field settings are determined by the frequency of erosion, transport and re-deposition of fluvially reworked aeolian sediments that may occur where fluvial floods transport aeolian sediments downstream to areas where floods discharge wanes or terminates.
6. The presence of increased vegetation cover on dunes at the eastern dune-field margin (e.g. Hoanib River) acts to limit the availability of aeolian sand that may be considered susceptible to re-working and transportation by floods.
7. Aeolian dune morphology within the studied region controls the shape of the adjoining interdunes. Fluvial systems are either able to penetrate along open interdune corridors (Fig. 3.24d and 3.24f) or are dammed by dunes that surround closed interdunes (Fig. 3.24c).
8. The nature of interdune flooding observed in this study is determined by morphological dune type, the orientation of dune bedforms relative to the main fluvial flow direction, the geometric nature of interdune corridors which are defined in terms of their width and length, and the rate of aeolian dune and interdune migration relative to the flood frequency, as well as by the nature of the flood events themselves.
9. This study is important because it serves as the basis for a series of predictive models underpinned by observed relationships that can be useful in predicting aeolian high-quality reservoirs. The outcome of this work is the construction of a series of qualitative facies models that can be used to assess likely subsurface reservoir prospectively in mixed fluvial-aeolian successions that range from aeolian-dominated inner erg-margin settings to fluvial dominated outer erg-margin settings.

6.1.2 The Helsby Sandstone Formation, Cheshire Basin, UK

1. The Helsby Sandstone (Anisian) formations is well-suited to sedimentary lithofacies analysis and to the development of an architectural-element-based approach with which to develop a set of detailed depositional models for water-controlled erg-margin systems.
2. The cyclicity of fluvial and aeolian architectural elements within the Helsby Sandstone Formation indicates that accumulation occurred in response to episodic ephemeral river incursion into the margins of an

aeolian dune field. This arose in response to one or a combination of the following mechanisms: 1) in response to large-scale autogenic factors such as episodic regional-scale river avulsion; 2) in response to climate change that caused repeated expansion and contraction of the margin of an aeolian dune field, possibly in response to Milankovitch-style cyclicity.

3. The widespread occurrence of fluvial multi-storey, multi-lateral amalgamated channel-fill elements (F1) suggests that the fluvial system undertook repeated avulsion and new channel development in response to a series of large-magnitude flood events that led to rapid channel incision, aggradation and avulsion.
4. The presence of lateral downwind continuation of intertonguing damp-interdunes with adjacent thin dune toesets in the Beacon Hill and Dunsdale Hollow aeolian successions indicates that the interdunes migrated coevally with on-going aeolian-dune migration.
5. The nature of the interdune surfaces that bound damp/wet interdune elements was controlled by the temporal changes in the ratio between the rate of aeolian-dune migration and rate of relative water-table rise.
6. In this study, the interpretation of the analysed aeolian strata demonstrates that the aeolian-dunes were water-table influenced and controlled.
7. The preserved form of aeolian dune and interdune architectural elements records the episodic preservation of a wet aeolian system, possibly resulting from an absolute rise of continental water-table in response to a long-lived temporal increase in climatic humidity due to environmental change during the episode of accumulation, though more likely due to the progressive subsidence of the accumulation beneath the absolute water table.
8. The outcrop data studied here can be compared to analogous examples of fluvial and aeolian interaction known from systems present in modern settings (e.g., Skeleton Coast erg, Chapter 3) and to subsurface reservoir successions (e.g. Triassic reservoirs of the East Irish Sea Basin, Chapter 5).

6.2 Future Work

The potential for identification of types of fluvial-aeolian interaction in modern systems through analysis of satellite imagery data such as that provided by Google Earth is increasing rapidly as new and higher-resolution datasets with near-global coverage become available. There remain a number of important research questions to be addressed in relation to the sedimentology of the world's deserts, many of which reflect global concerns about environmental change (Bullard & Livingstone, 2002). Many existing 'modern analogue' studies are primarily based on descriptive accounts and many lack contemporary process data (Bullard and Livingstone, 2002).

In the field of aeolian geomorphology, new techniques need to be considered as means to generate additional data sources. Such data sources will lead to the establishment of new study methods for better understanding geomorphic elements arising from fluvial-aeolian interactions, and their controlling factors. For instance, controls on interaction between ephemeral channels and flooded interdune corridors during flood events need to be historically recorded. This can be achieved by monitoring using real-time satellite tracking systems, rather than relying on a somewhat crude analysis of a time series of images of variable quality. This will improve our understanding of how aeolian dunes and neighbouring interdunes expand and contract through time. Thus, such approaches will enable models to be developed that more effectively explain the temporal sedimentary evolution of dune-field margin systems.

In addition, the diversity of styles of fluvial and aeolian system interactions is primarily determined by the angle of incidence of fluvial systems relative to the orientation of aeolian bedform crestlines. Numerical datasets characterizing the orientation of fluvial systems relative to the trend of aeolian bedform crestlines need to be considered in the future through both analysis of a broad range of modern system types, and also from analysis of preserved ancient outcrop successions, especially those from the Quaternary record where the preserved deposits can still be related to the formative fluvial-aeolian processes. This will be useful to predict and better understand the distribution of non-permeable units (e.g. fluvially flooded interdunes), which act as stratigraphic heterogeneities in subsurface reservoirs and water aquifers. The angle of fluvial-flow orientations of major floods – notably those that flow in a direction perpendicular to the trend of the crestlines of migrating aeolian bedforms – may change over time to become more oblique (e.g. due to channel avulsion) either gradationally or

abruptly, possibly until the original fluvial-flow orientation shifts to 90 degrees from its original orientation, thereby forming a new river pathway parallel to the trend of aeolian bedforms (i.e. along fluvially flooded interdunes). Recoding the history of the angle of fluvial-flow orientations in the future is also important to further develop our understanding of preservation mechanisms of low-permeable units at erg-margin systems.

Profiles of rivers in erg-margin settings commonly demonstrate how such rivers are affected by aeolian topography (Fig.3.2b). Quantitative elevation profiles of low-relief fluvial-incisions into erg-margins need to be measured using differential GPS, LiDAR and photogrammetry methods to assess how convex-up upstream-to-downstream fluvial profiles are influenced by aeolian bedforms from fluvial-dominated erg-margins, through to aeolian-dominated erg-centre regions. Such analysis will promote investigation opportunities that enable consideration of other external forcing factors, such as tectonic, climate, and fluvial sediment load size and type.

For preserved successions, traditional techniques such as trench excavations and GPR analysis of the Quaternary record of fluvial-aeolian interactions in erg-margin settings (and accompanied by OSL dating techniques) are also important and needs to be the focus of future attention. Such studies will record dune, interdune and fluvial architectural element geometries but will crucially show how such elements respond to well-documented episodes of climate change. Dating using OSL and related techniques will provide the necessary chronological framework. The development and widespread implementation of this technique could lead to more challenging and complex interpretations of the origins of aeolian and fluvial process interactions within accumulated stratigraphic sequences.

In general, analogous quantitative facies models, which are useful for predicting subsurface hydrocarbon reservoirs for fluvial-aeolian systems, currently remain too limited in their predictive capabilities. The construction of digital facies models (see Colombera et al., 2013) requires the compilation of sets of quantitative information from detailed outcrop datasets and results from ancient and modern case studies undertaken at different scales, and which include analysis of depositional elements, architectural elements and facies units (cf. Colombera et al., 2013). The compilation of such digital datasets for mixed fluvial and aeolian systems will help to identify subtle interrelationships in terms of facies and element distributions.

Finally, more detailed investigation of temporal environmental changes that drive shifts in the type of fluvial-aeolian interactions at the transitional zone

from one stable climate state to another is necessary (Bullard & Livingstone, 2002). This may be achieved by undertaking detailed studies to identify potential variations in tectonic regime or source catchments; or by petrographical and geochemical analyses of aeolian sandstones to explain climatic variations during the deposition of fluvial-aeolian successions.

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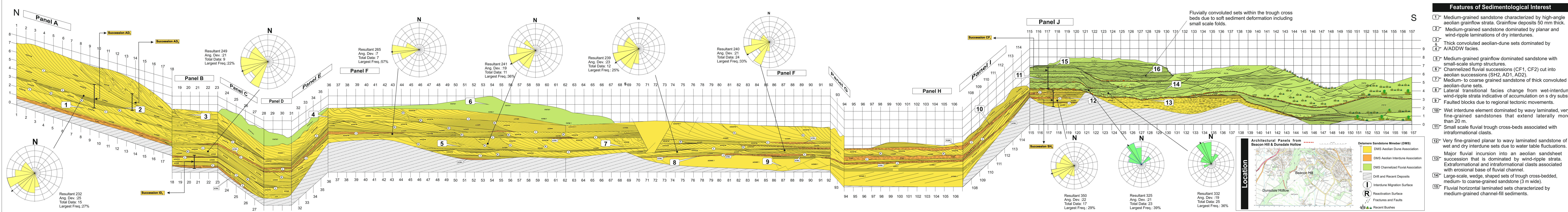
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Appendix

Appendix 1: Architectural panels depicting the stratigraphic architecture of the Helsby Sandstone Formation (HSF) as observed in Beacon Hill and Dunsdale Hollow outcrops.

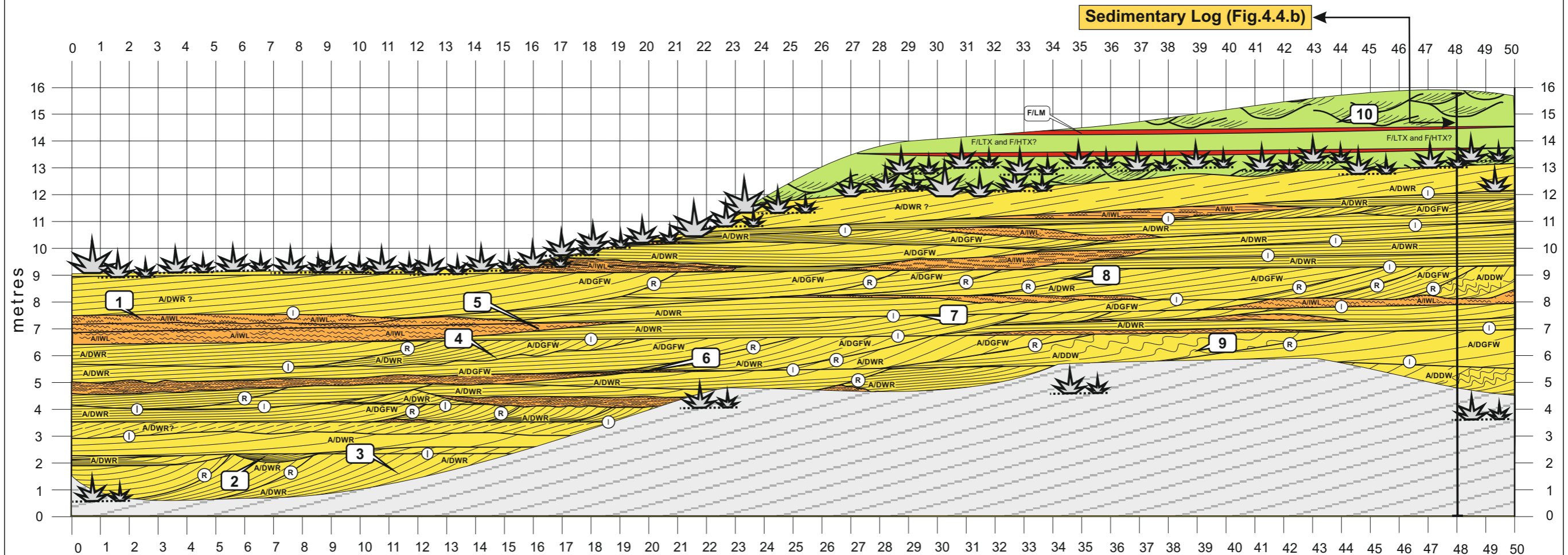
Figure 4.19: Stratigraphic Architectures of the Lowermost Section of Helsby Sandstone Formation in Beacon Hill, Frodsham Town, UK. See figure 2 and 29 for location.



Features of Sedimentological Interest

- 1 Medium-grained sandstone characterized by high-angle aeolian grainflow strata. Grainflow deposits 50 mm thick.
- 2 Medium-grained sandstone dominated by planar and wind-ripple laminations of dry interdunes.
- 3 Thick convoluted aeolian-dune sets dominated by A/ADDW facies.
- 4 Medium-grained grainflow dominated sandstone with small-scale slump structures.
- 5 Channelized fluvial successions (CF1, CF2) cut into aeolian successions (SH2, AD1, AD2).
- 6 Medium- to coarse grained sandstone of thick convoluted aeolian-dune sets.
- 7 Lateral transitional facies change from wet-interdune to wind-ripple strata indicative of accumulation on a dry substrate.
- 8 Faulted blocks due to regional tectonic movements.
- 9 Wet interdune element dominated by wavy laminated, very fine-grained sandstones that extend laterally more than 20 m.
- 10 Small scale fluvial trough cross-beds associated with intraformational clasts.
- 11 Very fine-grained planar to wavy laminated sandstone of wet and dry interdune sets due to water table fluctuations.
- 12 Major fluvial incursion into an aeolian sandsheet succession that is dominated by wind-ripple strata. Extraformational and intraformational clasts associated with erosional base of fluvial channel.
- 13 Large-scale, wedge, shaped sets of trough cross-bedded, medium- to coarse-grained sandstone (3 m wide).
- 14 Fluvial horizontal laminated sets characterized by medium-grained channel-fill sediments.

Stratigraphic Architecture from Dunsdale Hollow, Frodsham, UK



- Features of Sedimentological Interest**
- 1 Stacked beds of wet interdune origin dominated by wavy laminated, very fine-grained sandstones that extend laterally more than 13 m. Indicates cyclic deposition within interdune depressions.
 - 2 Fine-grained sandstone dominated by planar and wind-ripple laminated sandstone; dry interdunes.
 - 3 Aeolian grainflow strata characterized by high-angle stratified, medium-grained sandstone. Grainflow deposits are each up to 60 mm thick.
 - 4 Erosive base of aeolian dune dominated by grainflow strata.
 - 5 Lateral transitional facies change from wet interdune to wind-ripple strata indicative of dry interdune.
 - 6 Lateral thinning of wet-interdune facies elements.
 - 7 Stacked aeolian dune-strata separated by interdune surface.
 - 8 Stacked aeolian dune-strata separated by interdune surface.
 - 9 Deformed aeolian-dune sets (slumped) as a result of wetting of advancing-dune.
 - 10 Fluvial succession cuts into older aeolian strata.

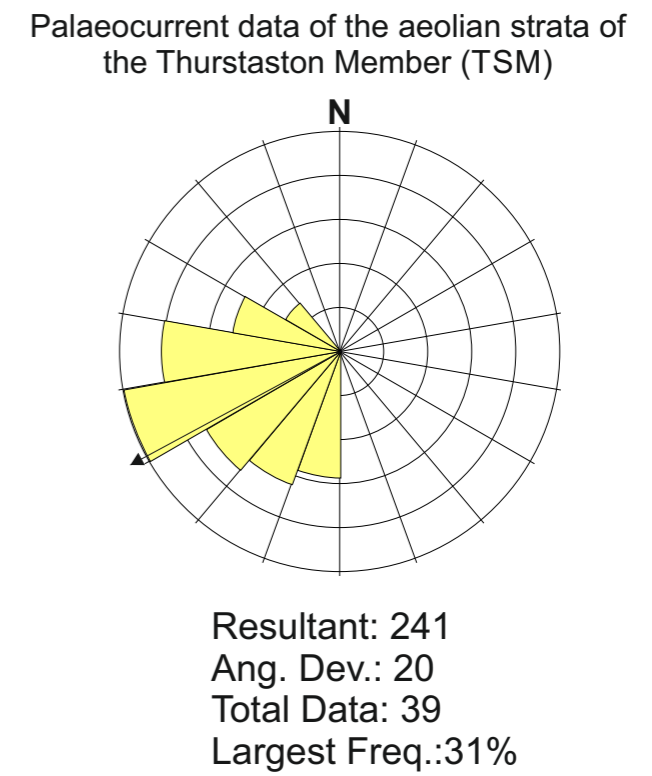
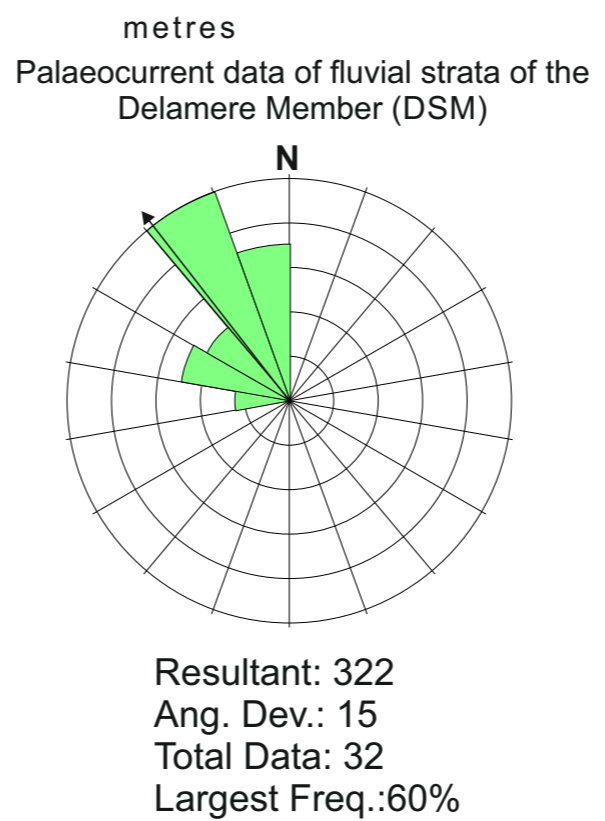


Figure 4.20: Architectural panel depicting the stratigraphic architecture of the aeolian strata of the Thurstaston Member (TSM) at observed in Dunsdale Hollow. See figure 4.2 and 4.29 for location.