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Aeolian sands and sandstones

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AEOLIAN SANDS AND SANDSTONES

by

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ABSTRACT

Models for the development of ergs and aeolian cross-bedding are deduced. These theories are partly deductions from the modern, partly explanations of the ancient. Ergs may be non-aggrading without complete sand cover or aggrading with or without complete sand cover. Aeolian cross-bedding generally comprises 2 superimposed orders of nested trough-shaped sets. One is developed by migrating draa, one by migrating dunes. Four orders of bounding surface exist: 2 migration surfaces and 2 modification surfaces.

The Yellow Sands of N.E. England represent longitudinal draa. The draa were initiated as sand patches, grew by the migration and climbing of sinuous transverse dunes, and stabilised at equilibrium by the development of capping sand-sheets. A bimodal wind regime pertained. A modern analogue of the Sands at equilibrium is provided by 'whalebacks' in the Egyptian Western Desert. The streamlined shape of the draa, well-packed surface, disposition as a series of baffles and distance from the influx minimised reworking in the Zechstein transgression, which drowned the draa in a matter of years.

The Bridgnorth Sandstone accumulated from slipfaced and slipfaceless transverse draa. Dune cross-bedding forms sets 2-4 m thick, 20-40 m wide. Slipfaced draa sets are 6-10 m thick, at least 100 m wide.

Characteristics of the Bridgnorth Sandstone, Yellow Sands and other formations of aeolian sandstone are explained and illuminated by the erg and cross-bedding models developed.

The Bridgnorth Sandstone is slightly finer than the Yellow Sands, and contains more sandflow lamination, less wind-ripple. These features follow from the contrasting erg and bedform types.

It is suspected that sandflows thicker than 80 mm may be confined to draa-sized slipfaces.

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PREFACE

In recent years there has been something of a revolution in the study of ancient aeolian sandstones. Concerning bedforms, the implications of Ian Wilson's Ph.D. thesis (1970) were partly translated to the ancient by Brookfield (1977). The lamination of aeolian sands was illustrated and explained by Hunter (1977). In the past these two principal features of aeolian sandstones tended to be dismissed in a few words or sentences: the lamination as good, the cross-bedding as large-scale. Minor features, such as raindrop imprints, wind-ripple form-sets and fossil footprints attracted much greater attention and carried great weight in the identification of aeolian sandstones. This is no longer the case: the major features can now be examined and analysed in detail, and confident interpretations of an aeolian origin made on features present throughout the mass of rock. This new approach has been applied to the Entrada Sandstone in the U.S.A. by Kocurek (1981a,b) and to several U.K. formations in this thesis. It is also expounded in Kocurek and Dott (1981) and Hunter (1981). A gratifying feature of the present work, symptomatic of these developments, is that most of the key references date from the past decade.

The new approach is extended and developed in this thesis, by a more sophisticated and general analysis of cross-bedding, by the analysis of grain size character- istics with reference to lamination, and by the acquisition of experience and data. Similar achievements were made by Kocurek (1981), especially with reference to the internal anatomy of ergs and the behaviour of bed- forms.

The first 3 chapters of this thesis are concerned with features of modern ergs relevant to the analysis of the ancient: the characteristics of ergs, bedforms and cross-bedding, and grain size characteristics and lamination. Chapter 4 leads into the ancient by reviewing the Early Permian setting in Britain. Chapters

5-9 are occupied with descriptions of U.K. aeolian sandstones. The characteristics of aeolian sandstones are then reviewed in Chapter 10.

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DECLARATION

The content of this thesis is the original work of the author (other people's work where included is acknowledged by reference). It has not been previously submitted for a degree at this or any other university.

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Durham

December 1981

ERRATA

There is no page 41, 54, 203 or 274.

CHAPTER   1

MODERN   ERGS

SECTION 1.1   CLIMATE AND VEGETATION

The Definition of the Term   Desert

Geographers and biologists have developed precise definitions of the term   desert .Geologists, being confined to an examination of fragmentary final results rather than a contemporary view of causes, effects and processes, are necessarily much more subjective.

In geography, Cooke & Warren (1973) recommend the use of Thornthwaite's moisture index as adapted by Meigs (1953).   By considering potential evaporation and water balance, the terms semi-arid, arid and extremely arid are precisely defined.   Deserts comprise the arid and extremely arid areas.

Lamb (1972) reports Köppen's classification of climate as the most widely used.   This attempts to match climate to type of vegetation.   Five main groups are distinguished: tropical rain, arid, warm temperate rain, boreal forest and snow, and cold snow climates. The groups are subdivided by juggling various characteristics of precipitation and temperature, with each group and subdivision given a standard shorthand notation. For example, BS denotes an arid climate with bush or grass-land (steppe), BW a desert.   Further letters may be added for further subdivisions (see Lamb, 1972 and e.g. The Times Atlas of the World).   Köppen's desert subdivision is shown on fig.1.1 with ergs added.

The Causes of the Desert Climate

The meteorological reasons for the development
of deserts have been discussed and reiterated as many
times as there are textbooks concerning deserts or
meteorology. Frakes (1979, pp.18-19) gives a concise
account, here rearranged into a list. There are four
principal mechanisms whereby deserts are generated:
1) Aridity produced by the descending limb of the
   subtropical atmospheric circulation cell in the
   trade wind zone (between 20° and 30° of latitude
   over land).
2) Aridity developed by the rain shadow effect in
   the lee of high mountains.
3) Aridity due to extreme continentality, ie. great
   distance from the ocean (the moisture source).
4) Aridity associated with subtropical continental
   west coasts, adjacent to upwelling cold ocean
   currents and at the distal end of the paths of
   continental trade winds.

Modern deserts can be attributed to these causes,
in combination more often than not: for example NW
Africa (factors 1 and 4), western South America (1,2
and 4), Patagonia (1) and the Chinese deserts (2 and 3).
Referring to Koppen's classification of climate (fig.
1.1), permutations and combinations of these mechanisms
lead to officially designated deserts occurring at
anything from 48°N (near the Caspian Sea) to the
Equator (in east Africa), to 50°S (in Patagonia).

Vegetation in Ergs

Vegetation prevents the wind moving sand by reducing the wind velocity at ground level. If vegetation can become sufficiently established it will kill an active sand bedform; this is the basis of many sand control programmes (see, e.g. Gore, 1979). Removal of vegetation has the converse effect; wind erosion and transport being greatly enhanced. This process is presently occurring in the Sahel and occurred in the U.S.'Dust Bowl' in the 1930's (see also Armstrong, 1976).

Vegetation within ergs is usually thought of as being non-existent, though Wilson (1971a), describing a journey across the Erg Oriental in Algeria, reports encountering nomadic shepherds complete with flocks finding enough food to survive on deep in+o.-the sand area.

McGinnies (1968, pp.381-566) gives an extensive review of desert flora. Those of concern to sediment transport by the wind are the perennial types which by their persistence may substantially influence sand bedforms. Vegetation is concentrated in inter-bedform areas, and when present in sufficient quantity its effect is to subdue active bedforms altogether or generate parabolic dunes. These bedforms take the shape of a parabola with the closure downwind, the tails being anchored by clumps of vegetation. Since they also seem to require incomplete sand cover they are probably not important geologically. McKee (1966) trenched one at White Sands National Monument, New Mexico.

Vegetation may be significant as a chemical source or influence in local diagenesis and as generators of trace fossils. The formation of plant root moulds in modern desert sands (dikaka) is described by Glennie and Evamy (1968) and Glennie (1970). Dikaka is most prevalent near sources of water, coastlines or wadis and is extensive for example in Lower Pliocene aeolian sands on the Trucial Coast. The dikaka may be preferentially cemented by gypsum.

On the whole it is almost axiomatic that the accumulation of large quantities of wind-blown sand in actively migrating bedforms precludes any significant floral presence: the two are mutually exclusive. In peripheral areas of ergs where water is more accessible and sand cover less complete, plants are more common, but these are areas where thick aeolian sequences will not develop. Also, as one goes back into geological time, xerophytic plants must become more and more rare. Active ergs may therefore have been more widespread in the past, being circumscribed only by the effects of fluvial erosive processes and plants less specialised and tolerant than their modern descendants.

SECTION 1.2 THE DIMENSIONS OF ERGS

'Erg' is an Arabic word used in the central and N.W. Sahara to describe an area of wind-laid sand. It was adopted and more precisely defined by Wilson (1973) as "an area where wind-laid sand deposits cover at least 20% of the ground, and which is large enough to contain draa". This definition will be adhered to in this

thesis. In most modern ergs blown sand does not cover all the ground and the underlying substrate is visible between the bedforms. Because of this, and the fact that sand cover often decreases gradually towards the margin of an erg, the arbitrary specification of 20% sand cover becomes necessary in the definition.

Ergs where bedforms are separated by smooth sand-sheet or sabkha surfaces may either be areas where aeolian bedforms are migrating across non-aeolian material and leaving no deposits, or both the interbedform areas and the bedforms may be migrating in concert and depositing a sequence of interbedded sediments. Unless the interbedform areas are bedrock or deflated areas of older water-lain sediments, the distinction cannot be made without some subsurface exploration. The distinction is geologically crucial: aeolian sand contained in a veneer of bedforms has a very low preservation potential. Wilson (1973, p.93) states, "During the present study, no large sand region with 100% sand cover and bedforms overlying buried (aeolian) deposits was found, although McKee and Tibbits (1964) claimed that 1000 ft (300 m) of aeolian sand underlay some ergs in Libya". It is not clear on what data this assertion is made, though, from his writings, it seems likely that Wilson regarded incomplete sand cover as precluding the presence of aeolian deposits at depth. This is the conclusion that would be made from his erg model (Wilson, 1971b). However, this is shown and section 1.3 to be incomplete: ergs with incomplete

sand cover can aggrade and accumulate aeolian deposits.

The predominance of ergs with incomplete sand cover at the present day can be verified by examining the many magnificent satellite photographs in McKee (1979) and Short (1976).

Having laboured the vertical dimension of ergs, what of their lateral extent? The Algodones Dunes in California, the only active U.S. erg bearing draa, have an area of ~ 460 km^2. The Rub' al Khali of Saudi Arabia covers 560,000 km^2. Wilson (1973) compiled a list of 58 ergs exceeding 12,000 km^2 in area, with additional information on the extent of vegetation, state of sand cover and whether or not draa are developed. A size-frequency histogram of all the world's ergs is redrawn from Wilson (1973) in fig. 1.2.

Ergs are confined to topographically low-lying areas, of varying absolute height. This is partly because the most easily deflated sand sources are usually found in those areas and partly because the airflow accelerates over high ground. Erg margins may be strongly topographically controlled (e.g. Erg Chech-Erg Iguidi, Algeria; Takla Makan, China), may be aligned to the sand transport path (Calanscio Sand Sea, Rebiana Sand Sea, Libya; Great Sand Sea, Egypt) or may be difficult to attribute to any particular cause merely by inspecting a good atlas.

SECTION 1.3 ERG DEVELOPMENT

Constraints on the Growth of Ergs

The aim of this section is to describe and assess the importance of the various factors that influence the development of ergs.

The minimum necessary condition compelling the aggradation of an erg body (i.e. for net deposition to take place at the bases of the bedforms migrating over the erg surface) is that the net sand-drift rate (Q) should decrease with distance in the direction of flow (x), i.e. that $\frac{dQ}{dx}$ is negative. The net sand-drift rate measures the amount (by mass or volume) of sand blown past a point over a certain length of time. It is usually expressed in units such as kilograms or cubic metres per metre width (normal to net drift) per year. If during the formation of an erg $\frac{dQ}{dx}$ is still negative after the bedforms have become fully developed, deposition must continue and the bedforms will climb, regardless of whether the basin in which the erg rests is subsiding. Subsidence will serve to counteract the constriction and acceleration of airflow that must otherwise accompany the vertical growth of an erg (see fig. 1.3). Brookfield (1977) dismissed the notion of net whole erg sedimentation at a rate any greater than 10^{-5} m/year in order to discard simple climbing of bedforms as a mechanism of erg aggradation (op.cit., p.317). He went on (p.319) to advocate subsidence of 3×10^{-4} m/year as a suitable control of sedimentation. However it is implicit in his argument that the erg

surface is maintained at a constant level during this
subsidence. This could only be accomplished by the
climbing of bedforms at net sedimentation rates orders
of magnitude greater than those he previously rejected.

In this context it is possible to reduce the
genesis of an erg to a consideration of three factors:
the development of a suitable topographic basin; the
onset of aridity; and an abundant sand supply. Given
that these requirements are fulfilled, in areas where
$\frac{dQ}{dx}$ is negative ergs must develop. These ergs may then
be able to build up a considerable sediment body thick-
ness before restricting the airflow sufficiently to
accelerate the wind to the extent that deposition ceases.
It is only at this stage that the rate of subsidence of
the basin need be considered.

Wilson (1971b) also reached this conclusion, describing
an erg significantly constricting the airflow as being
in first order equilibrium. In this state the amount
of sand entering the erg equals that leaving it, and the
erg no longer changes under the regional sand-drift.
As an example he suggested that the Erg Oriental might
reach this situation when its centre is 2000m thick
(present mean sand thickness 26m). From estimated
sand-drift maps this would require a further 120 Ma
of active deposition, or 12 Ma to reach 200m, a
perfectly respectable thickness for an ancient aeolian
sandstone formation. (Wilson (op.cit.) quotes 50 Ma
to reach 2000m but his calculation is incorrect, also
his formula is in error as printed; bulk density should

be in the numerator, not denominator).

Further advocating the primary influence of sub-
sidence, Brookfield (1977) cites the interfingering
of aeolian and fluviatile (mainly alluvial fan)
sediments of ?Lower Permian age in the Dumfries Basin
of SW Scotland as evidence of a rise in base level
controlled by subsidence. However drainage in the
basin is likely to have been endoreic and largely
by infiltration. With basin centre sedimentation wind
controlled, the fluvial base level would probably have
been independent of external factors.

The main problem on this singularly academic point
is that it is almost impossible to gauge how far an
erg might grow vertically without accelerating the first
order flow sufficiently for deposition to cease. Over
an area the size of an erg the acceleration of the wind
must come about by constriction of the whole troposphere
rather than by projection into the atmospheric boundary
layer. The flow then is $\sim$15 km deep and as wide as a
cyclonic or anticyclonic system. Brookfield's arguments
require these forces to respond to subsidence measurable
in decimetres per millenia.

Subsidence and tectonics probably have a greater
indirect than direct influence on aeolian sedimentation.
In an actively subsiding area the basin to highland
relief will constantly be rejuvenated, maintaining
bedrock exposure, promoting weathering and allowing
active alluviation. In stable areas relief will
slowly decrease, precipitation will generate little
new sediment and potential sand source areas will

quickly develop an armour of lag gravel, minimising

deflation and aeolian sand supply.

In many instances the rate of aeolian sand deposition

is immediately controlled by the rate of deflation of

pre- or co-existing alluvial sediment, whether on

fans, braidplains, bajada or elsewhere. If this

deflation rate is greater than the rate of fluvial

deposition, or if no such loose sediment is available

to the wind, aeolian deposition will be limited by the

rate of rock weathering.

Numerical values of weathering rates are given on

fig. 1.4. Fig. 1.4a is taken from Wilson (1971b) and

represents the most pertinent data available among a

large geomorphological literature on the subject.

Wilson derived the data by following his sandflow

(= sand-drift) lines from peaks (where Q = O) to the

nearest erg margin (where Q reaches saturation),

observing the types of terrain traversed and thus

evaluating deflation rates. It should be stressed that

this information is grain size specific, for 0.1 - 0.5

mm sand. Other work (e.g. Corbel, 1959; Langbein &

Schumm, 1958, shown in fig. 1.4b; Young, 1974),

whilst useful, considers sediment of all grain sizes

and a guess must then be made of the contained

proportion of sand susceptible to aeolian transport.

Also, in being derived from Saharan data, fig. 1.4a

applies to very large areas of generally quite low

relief and stable, intracratonic tectonics. In post-

Variscan N.W. Europe, home of all the rocks described

later in   this thesis, relief may well have been

greater and the tectonics more active.

A Model for the Development of Ergs

Wilson (1970 & 1971b) gives a theoretical model
for the development of ergs, expressed in terms of
deflation, sand-drift, and deposition. A few salient
points will be summarised, with occasional modification.

Initially it might be supposed that bedforms and
ergs will only develop where the sand-drift is saturated
and decelerating. Where this occurs sand cover should
be complete. However, since isolated bedforms can form
and persist on otherwise sand-free ground it is necessary
to introduce the concept of metasaturated flow. In
metasaturated zones, sand will only accumulate on already
sandy areas; this comes about because sand can be trans-
ported more easily across hard ground than a loose sand
surface (Bagnold 1941, p.72). Thus any fully developed
erg should be ringed by a 'metasaturation line' within
which is a concentric 'saturation line'. The erg
begins to develop at the metasaturation line by bedform
genesis. If the bedforms are efficient sand-trappers
(i.e. transverse) the erg spreads by the repeated
appearance of bedforms at the metasaturation line and
migration of those bedforms downwind. As this happens,
an area of saturated sand-drift may take shape and it
is only within this that complete sand cover may be
generated with subsequent erg body aggradation. With
sand-passing bedforms (i.e. longitudinal) the erg
spreads by growth at the downwind end of the bedforms:
there is no bedform migration.

Expressing this more rigorously, let V be a function representing the wind regime. If x,y,z are the 3 spatial variables and t is time,

$$V = f(x,y,z,t)$$

The carrying capacity of the wind, or potential or saturated sand-drift, $Q_S = f(V)$. The metasaturated sand-drift Qm is also a function of V.

Let $Qm = Q_S - \Delta_Q$ $\qquad \Delta_Q \geqslant 0$

Δ_Q will vary with V and the nature of the ground surface. According to Wilson (1971b) 100% sand cover can only be achieved where $\Delta_Q = 0$.

The only condition compelling deposition is that the sand-drift rate decreases downwind, ie. if $\frac{dQ}{dx} < 0$ deposition occurs, where x is fixed as the wind direction and Q is the actual sand-drift. The only other necessary condition for deposition is that $Q \geqslant Qm$. This is not a sufficient condition though: if $Q \geqslant Qm$ and $\frac{dQ}{dx}$ is not less than zero deposition is not inevitable (this differs from Wilson (1971b, p.194)). $Q \geqslant Qm$ merely denotes that bedforms are stable, not that they must occur. In conditions of steady flow over uniform ground no sand would be deposited. In real life the flow fluctuates and the ground is rough. Sand will be deposited from time to time and the fulfillment of the condition $Q \geqslant Qm$ means that on balance these accumulations will survive and grow into bedforms.

The foregoing is illustrated for 3 hypothetical
ergs in fig.1.5. In fig. 1.5a Q_s & Q_m, following V,
rise to a peak and tail off. Since in the left hand
part of the graph $Q < Q_s$, deflation occurs and Q
increases with x, $\frac{dQ}{dx}$ being controlled by surface
characteristics. Because deflation does not keep pace
with the competence of the wind in this instance, Q
reaches Qm on the descending curve at x_m. At this
point bedforms may develop and unless there is deflation
in the inter-bedform areas Q levels off. As sand cover
increases, Q_s must approach Q_m, the two becoming equal
when complete cover is attained. This is because the
wind is capable of carrying more sand over hard, bare
ground than over a loose sand surface (Bagnold, 1941,
p.72). Wilson (1971b) does not make this point. This
reduction of Q_s has the effect of sharpening the trans-
ition from incomplete to complete sand cover, which is
reached at x_s in fig. 1.5a. Beyond x_s, $\frac{dQ}{dx}$ is negative
and the bedforms must climb, thus accumulating an erg
body.

Fig. 1.5b illustrates the development of an erg with
incomplete sand cover. V and thus Q_s and Q_m decrease
initially, level off for a distance and then increase.
Such a situation might develop where the wind flows off
a mountain range with deceleration due to flow expansion,
crosses a flat plain, then rises and accelerates up the
next mountains. Where $x < x_{m1}$, Q is less than Q_m and
therefore deflation occurs to increase Q; across the
fringing alluvial fans for instance. At x_{m1} the rising

Q meets and exceeds the falling Q_m. Bedforms may
therefore develop and Q_s will be reduced as the proportion
of sand-free ground decreases. Beyond x_{m1}, V remains
constant and, if no deflation occurs between the bedforms, Q
levels off and maintains a value between Q_s and Q_m.
Because dQ/dx is zero, no further deposition takes place
after the bedforms have reached an equilibrium size, and a
non-aggrading erg with incomplete sand cover is maintained.

An aggrading erg with incomplete sand cover is shown in
fig. 1.5c. Here $dQ/dx < 0$, compelling deposition, but
$Q_s > Q > Q_m$ over the erg. Thus, since the actual
sand-drift does not equal the saturated sand-drift, bedform
cover is incomplete. Both bedforms and interbedform areas
aggrade leaving an alternating sequence of bedform and
interbedform deposits. If the bedforms and interbedforms
have different angles of climb, that with the lowest angle
is obliterated by the other. The relative thickness of
their respective deposits, t_D and t_I, is in proportion
to the length of the depositing areas, l_D, l_I. That is,
$$\frac{t_D}{l_D} = \frac{t_I}{l_I} = \text{the sine of the angle of climb.}$$

SECTION 1.4   ERGS AND ASSOCIATED FACIES

The various facies and sedimentary environments that
may be associated with ergs are not difficult to divine and
have been well recounted recently by Collinson (1978).

A first simple classification of facies patterns can be
made on the basis of two tectonic and two geographic
settings:

   (1) tectonically stable inland ergs,

   (2) tectonically stable coastal ergs,

   (3) tectonically active inland ergs,

   (4) tectonically active coastal ergs.

A selection of examples can be given: for (1), many
Saharan, Arabian and Australian ergs; (2), the western Great
Sandy Desert, Australia and the Namib Desert; (3), the Takla
Makan, China and many other Chinese and Iranian ergs and
Great Sand Dunes, Colorado; and (4), the Atacama Desert,
South America and Gran Desierto, Mexico.

The four divisions are illustrated in figs. 1.6 and

1.7. From the diagrams it is apparent that aeolian sands may be associated with almost any kind of sediment. If fairly coarse labelling can be tolerated, their nearest neighbours, both laterally and vertically, are likely to be some part of a fluvial system or anything from an inland sabkha to a saline perennial lake in the inland cases. Ergs are likely to occupy the centre of an endoreic basin or be situated between marginal areas of fluvial deposits and a central area of lacustrine, playa or sabkha sediments. Stable tectonics will favour the development of sediment bodies of large lateral extent. Relief will be eroded without rejuvenation by relative uplift, and sediment supply is likely to dwindle eventually, therefore limiting the vertical thickness of ergs.

With active vertical tectonics this constraint will be removed though the lateral extent of ergs may be limited by the reduced distance between highlands compared with a stable area. Note however that the Takla Makan, an erg in a tectonically active area (N.W. China), has an area exceeding 250,000 km^2.

In coastal situations aeolian sand may interfinger with any nearshore environment. The direction of the wind relative to the coastline may have important consequences. An onshore wind is only likely to lead to the development of a significant erg in the case of a clastic shoreline amply supplied with sediment by rivers, whether directly or indirectly. A carbonate shoreline might prove an unproductive or short-lived

source of sediment though the high probability of early cementation in carbonate dunes might enhance their subsequent preservation. In the Namib and Atacama Deserts, winds blow along or off-shore.

The influence of tectonics on coastal ergs will be similar to that in the inland case, though an actively subsiding coastal erg is going to be very vulnerable to transgression if sand supply is restricted for some reason.

Summary

There are many different precise definitions of the term desert, though the subtleties have little relevance to the geologist working on the ancient. Deserts have a limited number of meteorological causes but acting in combination these factors have produced deserts at latitudes ranging from 48°N to 50°S in the modern world.

Areas within deserts where aeolian sand has been deposited in and by active bedforms are called ergs, and vary in size from just a few square km to more than 500,000 km^2. Most modern ergs seem to consist merely of a veneer of bedforms.

The development of ergs is dependent on the coincidence of a suitable location, climate and sand supply. The vertical aggradation of ergs is believed to be governed more by sand supply than subsidence. A geomorphological model for the initiation and growth of ergs may be developed by consideration of the airflow and potential

and actual sand-drift rates. The model allows for the development of ergs with and without complete sand cover.

CHAPTER 2

THE WIND, BEDFORMS AND CROSS-BEDDING

Introduction

Some mention of the wind as a transporting medium
must be made in any account of aeolian sediments. It
is the aim of this chapter to give a brief geologist's-
eye view of the wind, and to follow this with a description
of the bedforms found in aeolian sand, with the intention
of then being able to present a generalised picture of
the cross-bedding of wind-laid sands.

SECTION 2.1 THE WIND

The overall geostrophic pattern of the wind is
unalterably determined by differential solar heating
at equator and poles and the Coriolis force. Within
those constraints the wind is free to proceed wherever
caprice takes it and is susceptible to the influence of
countless minor modifiers and considerations. The
theory of atmospheric motion is enormous, immensely
complex and incomplete.

Whereas weather systems involve the motion of the
whole troposphere, the effects of friction at the
earth's surface create the planetary boundary layer
1 or 2 km deep within which the wind diverges from the
geostrophic flow by between 10 and 30° anticlockwise.
According to Dutton (1976, pp.447-8) the theory of the
planetary boundary layer "has probably received more
attention and filled more books than all the rest of
dynamic meteorology." The theory, however, is still
"quasi-empirical". The structure of this layer over

a granular bed seems to consist of (from the bottom upwards):

1. A layer of still air 1/30 of a grain diameter thick. In sand-driving conditions this zone thickens (see Bagnold, 1941, pp.57-64).

2. The Ekman Layer, in which the wind profile is approximately logarithmic and the wind spirals with height to the geostrophic flow.

   Dutton (1976) distinguishes a "constant stress layer" in the basal 100m in which the eddy stress remains approximately constant. None of the geological literature on aeolian bedforms contains any specific reference to work of atmospheric physicists on whether or how this pattern might be altered in the presence of an extensive field of large bedforms.

   Contrasted with water as a transporting medium, the wind is far less circumscribed in terms of directional variability but far more so in terms of carrying power. The physics of sand movement in air are comprehensively dealt with by Bagnold (1941) and reiterated in Cooke & Warren (1973, p.258-266). Sand movement begins at a shear velocity of 0.15 ms^{-1} (equivalent to 4 ms^{-1} or 9.5 mph measured at a height of 2m) in grains of diameter 0.07mm. For all sand moving velocities the wind is turbulent. Once sand movement has been initiated the amount of sand carried by the wind is proportional to the cube of the difference between the actual and threshold velocities. To give an idea of the orders of

magnitude involved, a wind of 10 ms^{-1} or 27 mph
(measured at 1m) will move 120kg of 0.25mm sand across
a one metre width in one hour.

SECTION 2.2 AEOLIAN BEDFORMS AND THEIR ORIGINS

General

Wilson (1970) recognized and systematically described
4 distinct size orders of aeolian bedform: ripples, mega-
ripples, dunes, and draa. Megaripples are rare features
restricted to sands with a significant coarse or pebbly
fraction and were subsequently included with ripples
to give 3 size orders (Ellwood et al., 1975). This is
illustrated in fig. 2.1.

The obvious question is simply, why? Why does the
wind create bedforms in the first place and why are
there 3 distinct coexisting size orders of constant
dimensions in any given area of constant grain size,
whether sand cover be complete or partial?

Aeolian bedforms have been the subject of very
little wind tunnel experimentation. Given the
dimensions involved nothing other than ripples are
ever likely to be subject to such controlled study.
There is no evidence that a sequential development of
bedforms might prevail with increasing current
velocities as in unidirectional aqueous flows, but
such a system would be next to impossible to infer
from fieldwork.

Bagnold (1941) draws a distinction between "gentle
winds" in which pre-existing flat patches of sand
surrounded by a pebbly surface are degraded, and

"strong winds" in which such patches aggrade given a continuing sand supply (ibid., p. 172ff.). Bagnold suggests quite reasonably that this observation may be extended to dunes. Given a suitable succession of strong winds a dune could probably be created in this manner, provided the surrounding surface remains pebbly and the sand-drift above metasaturation. Given a large enough sand patch and a suitable wind for several thousand years (an average draa reconstitution time, Wilson, 1970, fig. 68) a draa might form.

Allen (1968, Ch.6) provides a review of ideas for the origins of bedforms formed by flowing water. All involve generating streamwise variations in sediment transport by some means or other. A suitable systematic disturbance in the flow may be postulated which pre-exists the bedforms and determines their wavelength. This pre-existing rhythm is merely labelled 'turbulence', or is procured at a discontinuity in a stratified flow. Alternatively, initially small, chance mounds or hollows are said to occur on the bed and set up a flow separation from which bedforms are generated and grow. Thus there is a chicken and egg problem, though the latter theory seems to be à la mode at present (Leeder, 1977).

Wilson, in his thesis (1970, pp. 107-120), gives the origin of aeolian bedforms a lengthy consideration along the lines described above, though acquiescing to the ballistic theory for ripples. In his 1972b paper he settles on having a pre-existing nucleus with which to generate or fix a rhythmicity in the flow (note

the ambiguity). To develop distinct orders of bedform,
dune nuclei must be bigger than fully grown ripples and
draa nuclei bigger than fully grown dunes. This seems
to be the state of the art for aeolian bedforms (with
one exception; see below, pp.24-29). It does not explain
why ripples do not grow up to become dunes and dunes
to draa: the theory may well generate a bedform but
does not seem to allow for 3 orders of bedform.

Folk (1976), and in discussion (1977a), attempts to
explain aeolian bedforms (ripples, megaripples and
dunes) with 3 orders of roller vortices in the airflow,
larger vortices corresponding to higher velocities.
Despite citing 2 of Wilson's papers he fails to mention
draa. To correct this all that needs to be done is to
slot megaripples in with ripples as low order features,
then label dunes as mid-order and draa as high-order.
The problem with this idea is envisaging successive
orders of bedform being formed at increasing wind
velocities, with "zones of chaos" developed in the
transition from one order to the next. Reconstitution
times could be played with, but the permanent coexistence
of all 3 orders of bedform irresistibly forces the pre-
conception that at the very least their stability
fields overlap considerably. Further, Folk's ideas
do not seem to be capable of explaining the existence
of draa lacking slipfaces. These have dunes migrating
down the leeside in what should be a zone of reversed
flow, according to Folk. Also, no-one has ever detected
rhythmic roller vortices in the wind.

Ripples

Typical aeolian ripples have wavelengths of the order of 0.1-0.2m and heights of a few millimetres. They are usually very straight crested and transverse to the wind; more sinuous varieties occur, but seem to be rare. They are asymmetric, with long windward and short leeward slopes, though lee slopes seldom attain the angle of repose. Sand is sorted on ripples such that coarser grains collect on the crests with finer grains in the troughs. They are ubiquitous on sand surfaces exposed to the wind (fig. 2.2), being suppressed only at very high wind velocities.

Bagnold's (1941, pp.62-64 and 146-153) ideas on the origin of wind ripples, generally referred to as the "ballistic theory", have held sway for 40 years. He suggested that in a population of sand grains undergoing saltation there would be a "characteristic path length" or modal length for a single jump in salt-action. Bagnold calculated this path length for a laboratory sand at given wind velocities and showed that the results corresponded very closely to ripple wavelengths produced in the wind tunnel. Thus on a rippled surface all the sand grains are deduced to be saltating in phase.

Folk (1976) attacked the ballistic theory and much of his criticism seems irrefutable, despite Leeder's (1977) reply. The thrust of Folk's argument is that though there may be a definable characteristic path length, it is the standard deviation of this length

that is important. He suggests that to produce coordinated ripples this deviation could not exceed about 25% of the path length. This is intuitively difficult and not borne out by Bagnold's photographs of migrating ripples in the wind tunnel. Also, aeolian ripples, though straight-crested, do show transverse junctures and other variations. These cannot be explained by the ballistic mechanism (see fig. 2.3).

Set against this, the work of Ellwood et al.(1975) would at first sight seem to confirm the validity of the ballistic mechanism, which is in fact the aim of the paper. The behaviour of uniform sand grains rebounding from a uniform sand surface was explored experimentally. To simulate saltation, sand was propelled onto the surface at an angle of 14° from the horizontal, and collected on a greased plate positioned at varying heights. Thus the percentage of grains rebounding to any given height was obtained and expressed as a "statistical 'rebound probability matrix'."

The results of these experiments are redrawn in fig. 2.4. This curve, of percentage attaining a given rebound plotted against that rebound, provided the "rebound probability matrix" (expressed as $R_p =$ $(R_5, R_{10}, R_{15} \ldots \ldots R_{100})$, where p% of the grains attain the rebound Rp). Modelling the saltation process on a computer, each run was initiated with an arbitrary wind shear which with Rp gave Vy, a matrix expressing the initial upward velocities after

the first bounce. The grain paths on rebound and
acceleration in the wind were computed to derive the
matrices Vp of the impact velocities and Lp, the jump
lengths. For the next jump, the initial velocity was
set at V corresponding to the mean path length L. The
iteration was continued until V converged to a
limiting value and the corresponding mean path length
was taken to be the wavelength of the ripples
produced. The paper claimed to achieve 2 ends;
firstly to show that a mean path length existed and
secondly to show that this path length could attain
the magnitudes ($\leqslant 50$ m) necessary for application to
granule megaripples. Megaripples were thus adjudged
to form a continuous bedform order with ballistic
ripples.

There are 2 flaws in this, both fatal but one
more fundamental than the other. Firstly, consider
the notion of adopting the mean velocity of the
previous jump as the initial condition of the next in
the iteration. The effect of this is to allow one
bounce to occur with all its concomitant dispersion of
the grains. Then in mid-air all the grains with their
different trajectories and velocities are collected
back together and given identical velocities and
trajectories ready for the next bounce. It is
therefore inevitable that V converges to a limiting
value. They are not modelling a sequence of limited
jumps but a series of totally independent events with
arbitrarily defined initial conditions.

The second flaw is that the curve of rebound versus percentage attaining that rebound is a cumulative curve (fig. 2.5). It declines monotonically with no points of inflexion: Ellwood et al. (1975) fit a logarithmic function to it. The frequency distribution of attained rebound is the derivative of this curve and since the cumulative curve has no point of inflexion the frequency distribution will have no peak. It will in fact be of the form $p = \frac{1}{R}$. Thus, though there is certainly a mean rebound there is no mode - no co-efficient of rebound attained by more grains than any other. If one imagines firing spherical particles at a bed of spherical particles one would intuitively expect this result anyway. The frequency curve of heights reached by the rebounding grains must therefore be of the form $n = 1/y - 1/c$ where n = the number of grains bouncing to height y, and c = the height attained by the most vigorous grain. There is then a mean height of bounce but no mode. Since wind velocity increases with height, the height to which a grain jumps determines the velocity to which it is accelerated, and consequently the horizontal path length. It follows that though there is a mean path length there is no mode; the mean is totally meaningless.

It therefore follows that any small disturbance in the bed surface, any chance mound or hollow, will not serve as a nucleus from which a ballistic ripple fan may propagate. Any such disturbance will be damped - a plane bed is stable under the pure ballistic mechanism.

This argument can be extended to a hypothetical case in which a well defined characteristic path does exist. This is illustrated in fig. 2.5. Given a point source of saltation at O and a characteristic mean grain path l with a standard deviation, the number of grains n arriving at any point after a single saltation will be represented by a normal distribution about the point l , as shown. Allowing a second set of jumps the landing grains will be distributed as shown by the second wave, centred on the point $2l$. Each point on the ground beneath the first curve acts as a point source for the second curve. Thus the second curve will have a larger standard deviation than the first and any succeeding curves will be still broader and flatter. The initial concentration is damped out, the process acting alone is dispersive and will be so for any finite value of the standard deviation of path length.

Bagnold (1941, pp.146-8) suggested from his experimental work that in saltation, the angle of approach for landing grains was remarkably uniform, from $10^{o}-16^{o}$ measured from the horizontal. Given a uniform rain of particles landing at this angle it is evident that a small area of bed tilted at just a slight angle to leeward will be substantially sheltered from the bombardment of saltation. A corresponding incline to windward receives an increased bombardment. Therefore grains in surface creep move rapidly up windward slopes and slowly down the lee, and a ripple formed of coarse

grains may build up. Therefore under a uniform
bombardment a plane bed with surface creep is unstable.
Bagnold postulates ripple train generation by this
mechanism in conjunction with the characteristic grain
path.

Sharp (1963), on the other hand, points out that each
leeward slope creates a zone shadowed from bombardment.
The length of this zone (determined by saltation approach-
angle and ripple height) will in itself fix the ripple
wavelength without resort to a characteristic grain
path.

The validity of this mechanism could be tested
either in a wind tunnel or by calculation of a set of
grain paths over a minor disturbance of the bed.
Bagnold's (1941, p.146) representation of the saltating
particles as a uniform rain descending from infinity
cannot be valid in the light of the deductions about
characteristic grain path made from Ellwood <u>et al</u>.
(1975). Nevertheless, a leeward slope must provide some
shelter from the saltation load, and a shelter prefer-
entially from the more vigorous grains. The least
energetic saltating grains will land in any shadow
zone as they hop over the crest of the ripple, but these
grains will be least capable of impelling the surface
creep. Therefore the surface creep will decelerate
and a small disturbance in the bed may grow.

What is not clear is how ripples propagate (c.f.
migrate) by this process. Deductive reasoning can only
go so far - sooner or later experiment becomes necessary.

Furthermore Folk's (1976) criticisms of ripples generated by characteristic path length would seem equally applicable to Sharp's (1963) ideas.

To summarize, nobody seems to know how wind ripples form. The characteristic path length idea can safely be rejected, leaving Sharp v. Folk: shadow zones and ballistics v. vortices. Sharp (1963) shows that vortices do not exist and Folk (1976) that ballistics cannot generate observed ripple patterns. Recent work by Seppala & Linde (1978) suggests that ripples are initiated in the manner suggested by Sharp. A composite compromise might be proposed whereby a Sharp initiated ripple generates some kind of damped wave motion in the atmosphere, so building other ripples downwind and extending the influence of the wave motion. Ripple wavelength could be determined by the most stable wavelength of the atmospheric wave. This idea has no basis other than that ripples exist - it just sounds reasonable. It is little removed from suggestions for the origin of many bedforms.

Megaripples

Turning now to larger bedforms, Ellwood et al. (1975) regarded megaripples as homologous with common sand ripples, as has already been stated. They reached this conclusion after studying many field examples and finding a continuous variation in all measured parameters from 0.02m wavelength sand ripples to 8m granule megaripples. Sharp (1963) also grouped the two together, both morphologically and genetically. Bagnold (1941)

saw megaripples (his "ridges") as almost stable bodies
of piled-up coarse material confined to areas of net
erosion, trapping part of the passing saltation and
unlimited in size. He found "ridges" of wavelength
20m and height 0.6m in Libya. He did not account for
the rhythmic spacing of these features. Wilson (1970)
regarded megaripples as initiated by regular secondary
perturbations in the airflow.

The only concensus seems to be on their coarse
grain size and large heights and wavelengths relative
to ordinary sand ripples. The evidence of Ellwood
et al. (1975) implies that ripples and megaripples are
homologous up to wavelengths of 2m (their data is
sparse for greater wavelengths). Acceptance of this
implies formation by the same mechanism.

Though this may be so, more than one type of
bedform appears to have crept in under the term
"megaripple". The ridges of Bagnold (1941, p.145 and
plate 5, lower picture) are very different in appearance
and occurrence to coarse ripples (ibid., plate 4, lower
picture and plate 6a; also plates 1D and 2 of Sharp,1963).
Bagnold links his ridges with deflation; this cannot
be said of any other of the examples given. Being
erosive forms, ridges are unlikely to be found in the
geological record, though features described by Piper
(1970) from the New Red Sandstone of the Isle of Arran
may be of this type.

The 'coarse ripple' type take the form of a
straight or sinuous crested, actively migrating,

asymmetric bedform (gentle windward slope, steeper lee), developed only in coarse sand or fine pebbles, with heights measurable in centimetres and wavelengths in decimetres or metres. Because of their coarse grain size they are found only near sources of such sediment and must be expected to be rare in ancient aeolian sandstones. The nature of their geological relics is not clear - none having been described. The most likely candidate for their fossil remains must be examples of small scale cross-lamination which are occasionally to be found in sand-sheet deposits (see Ch.3) in some formations.

Dunes and Draa

Dunes and draa constitute nature's largest, most prominent, widespread and complicated series of coexisting bedforms. They must also rank as the least analysed and understood. They form geology's largest and most spectacular sedimentary structures. Dunes range in wavelength from 3 m to 600 m (typically 50-200 m), and in height up to about 50 m. Draa range in wavelength from 300 m to 5 km, reaching 450 m in height (Wilson, 1970). You could bury the Empire State Building or most of the city of Durham in a single draa.

In the geological literature dunes and draa have been classified with great avidity. The most systematic and rigorous analysis lies in Ian Wilson's Ph.D. thesis (1970), "The External Morphology of Wind-

Laid Sand Bodies". A recent U.S. treatment (McKee, 1979a) makes no distinction between dunes and draa, despite use of the most extensive resources ever put into any study of aeolian bedforms, the most advanced available technology, and probably the largest ever data base. Many of the 'dunes' and all the 'compound dunes' featured on the hundreds of satellite and aerial photographs in the book (McKee, op.cit.) are draa.

The fundamentals of Wilson's work concerning bedforms may be divided into 2 separate conclusions, the first being the recognition and investigation of dunes and draa as distinct bedform orders. Secondly, the dissection of bedform patterns into their basic constituents or "bedform element continua". For both dunes and draa he suggested that 2 pure continua could be abstracted; of longitudinal and transverse elements (parallel and perpendicular to the resultant wind direction, respectively). By combination in varying proportions and emphasis governed by wind regime, any aeolian bedform may be produced.

Dunes and draa seem to differ only in scale, not morphology. Presumably it follows that the same can be said of their causative mechanisms. Wilson (1972b) discusses their origins without reaching any firm conclusion. Cooke & Warren (1973) favour some kind of rhythmic wave motion in the atmosphere. Helicoidal vortices have been suggested as the cause of longitudinal bedforms by Hanna (1969) and many subsequent authors. The alternative notion of an origin due to a bimodal

wind regime with the bedforms aligned along the resultant has also been postulated (e.g. Bagnold, 1941).

The only systematic work to date on correlating wind regime with dune form is that of Fryberger (1979a). He concludes that for any given sand drift potential calculated from wind records, barchanoid 'dunes' occur in the least variable wind regimes and star 'dunes' in the most variable, with linear 'dunes' intermediate (though with much overlap between dune types). Despite the enormous qualifications which might be put on this work (e.g. the reliability and validity of the wind records, the distance between the weather station and the bedforms, the state of equilibrium of the bedforms relative to the modern wind regime, effects of grain size on determining the effective wind regime and lack of distinction between dunes and draa) the results are by no means offensive to intuition and must be accepted as very important progress.

Acceptance of these results immediately casts a major doubt over the helicoidal vortex theory propounded by Hanna (1969), previously well established and widely accepted. Obviously no conclusions can be drawn until more substantial evidence becomes available. The confusion on this point serves to emphasise that this aspect of aeolian sedimentology-cum-atmospheric physics is still at the hypothesis stage. Fryberger's work represents the first movement to measurement, refinement and proof.

The deductions, conclusions and hypotheses that
in the author's opinion offer the best explanation of
the observed facts can be expressed in the following
list:

1.   There must be some rhythmic structure in the
atmosphere (whether fixed or initiated by bedforms),
which can be characterised by a particular wavelength.
This wavelength probably increases with velocity.
Dutton (1976, p.452) states: "the high resolution sounding
techniques that became available in the 1960s produced
the surprising result that the horizontal components
of the wind often have vertical oscillations both in
speed and direction, again with a regularity suggesting
that wave motion is present."

2.   This rhythmic structure has two stable wavelength.
modes probably of similar configuration at any given
wind velocity. These modes differ in size by an order
of magnitude or more and determine the wavelength of
the dunes and draa.

3.   Fryberger's (op. cit.) work suggests that in a
unimodal wind regime, simple transverse or barchanoid
forms are the most stable. As indicated by Wilson
(1970, 1972b) & Cooke & Warren (1973), secondary flow
patterns induced by the bedforms lead to the stability
of elements oblique to the flow, as is the case under
water. These are indeed important components of modern
ergs. Longitudinal roller vortices are a popular
feature of atmospheric flow but it appears that longi-
tudinal bedforms possibly require something more than

this to form, in the shape of a wide unimodal or
bimodal wind regime. If this is the case, Wilson's
abstraction of longitudinal and transverse end-members
as 'pure' bedform elements may not be valid, implying
as it does that both may form in a unidirectional
wind. Perhaps oblique forms should replace
longitudinal as the second end-member. The concepts
of longitudinal and transverse elements are neverthe-
less useful in describing bedforms.

4.   In any erg, coexisting dunes and draa do not
necessarily take on the same form. This is because of
the differing reconstitution and response times of the
different bedform orders. Draa represent a much
longer period time-average of the wind than dunes.   A
wind that appears bimodal to dunes may be unimodal to
draa (and vice versa).

5.   In a field of simple longitudinal bedforms no
lateral bedform migration takes place and all sand
transport is extra- or inter-bedform.   In a field of
simple transverse forms all sand transport is intra-
bedform and the mass movement of sand is accomplished
solely by bedform migration.   In intermediate bedform
types the processes mix.

6.   It follows that when net deposition is occurring
over an erg, the highest bedform order cannot be
longitudinal, whatever the wind regime.   Longitudinal
bedforms do not migrate laterally, therefore cannot
build up an erg body, and consequently have a low
preservation

potential. Ergs of longitudinal bedforms are sand-passing, not sand-trapping.

The tremendous variety of aeolian bedform shapes is well illustrated in Wilson (1972b), McKee (1979a), Short (1976) and Cooke & Warren (1973). According to Wilson (op. cit.), transverse and longitudinal elements of both dunes and draa may combine in three ways: with all elements in phase, with transverse elements out of phase, or with longitudinal elements out of phase. These were termed gridiron, fishscale and braided patterns, respectively (see fig. 2.6). This distinction forms part of Wilson's (1970, pp.169-171) own classification of aeolian bedforms. His system goes on to consider linguoid and lunate patterns, symmetry, reversibility, coexistence of two elements from the same continuum, and the amount of deposit cover.

The categorisation inherent in this classification is to a large extent artificial. This is not to say that the various labels are spurious or invalid, but a reflection of the fact that the natural system is neither uniform nor perfect. The picture is never clearly defined: there is always some variation; some randomness. Only over very small sampling areas do bedforms conform exactly to idealised patterns. The larger the area examined, the more confused, variable and inconsistent the picture becomes.

Breed & Grow (1979) listed 'dune' types (no distinction made between dunes and draa) in order of abundance as linear, crescentic (= barchan, barchanoid, transverse, sinuous transverse), star,

parabolic, and dome-shaped. Normal to their trend, the spacing of longitudinal 'dunes' ranges from 0.2 to 4.8km, the wavelength of crescentic 'dunes' from 0.06 to 5.5km, of star 'dunes' from 0.1-6.7km, and of dome-shaped 'dunes' from 0.8-5.4km. These measurements were made on a large number of satellite photos of localities from all over the world. All the data must include measurements from both dunes and draa. Wilson's (1970) data suggest that the dune-draa division occurs at a wavelength of ~500m, though with some overlap.

Brookfield (1977) summarises the limited available data on bedform migration rates. Dunes move at between 8 and 26.6m/year, though all these measurements are from barchans, which occur only on a hard floor, and not on preservable accumulations of blown sand. They may therefore not be geologically relevant. Barchanoid or transverse draa advance at 0.016-0.5m/year. Extending this, if $10m^3/m/year$ is taken to be a reasonable average net sand drift rate, a 20m high transverse dune should migrate at 0.5m/year, and a 200m high transverse draa at 0.05m/year, assuming no interbedform sand transport.

SECTION 2.3 CROSS-BEDDING

General

This section aims to predict and document the likely styles and features of cross-bedding produced by aeolian bedforms and preserved in the geological record.

The course of argument in the following is strongly influenced by the author's experience of numerous British aeolian sandstone formations and the published literature concerning such rocks in other parts of the world. This illustrates the nature of the problem: the answer is known (the cross-bedding of aeolian sandstones), and so are the initial conditions (modern aeolian bedforms). What has not yet been deduced and concisely presented is the set of processes and constraints whereby the ancient may be obtained from the modern by a uniformitarian approach. It is hoped that the following accomplishes an answer to this issue. The model is as much an explanation of the rocks as a hypothesis from first principles. As our knowledge of modern aeolian bedforms increases, and as more aeolian sandstones are systematically studied, so the ideas presented here will be extended, refined and corrected.

The Generating Bedforms

It was shown in Section 1.3 that aggrading ergs could have complete or incomplete sand cover. The aggradation of ergs and the preservation of cross-bedding is accomplished by bedform climb. The geometry of the cross-bedding reflects the geometry of the lower leeside of the generating bedform because only this part is preserved. To follow the mechanics of the generation of cross-bedding, it is therefore necessary to determine the configuration of the lower leeside of climbing bedforms in large aggrading ergs. This is where the bulk of preservable aeolian sand must be deposited. Small, thin ergs and erg-margin areas are less likely to survive into the stratigraphic record.

Wilson (1970, 1972b) asserts that draa should develop on all ergs with a mean spread out sand thickness of greater than 3 m, given sufficient time. It is obvious therefore that any ancient erg preserved as an extensive formation of aeolian sandstone must once have borne draa. Since draa always carry

superimposed dunes, the cross-bedding of any aeolian sandstone can only be analysed by considering the effects of these 2 bedform orders in combination.

To produce a set of cross-bedding, an efficient sand-trapping bedform with a steep leeside must migrate in the direction to which the leeside faces. Transverse bedforms best fulfill these requirements and are therefore the bedform type most likely to generate cross-bedding. Bedforms which stray from this ideal shape will be less efficient at generating cross-bedding. Longitudinal forms are sand-passers, not sand-trappers; their shape makes them inefficient at depositing sand. They are therefore less effective at preserving cross-bedding and may be restricted to areas of low net deposition, as discussed previously.

Parabolic dunes require a patchy vegetation for their development, and modern examples seem to be restricted to marginal erg areas. They therefore have a relatively poor preservation potential, and are likely to be less relevant as one proceeds back into geological time.

Star-shaped bedforms are an important type in modern ergs, and include the enormous draa massifs of the Rub'al Khali and environs (e.g. Glennie, 1970). These are in many cases amalgamations of more than one transverse pattern which have developed into peaked networks. Such patterns become truncated from the bottom upwards as sand cover decreases (Wilson, 1970). These have previously been imagined as completely static forms, though this may be a misconception prompted by a very low, but finite net movement (Wilson, 1970). Any net aggradation of sand as cross-bedding beneath these bedforms must be matched by sedimentation over the interbedform areas. This situation would lead to shoestrings of cross-bedded sand enveloped in interbedform deposits. Such a circumstance has not yet been reported from the geological record.

Thus the transverse bedform should dominate aeolian cross-bedding. To examine its flexibility to variation in wind regime, use may be made of "some principles that can be used in constructing qualitative models of bedform shape for a given set of oscillating wind conditions," listed by Wilson (1970, p.150). It is possible to apply these to the case of networks in a complex wind regime with complete sand cover, as shown on fig. 2.8. Wilson (loc.cit.) also states that if two winds differ by less than 45°, they usually combine to form a single ridge. The most complex wind regime possible is therefore one with a mode of the wind at every 45° round the compass (extraordinarily improbable).

As shown on fig. 2.8c this can be accommodated by just two ridge trends. If the critical angle for sharing is reduced to 30°, the most complex possible wind regime, with 12 modes, can be accommodated by just 3 trends (fig. 2.8D). Sinuous transverse networks can incorporate 2 or 3 trends and may therefore be admissible under wind regimes probably more complex than anything nature is capable of realising.

Symmetry, peakedness, the development of slipfaces and emphasis of individual ridges is decided by the relative strengths of the wind's various modes. Net migration of the system parallels the resultant sand drift. Consequent on the stability of oblique forms this _need_ _not_ be normal to the dominant trend in the bedform.

The sinuous transverse form is a magnificent compromise, robust and adaptable to change with minimum effort, capable of moulding many different winds to its own form.

In unidirectional aqueous flows, linguoid forms occur in small-scale ripples at higher velocities or shallower depths than straight or sinuous forms. (Allen, 1968). No such velocity dependence has been demonstrated in air. Wilson (1970, p.145) describes draa as usually lunate; sinuous aeolian ripples, megaripples, and slipfaceless dunes as linguoid, and other dunes as either lunate or linguoid. He was not able to give any quantitative ideas as to why this should be so.

Fishscale patterns dominate over gridiron except in some ripple and dune patterns (ibid., p.144). This is because in sinuous bedforms the separation bubbles in the lee are open and the helicoidal vortices thus generated determine the location of the next longitudinal element downwind, producing a half wavelength displacement at each transverse ridge. In less sinuous bedforms any longitudinal elements will be in phase, protruding abruptly from the lee of the transverse ridge to keep the separation bubble closed.

The symmetry of bedforms in a direction parallel to the flow is determined by the symmetry of the flow. Thus bedforms become more symmetrical as the frequency of wind reversals increases, as illustrated by the reversing dunes of McKee (1979a).

Flow remains attached over draa-sized features at lee-stoss slope angles of up to 25° (Wilson, 1970, p.130) thus providing a limit to which dunes may form on their lee-side. Separation over dunes takes

place at much lower angles, slipfaceless forms consequently being rare in normal sand. Cooke & Warren (1973, p.282) regard draa without slipfaces as being more common than those with.

To recapitulate, under almost any wind regime the relevant bedforms are likely to be linguoid or lunate dunes, probably in a fishscale pattern, migrating down the leeside of a lunate draa, also in a fishscale pattern. In the more unidirectional wind regimes the bedforms might be less sinuous and occasionally the draa will be fully slipfaced so that dunes are absent.

The Cross-Bedding Pattern

Having distilled what is preservable and geologically relevant out of modern aeolian bedforms, a model of aeolian cross-bedding can be devised. The exposition will begin in two dimensions (two spatial dimensions; time is always included) with bedforms behaving entirely deterministically. Subsequently the third dimension will be added, also with deterministic behaviour. Matters will then become more complicated (and realistic) as the random element of bedform behaviour is introduced, first in two dimensions, then three. Much of the groundwork and inspiration for what follows lies in Allen (1968, 1973, 1980) and Brookfield (1977, 1979).

The simplest two-dimensional view of cross-bedding is obtained in sections parallel to the net bedform migration direction. In the aeolian case there are 2 separate bedform types to be considered: draa with a slipface and draa without a slipface; These are shown on fig. 2.9. Migrating slipfaced draa develop potentially enormous foresets strung between major first order bounding surfaces. In slipfaceless draa the preserved cross-bedding is deposited from dunes, sets being bounded by surfaces dipping leeward at an angle determined by the slope of the draa leeside and the relative angle of climb of the dunes. These are Brookfield's second order bounding surfaces, labelled third order (3°) here for reasons to be explained later. The third order surfaces are terminated by first order surfaces identical to those formed by slipfaced draa. First and third order surfaces may be termed migration surfaces.

This pattern of a hierarchy of cross-bedding and bounding surfaces is the basic framework of aeolian cross-bedding. All other features are supplements, minor modifications or quirks of fate.

Since at all reasonable angles of draa climb it is only the very lowest parts of the leeside that are preserved the presence or absence of dunes here determines the prospective relative importance of the two prototypes. Following from the preamble to this model the slipfaceless type is probably predominant. Also, because perhaps only 10% or less of a draa is likely to be preserved, only completely slipfaced

varieties will be distinguishable in the geologic record. Even slipfaced draa with small, dune-sized longitudinal elements at the base of the leeside may be preserved looking like slipfaceless types.

Assuming deterministic behaviour of the bedforms (i.e. consistent migration in one direction), it is possible to compute the variation in preserved set thickness for the two categories in question. This was attempted by Brookfield (1977) in a somewhat fragmented fashion.

If the average rate of deposition in a particular part of an erg is δ m/year and if a draa of wavelength λ_D m migrates at D m/year, then the thickness of the cross-bed set deposited by that draa, or the distance between preserved first order bounding surfaces in the area, $H' = \dfrac{\lambda_D}{D} \times \delta$. A graph of this is plotted in fig. 2.10. First order surfaces in both slipfaced and slipfaceless draa will obey this relationship.

The behaviour of dune set thickness, or the spacing of third order bounding surfaces is drawn in fig. 2.11. The set thickness, h' is given by $h' = \lambda_d \dfrac{D \sin \theta}{d + D}$ where λ_d is the dune wavelength, d the dune migration rate, and θ the angle of the draa lee slope.

Any attempts to plug numbers into these formulae run into the major problem of selecting realistic values. The only quantities involved that have been systematically and reliably studied are the bedform wavelengths. Next to nothing is known of overall rates of deposition (δ), bedform migration rates in complete sand cover (d, D) or draa lee slope angles.

Ignorance apart, the equations may be rephrased as

$$\frac{H'}{\lambda_D} = \frac{\dot{\delta}}{\dot{D}} \quad , \quad \text{and} \quad \frac{h'}{\lambda_d} = \frac{\dot{D} \sin \Theta}{\dot{d} + \dot{D}} \quad .$$

Given geometrically similar dunes and draa in any single array, bedform celerity should be inversely proportional to cross sectional area parallel to the wind, and hence wavelength. A representative value of $\frac{\dot{D}}{\dot{d}}$ would therefore be say 1/20. Wilson's (1972b) values of $\dot{D}$ range from 0.016 to 0.34 m/year. Alternatively, pick a typical sand drift rate from Wilson (1970) or Fryberger (1979a) of $10 \text{m}^3/\text{m}/\text{year}$ and an average draa height of 100m to give a migration of 0.1m/year. For $\dot{\delta}$, be generous and say that thick formations need an abundant sand supply of 0.001m/year. If $\sin \Theta$ is put at 0.2 ($\Theta \simeq 12°$) both $\frac{H'}{\lambda_D}$ and $\frac{h'}{\lambda_d}$ equal $\sim 1/100$, though the heights of armwaviness that this analysis has reached cannot be overemphasised. It would not be at all difficult to justify altering this result by an order of magnitude either way, the most dubious estimate being for $\dot{\delta}$. Note that both ratios turning out as 1/100 is mere engineered coincidence; individually they are governed by quite independent quantities. The calculations and results are, however, illustrative.

As an introduction into the third dimension, that normal to the flow and horizontal, it must be remarked that a single draa alignment may include both slipfaced and slipfaceless leesides. Being lunate patterns, slipfaced elements will dominate at

backward pointing or barchanoid lobes while forward
pointing or linguoid portions of the bedform may well
lack a slipface. Without considering preservation
potential it appears that both of the described
structure types could form in the same bedform array
at the same time.

In considering the third dimension we have to
take into account the sinuosity of the bedform and
the topography at the base of its leeside.

Aspects of this are illustrated on fig. 2.12 for
both linguoid and lunate fishscale networks. We
yet again tread on dangerous ground here - this is
another item on the long list of features of aeolian
bedforms that have never been studied, despite its
importance in determining the style of cross-bedding
developed.

It is necessary to know the state and disposition
of the ground at points such as a & A in the linguoid
bedforms on fig. 2.12 and at a' & A' in the lunate net.
If A & A' are concave, scoop, or basin-shaped erosion
hollows, cross-bedding is generated by the migration
and infill of these areas. Trough-shaped sets will
result. If the area to the lee of each ridge is flat,
tabular sets result with curved internal laminae.

To decide which of the two cases is most probable
in nature, recourse must be made to the diagrams of
secondary flow over fishscale patterns provided by
Wilson (1970, figs. 60 & 63) and redrawn in Cooke &
Warren (1973, p.298). The deductions they presented

were made from logic, topology, and analogy with aqueous forms rather than direct experiment. The patterns envisaged all have a basically helicoidal airflow across the bedforms, with flow attachment, and therefore maximum shear and sand drift divergence, to the lee of backward pointing or barchanoid elements such as A & A' in fig. 2.12. To the lee of linguoid or forward pointing lobes, the flow separates from the bed, shear is at a minimum and sand drift converges. The net effect, for both linguoid and lunate fishscale networks is erosion at A & A' in fig. 2.12 and deposition at a & a', producing relative hollows and ridges, respectively.

The bedform profiles deduced from this, depicted in fig. 2.12, show that below a certain angle of climb, in both lunate and lingoid networks, loci of erosion such as A, A' entirely remove the deposits of the preceding linguoid element to leave cross bedding as a series of nested troughs. Fig. 2.13 illustrates the expected 3-dimensional character of the cross-bedding. If linguoid elements are removed by erosion linguoid and lunate patterns are not distinguishable (note the difference between a linguoid element and a linguoid pattern). Any preserved linguoid elements are present as convex-down current segments in horizontal section and convex-up in sections cut vertically across the movement direction. A convex-up form parallel to the current might develop in lunate patterns if the preserved linguoid elements lacked a

slipface. Such occurrences are rare in the ancient
(in the author's experience), suggesting that it is
only the trough and lunate segments that get preserved.

The 'troughiness' of the cross-bedding is deter-
mined by the overall sinuosity of the formative
bedform. Straight-crested varieties obviously produce
a more tabular and planar structure. Sinuous oblique
trends lead to asymmetric troughs. Forms with
exaggerated longitudinal elements preserve more
elongate troughs.

Discussion so far has dealt with only one order
of bedform and is therefore appropriate for slipfaced
draa. Superimposing dunes on draa, the basic structure
remains the same, though the internal layering of the
sets must shallow in angle. This layering then becomes
third order bounding surfaces enclosing dune cross-
bedding, with the result displayed in fig. 2.14.

Following from the similarity of dune and draa
shapes, all the arguments that led to fig. 2.13, and
the qualifications and possible modifications imposed
on it, apply to fig. 2.14 on two scales. The result
is a two-level hierarchy of nested troughs; one,
representing draa, enclosing the other, representing
dunes.

Considering the physical size of the sets so
represented, their thickness should be governed by
the relations already discussed. Their lenticular
shape will, however, give a widely varying apparent
thickness in any current-parallel section. Trough

width is limited by the wavelength of the sinuosities
in the original bedform. Wilson (1970) concluded that
this value was in general very similar to the wave-
length of the bedforms. The upper limit for dunes
therefore lies at about 500m, for draa at 5000m. These
limits are largely redundant since both figures are
much larger than both the average exposure and any
trough the author has ever seen. Trough width will
obviously be reduced as bedform climb decreases and
a smaller proportion of each bedform is preserved.

At this stage we have a reasonably detailed
picture of cross-bedding that is consistent with what
the author knows of aeolian sands, modern and ancient.
However, the treatment so far has been of systematically
migrating bedforms and is therefore not realistic, as
Allen (1973) has indicated. Randomness must be intro-
duced.

An immediate conceptual problem arises in extending
Allen's results on small scale subaqueous ripples in
a flume to aeolian dunes and draa tens or hundreds of
metres high. To affect the preservable parts of a
large draa most of Allen's quirks of fate need to
persist for longer than you or I are likely to live,
and some for longer than human civilization has
existed. Hence the conceptual difficulties.

Nevertheless the rocks are there to be seen and
to testify, and as Brookfield (1977) has pointed out
aeolian sediments do contain bounding surfaces which
cannot be attributed to any systematic bedform behaviour.

McKee (1966) discovered similar features in trenched dunes, though applying h'/λ_d to his transverse dune for instance, indicates that an identical climbing bedform will preserve no recongisable bounding surfaces (assuming deterministic behaviour!) - they all terminate before reaching the base of the dune.

To produce a preservable bounding surface some abrupt change in configuration is required at the base of a bedform, whether simply by deposition or by erosion and subsequent deposition. The surfaces so produced are here labelled second order where a draa is being modified and fourth order where a dune is affected. This is an extension by the addition of one new order (the second) of the system proposed by Brookfield (1977). Second and fourth order surfaces should be similar in shape but not scale. Second order effects on slipfaceless draa will be taken up by third order surfaces. Except in the case of a slipfaceless draa producing a second order surface by developing a slipface and therefore tending to truncate previously developed third order surfaces (confirming the hierarchy), the two orders are probably mutually exclusive, or at least indistinguishable. Second and fourth order surfaces may be termed modification surfaces.

Randomness in the first or third order is manifest as the creation or extinction of bedforms. Allen's (op. cit) ripples had finite lifetimes and presumably this goes for all other bedforms too. Further, bedforms apparently do not continuously produce a set of cross-bedding throughout their lifetime. Mean angles of

climb of 5-10° were necessary to negate this tendency
(a 1000 m wavelength, 100 m high draa moving at 0.1
m/year, climbing at 5°, will deposit 90 m thick sets
at an overall deposition rate of 9 mm/year; very
unreasonable). In arrays of climbing ripples the
tangent of the angle of climb was found to vary over a
threefold range.

Second and fourth order random effects, including
those described by Allen (op. cit) are illustrated in
fig. 2.15. Note that different processes may produce
the same, or very similar, results.

As Allen pointed out, the changes that bring
about modification surfaces arise from 2 independent
factors; the essential non-uniformity of the fluid
flow, and the inherent randomness present in the
behaviour of the bedforms as they react to the fluid.
Perhaps the most likely scenario for the production of
erosion surfaces preservable in aeolian cross-bedding
is by the action of relatively minor perturbations in
the topography producing a much magnified deviation of
the secondary flow pattern, which then persists for a
period significant compared to the bedform
reconstitution time. This, and perhaps other
situations, could lead to such events as the raising
or lowering of the bedform height, the lateral
migration of the peak of the bedform, a fluctuation in
the direction of migration or a hiatus in the
migration. All these events could change the
configurations of the leeside of the bedform and hence
generate modification surfaces. The configuration
shown in fig. 2.15A and B could form in this manner.

Storms and other meteorological disturbances
might have substantial effects on dune morphology,
though would be unlikely to influence the preservable
parts of draa. Depending on the vigour, duration and
direction of the storm, almost any configuration of
fourth order bounding surface might result (see, e.g.
fig. 2.15C).

Long-term changes in flow conditions (whether in the primary or secondary flow) might cause a slipfaceless draa to develop a slipface. The third order surfaces of the dunes on the draa leeside would then be truncated by a second order surface (as shown in fig. 2.15D), confirming the hierarchy. As a final step in the derivation of the model, all the random features must be incorporated into fig. 2.14. The two scales of nested troughs must be replaced by a bimodal size and shape distribution. The troughs must be given a finite length and within them must be contained a selection of second and fourth order features. Then all the qualifications previously made concerning linguoid or lunate elements and patterns, oblique trends, bedform symmetry and ridge sinuosity must be reemphasised and reapplied. The result is fig. 2.16.

In aggrading ergs with incomplete sand cover, draa migration surfaces (1st order bounding surfaces) will expand to become horizons of interbedform deposits. These might include deposits of migrating solitary dunes, zibar, sand-sheet deposits, water-lain sands or even the record of inland sabkhas. The draa migration surfaces will then be planar, rather than curved. These features are not incorporated in fig. 2.16. Their recognition and characteristics , and an ancient example, are described by Kocurek (1981, 1982).

In areas around the margins of ergs the record of a more diverse suite of bedform may be preserved, for example, longitudinal and parabolic types. Such deposits are, however, likely to be relatively thin, and unimportant in the geological record.

SECTION 2.4   IMPLICATIONS FOR PALAEOCURRENTS

Rose diagrams derived from measurements of cross-bedding azimuths for a given formation or member measure the relative volumetric contribution to that body of rock made by particular orientations of laminae. The resulting pattern is assumed to reflect the wind or water currents which deposited the sediment of which the rock is formed. In fig. 2.17 some bedform patterns and their formative wind regimes are redrawn from examples given by Wilson (1970). Potential palaeocurrent diagrams predicted from the orientations of slipfaces are added: note the disparity. Most bothersome is the bimodal pattern produced by the lunate network. The oblique trend might produce an error of a mere 45° and the pattern as drawn is probably less stable than one oriented normal to the

resultant of the two winds. According to the reasoning
given in the previous section, a pattern very similar
to the lunate network of fig. 2.17a might indeed be
expected to form in response to a wind regime corresp-
onding to that indicated by the palaeocurrent diagram:
the bedforms lie in a two-trend network.

There is therefore a potential ambiguity apparent
in palaeocurrent diagrams, in that a bimodal result
may be due to either unimodal or bimodal winds. The
effect is not just conjectural as reference to
Chapter 6 will show.

Stretching the imagination further, the single
wind bedforms of fig. 2.17c might produce a palaeo-
current pattern with three modes.

The confusion hinges on the interpretation of
a cross bedding azimuth as a measurement of the
primary airflow. As emphasised by Wilson (1970, 1972b)
and Cooke & Warren (1973), bedform elements are often
oblique to this, reflecting secondary airflow and
confounding reconstructions of the wind regime from
cross-bedding orientation.

In terms of preservation of bedforms the
magnitude of the effect is determined by the amount of
each trough limb that is buried as cross-bedding. If
only the central, curved segment is preserved, a uni-
modal pattern should result which may or may not
represent the balance of the depositing winds.
Judging from the published results from ancient aeolian
formations this situation normally prevails.

Mention must be made of the record likely to be left in the geological record by diurnal, seasonal, and longer term climatic variations in wind direction.

All consistent diurnal and annual fluctuations will probably be reflected in the bedform pattern as ridge trends such that each change may be accommodated with the minimum of modification of the pattern. This is so because all draa and most dune reconstitution times are much greater than one year (Wilson, 1970, fig. 68), preventing wind variations of that time period from significantly affecting the bedform pattern by seasonal erosion or construction. Fluctuations over longer periods are likely to be so subtle as to be buried in the natural background randomness of bedform behaviour. The result is that the vagaries of weather (including storms) and season are preserved only as subtleties of the cross-bedding such as fig.2.15c, if at all. Features of the ilk of herringbone cross-bedding are most improbable.

The coincidence of more than one size or orientation of dune was noted in parts of some ergs by Wilson (1970). He attributed these instances to a contrast in grain size between the two dune sets, the coarser-grained variety responding to a set of stronger winds than the finer-grained dunes. This situation is therefore not an example of the multiple development of bedforms in cyclically varying aqueous flows described by Allen & Collinson (1974). Possible large, cyclic variations of wind velocity are probably restricted to either diurnal or seasonal periods. The distances moved by

all draa and most dunes in these times are less than
their wavelength, thus precluding, or at least severely
restricting, multiple bedform development by this
mechanism in the aeolian situation.

SECTION 2.5 THE ANALYSIS OF CROSS-BEDDING

Having dwelt at length on the derivation and
construction of aeolian cross-bedding, the purpose of
this section is to suggest the lines and limits of
analysis that may be applied to such features preserved
in the geologic record.

Deductions upon the shape of the bedforms are
relatively straightforward and obvious, provided
preservation potential and randomness are borne in
mind.

An indication of bedform sinuosity can easily be
gained from the configuration of the cross bedding in
a horizontal plane; presumably most geologists can
distinguish trough and tabular cross-bedding. The
nature of the lower leeside of the bedform is obviously
recorded in every cross-lamina. Abruptly based sets
result from less 3-dimensional forms.

The size of the bedforms, both in order and
absolutely, presents a more difficult problem. It has
been shown that on average $\dfrac{H'}{\lambda_b}$ and $\dfrac{h'}{\lambda_d}$ both come
out to 1/100 give or take an order of magnitude.
Disregarding the orders of magnitude, the "on average"
is the problem. Obviously it is not practical to
pick up a set of cross-bedding and measure its
maximum thickness. Also, in this country the size

of exposures is a problem, not to mention the physical
limits on the height of geologists (i.e. we can't reach
the top of good exposures to measure things anyway).
And, of course, random bedform behaviour. Thus it may
be very difficult to define an average set thickness
for a formation.

Despite these qualifications a formation with sets
consistently exceeding several metres in thickness is
evidently more likely to have been deposited by slipfaced
draa than dunes. As a further indication of scale,
a slipfaced   draa should be depositing troughs with
a width of hundreds or even thousands of metres;
corresponding features in dunes being in tens of
metres. Note that a draa in an area of complete sand
cover bearing dune-sized longitudinal elements at the
base of a large slipface may deposit sets of draa-scale
thickness in troughs of dune-scale widths. Photographs
of such bedforms are provided by Wilson (1972b, plate
IID) and McKee (1979a, fig. 175).

The size of a slipface might influence the thick-
ness of the sandflows developed on that slipface -
bigger faces developing bigger flows because there's
more sand available. If so, this tendency may be
recorded by the laminae in preservable parts of a
bedform. However a gamut of other factors are also
involved; grain size distribution, grain shape, sand
supply to the leeside, overall configuration of the
leeside, and even the amount of early morning dew.
Nevertheless it's a possibility and will be explored

in ancient examples later in this thesis (Chapters 8 and 10).

The most reliable method of distinguishing dune and draa cross-bedding in the ancient must be to use the bounding surface hierarchy, in the manner of Brookfield (1977).

In what ranks as an important landmark for the analysis of aeolian sands, despite the criticism of some aspects voiced in the present work, Brookfield (op.cit) observed a hierarchy of bounding surfaces in quarry exposures of an aeolian formation. These correspond to the first, third and fourth order surfaces of Section 2.3, and were attributed to the passage of particular bedforms across the area. Unfortunately a primary requirement of this technique is large exposures, preferably in sections parallel to the direction of bedform migration. Whilst such a requirement is of small consequence in the western U.S.A. (e.g. photos in Stokes,1968,and McKee, 1979b), it causes real problems in the U.K., as related in Chapters 8-10.

According to Wilson (1970, 1972b), the wavelength of a bedform depends in part on the grain size character-istics of the sand of which it is constituted. This offers an independent handle on bedform size. Unfort-unately the diagrams displaying this relationship are very much scatter-plots,allowing a considerable range in bedform size for any given grain size.

To conclude, it is relatively simple to deduce

overall bedform shape from preserved cross-bedding.
The main problem lies in selecting the first and third
order bounding surfaces on whose enclosed sediment
the observations must be made: analysing the packets
of sand between second or fourth order surfaces is
likely to produce erroneous results.

Deciding the scale of the bedforms; whether or not
there were dunes at the base of the draa leeside, is
a fairly subjective process in most cases. It is a
matter of deciding whether various aspects of the
apparent structure transgress vaguely defined limits
of size. As should be habitual to geologists, don't
make a decision until you've noted everything possible,
plus some.

Summary

There is no wholly acceptable theory for the
origin of wind ripples. It is concluded that the
mechanism of wind-ripple formation remains an open
question. Most megaripples are homologous as a coarse
grained counterpart with normal ripples, presumably
with a similar origin. Their grain size restricts
the occurrence of megaripples in both modern and
ancient deserts.

The origin of dunes and draa is also open to
question. They form two distinct orders of bedform,
differing by about an order of magnitude in size, but
adopting a broadly similar range of shapes. A uni-
directional wind produces basically transverse bed-
forms with secondary flow patterns rendering oblique

trends also stable. Recent work suggests that longitudinal forms, long accepted as due to spiral vortices in unidirectional flow, may in fact require wide unimodal or bimodal winds to develop. The situation must be regarded as undecided at present.

The cross-bedding of ancient aeolian sandstone formations must always have resulted from basically transverse patterns of superimposed dunes and draa. Straight-crested bedforms are probably confined to relatively unidirectional flows. It is not clear whether more sinuous forms directly correlate with increasingly variable winds: strongly sinuous bedforms develop underwater in constant direction flows.

On average, only very small fractions of bedforms are preserved in cross-bedding. The resulting pattern (from sinuous transverse bedforms) is basically of two scales of nested trough-shaped sets, one size corresponding to dunes and one to draa. External and internal bounding surfaces are complex but may be organised into a four-tiered hierarchy attributable to bedform migration or bedform modification, each acting on the two relevant size orders of bedform. With allowance for the random element of bedform behaviour the total resulting pattern of cross-bedding is very complex.

It is, however, susceptible to analysis when preserved in the ancient, granted favourable exposure and bearing numerous qualifications in mind.

The palaeocurrents recorded from ancient aeolian sandstones may be ambiguous with bimodal patterns

resulting from unimodal winds or vice-versa.   There's

not a lot that can be done about this.

CHAPTER 3

THE LAMINATION AND GRAIN CHARACTERISTICS OF

AEOLIAN SANDS

Introduction

This chapter comprises a review of the lamination, grain size characteristics, grain shape and surface texture of aeolian sands, with emphasis on features likely to be preserved and observed in the ancient. It is in this emphasis on the ancient that the innovative aspect of this chapter lies.

SECTION 3.1   LAMINATION

A characteristic and striking feature of aeolian sands is their lamination. This is permanently on display in the ancient, though less accessible in the modern. In general sedimentologists have been content to consider features encompassing many laminae (e.g. a set of cross-bedding), or to investigate at length the details of single sand grains or the size characteristics of a population. Lamination falls between these two stools, though obviously in modern unconsolidated sediments it is difficult to observe at will.

In the case of aeolian sands this position was rectified by Hunter (1977) in a very important paper. From studies of coastal dunes in the U.S.A. he distinguished three principal types of stratification resulting from three types of deposition.

Migrating and climbing wind ripples (ripples of 'normal' size, i.e. in sand ranging from fine to coarse

in size; not 'megaripples' in any sense) produce various
versions of "climbing translatent stratification" (a
term semantically faultless but here replaced by wind-
ripple lamination). This consists of very thin (from
1 grain thickness up to 3 or 4 mm, typically 1-2 mm),
laterally extensive (dm or m) concordant layers
alternating in grain size (e.g. between·fine and
medium sand) of fairly well packed sand. An example
from the ancient is shown in fig. 3.1. Hunter mentions
porosities of 38-41% measured on six samples (average
39%). Possibly the single most important result of
Hunter's work is his proof that ripples, an ubiquitous
surface feature of blown sand, can climb and produce
this distinctive sedimentary structure.

Following Sharp (1963), Hunter also reported that
preserved foreset layering within wind-ripple deposits
is extremely rare. Wind-ripple lamination seems to
correspond to Bagnold's (1941, p. 127) accretion
deposits. The degree of packing is manifest as a
firmness underfoot - unyielding even under the weight
of a motor vehicle (Bagnold, op.cit., p. 236).

The second type of lamination is produced by
grainfall deposition. This is defined by Hunter (op.
cit.) as the settling of grains blown over the brink
of a dune (or draa) through the relatively still air
of the leeside separation bubble. This is equivalent
to the 'sedimentation' of Bagnold (op.cit., p. 127).
In the coastal dunes Hunter studied, the majority of
cross-bedding dipping at angles of 20-28° was formed
of this grainfall lamination. The laminae formed are
vague and gradational with poor grain size segregation

compared with wind-ripple laminae. Where best
displayed laminae with thicknesses of a few milli-
metres or less are distinguishable. They are laterally
extensive, conforming to and draping pre-existing small
scale topography. Six measured porosities ranged from
38-42% (average 40%).

Sandflows form the third stratification type. It
should be noted that in this context the term sandflow
refers to a cohesionless flow of grains on an over-
steepened dune leeside, contrasting with the usage of
the same term by Wilson mentioned in Chapter 1. In
subsequent chapters, sandflow in Hunter's sense will also
be described by such terms as avalanche stratification
or avalanche laminae. Lowe (1976) used the term grain-
flow for the same feature.

Individual sandflows on Hunter's dunes range from
2-5 cm in thickness at their thickest part. They form
lenticular bodies enclosed between grainfall laminae
high on slipfaces or stretching the length of small
slipfaces. Downslope the lenses coalesce, becoming
more tabular with slightly wavy boundaries. When a
flow is set in motion others tend to be initiated on
either side. An example from the ancient is shown in
fig.3.2.

Other wind-formed laminations are described by
Hunter (1973 and 1980) again from coastal dunes and
Fryberger et al., (1979) from an inland dune field.
Hunter (1973) records a "small-scale pseudo-cross lam-
ination" formed by climbing adhesion ripples. It
consists of thinly and irregularly laminated tabular

cross-stratification abruptly based and slightly
convex upwards. Each 'set' marks a continuous
depositional event. In sand damp, but not sufficiently
so for the formation of adhesion ripples,"quasi-
planar adhesion stratification" develops (Hunter,
1980), characterised by faint, fine and flat-lying
lamination with some small scale crenulation.

Fryberger et al. (op. cit.) illustrate a number
of lamination types and sedimentary features from
duneless, low angle sand sheet deposits at Great Sand
Dunes National Monument, Colorado. Of their two basic
varieties of deposit "type 'a'" would appear to
correspond to Hunter's (1977) planebed laminae, or
more likely to wind-ripple laminae. "Type 'b'"
includes coarser material (up to very coarse sand)
forming isolated laminae, lenses or ripple trains within
more irregular stratification, though still millimetre
scale. The coarse ripples may show foreset layering.
Faunal and floral bioturbation is common in these sand
sheet deposits, which are peripheral to and upwind of
the main dune mass. Though not explicitly stated by
Fryberger et al. (op. cit.), the sand sheet topography,
lacking slipfaced dunes, must owe its development partly
to the coarse grain size of the sediment and partly to
grass cover.

As will be shown later in this thesis, all of
the above lamination types, with the exception of Hunter's
damp sand varieties, are present in the U.K. aeolian
sandstones studied.

Contorted, disturbed and broken laminae in modern aeolian sands have been discussed at length by McKee et al. (1971) and McKee and Bigarella (1972). These are surface features formed during the normal processes of sand deposition and movement on the bedform leeside. However, the examples described are from gypsum and coastal dunes. The slight cohesion of the sand required for most of the structures is readily acquired and retained in the gypsum dunes by slight cementation. In the coastal dunes, located in southern Brazil, annual rainfall amounts to 1.5 m per year, at least ten times that of most deserts, again promoting cohesion. Thus in both the described areas the dunes are unusually prone to surface deformation features. In the author's experience, such features are extremely rare in ancient aeolian sandstones and also, presumably (by reverse uniformitarianism), in inland quartz sand deserts.

Doe and Dott (1980) discuss and analyse contorted cross-bedding in aeolian sandstones. Zones of intense convolution of the cross-bedding many metres thick and extending laterally for kilometres are common in the Navajo Sandstone exposures of the Colorado Plateau. The completely contorted strata are attributed to liquefaction of the sand below the water table, aided and abetted by high initial porosities. These features, generated in the shallow subsurface, are probably more likely to occur in inland quartz sand deserts than the minor deformations of McKee and Bigarella. Certainly they have a much greater preservation potential and occur sporadically in various British formations.

For comment and geological perspective on lamination types, the physics of sandflows has been dealt with by Bagnold (1954), Lowe (1976), and inevitably, Allen (1970, 1972). The most comprehensive analysis lies in Allen's 1970 paper.

The unusual features of the flows preserved in the ancient are their thickness (up to 0.12 m), angle of dip, ranging from 34° down to 20°, and wide lateral extent, being usually traceable for several metres. If these flows behave as described by Bagnold (1954) their surface velocities during emplacement would have been as much as 30 ms^{-1} (70 mph) - far in excess of the terminal velocity of the individual grains. Because no-one has studied processes on the lee side of large desert dunes and draa (c.f. Hunter's small coastal dunes), any constraints sandflow thickness and extent put on slipface size and sand transport rate are impossible to assess. The broad, thick flows observed in the ancient are certainly very different to Hunter's (1977) narrow tongues of sand. The preserved ancient flows must have involved large quantities of sand (many tonnes) and large areas of slipface (tens, possibly hundreds of square metres). The gentle slopes (even allowing for compaction and error in dip readings) reached by individual flows may reflect the momentum such large events were able to build up. Hunter describes flows propagating laterally by the initiation of other movements. This ramification, with mixing of

adjacent flows must be responsible for the laterally homogeneous and extensive laminae we see now.

Allen (1970) suggests that lee slopes may advance by the action of frequent small flows coming to rest on the upper reaches of the slipface and building up a considerable wedge of material to be swept away in infrequent large events. Thick beds might also come about by virtue of a large difference between the yield and residual angles of a particular sand (values up to $10-15^{\circ}$ were obtained by Allen (op. cit.) in natural sands), or by involving more of the slipface, or by a flow telescoping into itself from a long fetch. Possibly there may then be some broad correlation between slipface size and the thickness of the largest possible avalanche.

The occurrence of grainfall lamination also requires qualification before application to the geologic record. The problem lies in evaluating the effectiveness of the process and preservation potential of its deposits on large bedforms. Again analysis suffers in having only Hunter's work to refer to. Grain paths on the leesides of bedforms have been explored in detail by Allen (1968), but only for small scale subaqueous forms and the results can bear no relation to the problem in hand.

It is not possible to calculate grain paths over the leeside of a bedform - even assuming a situation where grains are projected horizontally from the brink into calm air, the resulting equations of motion

cannot be solved analytically. However there is no
reason to suppose that the energy of a grain's leap
into the unknown over the brink of the slipface should
be any greater than an average hop in saltation. There
seems to be enough validity in Bagnold's (1941, pp.
62-64) calculations of characteristic path length to
suggest that a grain in saltation will cover a distance
similar to the wavelength of a ripple (i.e. ~ 0.1m) in a
single hop. By this reasoning it would seem that
grainfall deposits should be very heavily concentrated
in the topmost parts of the slipface. The finest
grains however, might go into partial suspension and be
deposited some distance down the leeside. Thus it
appears that the pattern of deposition implied by
grainfall militates against its widespread preservation
in the cross-bedding of large bedforms.

The sand-sheet types of stratification described
by Fryberger et al. (1979) rely on a wide spread of
grain sizes for their development. They must therefore
have a tendency to be developed in the vicinity of
sources of sediment other than blown dune (s.l.) sand
or in areas where large quantities of sand have been
removed by deflation, leaving a coarser lag. An example
from the ancient is illustrated in fig.3.3.

Essentially the purpose of this section has been
to justify the lamination of ancient aeolian sandstones
as observed and interpreted in various U.K. formations.
These rocks contain, in order of abundance, wind-ripple,
sandflow, sand-sheet and grainfall laminae. The latter
two are rare, grainfall exceedingly so. No adhesion

ripple types have been detected though they are reported
to be quite common in North Sea gas wells (Glennie, 1972).
The onshore areas may have been too far from their
respective basin centres for the water table to have
approached the surface closely enough to form adhesion
ripples extensively. Surface deformations of McKee _et al._
(1971)   are next to absent _and_ contortions as per
Doe and Dott (1980) are found at only a handful of
localities.

SECTION 3.2 GRAIN SIZE

The standard description of ancient aeolian sand-
stones has always stressed their fine to medium sand
grain-size and good sorting. These characteristics
are often regarded as a major diagnostic feature of
such sediment. In British literature, especially in
the publications of the Geological Survey, the presence
of "millet-seed grains", meaning well-rounded coarse
sand, has also been used as diagnostic of aeolian sand.
This is a refinement of the oft stated view of all
aeolian sand as well rounded.

The purpose of this section is to examine these
statements in more detail.

The identification of sedimentary environments
by the analysis of grain-size characteristics has
received a great deal of attention over the past 20
years. Coastal and inland dunes have featured
prominently in the list of environments examined. The
success of these methods has been somewhat mixed and
opinion is divided on their effectiveness.

Typically, beach, coastal dune and river sands have been investigated, labelled in just those terms with no further specification. To the geologist confronted with an enigmatic outcrop, comparisons of aeolian sands with material from, say, shallow marine sand waves or sandy braided streams would be more relevant, those being environments capable of producing large scale cross-bedding in pebble and clay free sand.

It is unlikely that the interpretation of a whole formation of aeolian sandstone will ever hinge on grain-size analysis alone - there are many other features on which a decision may be more easily and soundly based. Features of the grain-size distribution weighty enough to confirm or deny an aeolian inter-pretation should be visible to the naked eye at outcrop, e.g. pebbles and clay layers. Further, there are practical problems in carrying the method of grain-size analysis from modern to ancient. Ancient sandstones have, by definition, suffered diagenetic modification from their original state. Apart from problems introduced by mechanical disaggregation, diagenesis is capable of substantially altering the grain-size characteristics by the formation of grain overgrowths, by leaching of feldspars and other unstable minerals, by pressure solution, and by the addition of authigenic clays. Such factors are minimised in many aeolian sandstones by the lack of cement typical of such rocks, but grain dissolution and authigenic clays are also characteristic.

These drawbacks can be reduced by careful experimental technique - gentle disaggregation, thin section assessment of diagenesis and examination of each sieved fraction under a powerful binocular microscope for example - but nevertheless remain a major barrier to the acceptance and development of grain size analysis as a diagnostic tool. A comprehensive, and,on the whole, balanced, review of the method is given by Friedman (1979).

In addition to the plotting of scatter diagrams of various statistical moments of grain-size distributions, the visual dissection of various graphic representations of these distributions is also a popular method of analysis, as exemplified by Visher (1969) and Folk (1968, 1971). Visher found that cumulative curves from many sedimentary environments, when plotted on log-probability paper, could be divided into several supposedly straight-line segments. Each segment was said to represent a separate log-normally distributed sub-population, and the sub-populations were interpreted as due to contributions from different transport processes: suspension, saltation or traction. Certainly a consistency of curve shape and characteristics within each environment was found by Visher. The basic sedimentology encapsulated in Visher's idea seems sound - it is difficult to fault the assumption that each transport mode in a sediment would show a log-normal distribution (or something of that type) if it could be separated off. However, the limits to each mode cannot

be clear-cut; only at extreme sizes will a grain move
by one means alone. For instance in an aeolian sand
the majority of grains will move mostly by saltation,
but also sometimes in surface creep and sometimes in
suspension. Coarser grains will move more often by
surface creep, finer grains will spend more time in
suspension. From this viewpoint it would be more
appropriate to join data points on a probability plot
with a smooth curve than with a series of straight
lines.

A more serious objection to Visher's method has
been put forward by Walton et al. (1980). This questions
assumptions about the mathematics of log-probability plots
implicit in Visher's work. Walton et al. (op. cit.)
examine the ways in which a cummulative curve showing 3
linear segments interpreted as suspension, saltation
and traction loads might be made up. To quote, "The
only perfect fit to a distribution of this kind is
obtained by percentile truncation of three sub-populations
of equal area". In this case the so-called traction
load is in fact the coarse tail of the finest sub-
population and vice versa. The thrust of the paper can
be best stated simply by quoting its conclusions:

"(1) Similar-looking cummulative frequency curves
      may be generated by a variety of truncation and
      mixing models.

 (2) The assumption that pseudo-linear segments
      composing some grain-size curves represent
      adjacent truncated distributions may be

incorrect. According to Clark (1976) no example in sedimentology of a truncated single log-normal distribution has been published.

(3) Estimation of graphical moments from pseudo-linear segments can often give badly erroneous results for sub-populations.

(4) Three adjacent truncated distributions of equal area may produce linear segments on the ϕ-normal probability plot but these segments are not consistent with the common interpretation of truncated sub-populations of suspension, saltation and traction."

Folk (1968, 1971) advocates the dissection of frequency curves as a means of analysing the sub-populations present in a sample of sand. Working on aeolian sands from longitudinal dunes of the Simpson Desert in Australia, he drew up frequency curves of grain-size distribution and split these curves into contributions from numerous smaller log-normal sub-populations combined in varying proportions. These sub-populations were placed and fitted by eye. The composite nature of his grain size plots led Folk to propose a quantum model for aeolian deposits whereby each quantum of well-sorted sand is carried by a separate gust of wind and deposited in a particular "micro-locality". The capriciousness of the wind leads to the deposition at any one spot of quanta from many different directions and sources. Each packet may have a different mean grain size, depending on wind

velocity, but all have similar sorting coefficients, reflecting the selectivity of a common transport process.

It is specious to attempt to criticise Folk on the grounds that many of his sub-populations may be pure flights of fancy. They are obviously present in the more polymodal curves and it is only logical to extend the dissection to apparently unimodal types. One might, however, suspect that the allowance of a little more latitude in the standard deviation of the sub-population in certain cases would produce solutions containing fewer quanta.

Bagnold (1937, 1941) and Bagnold and Barndorff-Nielsen (1980) advocate plotting the logarithm of grain-size against the logarithm of the frequency; essentially the logarithm of an ordinary frequency curve. This improves the resolution of the extremes of the distribution where quantities are small and cannot be adequately represented on an arithmetic frequency curve or histogram. It is argued that equal significance should be attached to all grades and that this is facilitated by the log-log curve. In addition to the dialectic aspect, Bagnold and Barndorff-Nielson (op. cit.) re-state and extend Bagnold's earlier assertions that the log-log curve reveals a fundamental property of natural grain-size distribution. This is that they assume a hyperbolic shape on log-log paper, whereas a Gaussian distribution plotted the same way forms a parabola. The hyperbolic form allows for differing efficiencies of sorting in the fine and coarse

limbs of the curve manifest as different gradients. This would be seen as skewness in a Gaussian distribution.

Bagnold and Barndorff-Nielsen (op. cit.) give various practical and mathematical illustrations of how the hyperbolic shape might be developed. They point out that in fluids transporting sediment, coarse grains get left behind and fine grains are carried on ahead, two processes which are likely to be largely independent and therefore allowing independent variation of the limbs of the grain-size distribution.

It will be some time before sufficient work is done to fully assess the applicability of log-log distribution. Their use is certainly logical, the hyperbolic form of natural curves another matter. As a start, all the grain-size analyses presented in this thesis include log-log curves. One practical problem that arises, particularly in slightly cemented ancient sands, is that the extreme tails of the distribution are most open to error. Aggregates of grains are present at the coarse end and spurious matter of authigenic or disaggregation origin in the fine fractions. One point that is fully endorsed here is Bagnold and Barndorff-Nielson's dislike of the cumulative curve as a means of presenting data. It is a method suited only to the concealment or glossing-over of significant tendencies.

A significant fact not mentioned by Bagnold (op.

cit.) or Bagnold and Barndorff-Nielson (op. cit.) is
that a hyperbolic log-frequency curve leads to a
slightly sigmoid probability curve. The hyperbolic form,
with straight limbs, has a slight excess of the coarsest
and finest grades over a parabolic curve, with downward
curving limbs. It follows that a cumulative curve on
probability paper has a steep central portion and more
gently sloping tails, the extent of each segment being
governed by sorting and skewness. Such a probability
plot might then be construed as being composed of 3
linear segments, in the fashion of Visher (1969).

A limitation inherent in all previous work on the
grain-size characteristics of aeolian sands is that none
has ever been related to the lamination of the sediment.
The lamination reflects the process of deposition, a
factor absolutely fundamental to the shaping of a grain-
size distribution. It is to be expected that sands
deposited by grainfall, avalanches, climbing ripples or
sand sheets will occupy different portions of the
spectrum. Their parameters may overlap but these
different processes must influence the grain size
characteristics of the final deposit. The limits within
which this shaping may be enacted are determined by the
place of deposition and immediate upwind source : the
path the sand has taken over the bedform (and the type
of bedform). Overall control is exerted by the
competence of the wind regime and ultimate source of
of the sand. An appropriate course at this juncture is
to examine these factors.

Provenance, the ultimate control, is the least definable element. Wilson (1971b) and Fryberger and Ahlbrandt (1979) have shown that wind-blown sand may travel enormous distances, hundreds of kilometres, before finally coming to rest. Thus there is no problem in depositing quartz sand on a limestone or shale terrain for example. (Collinson, p. 81 in Reading, 1978), however, makes the interesting point that in the Sahara there is some association between widespread sandstone outcrops and ergs. In the same vein one might note that pre-Permian subcrop maps of the U.K. (e.g. Ziegler, 1981) show vast expanses of Devonian and Carboniferous rocks, and both of these systems contain a good deal of sand-stone. On balance provenance probably is not important in determining grain-size. Given the vast areas from which sand may be derived the wind has a fairly free choice of what it moves and deposits, unless of course the erg in question lies a short distance downwind of a sand-drift divide (sandflow divide of Wilson). Whether the source be solid or drift it is clear that the grain-size distribution assumes an aeolian character in a very short distance, records in the literature varying from 0.34 km to 20 km (Cooke and Warren, 1973, pp. 311-312).

There is little in the literature specifically devoted to the competence of the wind. Bagnold (1941, pp. 38-95) deals precisely with the physics and mechanics of the relations of threshold velocities to grain-size and wind velocity to sand-carrying power. More practically Fryberger (1979a), in a unique and valuable study, documents the power of wind regimes from many

parts of the globe. There is nothing, however, which
covers the question of the spread, proportion and
velocities of grain-sizes that a given wind transports.
Whatever the ins and outs of the means, the end is
nevertheless abundantly clear. Countless records (e.g.
506 samples analysed by Ahlbrandt (1979), review of
topic in Folk (1971)) of aeolian sand and sandstone
from around the world and throughout geologic time
demonstrate that dunes and draa are made only of sand,
with minimal overlap into fine pebbles and coarse silt.
The limits are from -2 to 5ϕ, the bulk from 1 to 3ϕ;
medium and fine sand. Coarser grains are left behind,
finer race on ahead in suspension to be deposited as loess
or oceanic muds and silts. Reading from Ahlbrandt (op.
cit.) and Folk (op. cit.) aeolian 'dune' sand also tends
to be well or moderately well sorted and slightly fine
(positively) skewed.

A plethora of sorting processes on and between
bedforms has been proposed and are reviewed in Cooke and
Warren (1973, pp. 306-312). Reducing these to those
relevant to the lower lee side of a transverse bedform,
we are left only with the observation that during
avalanching the coarser grains travel furthest and are
thus concentrated in the lower part of the face. The
sorting at any spot within the avalanche will therefore
improve and we have a lamination-dependent distribution-
shaping process. Avalanche laminae on the lower lee side
of a bedform (dune or draa) will tend to be coarser and
better sorted than the average.

Ergs of transverse forms will tend to be finer and better sorted than ergs of longitudinal bedforms. This is because transverse forms are sand-trappers; longitudinal, sand-passers. In an erg of transverse bedforms undergoing net deposition, coarser grains will accumulate at the base of bedforms, partly because they are more difficult to move uphill onto the next bedform, and partly due to the avalanche sorting noted above. Hence they will be buried quickly and efficiently at the upwind margin of the erg, to leave the remainder of the field finer and better sorted. By contrast, such trapping will not occur among longitudinal bedforms and coarse grains are free to permeate the whole length of the erg, limited only by their lesser celerity. Even if the longitudinal forms are draa surmounted by transverse dunes, the corridors between the draa are still free for the passage of coarse grains, and the process of selective trapping and burial will be only partially effective.

Wind ripples produce laminae at most a few millimetres thick, and it is therefore not practical to sample them individually, except in thin section. Since these laminae alternate in grain size (if they didn't they'd be invisible), or are inversely graded, it is to be expected that the grain-size distribution of such sand will show some bimodality. If the two modes are very close together a sensibly unimodal curve will result. This will obtain whenever the separation of the modes is less than twice their individual standard deviation (Folk, 1971, p. 39).

As the two modes become more separated in size,the composite curve will vary from unimodal through platykurtic to obviously bimodal. Sharp (1966) suggested that during ripple migration saltating fine sand may penetrate and fall between the grains of a coarse surface creep and thus be sheltered from further bombardment and come to rest. This process obviously depends on having a closely packed carpet of surface creep, which is not to be expected in most ergs. However, where such a mechanism has been active, a quite strikingly bimodal grain-size distribution should be encountered: two perfectly sorted modes will differ by 2.7ϕ if the coarse is rhombohedrally packed, by 1.27ϕ if it is cubically packed. Note, however, that interpenetration acting in isolation cannot produce a lamination.

Elsewhere, the grain-size distribution is determined by the selectivity of the process of deposition in migrating ripples. It may be the case that wind ripples preferentially deposit the extreme sizes of grain in transport and pass on those most susceptible to the influence of the wind. They may develop bimodality in some other fashion. Alternatively their deposits may faithfully reflect the grain-size distribution of the entire wind load. Further speculation on this point would be casuistic in the dim light of our knowledge of the workings of wind ripples.

Sand-sheet deposits are a variation in degree rather than type of wind-ripple laminae. The difference lies only in the coarseness and proportions of the coarse mode. The well developed bimodality illustrated by Fryberger et al. (1979, Fig. 7, A-C) is to be expected. The total of 9 histograms presented (ibid.) comprise surface and sand trap examples, the latter representing direct sampling of the wind load. It is interesting to note that the fine mode in the bimodal surface samples corresponds quite closely to the mode of the sand trap samples. These are all unimodal, at 2 - 2.5ϕ, and markedly negatively skewed (no figures are given - visual assessment only). The coarse mode of the bimodal deposits lies between 0.5 and 1ϕ. Perhaps this is an example of Sharp's (1966) interpretation sorting, with the wind load losing as much sand as can filter into the surface creep as it traverses the sand-sheet area.

Grainfall deposits, if you can find them, should be individually very well sorted. The grainfall process operating in a constant velocity wind over a dead-calm lee side with uniform grain trajectories produces perfect sorting. This is reflected in the indistinctness of grainfall laminae reported by Hunter (1977). If Folk's (1971) quantum theory outlined above is construed as representing grainfall deposition (it is not clear exactly where Folk's ideas fit into Hunter's pattern), the polymodality, cryptic or apparent, of his curves provide the only available illustration of grainfall sorting. Folk's (op. cit.) samples, taken from the

crests of the dunes could indeed be grainfall, those from the flanks are more likely to be wind-ripple deposits. The only photo of lamination provided by Folk (op. cit. Pl. 2A, p. 24) appears to show wind-ripple laminae.

Grainfall is the principal process by which sand is carried to the bedform lee side. This may have implications for the sorting of avalanches, though no work has ever been done to explore the possibility. If avalanches originate in localised areas of the upper slipface they may inherit previous grainfall sorting, sorting which is potentially very good but which must be blurred by the varying trajectories of grains of any given size, and mutable through the inconstancy of the wind.

SECTION 3.3   GRAIN SHAPE AND SURFACE TEXTURE

The shape of aeolian sand grains is a topic to which a great deal of attention has been directed over many decades. The literature and conclusions of this attention are comprehensively reviewed by Folk (1978) so only a brief account is necessary here. Intertwined with the shape of the grains is the subject of their surface texture, which has also received much attention, especially since the advent of the scanning electron microscope. This too is reviewed by Folk (1978). The points at issue are the range of grain-sizes which are significantly rounded, the efficiency and rapidity of the rounding process, and the relative importance of mechanical abrasion and chemical solution. Apropos of

surface texture, geologists have contemplated and
argued over its nature, origin, the relative effect
of mechanical and chemical processes, the probability
of an original texture being preserved through
diagenesis, and whether or not the texture is diagnostic
of transport by the wind.

Coarse aeolian sand is always well or very well
rounded, provided it has travelled far enough. The
effectiveness of rounding decreases with grain size,
its gradual waning probably being responsible for the
lack of agreement on the lower limit. I know of no
systematic studies of rounding in relation to grain size
in aeolian sand - no-one seems to have gone to the
obvious length of plotting mean roundness in each
fraction vs. grain-size. Grain rounding is a mechanical
process; chemical effects are minor in comparison.

Solution and re-precipitation are, however,
important in the development of the surface character-
istics of grains. The famous frosting of desert sand
grains has been attributed to both mechanical abrasion
and chemical etching. The frosting is seen to consist
of a variety of textures when viewed under the scanning
electron microscope. That frosting is observed even
in re-entrants - areas sheltered from impact - indicates
that chemical action certainly does play a part.

Of the many textures illustrated and described
from aeolian sand grains (see e.g. Margolis and Krinsley
(1971), Krinsley and Doornkamp (1973), Krinsley (1978)),
upturned plates or cleavage plates are perhaps most

notable. These are parallel ridges 0.5 μm to 10 μm
in length, probably owing their orientation to a cleavage
direction within the structure of each grain. It is
claimed that the spacing of these plates is dependent
on the velocity of the wind under which they were
formed, and that past wind velocities might therefore
be estimated from examination of the grain surface
texture (Krinsley and Wellendorf, 1980). The surface
texture undergoes detectable chemical modification by
the action of desert dew promoting small-scale
solution and precipitation tending to smooth out
mechanical features. Folk (1978) reported a greasy,
rather than frosted, lustre to Simpson Desert sand
grains, due to the development of a thin "turtle-skin"
coat of silica precipitated from dew and rain, the
silicon being derived from abundant opal phytoliths.

Despite the evidence of very early chemical action
and the much greater potential destruction by diagenesis,
original grain surface textures have been found in
aeolian sands as old as the Permian (e.g. Waugh (1965);
Penrith sandstone and Yellow Sands, Krinsley and Smith
(1981); Yellow Sands,

Marzolf (1976); Navajo Sandstone).
Krinsley and Wellendorf (1980) have determined and
palaeowind velocity of 29 ms^{-1} (65 m.p.h.) from 5 grains
of the Permian Yellow Sands of N.E. England.

Summary

The lamination is an important and characteristic
feature of aeolian sands. Hunter's (1977) descriptions

of the types of lamination present and their distribution require considerable qualification in application to ancient aeolian sandstones. In these sediments wind-ripple laminae seem to dominate with sandflows or avalanches abundant. Grainfall and sand sheet types (Fryberger *et al.*, 1979) are rare. Deformation features in the ancient are principally of the sub-surface liquefaction type described by Doe and Dott (1980). Minor surface deformation features require special conditions and have a low preservation potential.

The presentation and analysis of grain-size distributions is controversial. In aeolian sands, grain-size characteristics are strongly influenced by the lamination since this is a result of the depositional process, and each type has its own tendencies in sorting the sediment. These factors have not been considered in previous published work.

CHAPTER 4

THE LOWER PERMIAN SETTING IN BRITAIN

Introduction

The bulk of the remainder of this thesis is an
examination in varying degrees of detail of several
formations of aeolian sandstone. Most of these are of
Early Permian age (the Yellow Sands of N.E. England,
the Bridgnorth Sandstone of the W. Midlands, the
Penrith Sandstone of N.W. England, and sandstones in
S.W. Scotland). This chapter therefore sets out the
present view of the tectonics, climate and geography
of the British area in the Early Permian, to place the
described aeolian formations in a wider context.

SECTION 4.1   STRATIGRAPHIC TERMINOLOGY

The terms used to describe the various divisions
of the Permian system are set out diagramatically on
fig. 4.1, in conjunction with the time scale given by
Smith (1964) and the apparent duration of the southern
hemisphere glaciation. Rotliegend, Saxonian, and
Autunian have a confusing usage, being basically
lithostratigraphic terms sounding deceptively like
biostratigraphic stage names. For instance Falke
(1972a, p IX) indicates that these terms may be
applied to Upper Permian rocks of continental type
representing lateral equivalents of marine Zechstein

deposits. This would appear to contravene all the known laws of stratigraphic practice. The most precise definition applicable to the term Rotliegend (literally: red-lying; red beds) seems to be N.W. European non-marine deposits, usually red beds and unfossiliferous, younger than proved Carboniferous and older than proved Triassic. Normal practice in Britain is to restrict usage to non-marine sediments beneath the Zechstein (literally: mine-rock) which is itself a lithostratigraphic term for N.W. European carbonates and evaporites of Upper Permian age beginning with the Kupferschiefer/Marl Slate. 'Rotliegend' then becomes part lithostratigraphic, part time-stratigraphic describing a particular group of facies of a particular age, and though lacking semantic elegance nevertheless has a certain value. The distinction of Upper and Lower Rotliegend is even more convoluted (see Falke, 1972b, 43-45 for a brief summary) and is essentially inapplicable and meaningless in the British area.

SECTION 4.2 THE PERMIAN CLIMATE IN BRITAIN

Linking Permo-Triassic climates, palaeowinds and continental drift is a perennial habit of geologists (e.g. Shotton, 1956; Nairn, 1961: Holmes, 1965). More up to date and elaborate climatic accounts are given in Robinson (1973; reiterated in Lamb, 1977) and Frakes (1979; includes an extremely useful account of the Carboniferous-Permian glaciation in the southern

hemisphere). Scotese et al. (1979) provide the most recent reconstruction of the continents for the Permian (reproduced as fig. 4.2).

Shotton's (1956) map of early Permian palaeowind results for the U.K. is re-drawn with additions from more recent work added in fig. 4.3.

That Permian aeolian cross-bedding in the U.K. always dips to present west indicating a N.E.'ly Permian trade wind exactly compatible with our Permian latitude is well known. That latitude is generally given as between 10° and 20° N. By comparison with the present day pattern this puts the U.K. on the southern edge of the desert belt. Coupled with the great stretch of land windward of the U.K. this would have minimised any chance of rainfall. The unimodal palaeowind pattern indicates that the U.K. lay to the north of the Inter-Tropical Convergence Zone (I.T.C.Z.) in both winter and summer. Robinson's (1973) suggestion that the I.T.C.Z. moved north of the U.K. in winter would lead to a winter S.W.'ly wind alternating with the summer N.E.'ly, a pattern not backed up by the rocks.

The Permo-Carboniferous glaciation in the southern hemisphere appears to have begun as early as the Namurian in Paraguay and Bolivia (probably related to a considerable mountain range in the area at that time). It reached its zenith in the interval comprising the Stephanian, Asselian and Sakmarian stages (fig. 4.1). Marine glacial sediments persisted

into the Kazanian in Australia (summarising from Frakes, 1979).

Thus the evaporite cycles in the Zechstein sequences may feasibly be attributed to glacio-eustatic sea level variations, though correlation of the major transgression at or about the beginning of the Late Permian with the final melting of the southern glaciers does not seem feasible. It may reflect an inter-glacial event or some other cause. If an inter-glacial event it is perhaps curious that no others of this magnitude have been reported and that a major transgression corresponding with the final melting of the glaciers during the Kazanian has not been noted.

To return to British Permian aeolian sediments the relevance of the above precis lies in the indication that whilst most of the formations discussed in this thesis were being deposited, glaciers existed at the south pole. Global climate may have been more like the present epoch than at any other stage in earth history.

SECTION 4.3 EROSION, DEPOSITION AND TECTONICS

In most areas of Britain there is a break between the Carboniferous and Permian sequences marked by an unconformity. This records the cessation of Carboniferous deposition, a change in climate (to more arid conditions), changes in palaeogeography and the lithification, uplift and partial erosion of the

Carboniferous sediments (c.3000 m. removed in Lancashire (Edwards and Trotter, 1954)). Permian deposition commenced in a desert climate on a land surface formed of rocks at least several tens of millions of years older. These first preserved Permian sediments were, in many areas, aeolian (e.g. N.E. England, parts of S.W. Scotland, N.W. England, and the Midlands) and whilst their rigorous climatic requirements imply deposition at similar times, a close correlation cannot be made; deposition probably commenced at different times within the Early Permian in different areas.

The geometry of the land surface which received these sediments is not well known. Thick coarse clastics in S.E. England (Laming, 1966), S.W. Scotland (Brookfield, 1980) and the Outer Isles of Scotland (Steel and Wilson, 1975) imply a rugged and possibly tectonically active terrane, but there is nothing to compare with, for example, the 10 km. thickness of coarse, post-orogenic molasse preserved as the Lower Old Red Sandstone of the Scottish Midland Valley. The Early Permian saw a certain amount of rather subdued diastrophism over most of Britain with a lengthy period of uplift accompanied by gentle folding of the newly deposited Carboniferous sediments. There are no strong reasons for believing that this uplift ever gained very far on its concomitant erosion.

The tectonic regime of the Early Permian was of a Mesozoic, rather than Palaeozoic theme. The

Variscan/Hercÿnian/Armorican orogeny seems to have influenced sedimentation only in the S.W. of the British area. Elsewhere, E-W tension was the rule, heralding the much later opening of the N. Atlantic. In most areas in and around Britain there is no evidence that this tension held any dramatic control over sedimentation, for instance in the basins under the North and Irish Seas. Only in parts of Western Scotland and N.W. England are there indications of contemporaneous faulting of any rapidity and magnitude.

CHAPTER   5

THE YELLOW SANDS - INTRODUCTION AND SEDIMENTARY

STRUCTURES

SECTION 5.1   THE LITHOLOGY AND DEFINITION OF THE

FORMATION

Durham and North Yorkshire contain the most complete
and best known Upper Permian sequence in Great Britain.
At the base of this sequence is a thin and discontinuous
formation of incoherent sands and breccias. Over most
of the outcrop, which stretches from Nottingham north-
wards, these never reach a thickness of much more than
a few metres. In Tyne and Wear and the northern part
of county Durham, the sands, while still discontinuous,
reach their onshore maximum thickness of nearly 60 m
and are well exposed in a number of quarries.

The formation is officially named the Yellow
Sands and rests on various levels of the Carboniferous,
from Westphalian C in Sunderland, to the lower part of
the Namurian south of Darlington. It is overlain by
the Zechstein carbonate sequence, generally represented
by the Marl Slate, occasionally by Lower Magnesian
Limestone. These rocks are dated as earliest Upper
Permian.

To the west and north the Yellow Sands are
limited by outcrop, seaward our knowledge is restricted
to a belt within some 10 k.m of the coast which has been
penetrated by a sparse network of N.C.B. boreholes.
Onshore a good deal is known of the subsurface

distribution of the Sands and their associated rudites
from borings and shafts sunk for coal. The two
lithologies are more cr less separated along a sinuous
line running from the coast at Blackhall to the vicinity
of Chilton (Smith and Francis, 1967), as indicated on
fig. 5.1.

Lithologically the Yellow Sands at the surface
typically are yellow, pale yellow or brownish yellow
(terms used in the strict Munsell Soil Colour Chart
sense), medium-grained, moderately sorted, incoherent
sand, well laminated and displaying large-scale trough
cross-bedding throughout. The rudites, visible only
at East Thickley, Middridge and Cleasby, are yellow,
badly sorted, unstratified, silty and highly dolomitic.
Varying from conglomerate to breccio-conglomerate they
contain clasts up to 70 mm in size of locally derived
Carboniferous sediments, together with quartz pebbles
and well—rounded quartz grit and sand. Subsurface the
sands and rudites are pale grey and pyritic, though
clasts in the rudites may be markedly reddened. Out-
crop and borehole information shows the Sands to be
organised into a series of ENE-WSW ridges with sand-
free corridors between.

Much of our present knowledge of the Yellow Sands
and Durham Permian in general, is owed to D.B. Smith
in a long sequence of publications (Smith and Francis,
1967; Smith, 1970, 1972, 1975, 1979, 1980; Smith and
Pattison, 1972; Krinsley and Smith, 1981; Smith et al.,
1974). Brief histories of investigation into the

formation are given by Smith and Francis and
Krinsley and Smith. The suggestion of an aeolian
origin for the Yellow Sands was first made by
Dalglish and Forster (1864). This interpretation
was subsequently accepted by all but Pryor (1971b)
who suggested a shallow marine origin. Pryor also
put forward similar views on the Weissliegendes sand-
stones of Germany, a deposit similar to the Yellow
Sands in age, type and position relative to the
Southern Permian Basin. Smith (1971) gave a reply to
this paper.

The present view of the palaeogeography of the
North Sea during the Lower Permian is shown in fig.
5.2b. The Yellows Sands evidently lay on the north-
western margin of a large desert basin, the Southern
Permian Basin. The marginal position is reflected in
the meagre thickness of the Yellow Sands compared
with equivalent sediments in the basin centre (figs.
5.2a, 5.3).

The previous interpretation of the Yellow Sands
(e.g. Smith, 1980) views each ridge as an individual
seif dune. At the top of each ridge a layer of planar-
bedded sand a metre or two thick is the result of
reworking by the Zechstein transgression, during which
up to two-thirds of the original dunes were removed.
To account for the preservation of uncemented sand
ridges and the apparent lack of diachroneity in the
Marl Slate, the transgression is said to have occurred

extremely rapidly, in the space of weeks or months, by the breaching and flooding of a basin which lay below mean sea level. Immediately prior to the transgression the Yellow Sands were brown or red, subsequent reduction by groundwaters beneath the Zechstein Sea altering them to the grey and pyritic state found subsurface at the present day.

This picture has a certain elegance and internal consistency and no major departures are in store in this thesis. The present work has served to refine, modernise and add more precise data and interpretations, in parallel with our increasing understanding of aeolian sands, bedforms and diagenesis.

Exposures

Exposures of the Yellow Sands occur at Cullercoats (364713); Tynemouth (375694); Frenchman's Bay (389662); North Hylton (358577); Claxheugh Rock (363575); McCall's Quarry (also known as Field House Quarry) Houghton-le-Spring (354506) and Hetton Downs Quarry, Hetton-le-Hole (358484) in the County of Tyne and Wear. In County Durham the formation may be seen at Elemore School (352442); Crime Rigg Quarry (340418); Sherburn Hill Sand Pit (344417); Bowburn Quarry (327380); Quarrington Hill Quarry (331378); Ferryhill railway cutting (303323); East Thickley Quarry (240257) and Middridge Quarry (249252); and in North Yorkshire by the Al(M) near Cleasby (246123). Most of these are shown on fig. 5.4.

Of the quarries, McCall's, Hetton Downs, Bowburn and Crime Rigg are active and access is permitted to all but Crime Rigg, which is run by the Sherburn Hill Sand Company. This company also owns Sherburn Hill Sand Pit and again restrict entry, though fortunately it is unfrequented at weekends and readily accessible from the road. Bowburn Quarry is in the process of closing, Sherburn Hill Sand Pit is being reactivated. Cumulative production from all 4 quarries probably amounts to 10,000-20,000 tons per week (Crime Rigg in May 1979 was producing 5-6,000 tons per week), price £1.50-£2.00 per ton at the quarry. It is the principal local resource of building sand.

SECTION 5.2   FORM AND INTERNAL STRUCTURES

Distribution and Thickness

The outcrop of the Yellow Sands as indicated by Institute of Geological Sciences maps, together with all available borehole information is plotted on fig. 5.4. Between the Rivers Tyne and Wear the outcrop is mapped as being nearly continuous whereas to the south of the Wear it is fairly well segmented into 2-4 km lengths measured along strike, separated by sand-free areas $\sim$1 km wide. However, given the limited surface exposure of the formation this pattern should be treated cautiously.

It was formerly thought that the sand had accumulated in hollows eroded into the surface of the Carboniferous rocks (Lebour, 1902; Hodge, 1932). That this is not the case may be seen at McCall's Quarry, Sherburn Hill Sand Pit, Bowburn Quarry and Ferryhill railway cutting. At these localities the topography on the top of the formation is well exposed and the mound-like shape of the sand bodies is revealed (see individual locality descriptions below). Complementary variations in thickness of the Marl Slate are seen wherever there is relief on the top of the Yellow Sands.

Though most of the variation in thickness of the formation occurs by undulation of the top there are two localities where relief on the Carboniferous surface is evident. At grid reference 497526, 7 km ENE of Seaham, two boreholes 30 m apart were drilled by the N.C.B. (numbers D8 and D8B), the second being drilled after the first was abandoned before completion. In hole D8 the top and base of the Yellow Sands were encountered at -278 m and -296 m O.D. respectively, in hole D8B at -280 m and -310 m. At Bowburn-Quarrington Hill the manager, John Bell, reports a thickening of at least 20 m taken up on the bottom surface from Bowburn Quarry into Quarrington Hill Quarry.

That the formation is disposed in a series of ridges is confirmed by borehole evidence. 208 onshore holes and 55 offshore are plotted on fig. 5.4. This

is a compilation of information from exposures, the
Durham district memoir (Smith and Francis, 1967),
published I.G.S. 6" to 1 mile maps, and published and
unpublished N.C.B. records (Magraw, 1975, 1978;
Magraw et al., 1963). Information is densest around
Ferryhill and a detailed isopach map of this area is
included in the memoir (Smith and Francis, op. cit.
p. 98) revealing a ridge oriented 060°-240°. North-
wards to Sunderland 7 other ridges can be confidently
identified with possibly 2 further bodies poorly
defined to the north. The interpretation given fits
the data remarkably well and must be regarded as a
unique solution; given the density of coverage no other
orientation of the ridges is tenable. The map is a
vast improvement on that published by Magraw (1975).
Apart from that at Ferryhill there do not appear to be
any ridges south of a line from Blackhall to Bowburn.
In this area the formation seldom exceeds 10m in thick-
ness and breccias are frequently mentioned in borehole
records. In the corridors between the ridges there are
only 8 localities where more than 5 m of sediment is
reported, including 3 with more than 15 m. These may
be secondary protrusions from the main masses and give
no cause to doubt the basic interpretation. Data
coverage is not sufficient to delineate such features
precisely, and for the same reason no attempt has been
made to contour the ridges.

From the evidence available it is not possible to discern any systematic thickening of the ridges. The ridges have an average width of 2 km (range 1.5-3.5 km), thickness 24 m and detectable length up to 25 km, oriented consistently at between 230° and 240°. The corridors between the ridges have an average width of 1 km (range 0.7-1.7 km).

Descriptions of Localities

The internal structure of the ridges is drawn on Enclosures 1-8. These are complete diagrams of every available significant exposure (excepting Crime Rigg Quarry) of the Yellow Sands drawn on a scale of 1:200, with accompanying maps at 1:1000. All but 3 of the ridges are sampled. The diagrams were compiled by photographing and surveying each exposure and superimposing data on bounding surfaces, cross-bedding dips, lamination, other sedimentary structures, colour, grain size and cementation onto the photographs. The lamination making up the rock was assessed visually at intervals along each section by estimating the relative abundance of each lamination type over a 2 m thickness. The results are indicated in a quantitative shorthand on the sections.

Three types are recognized (see Chapter 3): sand sheet (no. 1 on the diagrams), wind ripple (no. 2), and avalanche or sandflow (no. 3). Where sandflow laminae were present the thickest lamina of this type in a 2 m thickness at each point was measured and noted.

In transferring cross-bedding information to
the diagrams the many minor irregularities in the
exposure surfaces were smoothed for the sake of
clarity· The surveying was carried out by prismatic
compass and clinometer (for which thanks are
extended to Mr. G.F.G. Garrard) and pacing. Face
heights were measured by triangulation, again using
pacing as a distance measure. While the angular
relationships on the maps should be accurate the
"metres" are approximate with 10-20% error possible.
All compass bearings on the sections are measured
in degrees clockwise from north. The orientation of
the exposures is denoted by numbers and arrows in
heavy script at the top of the sections. In describing
the sections a grid reference system is used in con-
junction with the vertical and horizontal scales
attached to each diagram. Thus "52, 17 at McCall's
Quarry" denotes a point 52 m horizontally from the
beginning of the section and 17 m above the baseline,
and may be easily and precisely located using the
scales.

A total of 1.66 km of section length and
16,000m^2 of exposure is presented including 840 m of
the top and 120 m of the bottom of the formation.

(i) Cullercoats (NZ 364713). Map and Section on Encl. 1.
This exposure is part of an outlier of Permian

strata lying to the north of the 90 Fathom Fault.
The section drawn forms sea cliffs skirted by a sandy
beach in Cullercoats harbour. Further exposure lies
to the south in the foreshore and small cliffs but
is cut by many minor faults and of little value for
the present purpose. The main section, some 110 m
long and oriented NW-SE, is broken up by a number of
shallow caves eroded along major joints and small
faults. The offsets produced by these irregularities
have been smoothed out in the presented section. The
position of the section relative to the boundaries of
the formation cannot be determined. A tectonic dip
of perhaps 5°-8° S has been imposed.

Eighty per cent of the exposure is occupied by
a single set, 9 m thick and at least 100 m long.
Where accessible this set dips consistently south
and were it not for the caves would appear to be a
simple, tabular, asymptotically based set. Because the
exposure is 3-dimensional this is seen not to be so;
the set is in fact a very large trough, the direction
of dip turning from 170° to 050° and beyond. The
lamination in this set is of sandflow and wind-ripple
types, ripple lamination dominating towards the base
of the set. At the NW end of the exposure the
lenticular form of the sandflow laminae in strike
section is well displayed. The laminae are up to
40 mm thick and may extend laterally for several metres.

A feature of this exposure unique within the formation is a monolayer of small pebbles exposed at the base of the cliff around the 30 m mark (fig. 5.5). Excepting the basal rudites these are the only pebbles in the formation. The pebbles lie at or near the top of a layer of sand showing a sand-sheet lamination and bearing a good deal of coarse sand and very fine pebbles of which up to 1 m is exposed. A small area of the horizon was excavated to reveal that the pebbles are of vein quartz, subangular to subrounded, evenly distributed in a single layer over a surface of fine sand, have an average size (long axis) of 7mm and a crude preferred orientation of their long axes between E-W and N-S (Enclosure 1).

(ii) <u>Tynemouth (NZ 385694) Map and Section on Encl. 1</u>

A second faulted outlier of Yellow Sands is exposed in the cliffs of the headland on which Tynemouth Priory is situated. 170 m of the best exposed part of the section is drawn on Enclosure 1. The locality displays both the top and bottom of the formation, though since the base occurs 4 or 5 m above the foot of the cliff the Sands cannot actually be touched. The information to be gathered here is therefore restricted to that which can be deduced from the foreshore with the aid of binoculars. The arches on the section are buttresses of the sea wall. The line of exposure swings round from SW-NE in the south to NW-SE in the north. Several faults are present, though with movements of only a few metres.

Cross-bedded Sands rest directly on patchily reddened sandstones and shales of the Middle Coal Measures (fig. 5.6). The plane of the unconformity is quite smooth and horizontal, save for later faulting, and there is no basal rudite or planar-bedded sand.

The formation thins from 11 m to 5 m northwards along the length of the exposure, the decrease occurring at the top and being obsured by the sea wall. This would suggest proximity to the northern margin of a ridge. The pattern of cross-bedding shows that the ridge in this area was built up by southward-migrating and climbing bedforms, sets 1-5 m thick being separated by horizontal or gently N-dipping bounding surfaces. There is some evidence of a bounding surface hierarchy with occasional 'hanging' surfaces developed, e.g. at the 30 m mark. Probably more exist but went undetected.

At the top of the formation a cemented and structureless bed 0.1-0.2 m·thick is present beneath c.1 m of Marl Slate. From 0-100 m on the section the structureless bed is underlain by 1-3 m of flat-lying, presumably wind-ripple laminated sand.

(iii) <u>North Hylton (NZ 355574). Map and Section on Encl. 2.</u>

This exposure forms low cliffs on the north bank of the River Wear on the outskirts of Sunderland. Opposite lies the better known locality of Claxheugh Rock, a high crag of Lower and Middle Magnesian Limestone with Marl

Slate and Yellow Sands visible at its base.
Claxheugh and nearby Ford Quarry are notable for
their display of facies associated with the Middle
Magnesian Limestone reef, and also for the evidence
they provide for submarine slumping and sliding
during   Lower and Middle Magnesian Limestone time.
This activity has removed all the LML and Marl Slate,
and part of the Yellow Sands, at the east end of the
exposure.   Otherwise the main part of the limited
Yellow Sands exposure is obscured by trees and no
section was compiled.

Meanwhile, back at North Hylton, the section is
both short and low but nonetheless interesting and
displays the Sands resting unconformably on reddened
sandstones of the Upper Coal Measures at its western
end (fig. 5.7).   Only 10 m of the unconformity is
exposed before it is lost in a boggy gully occupied
by a fault of unknown throw.   Consequently, the
position of the remainder of the section in the
sequence cannot be ascertained.   The unconformity is
smooth and dips gently east.   The sand immediately
above this plane is quite coarse and bears a lamination
of sand sheet type with wisps and lenses of coarse
sand typically 5 mm thick and 0.2 m long (fig. 3.3).
0.3 m of this is succeeded by more flat-lying sand,
still coarse but with a dominantly wind-ripple
lamination.   The rest of the section is dominated by
flat lying or gently dipping sand, the low dips making

the tracing of bounding surfaces difficult. The
exposure is not large enough to erect any hierarchy of
bounding surfaces.

(iv) McCall's Quarry, Houghton-le-Spring (NZ 354506).
Map and Section on Encl. 3.

This is the largest active and accessible quarry
in the Yellow Sands, containing over 4000 m^2 of
exposure. It affords a deep section cut near the
centre of a ridge which is drawn on fig. 5.4 as being
2 km wide. One set here reaches 11 m in thickness (at
the 220 m mark on Enclosure 3), the maximum recorded
within the formation.

The general cross-bedding pattern displayed in
the quarry is explicable as very large scale, trough -
or scoop-shaped sets cut at various angles relative to
the net sediment transport direction. The portions
from 0 m to 40 m and 140 m to 250 m appear to be most
nearly transverse to palaeoflow, displaying the
trough-shaped nature of the sets, for instance in that
centred at 160, 18. From 50 m to 110 m the section
parallels the palaeoflow of most sets and it is
interesting to note both flat and concave-upwards
bounding surfaces in this section. There is no
indication that any more than 2 orders of bounding
surface are present. In the upper reaches of the
faces the lamination becomes less easily detectable as
the rock weathers and hence minor bounding surfaces
may have been missed. The cross-bedding pattern
visible in the quarry is much simpler than that seen
elsewhere in the formation (e.g. compare with
Quarrington Hill Quarry, Encl. 8). The sets are
thicker than usual with less interruption by internal
bounding surfaces, though there is no obvious feature

of grain size or location to account for this.

The distribution of lamination types in the quarry is as would be expected, with wind-ripple type predominant at low and moderate angles of dip, sandflow at high angles. 13 measurements of maximum sandflow thickness give a mean of 40 ± 17 mm with a maximum of 80 mm at 64, 10. A 2 m thickness of sand-sheet type sand is exposed for a short distance at 74, 5. Current excavation is taking place along the face between 150 m and 200 m. Often the fresh sand there has a pale grey colour, otherwise the colour is normal and yellow throughout.

From 40 m to 104 m, 138 m to 158 m and 223 m to 246 m the top of the formation is underlain by flat-lying, wind-ripple laminated sand. Oversteepened foresets occur at 41, 25 and 155, 24 and examples of small scale surface deformation of laminae at 40, 7 (as per McKee et al., 1971: the only examples discovered in the formation; see fig. 5.8). The top is occupied by between 0.1 and 0.4 m of cemented and structureless sand. An extensive pavement on top of the formation has been excavated in the area from 130 m to 240 m (fig. 5.9). The pavement is littered with abundant disarticulated Lingula credneri (fig. 5.10, 5.11). The Marl Slate is exposed above the north face (40 m - 100 m) and is only 0.1-0.2 m thick (estimated through binoculars; the bed is not accessible).

A relief of some 4 m on the top of the formation is visible in the quarry, with the top surface dipping 10° N.N.W. in the western part of the quarry and 10° E.S.E. in the eastern part. This is partially evident on Enclosure 3. The crest of the mound is oriented approximately N.E-S.W.

(v) <u>Hetton Downs Quarry, Hetton-le-Hole (NZ 358483).</u>

<u>Map and Section on Encl. 4</u>

Situated 2 km south of McCall's quarry, this
excavation works the next ridge in the sequence,
lying towards its northern side. In common with
McCall's and Crime Rigg, access to Hetton Downs is
via a housing estate and all three quarries are
under a certain amount of pressure from their local
council planning departments. Hetton Downs works
on two levels, the upper extracting Lower Magnesian
Limestone, the lower working Yellow Sands, and
exposing $2700m^2$ of the formation in two main faces
totalling over 250 m in length. Further exposure
occurs in a small, old and partly filled quarry
immediately east of the present workings.

In describing McCall's quarry it was mentioned
that in the upper reaches of faces the lamination
tended to fade with weathering. That process has
been carried to an extreme in the present case where
in the majority of the north face (0-120 m) and much
of the east (120m-250m) no structure or lamination
can be detected. Fig. 5.12 is a photograph of the
face taken ten years ago, kindly provided by Dr.
D.B. Smith. The cross-bedding is clearly visible,
demonstrating that the lack of structures is indeed
due to weathering. Where detectable the lamination
in the north face is either horizontal or low angle
and of wind-ripple type.

The east face affords a quite beautiful section
of stacked trough-shaped sets cut transverse to
palaeoflow, this being particularly obvious between
140 m and 200 m. Individual troughs are typically
30-40 m wide and 4-6 m deep. Owing to an unfortunate
perversity of outcrop no single trough has both
limbs accessible for dip measurement. Thus the
horizontal curvature of the formative bedforms
cannot be ascertained.

The pattern of bounding surfaces may be
explained by invoking 2 orders of a hierarchy. Some
surfaces obviously belong to the higher order, some
to the lower, but in many cases the evidence is not
clear cut.

Throughout most of the quarry the sand shows
various shades of its usual yellow colour. The
freshest material is often grey (current excavation
is at the southern end of the east face). Isolated
irregular patches with dimensions of one or two
metres situated high in the east face show a faint
pale pink colour.

The distribution of lamination types conforms to
expectation. Of the sandflow laminae the thickest
noted reached 40 mm.

The top of the formation is only exposed from
0-30 m and just off the far end of the section in
Enclosure 4, rising by 5 or 6 m in this distance.
Throughout most of the quarry the top of the face

lies just below the top of the formation. At the
southern end of the quarry 0.4 m of Marl Slate
overlies gently inclined, wind-ripple laminated
sand with a thin structureless zone at the top.
For the first 6 m of the north face the top of the
formation dips 10° due west and is overlain by 0.3 m
of brown, clayey Marl Slate with abundant fish
remains. A specimen of the **gliding reptile**
Weigeltisaurus jaekeli has been recovered from the
Marl Slate at this quarry (Pettigrew, 1979; Evans,
in press). The cemented zone at the top of the
Sands is unfossiliferous and from 0.01 to 0.15 m
thick, overlying structureless incoherent sand.
The top surface of the formation was excavated for
a short distance to reveal a series of anastomosing
cylindrical ridges 5-30 mm high oriented 020°-200°
(see fig. 5.13).

In the old quarry to the east a thickness of
only a few metres of the top of the formation are
exposed in several short and weathered faces. 0.5 m
of Marl Slate, brown and clayey at the base,
overlies 10-20 mm of cemented and structureless
sand above uncemented structureless or vaguely
bedded sand 1-1.5 m thick. Disarticulated Lingula
credneri are found in the top 0.1 m of sand. Beneath
this lies cross-bedded, flat-laminated (wind-ripple)
or structureless incoherent sand. At one place a
20 mm rib of soft sand is included 30 mm above the
base of the Marl Slate.

(vi)   Sherburn Hill Sand Pit (NZ 344417). Map and
       Section on Encls. 5 and 6

With a total face area of nearly 5000m^2 this is
the largest single exposure of the Yellow Sands,
though it is advisable to be discreet about access.
A general view of the west face of the quarry is
shown in fig. 5.15. The quarry lies near the
southern margin of the next ridge south from Hetton
Downs. 1.5 km to the west a borehole is reported
in the Memoir (Smith and Francis, 1967) as
encountering only 3 m of Yellow Sands, suggesting
that the Sherburn-Seaham ridge terminates in this
area. The quarry has been inactive for some years,
formerly being worked by Amey Roadstone. It is
now owned, along with the adjacent Crime Rigg
            (Fig 5.15)
quarry, by the Sherburn Hill Sand Company. With
Crime Rigg having its sole access through a
housing estate, planning permission has been
acquired to recommence working of Sherburn Hill
Sand Pit. At present (summer 1981) overburden is
being removed and a new access road constructed.
Thus in the near future the lone geologist will
need to be even more circumspect in gaining entry.

     The west, north and east faces are drawn on
Enclosures 5 and 6. The south face and southern
part of the east face were obscured by infill
shortly before fieldwork commenced. A pavement
on top of these faces at the horizon of the Marl
Slate remains, displaying the relief on the top of

the sands.   At the left hand

end of the drawing of the main, west face 10 m

of exposure are missing due to a failed photograph.

Nothing of importance has been lost.

    If it were possible to enthuse over large-scale trough

cross-bedding, the west face of Sherburn Hill Sand

Pit would be a principal object of such sentiment.

Emotion apart, the face affords an excellent

exposition of the complex arrangements of laminae,

sets and bounding surfaces that may be involved in

this sedimentary structure. It is evident that

from 0 m to 70 m as the face curves from ESE to NNE,

the section is transverse to palaeoflow. Many

broad troughs are visible, of varying internal

complexity, for example those centred at 30, 16;

30, 21; 30, 26; 50, 30; 60, 23; and 45, 20.

The troughs are typically 40 m wide and 5 m thick.

Where accessible, considerable horizontal curvature

is evident. The compound set at 30, 16, has a

dip of 26° to 310° at 22, 13 and 18° to 160° at

34, 16; that centred at 45, 20 shows dips of 26°

to 290° at 38, 18 and 28° to 180° at 48, 19. The

set centred at 60, 23 gives dip readings of 22° to

280° and 30° to 210° at points only 6 m apart. (Encl.5)

    From 70 m onwards, where the face is straight

and oriented NNE-SSW, the section is more nearly

parallel to palaeoflow, though probably still

slightly oblique. Trough-shaped sets are rare, with

bounding surfaces largely sub-horizontal. The cross-bedding is apparently unidirectional and southerly, though in fact varying from NNW to SE. Maximum set thickness is 6m, more typically 2 - 3 m. Each set extends for several tens of metres laterally; one appears to run 75 m from 70, 16 to 145, 13. Owing to the size of the west face some bounding surfaces with little apparent effect on the cross-laminae may have escaped notice. For instance, due to the terrain, photographs beyond 150 m were taken from much closer range than those to the south, and as the ground rises to obscure the exposure a much closer view of the face could be obtained. This is the reason why the bounding surface pattern becomes more complex across the gully at 148 m.

The north and east faces of the quarry confirm the pattern seen in the west face, the north showing a more transverse section of the cross-bedding than the east. Though the surveying of the quarry was rudimentary and any faith placed in the levelling should be heavily qualified, it is likely that the thick, flat-lying, sand-sheet layers at the bases of the north and east faces may be on the same horizontal level. This level may also overlap with the similar layer at 10, 5 on the west face.

As with previously described sections, the bounding surfaces need to be grouped into no more than 2 orders of a hierarchy, and again there is ambiguity in many

cases over which a label should be given to a
particular surface.

The lamination does nothing unusual anywhere
in the quarry, all types being represented. Of 28
measurements of maximum sandflow thickness the mean is
41 ± 12 mm with a maximum of 70 mm. Grain size shows
its usual variability from fine to very coarse sand,
with very fine pebbles up to 3 mm present in the
sand-sheet layer of the east face. This layer is
grey when fresh, as is its counterpart at 10, 05 on
the west face. Otherwise the sand is yellow
throughout.

Oversteepened foresets are present at 2, 19 and
18, 20 on the east face and 26, 29 on the west face.
These are all within 3 m of the top of the formation.
180 m of the top is exposed and of this distance
150 m is underlain by up to 3 m of planar-bedded (wind-
ripple laminated) sand. Whether this lamination is
exactly concordant or slightly discordant relative to
the top cannot everywhere be determined with certainty
in these 2-dimensional sections examined from a range
of 50 or 60 m. Details of the accessible parts of
the top of the formation are shown on fig. 5.16 (see
also figs. 5.17, 5.18). To summarise, the Marl Slate
rests on 0.05-0.5 m of homogeneous or obviously
bioturbated, cemented, dolomitic sand, locally
containing disarticulated <u>Lingula</u>, overlying laminated
aeolian sand either concordantly or discordantly.
In most cases the cement terminates at the base of
the unlaminated sand. The top surface of the sand

is smooth, though one small ridge of the type
described at Hetton Downs was seen on a loose block
amongst the talus covering the south face.

Some 20 m of relief is visible on the top surface
of the formation, with a ridge running E.N.E.-W.S.W.
between the centres of the northern and western faces
of the quarry. From the northern end of the west face
the top of the sand dips away markedly into Crime Rigg
Quarry before flattening out. At the southern end of
the west face the top surface dips at c.15° south,
reaching a minimum in the centre of the south face and
rising slightly to a peak in the S.E. corner. Along
the E. face the top is horizontal.

(vii) Bowburn and Quarrington Hill Quarries
       (NZ 327382 to 334377). Maps and Sections on
       Encls. 7 and 8.

These two quarries are owned by Hepplewhites but
likely to change hands in the near future. They
provide a long but intermittent section through the
next ridge south of Sherburn Hill. Both are often
termed Quarrington Quarry; to avoid confusion the
eastern, disused excavation is here labelled
Quarrington Hill Quarry; the western, working area,
Bowburn Quarry. The manager at Bowburn Quarry,
John Bell, is a keen geologist and as receptive a
gentleman as one could hope to find in such a
position. Geologists are always welcome. Unfort-
unately filling of the quarry will commence in the
near future unless a buyer can be found. Quarrington
Hill Quarry has been disused for some years and is
being filled very slowly.

Bowburn Quarry has a long, low, L-shaped section
through the top part of the formation. 600 m W.N.W.
of the northern end of the quarry an exposure near
Heugh Hall shows Marl Slate resting on Carboniferous.
The top of the formation (otherwise smooth and
horizontal throughout the quarry) begins sloping down
towards this at the northern end of the quarry at
10°-15° to the N.W. (fig. 5.19, 5.20). The cross-bedding
throughout the face beneath this also dips N.W.,
broken up by several bounding surfaces, which are
difficult to trace in the

strike section visible. The cross-bedding pattern seems to be influenced by the margin of the ridge as far as the 130 m mark on the section. Between 130 m and 190 m lies the tallest face in the quarry. It displays a very complex pattern of laminae and bounding surfaces probably representing a transverse section of trough cross-bedding. The low face beyond 190 m constitutes the most accessible exposure of the planar-bedded zone in the whole outcrop of the Yellow Sands. It is coarse grained, bearing a sand-sheet lamination and largely but not wholly concordant with the top. It is overlain by 0.1-0.25 m of structureless and cemented dolomitic sand and about 1 m of Marl Slate (this is mostly obscured by talus). At the northern end of the quarry the structureless and cemented sand is commonly 0.5 m thick and contains disarticulated <u>Lingula</u> and casts and moulds of <u>Permophorus</u> (fig. 5.21) in the top 0.1 m or so and littered across the top surface of the formation. The Marl Slate is again nowhere visible in its entirety but probably 1.5 m thick. There is locally a prominent plane of parting 0.05-0.1 m below the top of the sand, the surface of which may be lineated. A few 10-20 mm diameter, 0.1-0.5 m long, irregular, cylindrical ridges adorn the top surface on loose blocks in the NW corner of the quarry. Gentle contortions of the laminae are present at 113, 11; 128, 10 and 224, 11 (figs. 5.22-24).

Lamination and grain size conform to the
expected pattern in the quarry, sandflows being more
common at high angles of dip, wind ripple at low, and
grain size varying from fine to very coarse sand.
19 measurements of maximum avalanche thickness give
an average of 36 $\pm$ 9 mm and a maximum of 50 mm.

In the face between 145 m and 190 m are
numerous areas of red sand, lenticular in shape
and oriented parallel to the lamination (fig. 5.25).
These zones may extend 1 or 2 m laterally and are
generally 0.1-0.2 m thick.   Elsewhere the sand is
yellow.

The 9 separate faces of Quarrington Hill Quarry,
distributed over a distance of 400 m WNW-ESE provide
sections of varying orientation and clarity through
the formation. Most of the faces are steeply sloping
rather than vertical and thus have accumulated hill-
wash, rendering the location of bounding surfaces
rather difficult. Faces B, C and D have suffered most
in this respect. The impression is gained that most
of the faces cut sections oblique or transverse to
the axes of trough-shaped sets: nowhere is a pattern
resembling Sherburn Hill's west face seen. This
impression is strongest in faces E, F and G. Once
again it is probable that only 2 orders of bounding
surface are present.

Contorted laminae are present at 16, 10m on face
A, at 46, 22 and 49, 25 on face F, and at 1, 13
and 16,11 on I. The latter two locations are
particularly interesting since they must lie at least

10 m below the top of the formation - all other
examples of contortion seen in the Sands are within
3 m of the top.

160 m of the top of the formation is visible
in the quarry. Though flat throughout, the top is
more variable here than at any other locality; details
are given on fig. 5.26. For most of the section the
usual situation of aeolian sand succeeded by
homogeneous sand under Marl Slate is adhered to.
However, between 15 m and 48 m on faces E and F the
middle member of this sequence is absent, and thin
Marl Slate rests directly on sand showing an aeolian
lamination. Also notable are 3 examples of poorly
defined small scale cross-lamination (20-50 mm sets)
within the middle member. Two-dimensionality
precludes any more precise statement of palaeocurrent
than that both southerly and northerly flow is
indicated. These features are illustrated in figs.
5.27, 5.28. Over faces B, C, D, E, F and H the
thickness of the Marl Slate does not exceed 0.2 m.
At the east end of face A, 1.5 m of the Marl Slate is
visible while at the west end of this face it thickens
to 2.45 m, filling a local hollow a few metres across
in the sand surface. Also at this point the top
$\sim$ 0.5 m of Sand includes several clay partings and
the Marl Slate contains a number of ribs of fine -
medium grained, moderately sorted, white sand from
monolayers to 5 cm thick (fig. 5.29). None of these
shows any sedimentary structures, save for a sinusoidal

waviness of amplitude 5-10 mm and wavelength 0.1-0.2 m on the upper surfaces of the thicker beds within the Marl Slate, and a similar but more subdued feature on the lower surfaces. Numerous clay-filled later fissures and small faults cut the rock at this locality, complicating matters somewhat.

Lamination and grain size behave normally throughout the quarry. The sand is yellow though sandflow laminae of coarse sand may be clean and white; this is a common feature in the formation.

(viii) <u>Ferryhill Railway Cutting (NZ 303323). Map and Section on Encl. 2</u>

The next ridge south of Bowburn crosses the coast at Horden to terminate near Deaf Hill before reaching the line of outcrop. The eastern end of the ridge exposed at Ferryhill is 5 km SW of this point. The Ferryhill ridge has a proven length of less than 10 km, its only exposure being that drawn.

The cutting is occupied by the main London to Newcastle railway line, but is wide enough for there to be no danger to geologists and there are no problems with access.

Much of the surface of the outcrop is unfavourably weathered, with the lamination badly obscured. This does not affect the determination of the overall pattern of cross-bedding, which once again is of obliquely or transversely cut trough-shaped sets. The

trough centred on 110, 3 is particularly well
defined, the dip direction within it varying from
330° to 190°.

In the northern part of the outcrop the Marl
Slate is obscured and its thickness cannot be
measured until the 50 m mark, where it reaches 0.3 m.
Between 100 m and 125 m it is absent, Lower Magnesian
Limestone resting directly on sand. 10 m beyond the
end of the drawn section 1.1 m of Marl Slate is
present.

In general the top of the sands conforms to the
normal pattern of locally fossiliferous, structureless,
reworked material, overlying flat-laminated aeolian
sand. In detail matters are considerably more
complicated and ambiguous. The poor preservation of
the rock poses a major obstacle to description and
interpretation - the lamination obscured by the
weathering being a principal diagnostic factor.

A planar-bedded zone 1-2 m thick is present along
the whole length of the outcrop at the top of the
formation. It is overlain by 0.1-0.4 m of cemented,
slightly mottled, otherwise homogeneous, dolomitic
sand with <u>Lingula</u> fragments. The principal problem
lies in elucidating and interpreting the nature of
the lamination displayed by the planar-bedded zone.
It ranges from clearly sand-sheet type, to clearly wind-
ripple type, to all but obliterated by weathering to
what is believed to be a spurious liesegang-type
diagenetic colour banding.

Contorted lamination occurs at 50, 7; 66, 8; 106, 7 and 116, 6 (fig. 5.30). The example at 66, 8 is particularly notable since it lies <u>within</u> otherwise planar-bedded sand.

The exposure is all yellow, grain size varies from fine to coarse as usual, and the lamination is of the usual types distributed in the usual manner. A rather greater proportion of the sand is cemented than in other exposures, the cemented ribs following the lamination or bounding surfaces.

(ix) <u>High Moorsley (NZ 334457)</u>

This exposure no longer exists. It lay on the south side of the road from High Moorsley to Pittington, and consisted of about 2 m^2 of rock straddling the boundary between the Yellow Sands and Marl Slate (fig. 5.31).

The locality showed 1.35 m of Marl Slate with thin sands (maximum 15 mm thick) within 50 mm of the base. This lies on a surface of cemented, coarse grained, dolomitic sand showing irregular relief of 50 mm. This cemented sand is 0.1-0.2 m thick resting on wind-ripple laminated sand, and is structureless except for some very well defined burrows found on a sample taken near the base (fig. 5.32). The burrows are straight, inclined and cylindrical with smooth walls, a diameter of ∼10 mm and visible lengths up to 60 mm. They have an infill of silt-grade dolomite.

The locality lies on the extreme southern margin of the Hetton-le-Hole to Ryhope ridge. In the Memoir (Smith and Francis, 1967, p. 99) the Sands are described as being 6 - 9 m thick, wedging out a short distance south of the exposure.

(x)   Elemore Hall School (NZ 352443)

Exposure here occurs in and old quarry amongst the woods a short distance east of the main school building. The quarry face is 25 m long, up to 2.5 m high and oriented approximately N-S. The top of the Yellow Sands is exposed: it is level and overlain by Lower Magnesian Limestone, though it is not completely certain that the LML is in situ. The uppermost 0.2 m of sand is cemented. 0.04-0.1 m of this is structureless, the remainder wind-ripple laminated and dipping $10^{\circ}-15^{\circ}$ E or NE. This bed is up to 1 m thick and overlies cross-bedded, yellow aeolian sand dipping between ENE and NNE. The exposure is 2 km SW of High Moorsley and close to the northern edge of the Sherburn to Seaham ridge. The absence of the Marl Slate suggests a considerable thickness of sand in the area, though the Memoir (Smith and Francis, loc. cit.) tentatively mentions only 6-9 m.

(xi)   Cassop Vale (NZ 339383)

This exposure is approached along a public footpath leading west from Cassop Colliery village. It is described in the Memoir (Smith and Francis, op. cit., p. 100) as showing 7 ft (2 m) of sand. At present the

Yellow Sands can only be found by grubbing around among tree roots; most of the quarry is in Lower Magnesian Limestone. The Marl Slate appears to be absent, though there is a gap in the exposure of a few centimetres between the lowest LML and highest Yellow Sands. The top 0.2 m of the Sands is structureless, and cemented with flat-lying, wind-ripple laminated material beneath.

Smith and Francis (loc. cit.) suggest a probable thickness in excess of 30 m for the Yellow Sands in this area, which is in the centre of the Bowburn-Easington ridge. Such a thickness would be consistent with the absence of the Marl Slate.

(xii) Old Quarry, near Quarrington Hill (NZ 332374)

About 6 m of the top of the formation is exposed in this small excavation lying about 300 m south of the eastern end of Quarrington Hill Quarry. The Marl Slate is present but incompletely exposed and rests on 0.2 m of hard dolomitic sand showing no structures. Below this is a limited length of exposure of 0.5 m of medium-scale cross-bedding directed southwards. Two or three sets are present and the lamination does not appear to be of an aeolian type. This rests on low angle cross-bedded sand with a lamination of uncertain type. Elsewhere in the exposure aeolian sand shows oversteepened foresets.

(xiii) <u>East Thickley</u> (NZ 240257), and <u>Middridge</u>

(NZ 249252)

Situated just a kilometre apart these two

quarries are in Lower Magnesian Limestone but also

expose the Basal Permian Breccia. East Thickley

quarry is partly filled with rubbish and only a very

small amount of relevant exposure remains in a dark,

dank corner. Middridge, however, provides a good

sized and fresh exposure in Marl Slate, Breccia and

the top of the Carboniferous. The locality is des-

cribed in detail by Bell <u>et al.</u> (1979) and is especially

notable for a rich Marl Slate flora and fauna. It has

consequently been designated as a Site of Special

Scientific Interest, though it still collects domestic

rubbish from time to time, no-one having taken the

obvious step of erecting a notice to indicate its

importance.

The description of the Breccia provided by Bell

<u>et al.</u> (op. cit., p. 111) is excellent and thorough,

though the rock is a dolomitic, silty breccio-con-

glomerate rather than a breccia. To summarise, the

deposit contains angular, locally derived, fragments

of sandstone and siltstone and more rounded clasts of

quartz and quartzite. Flaky clasts tend to be

oriented sub-horizontally, reaching 0.15 m in length; the

maximum size of the quartz pebbles is ~50 mm. Other-

wise the deposit is largely unstratified with a

matrix of dolomitic silt and quartz sand, including

many well-rounded grains. The colour is yellow but grey when fresh and the Breccia grades up into the Marl Slate. The Carboniferous surface is largely smooth and unfissured, with the sandstone beneath light brown in colour. Negative attributes of the Breccia include the lack of dreikanter, current indicators, trace fossils (e.g. roots), calcareous concretions, and indeed structures of any kind, save for the orientation of bladed clasts and intercalations of pebbly siltstone mentioned by Bell et al.

Palaeocurrents

To assess the wind regime under which the Yellow Sands were deposited, measurements of cross-bedding dip and dip direction were made on every accessible set. The 239 readings with a dip of greater than 10° were then divided into 20° class intervals, and plotted on a single rose diagram, giving the result shown in fig. 5.33. No correction for tectonic dip was necessary. The striking bimodality, with one mode directed to 180° and the other to 290°, has its only precedent among studies of ancient aeolian sandstones, in Clemmensen (1978). The vector mean, at 235°, bisects the two modes, and with a remarkable and striking internal consistency corresponds exactly with the trend of the sand ridges deduced from the isopach map.

Palaeocurrent distributions for those outcrops with more than 20 readings are shown in fig. 5.34. All 5 localities show both modes with varying degrees

of emphasis due probably to insufficient sampling rather than any real systematic variation. Fig. 5.36 shows all the data on a Glennie (1970)-type diagram. Fig. 5.35 is a histogram of 242 measured amounts of dip by 4° class intervals. Since this plot is derived from data acquired primarily for the elucidation of palaeocurrents it is strongly biased towards high angles. The portion of the graph to the left of the mode can have no physical significance. The value of the diagram lies in its illustration of the sharp curtailment of dip values at about 28° (values were measured to an accuracy of ± 1°). Only once (at Sherburn Hill, west face) was a dip of 34° encountered in undisturbed sand, this angle being commonly quoted as the angle of repose of dry sand.

Summary of the Sedimentary Features at the Top of the Formation

This subsection is included to provide a concise and separate account of the top of the formation in order to properly emphasize features which have considerable importance, both to the interpretation of the deposition of the Yellow Sands and the mechanics of the Zechstein transgression.

The topography of the top surface of the ridges is fairly subdued, with the most relief (~ 20 m) and the steepest slopes (~ 15°) exposed at Sherburn Hill (see fig. 5.14). Other exposures (Tynemouth, McCall's, Hetton Downs, Bowburn, Quarrington Hill and Ferryhill)

show broad, gentle mounds with slopes of 10° or less.
No relief of the size or shape of dunes is present and
nowhere does the structure of underlying cross-bedded
sand appear to bear any relation to the top surface.
(See Encls. 1 - 8).

Detailed measured sequences of accessible parts
of the top at Sherburn Hill and Quarrington Hill are
shown on figs. 5.1b and 5.26. Figs. 5.19, 5.20, 5.9 -
5.11 , 5.13 , 5.17 - 5.24, 5.27 - 5.32 illustrate these
features. The sequence is summarised and compared
with the Rotliegendes-Zechstein boundary in the
southern North Sea and Poland in fig. 5.37.

Over the 840 m of exposure of the top of the
formation inspected, 14 samples of gently contorted
and over-steepened cross laminae were spotted. The
most extreme example is at Ferryhill (fig. 5.30),
where contortions also occur within the planar-bedded
zone. Disturbed planar-bedded sand was also seen in a
loose block at Bowburn. The variability of the
planar-bedded zone should be stressed. It is not a
continuous, homogeneous whole; bounding surfaces and
cross-bedded sets are truncated by the top of the
formation and the planar-bedded sand does become
discordant with the top in places. Nevertheless it is
a real feature, occupying nearly 70% of the exposed
length of the top, whereas similar material occupies
less than 20% of the formation along a line 5 m below
the top. It consists of wind-ripple laminated sand,
with sand-sheet type developed at Bowburn and Ferryhill.

Reworked sand (including bioturbation in the definition of reworking) occupies 97% of the exposed length of top, being absent only over 25 m of faces E and F at Quarrington Hill. The reworked zone is the most consistent feature of the boundary, seldom varying much from a thickness of 0.1-0.3 m and always being either structureless or clearly bioturbated. The fauna is restricted to disarticulated <u>Lingula credneri</u> and <u>Permophorus costatus</u>, the latter as casts and moulds littered sporadically over the top surface. The only evidence of active reworking, where the sand was moved and redeposited in flowing water, is the three examples of possible waterlain cross-lamination at the east end of Quarrington Hill Quarry (fig. 5.27).

Other minor features of the top are the small anastomosing or isolated ridges seen at Hetton Downs (fig. 5.13), Bowburn and Sherburn Hill, and the lineations on a parting plane within reworked sand at the northern end of Bowburn Quarry. Clay partings are present in reworked sand on face A at Quarrington Hill and at the southern end of Sherburn Hill east face (figs. 5.1b, 5.26). These must demonstrate localised redistribution post-transgression, as must the thin sandstone ribs in the lower part of the Marl Slate seen at High Moorsley and in face A, Quarrington Hill. At a much later stage the Yellow Sands was partly involved in large scale submarine slumping and sliding during Lower Magnesian Limestone times (Smith and

Pattison, 1972). The sand in the slumped beds is always disseminated, never as coherent cemented blocks.

Summary

The Yellow Sands are disposed in a series of parallel ridges oriented roughly N.E.-S.W. and on average, 2 km wide, 1 km apart and 24 m high. The Sands do not rest in hollows on the surface of the underlying Carboniferous strata. They are overlain by the Marl Slate and succeeding Zechstein carbonate sequence, the Marl Slate thinning over eminences in the Yellow Sands ridges.

The formation is for the most part constituted of cross-bedded sand, in trough-shaped sets typically 3-6 m thick and 30-50 m wide. The top part of the formation consists of laterally extenive flat-laminated and structureless sand up to a few metres thick. The base of the formation is also made up of flat-laminated sand, though this unit is less laterally extensive than the upper; cross-bedded sand resting on Carboniferous strata in places.

Cross-bedding azimuths derived from the formation show a bimodal pattern, with one mode directed to the S., one to the W.N.W. The vector resultant is directed to 235°, parallel to the trend of the ridges.

CHAPTER 6

THE DEPOSITION OF THE YELLOW SANDS

Introduction

Having described the lithology, distribution and sedimentary structures of the Yellow Sands in Chapter 5, this Chapter proposes an interpretation of the deposition of the formation, including the effects of the Zechstein transgression. The petrography and grain size characteristics of the formation are discussed in Chapter 7.

SECTION 6.1   BEDFORMS, BOUNDING SURFACES AND WIND REGIME

The individual ridges shown on the isopach map are longitudinal draa. Their size and regularity of spacing and direction allow no other interpretation. The consistency of dimension, orientation and relative position compels their interpretation as bedforms. There is no evidence that they are fixed by features developed on or in the underlying Carboniferous. If the ridges are bedforms (in aeolian sand) they can only be draa - no other bedform has a 2 km spacing. The parallel symmetries of the draa and the palaeocurrent diagram suggests that the caveat raised in Section 2.4 can be ignored, (at least as far as the draa are

concerned, see pp. 136-139). The coincidence of the ridge trend and palaeocurrent resultant (interpreted as the net sand-drift direction) is definitive of longitudinal bedforms.

There does not seem to be any regular peaking of the ridges that might suggest the development of superposed stellate draa forms, though the thickness data is not dense enough to be certain. The average thickness of the ridges (24 m) is not great as draa heights go (up to 430 m, Wilson, 1970, p. 19) and indeed is less than the thickness of some individual sets of cross-bedding in other aeolian formations. However, next to nothing is known of the vertical thickness of modern longitudinal draa and in the case of the Yellow Sands it is possible that sand may have been eroded from the ridges immediately before, during, or after the Zechstein transgression.

The isopach map shows that there must be at least a long wavelength undulation on the top surface of the draa, assuming the surface of the Carboniferous to be plane. Variations of the order of 10 m per kilometre are recorded. Given the density of the data this may not mean much, serving only to demonstrate that the draa are not regular, flat-topped prisms. Dune-sized features are not detectable.

Since longitudinal draa can migrate only by extension at their downwind end and lack slipfaces of their own, the cross-bedding of the Yellow Sands must have been developed by migrating dunes. The shape and

behaviour of these dunes and the development of the draa is recorded in the cross-bedding. Certain features are immediately obvious : the broad trough-shaped nature of many sets indicating that the dunes were of the sinuous transverse variety and that only lunate elements are preserved. The smooth profile of the laminae in sections parallel to palaeocurrent shows that the preserved parts of the dunes had smoothly curving leesides. Problems open to further discussion and argument are : (i) the interpretation of the palaeocurrent diagram, (ii) the specification of the formative wind regime, (iii) the meaning of the sand-sheet deposits, (iv) assigning particular causes to bounding surfaces, (v) the size of the dunes involved and (vi) the rate of deposition of the formation.

The Palaeocurrent diagram

This (fig. 5.33) raises the question of whether there were one or two sets of dunes present on the ridges, each set corresponding to a mode of the palaeocurrent diagram. If it could be established that two sets existed, the modes of fig. 5.33 could be taken as representing real winds (on a dune scale, see Ch.2). If only one dune set existed then the modes might be due either to real winds or the secondary flows of a single, N.E.'ly wind, providing an example of the ambiguity highlighted in Section 2.6. Whichever interpretation pertains, the resultant of fig. 5.33 may still be taken as the net sand-drift direction, justifying the labelling of the draa as longitudinal.

The palaeocurrent diagram records the orientations
of laminae making up the formation. Those laminae
mark former positions of dunes. It is therefore
possible to reconstruct the plan shape of the 'average'
dunes capable of generating the observed distribution
of dip azimuths. This. leads to the results shown in
fig. 6.1 . The single dune set solution accounts for
93% of the data and requires that the preserved parts
of the dunes were very tightly curved, encompassing
170° of azimuth, with the dune facing at about 230°.
The two dune set solution involves two similarly shaped
leesides with much more gentle curvature, one set
facing just west of south, the other to the WNW.
These solutions carry the implicit and false assumption
of deterministic bedform behaviour. Any allowance of
a standard deviation of dune migration and orientation
would reduce the curvatures of the reconstructed lee-
sides. The diagrams do not represent real bedforms
in any sense.

A uniformitarian approach to the resolution of
this question can bear little fruit - information
about dune patterns on longitudinal draa is not
available. Two photographs in McKee 1979a(one oblique
aerial photograph, the frontispiece; and one ground
photograph, fig. 67   ) are all that is available
on which both dunes and draa are resolvable. Neither
of these look capable of producing a bimodal palaeo-
current pattern.

The theory expounded in Chapter 2 would favour the

existence of a single bedform set, whether developed
by secondary flows of a unidirectional wind or the
primary flow of a bimodal wind regime.

However, this is not an occasion for the
exercise of theory. The question should be answerable
by reference to the rocks. If the bedforms behaved
systematically, the previous presence of two different
orientations of dune should be easily detectable:
two distinct orientations of cross-bedding should be
detectable in most vertical sections. Extensive
horizontal sections, if they existed, would show the
curvature of the cross-bedding in plan. This could
also be detected by multiple dip readings from indiv-
idual sets. The unquantifiable random element
inherent in bedform behaviour and shape (and the
vagaries of weathering and outcrop) inevitably
blurs and confuses all these idealistic distinctions.

The question can best be resolved by examination
of the 3 large outcrops: Sherburn Hill west face
(Encl. 5), McCall's Quarry (Encl. 3) and Hetton Downs
Quarry (Encl. 4). The evidence suggests that only one
dune set was present. For instance, Sherburn Hill west
face beyond the 50 m mark is oriented 020°-200° and
would be expected to show some perfectly transverse
sections through troughs, some perfectly longitudinal.
Most sets appear to be oblique sections, becoming
transverse only where the face is oriented roughly
NW-SE. This fits the single dune set model of fig.
6.1 . The curvature of the troughs deduced from

multiple dip measurements, detailed in the previous section, also satisfies the hypothesis of a single dune set. However, this conclusion must lack a certain amount of conviction in the absence of horizontal exposure. The curvature of the cross-bedding is by no means uniform; the lowest exposed set in part of McCall's Quarry turns through only 30° of dip azimuth in a section 80 m long across strike. Nevertheless, the concensus of the outcrop data is that only one dune set was present during the deposition of the Yellow Sands.

Wind Regime

The distribution of cross-bedding dip azimuths labelled as a palaeocurrent diagram in fig. 5.33 might be interpreted as representing :

    (1) regional N.'ly and E.S.E.'ly winds, alternating on probably a seasonal or diurnal basis,

    (2) secondary flows on a draa scale, viz. the helicoidal vortices of Hanna (1969), Glennie (1970, p. 91) and others,

    (3) secondary flows on a dune scale, after Wilson (1972) and Cooke and Warren (1973, pp. 298-299).

The bimodal regional wind regime has to be seen in the context of other palaeowind results from aeolian sandstone formations of Lower Permian age elsewhere in Great Britain. These are all unimodal, ranging from N.E.'ly to S.E.'ly (fig. 4.3). The N.'ly wind contributing to the Yellow Sands is most incongruous. The preservation of longitudinal draa in the Yellow Sands

is also most incongruous; perhaps there is a link.
It may be that the (necessarily transverse) bedforms
developed on the extensive, continuous sand cover
of other areas filtered out the bimodality of the
wind. Examination of the Permian continent
reconstruction (fig. 4.2) and comparison with a map
of the modern world reveals the possibility that a
depression   may have developed over southern Russia
during Permian summers. This would have been caused
by heating of the landmass in that area generating
rising air, analogous to the modern climate of
southern Asia. Circulation around this depression could
have brought about a summer NNE'ly wind (Permian NNE
in Britain = modern N) over Britain. Rainfall would
not have been induced, the upwind circulation lying
entirely over land. In winter the SE trades would
take over. Such a wind regime exists over Iran in
the present day (Times Atlas of the World, Pl. 4).

   Alternatively the Yellow Sands N'ly might be a
Katabatic wind draining off the Mid N. Sea High (see
fig. 5.2b). Its alternation with the E'ly would then
have been diurnal.

   A true bimodality of the wind, derived by what-
ever method, conforms to the view suggested in Fryberger
(1979a) for the origin of longitudinal dunes and draa.
This is that a wide unimodal or bimodal wind is
necessary (see pp. 33-34).

   The alternative view, that of contra-rotating
spiral vortices developed from a unidirectional regional

flow oriented parallel to the draa can also be accommodated by the Yellow Sands. In this instance the two modes of the wind represent the influence of the opposing vortices situated either side of each draa.

The third possibility also involves a uni-directional regional wind, any draa-sized spiral vortices in this case having no influence on the dunes (hence their presence could not be confirmed or denied from the palaeocurrent diagram). In this case the extreme sinuousity of the dunes would be attributed solely to dune-sized secondary flows. The two modes are then spurious, as in the preceding instance and the real wind is a NE'ly (Permian E'ly).

Establishing a preference among these three interpretations is not easy - all the models suffer from a dearth of documented modern analogues, as is so often the case with the windies. This enhances Fryberger's (1979a) work as the only study of the relation between wind and bedform shape. His conclusions suggest that something more than a unidmodal wind is required to develop longitudinal bedforms. On this basis the validity of the two modes of fig. 5.33 becomes more compelling and a preference has to be stated for a bimodal regional wind. If the wind had been unimodal the Yellow Sands would probably have developed as barchan or transverse draa (this is an important point), such

as those figured by Glennie (1970, figs. 76, 78).
Establishing a preference for the cause of a
bimodal wind regime lies beyond the realms of
sedimentology (and fact).

Sand-sheet deposits

These make up approximately 4% of the exposed
parts of the formation. The example occuring at
the base of each face in Sherburn Hill Sand Pit may
be contiguous and hence laterally extensive. It
is 5 m thick in the west face and at least 6 m in the
east face (Encls 5 and 6). Other examples are
found in faces H and I at Quarrington Hill Quarry
(Encl. 8); at the top of the formation at Bowburn
(212, 11, Encl. 7); at 75, 05,   at McCall's Quarry
(Encl. 3); at 05, 02; 30, 01 and 80, 02 at North
Hylton (Encl. 2); and at 35, 01 in the Cullercoats
section (Encl. 1). Lateral extensiveness in all
these examples is possible but cannot be proven.
At Bowburn the sand-sheet material lies immediately
below sand reworked by the Zechstein transgression.
One example at North Hylton (05, 02) rests on the
Carboniferous. The vertical position within the
formation of the other examples cannot be determined
with confidence. Most, however, could quite
feasibly be only a few metres above the base.

Whilst the Cullercoats example contains fine
pebbles and several other examples contained very
fine pebbles only $\sim$ 2% of the sieved sand-sheet

samples exceeded 1 mm in diameter. In a sandflow
sample taken from 52, 06 at Bowburn, 20% of the
grains were coarser than 1 mm. The crucial feature
of the sand-sheet material is not its mean size but
its well defined bimodality, with two equally rep-
resented modes of coarse and fine or very fine sand.
The mean of most other samples, and the average of
the means of all the sieved samples falls neatly
into the gap in the sand-sheet distributions.

The sand-sheet deposits always show low or
negligible primary dips, demonstrating that they
must have accreted as flat, dune-less, but probably
rippled surfaces. This must be due to the effect
expounded by Bagnold (1941, p. 180) whereby a cover
of coarse grains over a surface increases its
sand-passing ability whilst decreasing its capacity
to trap sand. Fine sand is thought to accumulate
by interpenetration of the coarse mode on each
ripple. Once initiated, such a surface may be self per-
petuating, capable of trapping fine sand at a rate
governed by the supply of coarse grains whilst medium
grains pass on unhindered. The sieved samples show
that by mass the resulting accumulation is divided
roughly 50-50 between fine and coarse. The question
is: how might such a surface be initiated?

What is desired in a relative enrichment in the
coarse sand population. This does not involve the
sudden influence of a new supply of grains much coarser

than the rest of the formation, but merely a selective enhancement of one particular component in the normal supply. There are a number of ways of doing this, none testable by reference to the rocks.

The coarse mode, moving by surface creep, tends to be held up at the foot of inclines. Thus if the draa went through an initial stage of growth as sand sheets their accessibility to coarse sand would be increased whilst the relief of the mounds remained low. As the mounds aggraded, the coarse sand supply would diminish and true dunes would develop. This is a potential explanation of those developments of sand sheet material located at or near the base of the formation.

Alternatively the dune-less surfaces might develop by deflation if the wind load ever fell below saturation. This could only occur at the upwind parts of the erg because the ensuing deflation would load up the wind in a very short distance. However any sand sheets so developed, of whatever area, might then migrate downwind, just as any other bedform, for as long as it was not swamped and obliterated by medium sand. If deposition continued over the sand sheet during migration it could build up a planar-bedded "set".

If the draa reached equilibrium (i.e. ceased to grow vertically) at any stage, the superimposed dunes would, on average, cease to climb as they migrated. It is possible that they would still leave

a residue of the coarser, less mobile grains as they passed, thus armouring the surface and perhaps generating a sand sheet.

As a final possibility, particularly vigorous storms might sweep sufficient coarse grains onto the draa to suppress dunes for a while.

The demise of the sand sheets seems to come about simply by the resumption of dune migration across the area, the resulting cross-bedding being separated from the sand sheet by a bounding surface. Nowhere was a gradual transition upwards noted.

Circumstances that might bring this about are less easily definable than those for the initiation of a sand sheet. If the sand sheet areas are able to migrate they could conceivably do so at the same speed as the surface creep, and a couple of orders of magnitude faster than a dune. On the other hand they might be held up by an incline (e.g. the windward side of dunes) lying downwind. Thus it is not clear whether sand sheets can be overtaken and smothered, or are just naturally followed by migrating dunes. It is possible that an abnormally abundant supply of medium sand may be required to swamp them. Alternatively it might be stated that the normal behaviour of upwind transverse dunes would tend to filter out the coarse sand supply to the extent that any sand sheet with such dunes upwind is rapidly strangled.

A number of mechanisms for both the birth and
death of sand sheets can be imagined, though a few
sentences of untested hypothesis do not necessarily
constitute validity. Nevertheless the ideas presented
above suggest that sand-sheet deposits may be
preserved without signifying any specific radical
events or changes in conditions of deposition of the
Yellow Sands. These processes must, however, be
enhanced by the longitudinal nature of the draa,
whereby no area of Yellow Sands is more than 2 km from
an interdraa corridor. These may well have
functioned as sources or passageways for the transport
of coarse grains.

Bounding Surfaces

First order bounding surfaces should be absent
from the formation because longitudinal draa do not
migrate and climb over each other. Second order
surfaces may be present, but the picture should be
dominated by 3rd and 4th order surfaces : those due to
the migration and modification of dunes. In the
outcrop descriptions of the previous chapter it was
indeed repeatedly stated that the pattern of bounding
surfaces within the Yellow Sands could be explained as
two orders of a hierarchy.

The interpreted labels of the bounding surfaces
have been colour-coded onto the outcrop diagrams.
Extensive surfaces and those separating sets differing
in dip by a considerable amount were labelled 3rd
order. Other surfaces truncating these were also
labelled 3rd order. 'Hanging' surfaces, such as those
at 240, 12 and 245, 12 in McCall's Quarry were about
the only type that could confidently be labelled 4th
order. The principal instructive function of this
exercise is that it is a salutary lesson in the

complexity and intractibility of bounding surfaces
in aeolian sandstone. Exposures such as McCall's
Quarry and the west face of Sherburn Hill Sand Pit
are dominated by 3rd order surfaces, mostly
identifiable with a reasonable degree of confidence
and showing some regularity in arrangement. Smaller
faces, such as faces A, H and I at Quarrington Hill
and Sherburn Hill north face, examined from closer
range and where more detail was discernible, show a
dense and perlexing array of surfaces with little
regularity. Hierarchies may make for wonderfully
elegant theories but are not so great in practice.

The overall form of the lower surface of each
set is clearly trough-shaped, or in the form of a
longitudinal segment of an extremely prolate
ellipsoid; as if a saucer were considerably
stretched along one diameter. Cut transversely, the
resulting troughs are 30-50 m wide. In longitudinal
section they reach 90 m at McCall's Quarry (70 m at
Sherburn Hill) without either end being seen.
Measured at their thickest point, sets are typically
3-6 m thick, maximum 11 m.

It is possible that the lower boundaries of
some of the accumulations of sand-sheet type laminae
represent 2nd order bounding surfaces. The definition
(that the surfaces record draa modification) would be
satisfied for those examples induced by temporary
stabilisation of the draa. Whether the development
of limited sand sheet areas can be counted as draa

modification and hence merit a 2nd order surface is a moot point, though one which goes some way to push back the frontiers of hair-splitting.

On the whole the picture provided by the larger exposures is one of sinuous transverse dunes climbing and migrating in a fairly regular way along the draa. The deposits of each dune are only preserved over a finite distance though this distance is greater than can be assessed from the present exposure. Fourth order surfaces are generally not preserved, or are not distinguishable with confidence.

Dune Size

Wilson (1972b) suggests that in sinuous transverse bedforms the wavelengths of the longitudinal and transverse elements are similar. The trough width preserved in the Yellow Sands (up to 40 or 50 m in the best exposed samples at Sherburn Hill) sets a lower limit on the longitudinal wavelength. Thus the wavelength of the Yellow Sands dunes might have been 50-100 m. (The relationship $\frac{h'}{\lambda_d} = \frac{D}{d + D} \sin \theta$ worked out on pp. 44 - 46 applies only to dunes on the lee-side of a transverse draa). Kocurek and Dott (1981), in a recent paper, provide a graph of grainflow (= sandflow) maximum thickness versus slipface heights, measured at the Little Sahara dune field, Utah. Extrapolating this graph a typical Yellow Sands maximum sandflow thickness of 40 mm indicates a slipface height of 20 m, though the method should not be trusted too much.

Otherwise, the size of the dunes which contributed to the Yellow Sands remain indeterminate, except to say that they were generally larger than c.5 m (a typical set thickness).

The Rate of Deposition of the Formation

Those dunes whose deposits are preserved in the Yellow Sands left trough-shaped sets with an average thickness of the order of 2 m (spreading the volume of the trough out as a flat slab of the same width). Whether every dune or just a tiny fraction deposited sand cannot be determined. If the dunes had an unlimited lifetime, all deposited a 2 m equivalent slab, had a wavelength of the order of 100 m and celerity of about 1 m per year (plucking 'reasonable' figures out of the air) the draa would have grown vertically at 20 mm per year (given by equivalent set thickness - time taken to migrate one wavelength). This must be a maximum figure, given the assumptions of deterministic dune behaviour. It indicates a minimum growth time of the draa of 1000 - 2000 years, provided sand supply could be maintained at this rate.

Alternatively, consider a typical modern sand-drift rate of $10m^3/m/yr$, of which say 30% (a complete guess) is retained in the Yellow Sands erg during draa growth. The mean spread-out sand depth is about 15 m. On this basis each 10 km length of the erg (measured parallel to the ridges) would have taken 50,000 years to accumulate. This may be a high estimate in that a doubling of the sand drift rate would not be unreasonable, judging by figures given by Fryberger (1979a). This figure is felt to be more realistic than that derived from dune parameters.

SECTION 6.2   THE ZECHSTEIN TRANSGRESSION

Previous Work

Smith (1970, 1979, 1980) has postulated that the transgression which terminated deposition of the aeolian Yellow Sands and brought in the restricted marine Marl Slate may have occurred extremely rapidly, filling in a matter of weeks or months a former basin of internal drainage which had subsided considerably below sea level. The evidence, paraphrased from Smith (1979), centres around three points :

1.   There is good sedimentological and stratigraphical evidence for the existence of a large endoreic desert basin in the southern North Sea area during the Early Permian. Flooding of such a basin would yield a water depth equal to the pre-existing topographic closure.

2.   The preservation of the Yellow Sands as apparently uncemented ridges. The draa would certainly have been eroded by a progressively encroaching tidal sea in a more 'normal' transgression.

3.   The Marl Slate, though thinned by slumping off buried slopes, is "little thinner on the eminences than on intervening lower areas". Lithology, fauna and trace element composition are constant on and between eminencies, with the latter having a consistent vertical distribution suggesting a lack of diachroneity.

Point 3 is open to criticism. Examination of the larger exposures (McCall's Quarry, Hetton Downs Quarry, Sherburn Hill Sand Pit, Bowburn Quarry and Quarrington Hill Quarry) shows that the Marl Slate is considerably thinned, commonly to less than 0.2 m, in places to less than 0.1 m, over the crests of ridges. At Cassop Vale, Elemore and over parts of Ferryhill railway cutting the Marl Slate is absent. These variations are also recorded by Turner et al. (1978) who give a maximum thickness for the member of over 5 m near Rushyford (15 km south of Durham). Some of the variation may be due to slumping, e.g. at Ferryhill where the thickness changes across a gently southward-sloping Yellow Sands surface from zero to at least 1.1 m in a horizontal distance of only 15 m. However at Quarrington Hill Quarry the Marl Slate is thin over a near perfectly flat surface of Yellow Sands; slumping could not have taken place there. There can be little doubt that the relief of the Yellow Sands draa has induced complementary variations in the Marl Slate.

Nevertheless it remains a fact that the Yellow Sands should have been reworked in a normal, slow transgression and that the southern (and northern) Permian basin was closed and endoreic and may well have lain below sea level.

The Evidence of the Sedimentary Structures

Individually, the immediate causes of the various components of the top of the Yellow Sands, described in Section 5.2, are easily interpreted and should not be controversial.

The disturbed cross-bedding is a subsurface feature of the kind analysed by Doe and Dott (1980), developed in sand fully saturated with water by localised and short-lived liquefaction of highly porous sand. It must be post-transgression because the Yellow Sands draa, sitting isolated on bedrock and separated by sand-free corridors, could not possibly have retained anything more than capillary water whilst subaerially exposed. It is possible that the contortion may have occurred spontaneously on wetting, without the usual cyclic loading or earthquake shocks envisaged for initiating liquefaction. The initial porosity of the sands may have been over 45% and the simple replacement of air by water as the pore fluid might have initiated the instability. It is evident from the very gentle nature of all the contortions that the liquefaction was very short lived and therefore probably involved only a very small porosity reduction. Restriction of the contortions to the sand immediately below the top may be explained by porosities below being slightly reduced by very early compaction, or a small amount of infiltrated clay lending a slight cohesiveness to the sand. The planar-bedded zone generally escaped the slumping

because of its much lower initial porosity, which was probably in the region of 40%. That it did not completely escape the disturbance indicates that the contortion occurred after its deposition.

The planar-bedded zone has always previously been regarded as subaqueous in origin (e.g. Smith & Pattison, 1972, p. 70). Its lamination shows it to be aeolian; this is absolutely incontrovertible. There is so much material elsewhere within the Yellow Sands identical in every respect that a denial of the aeolian origin of the planar-bedded zone is tantamount to interpreting the whole formation as being of subaqueous origin.

It cannot be a coincidence that the planar-bedded zone covers 70% of the draa. More specifically, if the transgression had involved large-scale erosion of the draa it is peculiarly fortuitous that that erosion should cease at or within an unusually extensive bed of flat-laminated sand sitting at roughly the same height in every ridge (there is no great difference in thickness of the Sands from one ridge to the next). The lateral extensiveness of the planar-bedded zone must indicate that very little sand has been removed from the ridges : it must have been developed in atypical conditions (there is no other similarly extensive feature in the formation) influencing the whole erg before the transgression. It is possible to envisage a scenario which might lead to this situation, aeolian-wise.

The model for the growth of an erg with incomplete sand cover developed in Chapter 1 indicates that once the draa reached an equilibrium size, deposition in the Yellow Sands erg could have ceased, with sand-drift constant over the Durham area during an indeterminate period preceding the Zechstein.

It is during this period that the upper sand-sheet and related deposits of the Yellow Sands - the planar-bedded zone - must have accumulated. The origin of this unit can be explained using the ideas on pp. 139-143.

As explained previously, the deposition of sand-sheet and associated deposits may be contingent on the presence of a surface creep of coarse sand providing a permeable carpet into which sufficient finer material may penetrate and some to rest. The accumulation of such deposits may be independent of the rate of sand-drift, apart from requiring a non-depositional system for the finer sand (default on this condition would lead to the smothering of the surface creep by large bedforms of medium sand).

Low, flat-topped longitudinal draa at equilibruim in a relatively small erg are thus ideal - indeed inevitable - sites for the accumulation of sand-sheets. They are purely sand-passing bedforms, any superimposed dunes being (on average) non-depositional. This state would promote the first accumulation of a residual coarse-grained armour, slowly spreading over each draa and inhibiting the

passage of dunes. As successive layers of this armour advanced, the interstices between the grains would be filled by a suitably fine portion of the passing sand-drift. The fine lamination of the planar-bedded zone must show that deposition was from ripple-like bedforms, and not from nebulous, inexorably spreading sheets : the laminae of the planar-bedded zone are climbing translatent strata. The interpenetration process most likely takes place at the crests of the generating ripple-like forms.

In addition to the preservation of the planar-bedded zone, the lack of marine erosion of the draa is indicated by the complete absence of substantial quantities of reworked sand. The obvious place to look is in the inter-draa areas, yet here the average sand thickness is only 2 or 3 m. Given that the draa occupy roughly 60% of the area of the erg, this material can only account for a reduction of the draa by a maximum of 2 m.

The possibility that the Yellow Sands bore an early cement which prevented reworking merits a few words. Any cement would most likely have been of gypsum, derived from the dissolution and reprecipitation of clastic gypsum. A possible potential source of gypsum lies in playa sediments in the centre of the northern Permian basin (the well known sabkha evaporites of the southern Permian basin are not a candidate : they do not lie up sand-drift from the Yellow Sands). First of all it is doubtful whether a

mineral as fragile as gypsum could have survived the journey, secondly there is no petrographic evidence to support this idea. Nor is there any evidence at outcrop that the sands were anything but incoherent.

Mass Balance and Palaeogeographical Considerations

These matters, though largely speculative, are important considerations which need to be heeded in any suggested history of the Zechstein transgression.

Palaeogeographies of the North Sea area during the Early Permian show that both the northern and southern Permian basins were closed and endoreic (e.g. Ziegler, 1981). That being the case, any influx of the sea into those basins **must** have produced a very rapid transgression. The very fact that any transgression took place renders it impossible that these closed endoreic basins were anything other than below contemporary sea level. Such a transgression would inevitably have filled the basins to sea level, at a speed constained only by the characteristics of the feeding channel.

That channel is presumed to have led from the Boreal Ocean, the main mass of which lay some 1400 km N. of Durham during the Early Permian (palaeogeography in Ronnevik, 1981). At the last, just one barrier must have lain between the Ocean and up to 1000 km of channelway leading to the North Sea area deserts. Once this barrier was breached, whether by storm, earthquake or natural background erosion, the result-

ing flood would have been constrained only by the size and shape of the portal and the convolutions of the channel.

A possibly simplified version of events can be envisaged whereby the basin is filled via a steep waterfall from the Boreal Ocean, thus avoiding the assessment of any constraints posed by a feeder channel and minimising filling times. In this instance the net inflow is controlled by the mean depth of water at the head of the falls, the width of the falls, and the rate of evaporative loss.

The results of such a model are plotted in fig. b.2 , for a basin of area 10^6 km^2 (measured from Ziegler's (1981) palaeogeography) and guesstimated mean depth of 300 m. The problem now is to guess at the size of the waterfall. A minimum fill time of several years to a few decades seems to be indicated : the analysis serves the purpose of showing that the notion of a complete flood taking just weeks or months (Smith, 1979) is difficult to sustain.

Discussion

Being so far removed from the vicinity of the influx, the waters must have risen quite quietly over the Yellow Sands draa. However, it remains a worry that despite being partially submerged or submerged but above wave base for perhaps several years, the Yellow Sands show no sign of significant reworking : one might certainly have expected at least a few

storms during the period in which wave base was closing over the draa.

Their numerous presence and streamlined nature must have acted to reduce reworking. The rising waters would have been split into narrow, finger lakes, breaking up and baffling wave action. The topmost layers of the Yellow Sands, being formed of sand-sheet and wind-ripple laminae, would have been quite firm and relatively resistant to erosion : such surfaces can readily bear the weight of motor cars (Bagnold, 1941).

Some evidence of current reworking may well once have been present : typically the topmost 0.2 m of the formation is now preserved as bioturbated sand, all original structures having been destroyed.

There factors, acting in concert with the rapidity of the transgression, must have been sufficient to produce the preserved passive draping of the Yellow Sands by the Marl Slate.

For some few years after the transgression the bottom waters of the Zechstein Sea must have remained aerated. This is shown by the presence of the bioturbation and the Lingula - dominated fauna. Much of the thickness of the Marl Slate, however, indicates anoxic bottom conditions. It is interesting to note that in a silled basin, such as the Zechstein, anoxic bottom conditions are indicative of a positive water balance; net outflow of water to the open ocean (Demaison and Moore, 1980; Friedman and Sanders, 1978). This implies that during Marl Slate times, the freshwater influx to the southern Permian basin was

greater than the evaporative loss - surprising given
the previous desert conditions, surviving desert
hinterland and the subsequent evaporite sequence.

The History of Development of the Yellow Sands Erg

The Yellow Sands may be divided into 3 units :
(i)a lower flat-laminated zone, discontinuous but up
to a few metres thick; (ii) a cross-bedded unit,
occupying most of the formation, characterised by
trough-shaped sets 3 - 6 m thick and 30 - 50 m wide;
and (iii) an upper flat-laminated zone, laterally
extensive and up to a few metres thick.

These units represent (i) the initiation of
longitudinal draa as small dune-less sand-sheets and
the piling up of sinuous transverse dunes; (ii) the
growth of the draa into flat-topped structures c.30 m
high by the accumulation of sand from sinuous
transverse dunes migrating along the length of the
draa under a bimodal wind regime; and (iii) the
stabilisation of the draa as they reached equilibrium
size under a metasaturated sand-drift. These
processes (illustrated in fig. 6.3) may have
occupied several tens of thousands of years. An
excellent modern analogue of the Yellow Sands at this
stage is provided by the "Whalebacks" described by
Bagnold (1941, pp.230 - 231 and Pl. 13). These are
longitudinal draa described as "continuous plinths 1
to 3 km in width and perhaps 50 m high running in
straight lines for distances of the order of 300 km"

(ibid., p.230). The draa typically are 5 km apart, and most significantly bear widespread duneless sand sheets as well as areas partially or completely covered by dunes. Quoting Bagnold again (p.230 - 231),

"The general appearance, internal structure," (he does not mention digging any pits; this appears to be speculation; see also ibid., pp. 242 - 243) "and grain grading" (no figure provided) "of these huge formations suggests strongly that each is the residue which has been left behind by the march either of a single dune chain of a far greater size than any now existing, or of any age-long procession of successive chains of one present size which have travelled down the same line". These draa occur in the Great Sand Sea of Western Egypt. Their similarity to the envisaged interpretation of the Yellow Sands is remarkable, even to their having a surface of sand-sheet material.

The endoreic, sub-sea level basin in which the Yellow Sands lay was then engulfed by the invading waters of the Boreal Ocean. But for the rapidity of this transgression, which was over in a matter of several years to a few decades, the Yellow Sands would not have been preserved. Their position, as isolated, stable and static bedforms founded on a rock pediment, would normally have given them very little chance of survival into the geological record.

CHAPTER 7

THE LAMINATION, GRAIN SIZE CHARACTERISTICS AND

PETROLOGY OF THE YELLOW SANDS

SECTION 7.1 LAMINATION AND GRAIN SIZE CHARACTERISTICS

Introduction

Brief general descriptions of the distribution of
lamination within the Yellow Sands were given in
Chapter 5.   The following comprises a more
detailed and quantitative account of the lamination
of the formation together with its relationship to
grain size characteristics.

Lamination

Neglecting Ferryhill railway cutting, where the
lamination is generally obscure due to adverse weather-
ing, an assessment of the lamination type was made at
246 points in the Yellow Sands. It must be stressed
that these assessments were entirely estimates - no
measurements of proportion were taken. However, it
is hoped that the information collected is representative
and will provide some general indication of the ordering
of lamination within a formation of aeolian sand.

The distribution of wind-ripple laminae (labelled
as type 2) and sandflow laminae (type 3) with respect
to cross-bedding dip is shown in fig. 7.1. The diagram
also illustrates the scheme of notation used for re-
cording the relative proportions of each type of
lamination at any spot. The tailing off of recordings
below 12° is an artefact of the sampling bias mentioned
on page 127   . Sandflows, whether dominant or sub-

ordinate, are confined to angles of dip between 20° and 34° with little detectable variation across the graph. Wind-ripple lamination may be found at any angle of dip and makes up all the sand dipping at angles below 20°. Sand-sheet material is not marked on the diagram but always dip at angles less than 10°.

Except for the sand sheet at the unconformity at North Hylton and the planar-bedded zone there appears to be no systematic variation of the lamination either vertically or laterally within the formation. Adding all the data together reveals that 71% of the Yellow Sands is made up by wind-ripple laminae, 25% by sand-flows, and 4% sand-sheet material.

The Grain Size Data - Description

The grain size data comprises 9 samples sieved in a half-phi stack and 38 at a quarter-phi, all from -1 to 4ϕ. Each of the latter was done in two stages of 10 minutes each on a standard shaker. Only the least coherent samples were selected for sieving. These required only the gentlest of disaggregation which was accomplished by tentative pounding with the butt of a hickory hammer shaft or simply by finger pressure with the specimens in cardboard trays. Each fraction was weighed to 0.01g and checked for spurious grains under a binocular microscope. Aggregates were present in the coarsest few fractions (seldom more than 3) of about two-thirds of the samples. The proportion was estimated by eye and the appropriate corrections made. In no case were fines generated by

disaggregation observed, except for concentrations of fine silt-grade dolomite and iron oxide in some pan fractions.

The accumulated data was plotted as histograms, log-log frequency curves, arithmetic cumulative curves and probability cumulative curves. The standard graphical measures of mean size, sorting, skewness and kurtosis were derived from the arithmetic cumulative curves in the normal manner.

The statistical measures are listed for all samples on Table 7.1 along with the exact location, lamination type and cross-bedding dip. 20 samples were selected for graphical display in figs. 7.2-7.11. Cross-plots of mean v. sorting (fig. 7.12) mean v. kurtosis (fig. 7.13) and sorting v. kurtosis (fig. 7.14) are also presented. Three of the samples bore a sand-sheet type lamination, 18 were from sandflows, 21 were of wind-ripple laminae, 4 apparently structureless and one structureless, probably reworked during the Zechstein transgression. Whilst the wind-ripple and sand-sheet samples are obviously composites of many laminae an attempt was made to ensure that the sandflow samples were each from a single lamina unit. Collection of these was made either with a teaspoon, the point of a trowel or by carefully trimming a slightly coherent block.

The immediate impression gained from the graphs and statistics is the great diversity of the grain-size characteristics, within the constraint of it all being sand. Mean size ranges from 0.23-2.49 ϕ (average

1.78 $\pm$ 0.49 ρ), sorting from 0.29-1.29 ϕ (average

0.67 $\pm$ 0.26 ρ), skewness from -0.22-1.49 (average 0.16

$\pm$ 0.27) and kurtosis from 0.6-2.27 (average 1.07 $\pm$ 0.39).

Two are tri-modal, 15 bimodal, the rest unimodal.

In the author's opinion the Bagnold-type or log-log graphs are the most informative in investigating the various systematic tendencies and characteristics. More importantly, the curves strongly support Bagnold & Barndorff-Nielsen's (1980) contention that the log-frequency curves of natural grain size distributions are of hyperbolic form and thus not Gaussian. The majority of the log-log graphs show curves with straight limbs, whatever the lamination of the sample. Only a few are parabolic, with limbs convex upward (eg. fig. 7.2a) suggesting conformity to a Gaussian distribution. The most perfect hyperbolae (eg. figs. 7.2b, 7.3a) have a slight sigmoid form on the cumulative probability curves, a form which might be interpreted as consisting of 3 linear segments in the manner of Visher (1969). Because the apparently standard or median form is non-linear on the probability plots this mode of presentation is considerably devalued. Only particularly well marked tendencies should be trusted as being real.

The principal value of figs. 7.2-7.11 and the associated statistics lies in their confirmation of systematic differences between the grain size characteristics of the 3 types of lamination.

The sandflow curves (lamination type 3 on Table 7.1) typically are leptokurtic (average 1.31) with a low standard deviation (average 0.48). Some sampling bias must be admitted, coarse-grained sandflows being over-represented. These are more eye-catching at outcrop than their medium-grained equivalents and often totally incoherent or an unusual colour, further inducing sampling. Often the sandflows have a subsidiary fine mode or shoulder (eg. figs. 7.3b, 7.4a), this lying at 2.5-3.0 ϕ whether the sandflows be coarse- or medium-grained. This fine tail accounts for the high positive skewness values obtained and the dog-leg shapes of the probability plots. It seldom exceeds 2% in any fraction and typically comprises 5% of the sample. It is possible that it represents inadvertent contamination of the samples by the very fine-grained laminae which bound and define sandflows. These are . prominent in fig. 3.2 . Other sandflow curves lack this tail but nevertheless retain an asymmetry with the slope on the fine side less than the coarse. Since Bagnold (1941, pp.127-143) devotes considerable attention to these slopes their values have been computed. They are regarded as being related to the selectivity of the transporting process with respect to coarser and finer grains (Bagnold & Barndorff-Nielsen, 1980). Note that values taken directly from the graphs in this thesis need to be divided by log $\frac{1}{2}$ before comparison with Bagnold's (1941) figures. This is because Bagnold measured grain diameter in logarithms to base 10;

phi values are logarithms to base $\frac{1}{2}$. This treatment
has been applied to all figures quoted in the following.
The fine limit of the sandflow curves ranges from
-3.5 to -9.5 (average -6 $\pm$ 2), ignoring the subsidiary
fine modes and shoulders. The slope of the coarse
limit varies from 5 to 23.5 (average 11 $\pm$ 5).

By development of a flat or angled crown to the
distribution (eg. figs. 7.4b, 7.5b) or by an increase
in the magnitude of the fine tail (eg. fig. 7.5a) the
sandflow curve form merges into and overlaps with that
of wind-ripple laminae. These are characterised by
distributions which are platykurtic (average 0.88),
moderately or moderately well sorted (average 0.81)
and with a skewness variable but tending to be positive
(average 0.13). These figures are manifest as a
tendency to bimodality, varying from obviously so
(fig. 7.6a, 7.6b), to curves with a very flat crown
falling away abruptly on either side (figs. 7.7a, 7.7b),
to broad, sensibly unimodal distributions (figs. 7.8a, 7.8b).
The modes are usually equally represented but may be
extremely biased, as in fig. 7.9a and fig. 7.5a. Fig.
7.9b is tri-modal. More often than not the two modes
obviously, or may be inferred to peak at 1.75-2.0 ϕ and
2.75-3.0 ϕ . Slopes of the curves range from 4 to 18
(average 9.5 $\pm$ 3) on the coarse side and -2 to 15
(average -4 $\pm$ 1) on the fine side.

The curves from sand-sheet samples (figs. 7.10a,
7.10b) are markedly bimodal with the two modes at 0.5-1
ϕ and 3 - 3.75ϕ . The trough between bottoms at

between 1.75 ϕ and 2.25 ϕ . A gradation into a wind-ripple type curve may be traced through figs 7.11a and 7.6b. The proportion of each mode varies only from 60-40 to 40-60. The curves are all somewhat truncated at the fine end and the pan fractions (ie. finer than 4 ϕ) were indeed noted as containing a high proportion of quartz. However, the silt content is not more than 2% in any case. Because of this truncation slopes at the fine end cannot be determined. The slope of the coarse tail varies from 6 to 7.5.

Of the samples of apparently structureless sand four are from Hetton Downs (two from the old pit, including the possibly reworked sample) and one from Sherburn Hill west face. The examples are all well or moderately well sorted (average 0.53 $\pm$ 0.07 ϕ), symmetrical and mesokurtic (averages 0.06 $\pm$ 0.06 and 1.03 $\pm$ 0.08, respectively) medium or fine sand.

Turning now to the cross-plots (figs. 7.12-14), skewness is absent from the list of parameters used because of its ineffectiveness in discriminating the lamination types. The mean v. sorting (fig. 7.12) graph shows well the wide range of grain size and narrow range of sorting of the sandflows. The wind-ripple laminae are less restricted in sorting, more so in grain size and there is considerable overlap with the sandflows. The structureless samples fall within this overlap while the sand-sheet types are just distinguishable by poorer sorting. The mean v. kurtosis graph (fig. 7.13) produces a similar poor segregation.

Sorting v. kurtosis produces the best separation and is drawn in fig. 7.14. There is, however, still overlap. The grain size distributions presented should be regarded as showing a continuous spectrum of parameters, not as four discrete categories.

Since the orientations of the cross-bedding was noted at 28 of the sample points various parameters were plotted against amount of dip. The graph showed no detectable trends and this is not surprising given the limited data base and unsystematic sampling. Many specimens from a small number of cross-bed sets would be needed for such an undertaking. Also, given the present data there are no systematic tendencies either vertically or laterally within the Yellow Sands.

The Grain Size Data - Deductions and Interpretation

The continuous variation of grain size parameters from one lamination type to another is amply demonstrated by the cross-plots (figs. 7.12-14) and the sequences figs. 7.15 and 7.16. These continua, from sand-sheet to sandflow, developed in similar overall grain sizes show that the individual distributions come about by the action of two different sorting processes or sets of sorting processes acting on the same raw material. One process occurs during deposition from ripples and may develop into a sand-sheet type if circumstances permit. Its characteristic feature appears to be a tendency to develop bimodality. This supposition could only be tested by systematic sampling in a modern desert. How the bimodality develops is not clear, as

discussed on pages *81-82*. It may be the gradual influence of interpenetration sorting or perhaps a segregation of the saltation and surface creep fractions occurring during ripple movement. The process must begin acting as soon as ripples develop on sand newly arrived on the leeside of a bedform, and continue as sand leaves the bedform and migrates up the back of the next downwind. At the brink of this bedform the ripple sorting is destroyed by the grainfall process. The finest grains might travel far out over the slipface in a state of partial suspension, the coarsest grains will dribble passively over the brink. This segregation is potentially very efficient and is enhanced as the sand becomes unstable and avalanches. The result of these processes is the unimodal, well or very well sorted sandflow curve.

These processes are not specific to the Yellow Sands data. The only peculiarity of the formation is its content of coarse and very coarse sand in quantities sufficient to develop sand sheets from time to time. This may be attributed to the small size of the erg (potential sources of coarse sand at no great distance), the incomplete sand cover (potential source of sand immediately adjacent to each draa; the sand-free corridor might also provide a hard pavement enhancing coarse sand transport) and most of all to the longitudinal form of the draa (the erg was sand-passing, not sand-trapping, facilitating the penetration of coarse grains).

Finally, a note about grainfall laminae must be

added following a paper by Kocurek and Dott (1981)

published in the interval between the writing of the

descriptive and interpretive parts of this section

(see also Introduction, p.xx). These authors

confidently identify grainfall lamination in ancient

aeolian sandstones though it is subordinate to wind

ripple (their climbing translatent stratification) and

sandflow (their grainflow) types. Judging by the

photo on p.596 in Kocurek & Dott, if a mistake has

been made in the present study and grainfall laminae

overlooked, it has probably been labelled as very fine

and faint wind-ripple lamination. In terms of grain

size characteristics, grainfall deposits, on the basis

of the interpretations developed in the foregoing,

might be expected to have values of kurtosis and sorting

intermediate between those of wind-ripple and sandflow

type.

SECTION 7.2 PETROLOGY OF THE YELLOW SANDS

Introduction

The diagenesis of desert sediments, especially with

regard to the origin of red beds, has received a great

deal of attention in recent years (e.g. Walker, 1975,

1979; van Houten, 1973; Turner, 1980). Many new facts,

theories and interpretations have appeared in the

literature. A description of the petrography and

diagenesis of the Yellow Sands must benefit from and

be of benefit to these recent advances.

The petrology of the formation has been the focus of

only a limited amount of attention over the years.

The most recent accounts are those of Pryor (1971b)

and Glennie (1970, p.188-190). Pryor's work was fairly exhaustive, though carried out in the context of the formation having a shallow marine origin. Glennie was principally concerned with an explanation of the colour of the Yellow Sands.

Briefly, the present work has revealed that the diagenesis of the formation comprises syn- and post-depositional pigmentation, the precipitation or infiltration of small euhedral dolomite rhombs, the development of quartz overgrowths and subsequent partial cementation by calcite. Minor authigenic phases are pyrite, kaolinite and barytes.

Methods

The incoherence of the formation presents an obstacle to any examination of its petrography. In the present study samples were vacuum-impregnated with blue epoxy resin to facilitate thin-sectioning. Various resins were used with slightly differing viscosities and setting times. None had any great influence on the degree of penetration, which was normally good in samples which typically had high porosity and permeability with wide pore throats. The material in current use is Ciba Geigy Araldite resin MY778 with 3 parts in 10 of Permabond hardener E27B. The dye used is Dupont's Oil Blue A; 15g in 1kg of resin. The recipe is derived from one given by van den Berg (1973). Most thin sections were point counted, 300 points per slide, classified as straight quartz, poly-quartz, undulose quartz,

dolomite, feldspar, calcite, porosity, pigment, authigenic quartz, and rock fragments. Where appropriate and possible kaolinite, barytes and secondary porosity were also counted. Grain shape was assessed visually.

X-ray diffraction analysis was used extensively in the determination of cement types and the composition of the various pigmenting and grain coating phases. Concentrations of grain coating minerals were obtained by putting 20-30g of sand in water in an ultrasonic cleaner for 15-30 mins. This agitates the grains and removes part of the clay pellicles into suspension. The water may then be decanted and the clays filtered out. Discrete particles of supposed iron oxide adhering to grain surfaces presented a problem, these being too coarse to go into suspension in the ultrasonic cleaner. It was discovered that many of these were dislodged during sieving and accumulated in the pan fraction. When this occurred the samples were set aside and analysed by XRD.

The examination of hand specimens and loose samples under a powerful (x100) binocular microscope was found to be extremely valuable for studying grain surface texture, clay pellicles and other pigmenting material.

Twenty samples were analysed by X-ray fluorescence, principally to determine their iron content. Specimens were also examined under SEM, the instrument used being a Cambridge Stereoscan 600. This had an unsatis-factory performance, supposedly being capable of

magnifications up to 5OK but in practice could not be focussed at greater than 5K. It was not, unfortunately, fitted with an EDAX unit.

Colour

In respect of the colour of the formation, the fundamental observation to be borne in mind is that the sands are yellow only at outcrop (fig. 5.6). Subsurface they are grey and bear pyrite. Fig. 7.17 shows part of the cores from N.C.B. offshore boreholes D1 (NZ 530521), D8B (NZ 497526) and D2 (NZ 481504). A specimen of Basal Breccia from N.C.B. borehole Mainsforth D (exact location unknown; Mainsforth is 1km SW of Ferryhill railway cutting) has clasts of red limestone, siltstone and sandstone in a red matrix.

In the southern North Sea, Germany and Poland the bulk of the Rotliegendes sequence is coloured varying shades of red with the uppermost parts grey.

Using a Munsell Soil Colour Chart 38 colour determinations were made on hand specimens of the Yellow Sands. The specimens had not been sieved or treated in any way except for thorough drying at room temperature. Many specimens were mottled in some way, whether due to bioturbation, weathering or cementation and no determination was attempted. Others were very pale creamy yellow or very pale yellow in colour, ranges where the Munsell Chart becomes extremely subtle from one colour chip to the next, allowing equally convincing comparisons across a wide spread

of Hues and Chromas. These specimens were therefore also omitted from the survey. The same problem and consequence was encountered with grey and pale grey specimens both from subsurface and outcrop. These omissions represent only a small proportion of the formation at outcrop, and though unfortunate, are not thought to devalue the results presented to any great extent.

The numerical determinations made are listed in tabular form in Table 7.2. To summarise, the Hue of the majority is 10YR or 2.5Y; the Values (lightness) of all but 2 are 6 or 7 indicating a fairly light colouration, and the Chromas (strength: departure from neutral of the same lightness) of all but 3 are 4, 6 or 8. In words, one sample (from the prominent red areas at Bowburn Quarry, fig. 5.25) was red, 5 light red, 5 brownish yellow, 5 very pale brown, 14 yellow, and 8 pale yellow. Thus we are delighted to be able to confirm that the name of the formation is quite accurate and need not be changed to, for instance, the Light Yellowish Brown Sands.

Framework Grains

The Yellow Sands of the main outcrop may be classified as subarkoses, in the scheme of Pettijohn (1975, p.211), with some samples straying into the quartz arenite and sublitharenite fields (fig. 7.18). Two samples from near Leeds had 21% and 31% rock fragments, the latter therefore overlapping into the lithic arenite field. These results confirm Pryor's

(1971b) work, though he used a different nomenclature. Clays and pigmenting material were omitted from these measurements since they are regarded as largely authigenic in origin.

The quartz is mostly unstrained and unicrystalline (averaging 60% of the clastic components) with about 20% polycrystalline and 5% undulose. A wide variety of polycrystalline types are present and include the occasional ganister fragment (C.J. Percival, perc. comm.).

The suite of rock fragments is diverse but dominated by sedimentary types, a conclusion also reached by Pryor (1971b). These are chiefly fine, angular clay cemented sandstone and siltstone with chert also common. Shale clasts, often deformed, are rarer and concentrated low in the formation with identifiable limestone fragments exceedingly scarce. The sandstone fragments are often cemented by opaque material which is a rusty orange colour in reflected light. They may also be at the centre of developments of similar material acting as a cement to the surrounding few grains. All the sandstone grains are prone to disaggregation and destruction (figs. 7.19, 7.20).

Igneous rock fragments are sparsely distributed and when manifest as green (presumably yellow but for the blue resin) or brown aphanitic blobs only dubiously identifiable. The most interesting and prominent type is a fine to medium grained basic rock consisting of needle- to lath-like feldspars in an opaque (rusty)

groundmass. One or two grains of this are present in about half the slides examined (fig. 7.21). Occasional grains of microgranite were also seen.

Metamorphic types are rare and principally composed of quartz schist or meta-quartzite.

Feldspars are quite common in the Yellow Sands, comprising 20% of the clastic grains in one offshore specimen and averaging 9%. As usual, alkali types are very much in the majority. They vary from appearing very fresh to mere ghosts identifiable only because the sections are impregnated; some extremely delicate textures are preserved. Degrading feldspar may be seen in fig. 7.22.

The heavy mineral population of the Yellow Sands has fairly recently been described in great detail by Pryor (1971b). All that need be said here is to affirm the dominance of garnet, tourmaline and zircon amongst a sparse population. Garnets up to 0.75mm in size were seen; tourmaline and zircon are always medium or fine and well rounded.

Grain shape varies from well rounded in the coarser modes to angular in the finest. Only at the very base of the formation does the coarse mode become partly angular.

The modification of shape is completely gradational from one extreme to the other. The transition from subangular to subrounded grains occurs at about the medium-fine sand boundary, from say 1.5 ϕ to 2 ϕ . Very often feldspar grains show the best rounding in their size class, presumably reflecting a lower resistance to mechanical abrasion than the quartz.

Pigmenting and Associated Minerals

Being predominantly yellow or grey the Yellow Sands
are something of an oddball in the context of our
present perception of continental red beds;
one would expect them to be red throughout.

The most effective means of investigating the
pigment were found to be by examination of hand
specimens or loose samples in reflected light under
a binocular microscope (the immense value of this
procedure cannot be over-emphasised), and in reflected
light in thin section on a standard petrological
microscope. The most convenient set-up for the
latter process was to have a spare microscope light
in a clamp stand shining onto the thin section beneath
which a piece of white card had been placed. Thus
the true colour of the various minerals could be
determined and the expenditure of time and money
involved in manufacturing special polished sections for
episcopy was avoided.

The yellow colour of the formation has always
previously been attributed to a coating of limonite
on the grains (eg. Smith and Francis, 1967, p.97).
According to Deer, Howie and Zussman (1962), limonite
consists principally of cryptocrystalline geothite
or lepidocrocite with adsorbed water and the possible
presence of haematite. "The name limonite is now
retained as a field term or to describe hydrated oxides
of iron with poorly crystalline characters whose real
identity is not known." (ibid., p.444).

Hand specimens of outcrop samples viewed under the
microscope show 4 components to the colour of the
formation and these may be verified, though with less
facility, in thin section. They are:

(i)   A yellow staining over the entire surface of
most grains. It cannot be removed by scratching the
grains with a needle but is destroyed by boiling in
HCl. Its presence becomes obvious when an otherwise
apparently clean Yellow Sands grain is compared with
truly clean sand from another formation or cleaned
Yellow Sands grains. It is conceivable that it is an
optical illusion and merely the reflected and refracted
image of other yellow material elsewhere on the grain
or on other grains, but this is thought most unlikely.
Even when other pigmenting materials (see below) have
been mechanically removed this stain remains. It has
no identifiable material cause in either thin section
or SEM. Its composition must therefore remain an
enigma.

(ii)   A very thin and delicate (easily removed with
a fine needle) pellicle of translucent yellow material
evenly covering most grains though less well developed
on coarse modes. The coating is probably only 1 or 2μ
thick and because of this is not nearly so apparent in
thin section, on whose evidence the grains might be
adjudged to be much cleaner than they actually are.
It has, or enhances, a matte, frosted texture and is
absent at grain contacts which in hand specimens appear
as circular, elliptical or irregular areas where the
adamantine lustre of a fresh quartz surface is often
visible. The material appears to be very finely
crystalline and often develops into a slight ridge
around grain contacts.

(iii) Yellow or yellow-orange very fine-grained
(beyond the resolution of the microscope) material
concentrated in hollows and re-entrants on grains of
all sizes, though concentrated on finer modes where
such sites are more common. It is more strongly
coloured than (ii) and opaque rather than translucent.
It also forms small patches and wisps over exposed
parts of grain surfaces, though seldom covers more
than 20% of any grain except in the finest fractions
and most strongly coloured samples. It has the same
lustre as (ii) which may be an inherent property or
the result of it being covered by a thin film of (ii).
In thin section the material appears green where
backed by the blue resin used to impregnate the samples.
Its colour and grain size prevent the determination of
any optical properties though any birefringence it may
have appears to be very low. Its predilection for
hollows in grains is emphasized in thin section where it
lies in lenses typically 10μ thick. Over exposed
parts of grain surfaces it is $1-3\mu$ thick and is present
at some grain contacts, absent at others. It bridges
close approaches of grains and has sporadic suggestions
of a meniscus texture. It completely or partially
replaces many rock fragments; in basalt fragments it
occupies part of the areas between the feldspar laths,
presumably as an alteration product of pyroxene. One
or two areas may be seen in any one slide where a circle
or part circle of the material lies suspended in resin
recording the former presence of a now totally dissolved
grain.

(iv)   Granules, aggregates and accumulations of material ranging in colour from orange through shades of red and brown to black. Individual granules take the form of degraded rhombs, cubes or spheres or are less regular and from $5-40\mu$ in size. They adhere to grain surfaces individually or in accumulations up to 200μ across and show no special preference for re-entrants in grains. Seen in thin section, this material is red, orange or black in reflected light, opaque in transmitted light. Single particles lie on grain surface, wedged close to contacts or between grains. Accumulations and relatively thick $(15\mu$) films may develop on single grains or lying across two or more grains. It may fill pores and in such cases generally extends as a partial cement across several surrounding grains. Discrete and undeformed clastic grains also occur. It is found in and around rock fragments and former rock fragments, often in association with (iii). Individual granules may be situated in voids developed along the cleavage of feldspars. Of these four types or textures, (i) and (ii) are most important in colouring the formation, with (iii) a strong influence. The descriptions are illustrated in figs. 7.19, 7.21, 7.28-31.

Most of the variegation at outcrop occurs by variation in the development and relative proportions of these 3 components. In the red lenses at Bowburn Quarry (i) - (iii) are red; (iv) does not vary. The pale grey sands seem either to be free of all pigmenting

material or to contain only a white or translucent form of (ii).

All these components, colours and features must be fairly recent phenomena; merely an alteration and weathering product of the grey colour found at depth. In borehole samples direct counterparts of (ii)-(iv) above are found with very similar textures and a comparable range of development ((i) is either not present, or colourless and not detectable). The differences are that (ii) is white or colourless and transluscent and often better developed or more crystalline. Concentrations of material in hollows on grains comparable to (iii) similarly are white and transluscent. The counterparts of (iv) above are crystalline, metallic accumulations principally and visibly consisting of pyrite, resting on or between grains.

The deduction of the composition of these various components was principally based on X-ray diffraction analysis. The principal problems encountered lay in deciding which material was segregated by the individual concentration methods utilized (ultrasonic cleaning, sieving off the pan fraction, heavy mineral segregation). The filtrate derived from ultrasonic cleaning of both outcrop and borehole samples always gave very strong (usually off the scale) kaolinite peaks with indications of illite and mixed layer clays, strong quartz and feldspar, strong dolomite in some samples, none in others, and weak indications of goethite in about half the samples. This probably comprises both (ii) and (iii).

Pan fraction samples (outcrop only) produced all these peaks with dolomite, quartz, feldspar and goethite more frequent and better developed, kaolinite much weaker. These samples are relative concentrations of (iv). The pan fraction of red sand from Bowburn gave very good goethite lines but no indications whatsoever of haematite. This lack of haematite was common to all specimens.

SEM examination shows (ii) and (iii) to have a superficial textural similarity, supporting the suggestion of a kaolinitic composition for both. Also in SEM, (iv) may sometimes be seen to have excellent framboidal forms indicating derivation from pyrite. A selection of SEM pictures are shown in figs. 7.32 -7.38. Heavy mineral separates (neglecting obviously clastic species such as zircon) contain pyrite in subsurface samples and traces of goethite from outcrop. The goethite peaks on XRD are best developed in the red sand from Bowburn, subdued in normal yellow sand. This suggests that the iron oxides in the yellow samples are less crystalline than those in the red with the subdued XRD peaks either resulting from this directly or because the less crystalline material is less dense (goethite has a specific gravity of 4.3, limonite varies from 2.7 to 4.3, according to Deer, Howie and Zussman, 1962).

Thus pigment types (ii) and (iii) are identified as iron-stained kaolinite and (iv) as a mixture of geothite and amorphous iron oxides derived from pyrite.

Type (i) is a perplexing problem, being a most wide-spread but least tangible phase at outcrop. Worse still, it has counterparts in both the Bridgnorth and Penrith Sandstones (see Chapters 8 and 9). Further discussion of its possible nature is given on pp..

Table 7.3 shows part of the results of 17 X-ray fluorescence analysis of whole-rock samples of the Yellow Sands, numbers 1-9 being subsurface specimens and 10-17 from outcrop. The iron content is expressed as Fe_2O_3 which should be fairly accurate for the outcrop samples. In the subsurface specimens the iron is most likely to be in the form of FeO. To convert the Fe_2O_3 figures to FeO they must be multiplied by 1.113. Those subsurface samples containing more than 1% 'Fe_2O_3' also contain sulphur indicating the presence of pyrite. Not enough analyses have been made to properly compare the iron content of surface and sub-surface samples, but the figures do not deny the expected equality. Sample 11, a yellow sand from Sherburn Hill, demonstrates that as little as 0.42% Fe_2O_3 is needed to colour the rock, while sample 12, also from Sherburn Hill, shows that the rock can contain 0.47% Fe_2O_3 whilst remaining white or pale grey. The red sample from Bowburn (2.5 YR5/6), number 16, has 1.9% Fe_2O_3, an intermediate value.

The MgO and CaO values reflect the calcite and dolomite content of the rock. It is interesting that numbers 1,9 and 15, each containing over 12% CaO also have relatively high iron contents, suggesting

some correlation and that the calcite may be ferroan.
However, in the dozen thin sections treated with
potassium ferricyanide the calcite never took a stain,
indicating an iron content of less than 0.5% (Dickson,
1972).

Al_2O_3 and K_2O values measure the feldspar and
clay content of the samples.

Authigenic Dolomite

Dolomite (identified by staining and XRD) is a
common authigenic phase in the Yellow Sands, though
never in a form that could be described as cement. It
is developed as euhedral rhombs generally 20μ in size
(ranging from $5-50\mu$) lying on grain surfaces, forming
a complete monolayer over every grain, except at
contacts, where most abundant. The rhombs are generally
discrete and separate but occasionally where particularly
fine and well developed form a contiguous cover giving
a conspicuously sugary texture in hand specimen under
the binocular microscope. For illustrations of dolomite
textures see figs. 7.37, 7.39 and 7.40. Silt grade
euhedral dolomite in the same habit is involved in the
bioturbation immediately beneath the Marl Slate, filling
burrows (see figs. 5.20, 5.32), forming mottled sand-
stone and in rare instances acting as a true matrix.
At outcrop much of the formation is free or contains
only small quantities of dolomite which is concentrated
close to the top. In the borehole cores it occurs in
most specimens, though with core recovery as sparse as
it is this may not be significant. The dolomite occurs

within both quartz overgrowths and sparry calcite and therefore pre-dates these features (fig. 7.41). Subsequent disolution of the dolomite has occurred, as mentioned in the subsection on porosity.

In some slides dolomite bears pigment, usually a covering of type (iii) (fig. 7.42). SEM observations generally show the dolomite as having the same clay coatings (type (ii)) as the clastic grains.

Grain Overgrowths

Authigenic overgrowths on clastic grains are either absent or extremely small throughout most of the formation. They are best developed close to the Marl Slate though even there constitute only 1% or less of the rock. The overgrowths occur as small syntaxial crystal faces only on quartz and are never more than about 30μ thick. Greater developments occur in the Ferryhill exposure, the overgrowths here not being restricted to the top and constituting up to 2% of the rock (figs. 7.22, 7.41). The overgrowths generally avoid areas of grain surface which carry pigment, and are themselves free of pigment of any type.

Calcite Cement

The major lithifying cement both subsurface and at outcrop is sparry non-ferroan calcite, often poikilotopic, with crystals reaching 20mm in size. The broad pattern of cementation within the formation is best illustrated by borehole cores. Core recovery is usually limited to sand within a few metres of the top and bottom of the formation, the middle part being lost due to its

incoherence (Clowes, pers. comm.). At outcrop the variation on this theme is evident.

The most widespread form of the calcite is as localised patches of one or a few crystals to give cemented nodules of spherical or ellipsoidal shape 5-20mm in diameter. These are visible in figs. 5.12, 7.42 and 7.43. This habit is distributed widely and randomly through the formation and is most obvious when concentrated in particular zones which may be from 5 to 50% cemented in this manner and several metres in dimension. Even where the nodules are not evident a very small amount of cement is usually present. The cement shows no preference for any grain size, bedding type, or stratigraphic or geographic locale. The nodules are the only objects ever screened out of the sand before dispatch from the active quarries.

The calcite also has various degrees of development in more densely packed and homogeneous forms. In bimodal wind-ripple laminated sand in the floor of the NW end of Bowburn Quarry the calcite preferentially cements the fine-grained laminae, here forming 14% of the rock (measured by point-counting). The cement extends no more than a metre or so vertically.

Coarse-grained sandflow laminae commonly gain a preferential cement for part of their length (generally only 0.1-0.3m), especially immediately below the overlying set. The calcite here forms a network of intersecting, euhedral, poikilotopic rhombs cementing 50-100% of the rock. This habit is illustrated in

fig. 7.44 at outcrop and fig. 7.45 in thin section. The considerable minus-cement porosity of the latter suggests that the calcite may be displacive.

Occasional well cemented horizons within the formation may even merit description as rock. These generally follow the lamination, are sharply bounded and in dimension extend several or many metres laterally, a metre or less vertically. Apart from being particularly well developed in Ferryhill railway cutting there appears to be nothing else systematic about their distribution.

In thin section there is frequent local evidence of calcite dissolution manifest as lamellar pores (fig. 7.26), corroded quartz grains without cement and disconnected fragments of former poikilotopic crystals (fig. 7.27). However, the presence of perfect euhedral crystals 20mm long in the clusters such as those in fig. 7.44 and the coherent restriction of the cement to particular beds at several localities militates against any extensive decalcification.

The calcite is aggressive towards quartz in some specimens (fig. 7.46), even acknoledging McBride's warning on the interpretation of quartz-calcite relationships (Jonas & McBride, 1977, p.78). In rare cases it is displacive (fig. 7.45) but there is no evidence for wholesale corrosion of the quartz within or without areas preferentially cemented by calcite. Marzolf (1976) links the surface textures of sand grains from the aeolian Navajo Sandstone in Utah to

a possible former calcite cement. Applications of this method to the Yellow Sands is hampered by the possible effects of the authigenic kaolinite and dolomite. However, samples examined under SEM (boiled in HCl to remove carbonates and pigment) showed none of the features depicted by Marzolf. Some of those features seen are illustrated in figs. 7.47, 7.48. Pryor (1971b) regarded all the surface textures he found in the Yellow Sands as being of diagenetic origin. More recently, Krinsley & Smith (1981) studied Yellow Sands surface textures and came to the conclusion that original abrasion and later diagenetic features were roughly equally represented.

Textural relationships visible in thin section generally quite clearly show that the calcite post-dates the pigment, dolomite, quartz overgrowths and all other authigenic phases (figs. 7.22, 7.26, 7.41). There are exceptions to this rule, for instance where partially dissolved calcite crystals have a coating of pigment on some surfaces. Also, a specimen from 0.1m below the top of the formation near face H at Quarrington Hill Quarry, bearing a nodular calcite cement, has much less dolomite within the calcite than without. Dolomite is present, though sparse, within the calcite, abundant without. This would suggest that either there are two phases of dolomite formation or the calcite formed at roughly the same time as the dolomite.

Other Minerals.

The only other minerals seen in thin section were barytes and vermicular kaolinite. *Barytes* was found in two specimens from Sherburn Hill and two from Hetton Downs. Its most interesting habit occurs in a specimen of the top of the formation from a loose block below the west face of Sherburn Hill Sand Pit. The top surface here has a scattering of 1-2mm spherules of radiating barytes nucleated on and enclosing sand grains. The rock is dolomitic but no dolomites occur within the barytes, the spherules appearing displacive towards both quartz and carbonate (fig. 7.49).

The Hetton Downs sample containing most barytes (5%) lacks any pigment and is also cemented by calcite, though the calcite has been partly dissolved and every grain is surrounded by lamellar pores. Grains within the calcite are corroded. The barytes has crystal terminations against the calcite by which it is completely surrounded. Grains are also corroded within the barytes, which shows no signs of dissolution (fig. 7.26).

Vermicular kaolinite is present in all 4 thin sections taken from the Ferryhill outcrop, averaging about 3% of the rock. A tiny amount was seen in one slide from Sherburn Hill. It occurs in patches up to 300μ across and is usually closely associated with a partially dissolved feldspar grain (fig. 7.22). Alternatively it may be partially or wholly enclosed within a pellicle of pigment or fill an oversized pore.

Kaolinite patches may be totally enclosed by calcite and in one example there is a most peculiar texture where the kaolinite has been completely impregnated by calcite, the outlines of kaolinite books being present within and infused by calcite (fig. 7.50).

Porosity and Compaction

The average minus-cement porosity (remaining porosity + percent calcite, dolomite and authigenic quartz) in the wind-ripple laminae slides examined is 26.6% (measured by point counting, pigment not classed as cement). In sandflow samples the corresponding figure is 31.4%, giving a difference between the two types of lamination of about 5%. The contrast in packing is illustrated in figs. 7.23 and 7.24.

Secondary porosity is recognizable where feldspars, rock fragments or other labile grains have been dissolved (fig. 7.25). Complete grain destruction is obviously virtually undetectable, but in the slides measured the visible secondary porosity gained in this manner never amounted to more than 1.3%. Secondary porosity also occurs by the dissolution of both dolomite and calcite cements (figs. 7.26 and 7.27). Dissolution of dolomite is a minor feature only detectable when dolomite-sized and shaped holes are visible in calcite. Some calcite has definitely been removed from the formation, how much is difficult to assess (see p.184).

Through most of the formation there is little restriction of the porosity, this being enacted only by pigmenting phases. The dolomite can effectively block pore throats, where present, though it is often only a pore lining phase. The various developments of calcite cement must act as

a random series of baffles to permeability, being generally unsystematically distributed and present in widely varying quantities. In general, permeability must be strongly anisotropic, much favoured parallel to the lamination, restricted perpendicular.

The histogram showing the frequency of amount of dip of the cross-stratification (fig. 5.36) peaks in the 23°-27° class, falling away rapidly above 29°. Hunter (1981) gives the maximum initial dip of sand-flows in modern dunes as about 32°. In ancient aeolian sandstones these initial dips must be reduced by compaction. It should therefore be possible to assess the amount of compaction the Yellow Sands have suffered, though only if the present day equivalent of the initial 32° could be pinpointed. Fig. 5.36 is little help - arguments could be made for both 25° (the centre of the modal class) and 29° (the brink of the cut-off to high angles). These indicate 25% and 11% compaction respectively. The initial porosity of wind-ripple laminae would therefore have averaged either 52% or 38%, that of sandflows either 56% or 42%. Comparing with Hunter's data on the initial porosities of these lamination types (39% wind ripple, 45% sandflow) the lower figure of 11% compaction for the Yellow Sands seems more tenable.

SECTION 7.3 <u>THE PROVENANCE AND DIAGENESIS OF THE</u>

<u>YELLOW SANDS</u>

<u>Provenance</u>

Previous workers on the Yellow Sands have all pointed to the Carboniferous as the source of the formation (e.g. Versey, 1939; Hodge, 1932; Pryor, 1971b). The rock fragment suite is dominated by sedimentary types having for the most part direct analogues in Carboniferous rocks presently exposed in the County. Garnet, one of the most common heavy minerals in the formation is characteristic of the local Carboniferous (A.P. Heward, pers. comm.).

The palaeocurrents and draa orientation indicate a net sand drift from N.E. to S.W.. The pre-Permian subcrop map of Ziegler (1981, p.5) shows, proceeding offshore from the County in a NE'ly direction, about 30km of Westphalian, followed by a 40km strip of Lower Carboniferous, followed by nearly 300km of Devonian. His palaeogeography of the Rotliegendes has the Mid-North Sea High about 60km offshore. Thus if the Yellow Sands have not been reworked from earlier Permian sediments the rock fragments have not travelled far. The coarse size of much of the quartz in the

Yellow Sands precludes its derivation from many Westphalian rocks, but the Namurian grits offer good candidates; unfortunately little is known of the Devonian under the North Sea. Given the robustness of quartz and the great distances aeolian sand can travel further speculation would be meaningless (Wilson, 1971b; Fryberger and Ahlbrandt, 1979; Chapter 1). It should however be realised that just because the rock fragments and heavy minerals are derived locally it doesn't mean all the quartz was.

The ?basalt grains are an interesting component and not previously recorded from the formation. Basalts are present in abundance in the Devonian and Carboniferous of the Midland Valley of Scotland and these may well extend under the North Sea, though nothing is published. Lower Permian basalts are known from the Mid-North Sea High (Dixon et al., 1981). Derivation from either of these sources would involve transport of hundreds of kilometres. Whether sand-sized plagioclase-pyroxene aggregates could withstand this in a desert climate is open to question. Allowing for temporary entrapments in bedforms along the way the journey could take years, even decades. Closer to home the Late Carboniferous Whin Dykes offer a potential source (the grains are too finely crystalline to be derived from the Whin Sill itself).

The Diagenetic Sequence

The overall paragenetic sequence of the major authigenic minerals within the Yellow Sands is generally

unambiguous and consistent. That sequence is:

1. pigment (probably excluding part of type (ii), the thin kaolinite pellicles)

2. authigenic dolomite

3. authigenic kaolinite (including part of pigment type (ii))

4. reduction of iron oxides, precipitation of pyrite and development of quartz overgrowths

5. sparry calcite

6. oxidation of pyrite and some dissolution of carbonates.

Note that the presumably post-uplift alteration of the sulphides means that the formation has gone through two phases when iron bearing minerals were oxidised, one very early and one very late.

The details of the paragenesis are less easy to elucidate. The concentration of stained clays in hollows on grains, a texture shown by pigment type (iii), is a common feature of red beds and is regarded as indicative of formation by infiltration during influent seepage of rain or flood waters by, for example, Turner (1980, p.120). In a study of modern aeolian sands Walker (1979) also regarded this as an infiltration texture with material deposited on exposed parts of grains being removed by abrasion during transport. Walker set beyond doubt the previously controversial notion that aeolian sandstones could acquire pigment during transport, before final deposition. This should not be surprising given that a grain trapped in the

lee of a migrating transverse dune is doomed to be buried for perhaps 50 years before being exhumed on the windward size, 10,000 years if trapped in the lee of a migrating transverse draa. Walker regarded the composition of the infiltrated clay as being controlled solely by the composition of local clay sources (eg. bedrock, unconsolidated alluvium, soils, dune sand (s.l.) itself). The deduced kaolinitic composition of the material in question in the Yellow Sands can therefore be linked to the abundance of kaolinite in the local Carboniferous, where it is a common cement in sandstones, and to the breakdown of feldspars during and after transport.

Therefore pigment type (iii), the yellow stained (at outcrop) clays concentrated in hollows or grains, is adjudged the earliest diagenetic modification of the Yellow Sands, originating partly before and partly after final deposition by the mechanical infiltration of dominantly kaolinitic material carrying adsorbed ferric iron in some form. That part which was introduced after final deposition may well have formed complete pellicles on some grains but this is now largely obscured by later authigenic minerals. Since it is likely to have developed as a pellicle after final deposition it must also make up part of pigment type (ii), the yellow stained (at outcrop) kaolinitic films covering grains. SEM examination of these clays show a mixture of unambiguously authigenic and possibly infiltrated textures, though all are the same

translucent yellow colour at outcrop and translucent white subsurface.Euhedral kaolinite platelets are common.

The origin of pigment type (i), the intangible yellow staining of most grains can only be left to guesswork. Since it occurs underneath all other phases it is presumably earlier than, or at least coeval with them. It must therefore accumulate during transport and can only be supposed to consist of minute quantities of some form of iron oxide or iron oxide stained clay collecting in extremely small hollows in the surface. Techniques more refined than those used in the present study are evidently required to solve the issue.

The timing of the generation of pigment type (iv), the orange through red and brown to black granules with goethite as their major crystalline component, is uncertain. Their association with (iii) amongst degrading rock fragments suggests a common origin in oxidising desert conditions. The pseudomorphing of framboidal pyrite and pyrite cubes may indicate an origin by the recent oxidation of authigenic pyrite. However, since all iron oxides bar certain red sandstone rock fragments are pyritised subsurface the pyrite framboids and rhombs may themselves be developments on earlier oxidised cores. Looking at other (red) aeolian formation the geothite granules distributed over grain surfaces is an unusual texture. Pigment type (iv) is therefore regarded as derived from

pyrite formed in reducing groundwaters with the pyrite part authigenic, part derived from iron oxides formed round rock fragments degrading pre-Zechstein.

Were the sands once red? Because of the complete destruction of any original colour by later reducing groundwaters this question cannot be answered for certain by studying the formation in isolation. X-ray fluorescence analysis (Table 7.3) shows total iron concentration (expressed as Fe_2O_3) from 0.42% to 3.68%, subsurface and outcrop. This is certainly enough iron to produce a red pigment, iron content in most red beds being. only a few percent (eg. Van Houten, 1973). Note also that the single red sample (from Bowburn) analysed gave 1.9% Fe, falling within the range of the other groups. Given this, the position of the formation at the down-wind end of a very large desert and the likely antiquity of that desert at the time of deposition of the sands, it is quite likely that the Yellow Sands were once red. Walker (1979) showed that modern Libyan aeolian sands became red in transport distances from the Mediterranean coast, ranging from only a few to 400km, the longer course being more arid. Similar transport distances are credible for the Yellow Sands sand.

A critical factor in this discussion is the colour of the underlying Carboniferous. This is grey to a depth of 1 or 2 metres, red to ~8m (max. ~15m) below (Anderson & Dunham, 1953). The grey colour has been attributed to Zechstein reduction. Reddening extends to depths of hundreds of metres in Carboniferous rocks

below the Permian desert sediments of the Vale of
Eden (Trotter, 1939) and has been noted in every other
such area in Britain. The entire floor of the North
Sea area in the Permian, consisting of Carboniferous
and Old Red Sandstone sediments was therefore probably
red. Given this, it is hard to conceive the Yellow
Sands as being enything other than red as the first
breakers of the Zechstein transgression rolled across
the land.

To summarise, the Yellow Sands were probably
initially red by virtue of the now widely accepted
early diagenetic processes of the destruction of ferro-
magnesian and other unstable grains and the formation
of red grain coats by infiltration of clay minerals.
containing adsorbed iron oxides. The clay pellicles
are preserved as pigment type (iii) and part of type
(ii), consisting largely of kaolinite. The destruction
of unstable grains is fossilized in the grain-sized
concentrations of pigment types (iii) and (iv), the
latter now consisting of goethite. At some later date
pore waters became strongly reducing. Pigment type
(iv) was altered from its original (unknown) composition
to pyrite and further pyrite precipitated as a true
authigenic phase, some of it in framboidal form. Any
iron adsorbed on the kaolinite pellicles was also
reduced, rendering those pellicles white. Further
authigenic kaolinite formed at a later date (discussed
below). During Tertiary uplift the formation came to
outcrop and in the oxidising zone close to the ground

surface all the pyrite and ferrous iron was oxidised to goethite and other, amorphous forms to give the present yellow colour.

The dolomite rhombs must have been precipitated (possibly as aragonite or calcite) during the earliest history of the Zechstein Sea, before the Yellow Sands ridges were buried by any succeeding strata. This is demonstrated by the fact that the dolomite has been mixed with the sand by bioturbation in the reworked bed at the top of the formation. This does not unambiguously prove that all the dolomite formed at that stage but the assumption is obvious and not contradicted by any other evidence. Dolomite of similar habit and origin is described by Glennie et al. (1978) from the southern North Sea.

Turner et al. (1978) describe the formation of framboidal pyrite in the Marl Slate. The pyrite is envisaged as forming "by reaction of sulphide produced by reduction of sulphate by organic materials and micro-organisms with iron also released in the reducing environment" (ibid., p.256), organic material, sulphates and detrital iron oxides being readily available. At first sight there seems to be no reason why this mechanism cannot be extended to the pyrite in the Yellow Sands.

SEM examination (fig. 7.37) consistently shows that the dolomite rhombs bear a coat of authigenic clay (identified as kaolinite by XRD) whereas the pyrite does not. This suggests that the growth of

the kaolinite punctuated the paragenesis between the
dolomite and the pyrite. Now the most facile explanation
for the origin of the kaolinite is by destruction of
clastic feldspars (a relationship convincingly demon-
strated in fig. 7.22) in slightly acid pore water.
The most convenient source of acid pore water is the
underlying Upper Carboniferous. This feldspar
alteration would also release silica as a potential
source for the quartz overgrowths, though the increased
development of overgrowths near the top suggests that
clay mineral transformations in the Marl Slate also
contributed. These carry the impression of being
processes that would operate after a moderate amount
of burial. This therefore must call into question the
very tempting attribution of the reduction of the iron
oxides to early Zechstein groundwater. Perhaps the
earliest deposits of the Marl Slate effectively sealed
off the sands from Zechstein sea water or maybe
the pyrite in the Marl Slate could only form in the
immediate presence of decaying organic matter. Glennie
et al. (1978) tentatively link decolouration (from
red to grey) of parts of the Lower Rotliegendes under
the southern North Sea to the expulsion of acidic
formation waters from the underlying coal-bearing
Carboniferous during compaction and coalification.
The upper part of the sequence, actively reworked by
the Zechstein Sea, lost its colour by reduction in
groundwaters associated with that sea. The middle of
the sequence is red because neither process could
propagate completely through the sediment pile, being

restricted by permeability.

Glennie's work justifies the application of both processes to the Yellow Sands. The textural relations favour the latter: the reduction of the iron oxides and precipitation of pyrite by waters expelled from the underlying Carboniferous sediments. With impermeable carbonates and evaporites above, and being thin and permeable, the Yellow Sands must have been subjected to an undiluted influence of the groundwaters emanating from the Carboniferous. In view of the acidity of the pore waters invoked for destruction of the feldspars it is surprising that the dolomite has survived, but this nevertheless seems the most plausible scenario. Where the dolomite is visible through its clay coat in some specimens the rhombs might, with the eye of faith, be suspected of showing slightly etched surfaces (eg. figs. 7.41, 7.42), but there is no other evidence of dissolution at this stage.

Under the southern North Sea kaolinite is envisaged as developing during burial at depths of 1000-3000m, transforming to illite and chlorite beyond 3000m (Glennie et al., 1978; Nagtegaal, 1979). Since illite is subordinate and chlorite absent in the Yellow Sands the maximum depth of burial of the formation is therefore limited to 3000m or less, on this basis.

The sparry calcite cement was the last of the authigenic phases to precipitate though the presence of a limited amount of early calcite cannot be ruled out. Its development immediately below the Marl Slate

might indicate a concentration of the relevant ions by membrane filtration. It also shows some preference for coarser, more permeable horizons (fig. 7.44) but its reported affinity for the lower part of the formation does not seem to relate to any textural characteristic. The source of the calcite is not known. Dolomitisation of the overlying Upper Permian carbonates must have released vast amounts of calcium but this is believed to have occurred quite soon after deposition (Smith, 1981, p.179).

Glennie et al. (1978) postulate that most of the original carbonate (dolomite in this case) in southern North Sea Rotliegendes sandstones was dissolved and reprecipitated during Cretaceous uplift. A similar story might be applied to the Yellow Sands to explain the partial dissolution of the calcite. Alternatively the dissolution could have occurred at a later stage of uplift as meteoric water penetrated the formation. As stated earlier (pp. /87) the amount of material removed by that dissolution cannot be ascertained.

The final diagenetic modification of the Yellow Sands, the oxidation of the pyrite, could almost be termed weathering. Presumably it took place in oxidising meteoric waters during the later stages of uplift. Whether the yellow colour of the authigenic kaolinite is due to the oxidation of adsorbed ferrous iron compounds or the relatively recent adsorption of forms of ferric iron liberated from the pyrite is uncertain. Instances of pigment lying on authigenic

calcite crystal surfaces demonstrates that the process was not one of purely in situ oxidation: a certain amount of mass movement was involved. It is interesting to note that the oxidation has even penetrated areas apparently tightly cemented by calcite - normal pigment colours are seen within cement in thin section. The oxidising fluids must have propagated along very tight pores between grains and cement; the irreducible lamellar porosity of up to 2.5% described by Schmidt and Macdonald (1979, pp.184-185).

Little can be said of the barytes cement except that it is pre-calcite and, on the evidence of one specimen, apparently was not replaced by calcite. The barytes spherules at Sherburn Hill (fig. 7.49) are a most peculiar feature. Their textural relations do not offer any clues to dating and no explanation is offered to their origin.

Summary

The wide spread of grain size of the Yellow Sands can be attributed to the disposition of the formation in longitudinal draa, the access of coarse sand being expedited by the inter-draa corridors. The variation of grain size within the formation reflects the lamination of the sediment and sorting processes particular to each resulting lamination can be elucidated.

The rock fragment suite in the Sands is strongly influenced by the underlying Carboniferous sediments though this does not mean that all the sand was

similarly derived.

The main features of the diagenetic history of the formation are, in order of development:

1. Accumulation of probably red, dominantly kaolinitic clays in hollows on grains during transport and a red stain on most grain surfaces,

2. Infiltration of further clay, again dominantly kaolinite and probably red, by rainwater after final deposition. Also gradual destruction of unstable rock fragments.

3. Precipitation of silt grade, euhedral, generally pore-lining dolomite soon after the Zechstein transgression.

4. Precipitation of pore-lining (locally pore-filling) kaolinite derived from the destruction of feldspars in acid groundwater produced from the underlying Carboniferous strata. Also growth of quartz over-growths, reduction of iron oxides and precipitation of pyrite in the same groundwater, though probably slightly later.

5. Precipitation of the sparry calcite cement from an unknown source at an unknown date.

6. Dissolution of the calcite and dolomite to an unknown extent, possibly by meteoric waters during uplift.

7. Oxidation of previously reduced iron and pyrite with some redistribution of material, also in meteoric waters.

The preservation of kaolinite as the dominant
clay mineral in the formation suggests, by comparison
with southern North Sea Rotliegendes sediments, that
the Yellow Sands have never been buried deeper than
about 3000m.

CHAPTER 8

THE BRIDGNORTH SANDSTONE

SECTION 8.1   INTRODUCTION, STRATIGRAPHY AND DISTRIBUTION

The Bridgnorth Sandstone, as the "Lower Bunter Sandstone" was the subject of a classic paper by Shotton (1937). This work, classic not just in the study of aeolian sediments, but in sedimentology as a whole, dealt systematically with the cross-bedding and grading of the formation. Its premises and approach set a standard that has persisted for 40 years and it is only in the late 1970s and 1980s that a new generation in aeolian sedimentology (e.g. Brookfield, 1977, Kocurek, 1981 and this thesis) has developed. Little of significance about the formation has been published since Shotton's work.

In its type area, around the town of Bridgnorth in Shropshire (20 km WSW of Wolverhampton) the formation, ~180 m thick, is composed entirely of red aeolian sandstone. It rests unconformably on Upper Coal Measures and is overlain unconformably by what used to be called the Bunter Pebble Beds, now called the Kidderminster Conglomerate Formation (conventionally regarded as Lower Triassic in age). This formation seems to have escaped any detailed attention but appears to be a pebbly fluviatile deposit.

Despite recent comprehensive revisions (Smith et al., 1974; and Warrington et al., 1980) the organisation and nomenclature of the stratigraphy of the Bridgnorth Sandstone and associated rocks is extremely confused.

This is due to the lateral variability of the rocks
and the paucity of both fossils and exposure.

It appears that there is a contiguous body of
aeolian sandstone from the Stourbridge area northwards
into the Irish Sea Basin (see fig. 8.1). It is known
as the Bridgnorth Sandstone in Worcestershire,
Staffordshire and Shropshire, as the Kinnerton Sand-
stone Formation in most of Cheshire, and as the
Collyhurst Sandstone in north-east Cheshire, Greater
Manchester, Merseyside and Lancashire. Westwards
Permo-Trias outcrop terminates. To the east in
Staffordshire and the West Midlands the Bridgnorth
Sandstone thins to zero along a line from Stoke to
Wolverhampton. Whether this thinning is by non-
deposition or deposition and subsequent erosion is
unknown. The outcrop and subcrop area of this body
of rock (on land) totals about 4500 km^2.

This picture is complicated by the presence of
the Manchester Marls replacing the top part of this
sequence around Greater Manchester and in Lancashire.
These yield an early Upper Permian marine fauna from
beds near the base. The Manchester Marls are absent
in the Liverpool area though identified further into
the Irish Sea Basin. To the south they pass into
presumed aeolian sandstone. Also, in the northern part
of the Cheshire Basin,Poole and Whiteman (1966) report
the presence of "sporadic marl bands" within the Lower
Mottled Sandstone. These seem to increase northwards
until the entire aeolian interval shales out in the centre

of the Irish Sea Basin (Colter and Barr, 1975). A local sandy interval between the Manchester Marls and Chester Pebble Beds (= Bunter Pebble Beds), formerly named Lower Mottled Sandstone, is now regarded as a sandy variant of the Pebble Beds and included within that formation (Warrington et al., 1980). It is therefore transferred to the Trias. The stratigraphy of the Bridgnorth-Kinnerton-Collyhurst Sandstone is presented diagrammatically in fig. 8.2. The least confusing modern account of the subject lies in Colter and Barr(1975), though this uses the older (and simpler) nomenclature.

Despite the confusion and lack of exposure a coherent picture emerges of a contiguous body of aeolian sandstone, its present onshore outcrop being only an offshoot of a much larger basin under the eastern Irish Sea. Tributary to the northern margin of this basin are the Permo-Trias outcrops of S.W. Scotland and the Vale of Eden. Their evolution is in many ways analogous to that of the Irish Sea and Cheshire Basins (see e.g. Colter and Barr, 1975; Smith et al., 1974). Collectively, these areas form a small counterpart to the Northern and Southern Permian Basins of the North Sea. Were it not for the fact that most of the rocks lie beneath waves or drift the separate outcrops could no doubt be interpreted in a coherent regional pattern. The Collyhurst, Kinnerton and Bridgnorth Sandstones are almost certainly parts of the same erg. In subsequent sections these three Sandstones will be treated as one, referred to as 'the Bridgnorth Sandstone' or simply

"the formation". It is not impossible that this erg extended as far as the Vale of Eden and S.W. Scotland.

The age and duration of the aeolian interval are difficult to assess. The Manchester Marls signal a transgression which is taken on faunal avidence to be of the same age as the first Zechstein transgression east of the Pennines, i.e. earliest Late Permian. This terminated aeolian deposition in Lancashire, Greater Manchester, the Irish Sea basin and probably the Vale of Eden too. Near Formby in Lancashire 715 m of aeolian sandstone had accumulated prior to this event (Kent, 1947). Elsewhere the unit (typically 150-300 m thick) is terminated beneath the unconformity of the Lower Triassic Chester Pebble Beds/Kidderminster Conglomerate Formation/Bunter Pebble Beds. The unconformity does not mark any structural break, though the Pebble Beds overlap the aeolian interval by up to 30 km eastwards.

The Bridgnorth Sandstone is poorly exposed, out-crops being overwhelmingly concentrated in the immediate vicinity of Bridgnorth and also in the Kinver and Bewdley areas (NE and W of Kidderminster respectively). To the north and north-west of Bridgnorth, in the Cheshire basin, exposures are extremely sparse. The only lists of outcrops lie in the memoirs of the Institute of Geological Sciences which range from 15 to nearly 100 years old. During fieldwork it was frequently found that exposures no longer existed and in several instances rocks referred by the survey to

the Lower Mottled Sandstone were found to be water-lain
rather than aeolian and thus best placed in the Pebble
Beds (such examples were always at or very near the
mapped top of the formation).

SECTION 8.2   SEDIMENTARY STRUCTURES

Introduction

Scale diagrams of the cross-bedding in the 19
best, accessible exposures of the Bridgnorth Sandstone
are reproduced on Enclosures 9-14. These follow
exactly the same format as those drawn for the Yellow
Sands and are compiled from field sketches and
photographs. Surveying was done by pacing, crude
triangulation and estimate. "Metres" on the scales
should be read as "metres $\pm$ 20%".

All but one of the sections are from the Bridgnorth,
Kinver and Bewdley areas, reflecting the density of
exposure of the formation. The exception is a road
section from near Preston Brockhurst, 40 km NNW of
Bridgnorth. All the sections have a significant tectonic
dip, up to about 10°, generally to the north or east.
The exact value of the dip is impossible to assess,
the nearest reliable markers of the original horizontal
being in the overlying Pebble Beds.

Description and Interpretation of the Cross-Bedding

Shotton (1937) considered the cross-bedding of the
formation as formed by migrating and climbing barchan
(meaning transverse) dunes and complicated by bounding
surfaces developed by re-orientation of the dunes in
changing winds. This interpretation was way ahead of

its time and in current parlance allows for random
bedform behaviour and the development of 3rd and
4th order bounding surfaces. The resultant cross-
bedding dip azimuth is given by Shotton as 274°,
interpreted as indicating an easterly wind.

A first division of the outcrops can be made
into those parallel to and those normal to
palaeocurrent. In the Bridgnorth area most are
normal to the palaeowind, following the strike of
the Severn Valley. The exceptions to this rule are
the Worfe Bridge, A454 and Dudmaston sections
(Encls. 9, 11 and 13 respectively). On Encl. 14
the NE elevation of Holy Austin Rock, the south
elevation of Drakelow Depot, and the Preston Brockhurst
section are parallel to the wind.

One important exposure not included amongst the
diagrams is Blackstone Rock (marked on the locality
map on Enclosure 14) overlooking the River Severn
between Stourport and Bewdley. There is about 200 m
length of exposure here in a face up to 25 m high
(see fig. 8.3). Though the eastern half is much
cut up by faults the highest part of the main face
is made up entirely of the sandflow laminae of one
set of cross-bedding, dipping at up to 65° SW, of
which at least 30° must be tectonic. The faulting
makes it difficult to gauge the exact thickness of the
set but it must be at least 30 m (neither top nor
bottom are exposed). This is possibly the thickest

set of cross-bedding exposed anywhere in the U.K.

Blackstone Rock represents one extreme of the variability of cross-bedding style in the formation, being an enormous set lacking any internal bounding surfaces. The opposite end-member is manifest for instance in face C at The Hermitage, Bridgnorth (Encl. 10) and the southern cut by the A442 at Quatford (Encl. 12). These have a complex and intricate arrangement of bounding surfaces defining sets typically 1 - 3 m thick.

Thick, laterally extensive sets lacking internal surfaces may also be seen at Bridgnorth Golf Club Car Park (Encl. 9) where a 60 m long strike section has been cut through a single set showing little variation in dip. At High Rock, Bridgnorth (Encl. 9) the sets are consistently 6 - 10 m thick and extend laterally for at least 80 m without terminating. The set occupying most of the A454 section at Bridgnorth (Encl. 11) is about 20 m thick. In the Dudmaston A442 section (Encl. 13) a single set is broken only by 2 minor surfaces in a distance of 150 m and may extend the whole length of the section. Bewdley railway cutting shows sets extending laterally with little change in dip direction for 80 - 100 m.

To interpret the cross-bedding it is convenient to consider first the two tallest exposures: High Rock (Encl. 9) and Quatford Rock (Encl. 11). The latter provides a fine example of large scale trough

cross-bedding cut transverse to palaeocurrent. The
sets are typically 2 - 4 m thick with a lateral
extent of 20 - 40 m. There are only 2 orders of bound-
ing surface apparent. In High Rock the sets are
consistently 3 times thicker and much more laterally
extensive, stretching further than the exposure.

The cross-bedding in Quatford Rock must be
interpreted as having been deposited by sinuous
transverse dunes, exactly in accordance with Shotton's
(1937) ideas, if not his terminology. The next step
then logically becomes the attribution of the cross-
bedding in High Rock to slipfaced transverse draa,
an interpretation which is extended to all the other
thick, laterally extensive sets mentioned above.

This conclusion is founded only on the systematic
divergence in scale of the two types of cross-bedding,
not on any hierarchy of bounding surfaces, though
the pattern of bounding surfaces is consistent with
this interpretation. Indeed, given the 17 m height
of Quatford Rock, compared with the 6 - 10 m draa
set thickness in High Rock, the former should be
expected to contain at least one first order bounding
surface. No such feature is evident, presumably being
a victim of the scouring of the feet of draa lee-
sides by dunes as mooted in Chapter 2. The basic
argument is that there appear to be two contrasting
styles of cross-bedding, differing principally in
scale, and it is logical to attribute these to the
two available orders of generating bedforms. The
only alternatives would be to attribute all the

cross-bedding either to slipfaced draa or dunes.
However, it is difficult to envisage draa forming
the trough sets only 40 m or less wide, visible,
for example, in the Quatford sections. Likewise
the deposition by dunes of 20 and 30 m thick sets
in the A454 section and at Blackstone Rock, and
repeated 6 - 10 m sets in High Rock is singularly
improbable. It is therefore suggested that both
slipfaced and slipfaceless draa contributed to the
deposition of the Bridgnorth Sandstone.

Distinction of the two cross-bedding styles
is only possible in those exposures where the super-
position or lateral extent of several sets is
evident. This is reflected in the question marks
attached to the interpretations of bounding surface
order on the enclosures. For example in the A454
section (Enc. 11) the 4 1-2 m thick sets occuring
above the main 20 m thick set could either be dunes
or draa, though the contrast with the main set tempts
labelling as dunes. Since the section is parallel
to palaeocurrent the true thickness of the thin sets
is not known.

In labelling the bounding surfaces the greatest
problems arise over the distinction of 1st order from
3rd order (the migration surfaces) and 2nd from 4th
(the modification surfaces). This can usually only
be done on an arbitrary judgement of the scale of the
sets concerned. 1st and 2nd order surfaces, and 3rd and
4th order in pairs are more easily distinguishable,
since they tend more often to observe a hierarchy.

Also the modification surfaces generally only
involve small angular discordances above and below,
whereas migration surfaces have sharp truncations
below and asymptotic toesets above. Parallel to
palaeocurrent the former generally dip more steeply
than the subhorizontal migration surfaces.

According to the interpretations given,
modification surfaces (2nd and 4th order) are much
less common than migration surfaces, as was the case
with the Yellow Sands.

Comparing Enclosures 9-14 with McKee's (1966)
diagrams of sections through modern dunes highlights
the paucity of modification surfaces in the Bridgnorth
Sandstone. All the dunes McKee excavated were
densely cut by such surfaces though they are
significantly concentrated in the upper, least
preservable parts of the dunes. The diminutive
proportions of the Bridgnorth dune sets are also
emphasised by examining McKee's diagrams, which in
the case of his transverse dune (ibid., fig. 7) are
of a section 70 m long and 12 m high. The contrasts
between McKee's sections and the Bridgnorth Sandstone
exposures should be seen as an illustration of the
influence and action of preservation potential.

The Bridgnorth Sandstone exposures are a regrettably
indigent fount of enlightenment on the behaviour of
transverse aeolian bedforms during migration. Exposures
transverse to palaeocurrent, of which there are many,
record only a succession of discrete frozen moments
in the history of the erg. These exposures (e.g.

Quatford Rock, Encl. 11) show a sometimes random, sometimes apparently systematic stacking of trough-shaped sets generally containing relatively few internal surfaces. The systematic stacking is evident at the 70 m mark on Quatford Rock where 3 sets are successively and vertically superposed in a manner that suggests the regular migration of an array of sinuous transverse dunes in a fishscale arrangement.

Only exposures parallel to palaeocurrent provide a continuous, analog record of bedform migration and development. None of the sets so displayed is in length anywhere near the likely original wavelength of the generating bedforms.

Shotton (1937) gives two illustrative palaeocurrent roses for the formation, one from the Quatford area and one from Kinver. Since these probably contain data from both draa and dune-generated sets reconstruction of a composite bedform leeside shape would have even less actualistic relevance than it did for the Yellow Sands. The cross-bedding roses are however, unimodal, indicating probably much less sinuous dunes than those suggested for the Yellow Sands. Similar caveats as for the Yellow Sands must be given on the inference of a unimodal wind regime from the unimodal dip azimuth pattern (pp.136-139). The dip azimuth data can be used as a paradigm only for the sand drift direction in the erg. According to the ideas presented in Chapter 2, a pattern of sinuous transverse bedforms in an erg with complete sand cover is compatible with a wide

variety of wind regimes. The cross-bedding pattern merely denotes that the overall resultant of whatever regime prevailed was to present west (Permian WSW).

From the exposure available it is not possible to draw any firm conclusions about the distribution of slipfaced and slipfaceless draa deposits in the formation. Particular styles of cross-bedding do not seem to be confined to any particular area or stratigraphic level, but correlation within the formation is impossible anyway. It may be that the dune sets represent the cross-bedding of the slip-faceless linguoid elements of otherwise slipfaced draa, though it might then be expected that some evidence of slipfaceless draa would be apparent in High Rock for instance. Perhaps the contrasting styles signify the transformation of draa over the whole erg to and from slipfaced to slipfaceless types. This would presumably have to result from some change in flow conditions. Though there is almost no data to consult on the issue it is likely that slipfaceless draa signify more variable winds than slipfaced (inference drawn from Fryberger, 1979a).

Other Structures

At only 2 localities was anything other than sand seen in the formation. At Knowlesands (SO 719913), just south of Bridgnorth, the unconformity with Carboniferous Keele Beds was temporarily exposed in

the summer of 1980. The top of the Carboniferous
here consists of fissile,micaceous maroon siltstones
and fine sandstones. The base of the Bridgnorth
Sandstone is made up of 0.3 m of a clayey and sandy
fine conglomerate, pebbles being typically 4-5 mm.
This had to be dug for and only a very small area was
exposed revealing no lamination or other structures.
The pebbles are mostly of chert, sandstone or durable
fine-grained igneous types. There are potential
primary or secondary sources for all these in the
local Upper Carboniferous sediments.

Above this comes sandstone bearing an aeolian
lamination with fine and very fine pebbles confined
to the lowest 0.5 m of sand sheet material, overlying
which is cross-bedded sand of normal grain size.
Thus there are no sediments in the area recording a
gradual onset of aeolian deposition: the formation
begins as abruptly as it ends.

By the River Severn at Eyton (SJ 569059) is a
single exposure 10 m long of pale greenish grey
(Munsell designation 5Y7/2: light gray), fissile,
clayey and micaceous siltstone only a few millimetres
thick lying just above a migration-type bounding
surface and conformable with wind-ripple laminae above
and below. There are absolutely no sedimentary
structures (e.g. dessication cracks, roots, adhesion
ripples, or current indicators or fossils of any kind)
associated with the layer. Its composition (mostly
clastic quartz) rules out a pyroclastic origin and

since it is so different in grain size, composition
(being micaceous), lamination, and colour from the
rest of the formation it was presumably deposited in
standing water. Its isolation and position slightly
above rather than at a bounding surface militate
against an interdraa sabkha origin, though the layer
might represent a temporary lateral extension of a
nearby sabkha a short distance up the toesets of
neighbouring bedforms. The bounding surface
associated must be 1st order since standing water
is only to be expected in the topographically lowest,
and therefore interdraa, rather than interdune areas.

Preston Brockhurst road cut (Encl. 14) contains
the only sand-sheet deposits seen within the formation.
This occurs in a 'set' about 2 m thick and exposed
over only 50 m of the section. The sand has a very
strong coarse component and shows numerous examples
of small-scale cross-lamination similar to those
illustrated in modern deposits by Fryberger et al.
(1979). The sand-sheet material also contains a
back-filled burrow 0.32 m long and 15 - 20 mm wide,
straight and slightly inclined from normal to the
lamination, with a 50 mm diameter circular bulb of
'white' sand at the lower end (the sand in the cutting
is mottled red and 'white', see fig. 8.5). The burrow
resembles a fulgurite in morphology but fulgurites
generally are formed of a tube of fused silica which
is not present in this case. The perfection of the
backfill would also argue for a biogenic origin. The

burrow is not preferentially cemented, but there are 5 other vertical or subvertical rod-shaped (.2 - .5 m x 10 - 30 mm) features within the set which are (fig. 8.6). These have no backfill though they may be associated with some disruption of the laminae. To the author's knowledge no burrows have previously been described from any pre-Tertiary aeolian sandstone.

The trace is most likely an arthropod dwelling burrow. The vertical concretions present more of a problem since they lack any clear internal structure. They could be other burrows, or possibly dikaka, but may be completely spurious.

The sand-sheet material in which these features are preserved probably represents an interdraa area. This would be the most likely site for the accumulation of coarse sand and the possible growth of vegetation. It is unfortunate that the sand sheet is not more extensively exposed; this hampers any assessment of its implications for the internal geography of the erg, e.g. whether it is indeed at a 1st order bounding surface and is unique, or is repeated elsewhere in the sequence and marks the first stage of a progression to interdraa sabkhas in the basin centre.

SECTION 8.3   LAMINATION AND GRAIN SIZE

Methods

The lamination of the formation was assessed by visual estimation at outcrop in the same manner as for the Yellow Sands. The data is hamstrung by the same

possible misconception as the Yellow Sands: the
conviction that grainfall lamination was irrelevant
in the ancient. With this borne in mind the data
acquired is presented without further comment.

Only ten samples were sieved, partly because
this aspect of the formation has already been
investigated by Shotton (1937) and partly because
the exercise was intended to be merely complementary
and comparitive to the Yellow Sands data.

Lamination

On the basis of 201 estimates of lamination
proportion the Bridgnorth Sandstone is made up 55%
of wind-ripple laminae, 44% of sandflow cross-
stratification and 1% sand-sheet lamination (cf 71%,
25% and 4% respectively, for the Yellow Sands). One
hundred and sixteen measurements of maximum sandflow
thickness give an average of 62 $\pm$ 19 mm. Sandflows
100 mm or more thick were seen at Eyton (SJ 569060),
Habberley Park (SO 800781), Bewdley railway cutting
(SO 798747) and Holy Austin Rock (SO 836835), with
the thickest of all, 120 mm, at Blackstone Rock
(SO 793740).

Two factors can be envisaged to account for the
increased importance of sandflows in the Bridgnorth
Sandstone, as compared with the Yellow Sands. It may
simply be the result of the greater average set
thickness in the Bridgnorth Sandstone resulting from
the preservation of slipfaced draa sets. This could
enhance the proportion of sandflow-laminated foresets

relative to the more gently inclined wind-ripple
laminated toesets. Alternatively this might be
brought about by the lower sinuousity of the generating
bedforms averred in the previous section, since more
sinuous bedforms tend to have more concave leesides
and thus develop longer and thicker toesets. Probably
both mechanisms are valid.

With the cross-bedding of the formation inter-
preted as developed from both dune and draa slipfaces,
and a great deal of sandflow thickness data recorded,
we have all the necessary ingredients for assessing
whether the two bedform types may be distinguished
solely on the basis of sandflow thickness.

Firstly, the obstacles to this attempt should
be mentioned. Principal among these is that most
exposures admit the examination of only a very small
area of former slipface. An exposure cutting the
margin of a slipface will probably give a lower sand-
flow thickness than one through the centre. Similarly
sandflow thickness must vary up and down the slipface.
Also the identification of dune and draa sets is
subjective and not certain, though it does seem
reasonable and consistent.

Consideration will be given only to those
exposures where the cross-bedding can be most confidently
identified. Sets deposited by slipfaced draa are
exposed at Blackstone Rock, Eyton, High Rock (Encl. 9),
Bridgnorth Golf Club (Encl. 9), the A454 section
(Encl. 11), Dudmaston A442 section (Encl. 13) and
Bewdley Railway Cutting (Encl. 14). Respectively, the

maximum sandflow thicknesses measured at these local-
ities are 120, 100, 80, 70, 80, 85 and 100 mm from a
total of 29 measurements (average 68 $\pm$ 19 mm). The
dune cross-bedding data suffers from its 'type'
locality, Quatford Rock, being inaccessible. However,
the Worfe Bridge (Encl. 9), Worfe Mouth (Encl. 9) and
Quatford A442 north and south sections (Encl. 12)
yield maximum sandflow thicknesses of 80, 55, 60 and
70 mm from a total of 22 measurements (average 54 $\pm$ 14 mm).

Thus, on this limited data, there does seem to be
some basis for distinguishing dune from draa slipfaces
by the measurement of maximum sandflow thickness.
Sandflows thicker than about 80mm, and certainly those
thicker than 100 mm must be taken as a suggestion that
the set concerned was deposited on a draa slipface.
This can only be a tentative conclusion: data is needed
from more and better exposed formations, and more
especially from modern bedforms. The only modern data
available at present is the graph Kocurek and Dott
(1981) provide of a maximum sandflow thickness v. dune
height from the Little Sahara dune field, Utah.
Extrapolating this to the Bridgnorth Sandstone data
indicates bedforms 1000km high for a 120 mm sandflow,
200 m for a 60 mm flow. The extrapolation is evidently
an abuse of the available information.

Fig. 8.6 illustrates an interesting feature of
the lamination seen in a few thin sections. A single
lamina 1 mm thick containing much very fine sand is
visible separating two laminae of medium sand. The

very fine sand may either be the base of a sandflow or a single wind-ripple lamina. Some of the sand within it has fallen through the gaps between the coarser grains below, indicating the way up of the specimen.

Grain Size Characteristics

Shotton (1937) sieved 51 samples of Bridgnorth Sandstone and summarized the grain size characteristics of the formation as a whole. He divided his samples (most were from a single borehole core at Kinver) into uniform and laminated types, but these cannot be related to the current categorisation of aeolian lamination. Presumably though, the "laminated" samples will include a high proportion of wind-ripple material.

The results of the sieving carried out in the present study are shown graphically in figs. 8.7-8.11, in exactly the same format as the Yellow Sands data. Figs. 8.7-8.11 confirm the observations and deductions made from the Yellow Sands. The wind-ripple distributions are broad and flat-topped or bimodal, tending to be only moderately sorted, fine skewed and platykurtic. The sandflow samples are well sorted, fine skewed and meso- or leptokurtic. The only contrast with the Yellow Sands is that the formation is slightly finer in grain size. Using Shotton's (1937) data the mean of the medians of 27 samples is 2.16ϕ, comparing with 1.75ϕ as the mean

of 39 Yellow Sands medians. This is reflected in the coarse and fine tails of each formation: many Yellow Sands samples contain grains of 0ϕ whereas only 3 of the Bridgnorth curves extend beyond 0.5ϕ. The pan fraction (finer than 4ϕ) is more pronounced in the Bridgnorth Sandstone.

Part of this contrast is believed to be due to the difference in erg and draa type: the Yellow Sands being longitudinal draa in an erg lacking complete sand cover, whereas the Bridgnorth Sandstone is a thick erg body built up by transverse draa. In the Bridgnorth Sandstone coarse grains were efficiently trapped and filtered out as they entered the upwind part of the erg. The lack of complete sand cover in the Yellow Sands facilitated the penetration of coarse grains. This confirms the influence of bed-form type on grain size suggested in Chapter 3 (p. 81   ). The other possible contributor to the difference in grain size is the influence of the respective sources of the 2 formations, but this is unquantifiable.

The Bridgnorth Sandstone curves pass an equivocal verdict on the ideas of Bagnold and Barndorff-Nielson (1980) discussed in Chapter 3 (pp. 76-78) and supported in Chapter 7 (p. 161   ). Some limbs of the log-frequency curves are straight, suggesting conformity to a logarithmic distribution, some are contrarily convex-upwards, the most extreme examples being figs. 8.11a and 8.11b.

SECTION 8.4   PETROGRAPHY AND DIAGENESIS

These aspects of the formation were investigated
by the same methods as used for the Yellow Sands,
with the exception of X-ray fluorescence and heavy
mineral analyses. The petrography and diagenesis
of the Bridgnorth Sandstone does not seem to have
received any attention previously.

Colour and General Characteristics

In terms of diagenetic effects apparent at
outcrop, the Bridgnorth Sandstone is remarkably
homogeneous. With minor exception, it is red
throughout, varying little in hue, value or chroma
(colour determinations made on 46 hand specimens are
presented in Table 8.1). It is uniformly lacking in
hard cement, the many rock-houses of the Bridgnorth
and Kinver areas being held together only by a thin
pore-lining mixed clay cement. The only pore-filling
cements are minor amounts of vermicular kaolinite and
feldspar overgrowths.

Framework Grain Composition

Point-counting of thin sections shows the
formation to be a sublitharenite (Pettijohn, 1975; see
fig. 8.12), containing generally 10 - 20% rock fragments
and 5 - 10% feldspar. The rock fragment suite is
diverse, consisting of sedimentary, igneous and low-
grade metamorphic types, numbering about 2 to 1 in

favour of sedimentary species (averaging $\sim$11% sedimentary, $\sim$5% combined igneous and metamorphic). These are dominated by chert (occasionally fossiliferous) and fine sandstones and siltstones cemented by chlorite and other clays. Shale, quartz arenite and opaque-cemented fine sandstone rock fragments are subordinate. Igneous rock fragments comprise everything from rhyolite, and spherulitic rhyolite of Uriconian affinities to granite, granophyre, ?andesites or andesitic tuffs and fine grained basalt. The acid types are most common. Metamorphic rocks are represented by foliated chloritic sand- and silt- grade quartzites. The heavy mineral suite of the formation is dominated by opaque minerals with exceedingly rare green/yellow tourmaline, muscovite and biotite.

Feldspars (average $9.5 \pm 3.4\%$ of the clastic component) are exclusively of alkali types, with plagioclase extremely rare. Theyvary from fresh to dusty to sericitized to nearly completely dissolved, (see subsection on porosity); fresh and slightly altered types being most common.

Grain Shape and Surface Texture

Grain rounding varies with size, coarse grains being rounded or well rounded, medium grains subrounded to rounded, fine and very fine grains subangular to angular. The modal sizes (1.5 - 2.5ϕ) therefore tend to be subrounded. Sphericity is generally moderate to high.

Viewed in hand specimen under the miscroscope all grains appear frosted over their entire surface, in hollows as well as salients(fig. 8.13). SEM examination (figs. 8.14, 8.15) reveals a widespread preservation of upturned or cleavage plates, an aeolian surface texture. This must be responsible for most of the frosting. Hollows on the grain surfaces must be partly sheltered from the abrasion that produces the frosting but few are so deep as to be impenetrable to the process. The extensive preservation of this texture possibly militates against there ever having been a widespread hard cement in the formation (Marzolf, 1976).

<u>Porosity and Compaction</u>

The porosity of the formation is uniformly high. It varies with lamination such that the average minus-cement porosities (subtracting pore-filling kaolinite and authigenic feldspar; not including pigment) of sandflow and wind-ripple samples are $28.0 \pm 3.2\%$ and $23.0 \pm 3.8\%$ respectively.

Permeability must be strongly anisotropic, the fine laminae within wind-ripple laminated samples and the fine bounding laminae of sandflow samples markedly restricting any flow across the stratification.

Two sandflow specimens had sections cut both in the plane of and normal to the lamination. The minus-cement porosities in the plane of lamination were 33.3% and 30.7%; normal to the lamination 24.3% and 23% respectively, giving differences of 9% and 7.7%. This

is more than can be attributed to error though 2 pairs of samples is no great data base. It suggests an anisotropy of apparent porosity developed during the deposition of sandflows such that the packing of grains parallel to stratification is much looser than packing normal to stratification.

A small amount of secondary porosity has been developed in the formation by the dissolution of framework grains. This is principally manifest as hollow feldspars though other rock fragments are affected (fig. 8.6). Such dissolution porosity averages 1.4% of the rock, never making up more than 3% in any slide.

Shotton, (1937, p. 546) provides a frequency curve of the angle of dip of the cross-stratification in the formation (corrected for tectonic dip), compiled from nearly 800 measurements. This peaks at 26°, falling away abruptly to 1/12 of the peak value at 34°. Hunter (1981, p. 323) reports that the initial dips of sandflow cross-strata cluster tightly around 32° (note that this differs from the much quoted value of 34° as the angle of repose of dry sand). Reducing 32° to 26° requires 22% compaction, assuming that the selective preservation of lower, more gently dipping slopes has not influenced the angle of dip frequency curve.

Pressure solution has had only a very minor effect on the formation, limited to rare instances of

quartz grains penetrating rock fragments; quartz-quartz
solution was seen in only one slide. This being the
case, the 22% compaction suggested must have come
about mostly by grain rotation with an unknown
contribution from the destruction of labile grains.
Removing the compaction produces 50% initial porosity
in sandflows, 45% in wind-ripple lamination. These
contrast with porosities measured by Hunter (1977) on
modern (coastal) dune sands. He found the average
porosity of sandflow cross-strata to be 45%, and 39%
in wind-ripple deposits. This would suggest that
either the compaction figure is too high, that there
has been a great deal of now undetectable grain
dissolution or that inland and coastal aeolian sands
have different initial porosities. The first
explanation is thought to be most likely. If this is
the case, comparison of Hunter's porosity figures with
those for the Bridgnorth Sandstone indicate 17%
compaction. This would transform an original dip of
32° to a post-compaction 27.4^O and suggests that a
selective preservation of lower, more gently sloping
parts of bedforms has affected the dip distribution.

Pigment and Pore-Lining Clays

Several different components contribute to the
colour of the formation. Principal among these is a
red staining which covers every grain but cannot be
removed by scratching grain surfaces with a fine needle
(fig. 8.13). It is translucent and must be extremely

thin since it is often not detectable in thin section. Nor is any material origin of the colour evident under the SEM; otherwise clean grain surfaces showing an apparently unencumbered aeolian quartz surface texture (e.g. figs. 8.14, 8.15). Since the colour does not seem to be mechanically removable it is unlikely that the filtrate derived from ultrasonic agitation of the sand concentrates it to any great extent. The identity and origin of this red stain must therefore be left as an open question. The employment of a better SEM, probe and EDAX methods, and electron microscopy of sectioned grains might produce an answer.

This staining complicates the assessment by optical microscopy of the quantity of pigment present in the formation. In thin section, an oblique cut across the margin of a red-stained grain appears as a relatively wide band of colour - this most insubstantial component of the rock appears volumetrically important when projected into the two dimensions of a microscope image. Any point-count measurement of the pigment quantity must therefore be an overestimate. The error increases in finer grain sizes because of the increased ratio of (red) surface area to volume.

More tangible components of the pigment comprise accumulations of red-stained clay (often appearing nearly opaque in transmitted light) in hollows on grains, as partial grain coats and pore linings and in masses associated with the breakdown of rock fragments. These

are present in varying quantity throughout the form-
ation. Where most abundant, excellent and unambiguous
infiltration textures are common. These include the
draping of material across several grains, preferen-
tial concentrations on the upper surfaces of grains,
concentrations to one side of tightly packed wind-
ripple laminae and meniscus-bridging textures (figs.
8.16-8.20). The first three features indicate the
way up of the specimen and all are evidence for the
accumulation of the pigment from influent seepage at
or above the water table. Concentration of the
material in and around decaying rock fragments
illustrates the contribution from intrastratal sources.
This is a classic picture of red-bed diagenesis.

Any optical clues to the composition of these
varieties of pigment are masked by their colour,
though here and there the suspicion is aroused that
some of the material has a relatively high birefringence
suggesting the presence of illite or montmorillonite.
A specimen of the 'colourless' sand in the Preston
Brockhurst roadcut (actually 10YR7/4; very pale brown)
has grain coats with similar habits to those just
described, but colourless. This does indeed have an
illitic/montmorillonitic birefringence, with extinction
parallel to grain surfaces suggesting an influent
origin.

The composition of these phases is believed to
be best represented by the XRD analyses of the
filtrate from ultrasonic agitation of samples, viz.
kaolinite, illite, mixed layer illite/montmorillonite

and hematite, though at least part of the kaolinite must be derived from the pore-filling, non-pigment habit of that mineral. The heterogeneous suite of clays is to be expected from the heterogeneous rock fragment suite, and hence provenance, of the formation.

It is at least curious that the only non-red sand in the formation should occur in the Preston Brockhurst roadcut, the only locality to preserve any signs of biological activity in the formation. The obvious deduction is that some organic matter was preserved in the sediment either to reduce or prevent the oxidation of the pigmenting iron oxides locally. No data (e.g. measurements of the iron states and quantities in red and non-red sediment) is available to justify this statement. That the non-red sediment is very pale brown rather than the usual greenish colour of reduced red-beds (e.g. the clayey silt layer at Eyton) might be held against the theory, but this is speculation - the facts ran out 3 sentences ago.

The last pore-lining clay component to be mentioned is recognisable only under SEM. This is a limited development of boxwork-textured illite distributed over some grain surfaces (figs. 8.16, 8.20, 8.21). The crystallinity of this phase suggests that it is of authigenic rather than infiltration origin.

Feldspar Dissolution and Overgrowths

Degrading feldspar grains make a major contribution to the average of 1.4% recognisable secondary porosity. Feldspars (always alkali feldspar- no plagioclase is

present) are also the only grains in the formation to bear authigenic overgrowths, amounting to up to 1.3% of the rock (average 0.22%). The overgrowths are not affected by any dissolution and may be seen to have developed on and within the remnants of partially destroyed grains (figs. 8.22, 8.23). The overgrowths therefore post-date the dissolution. They vary in development, from being absent to scarcely visible hacksaw-like projections from grain surfaces to complete euhedral rims. In the early stages of development the overgrowths avoid thick accumulations of pigment which may be present on the grain surface. The most complete overgrowths surround the entire grain, irrespective of what lies on the surface. The overgrowths them-selves are always free of both pigment and authigenic illite which they therefore post-date.

Authigenic K-feldspar in British Permo-Triassic sandstones, the Bridgnorth sandstone included, has been studied by Waugh (1978). He found that the overgrowths in a number of formations were all of the same chemical composition; stoichiometric $K\,Al\,Si_2\,O_8$, and structural type; potassian intermediate sanidine.

Pore-Filling Clays

Vermicular kaolinite is the major constituent under this heading, though it averages only 0.4% of the rock. Each accumulation is confined to one or a few pores. It never shows any obvious affinity for a particular type of clastic grain, as for instance the association of kaolinite with degrading feldspar

in the Yellow Sands, and indeed its distribution seems entirely random. It is most abundant in an exposure near the base of the formation at SJ 286250, 4 km south of Oswestry. Here it makes up 3% of the rock, and here also feldspar is most abundant (19.4% of the clastic grains in one sample, average of 3, 14.3%), confounding their apparent independence in thin section.

Under the SEM the kaolinite is clearly identifiable, and it is noticeable that the platelets are not euhedral but tend to be somewhat ragged round the edges (fig. 8.24) perhaps suggesting slight dissolution at some stage.

The textures observed under SEM and in thin section admit only the conclusions that the kaolinite is post-pigment. Its relationship to the illite and the feldspar overgrowths are uncertain.

A second pore-filling phase, not noticed in thin section but seen in one SEM sample, is fibrous illite. It appears to be a development of the grain-coating authigenic illite (fig. 8.21).

Other Minerals.

This category comprises 3 mineral species, each detected at only one locality. A specimen from Blackstone Rock (SO 794740) near Stourport contained 2.7% calcite cement as isolated and degraded poikilotopic rhombs or rosettes up to 0.5 mm across, developed post-pigment. X-ray diffraction of 2 specimens from Hermitage Rock, Bridgnorth (SO 728933)

gave quite strong indications of gypsum, but this was never seen in thin section. Finally, Preston Brockhust roadcut (SJ 542252) contains numerous radiating aggregates 20-50 mm in diameter of poikilotopic barytes.

Paragenesis

There is clear evidence that the pigment minerals in the rock were developed soon after deposition by infiltration in downward percolating water, the dissolution, replacement and redistribution of labile rock fragments, and the oxidation of all this material to transform its iron content to the ferric state. XRD results suggest however that hematite is only a minor constituent of the pigmenting material. Much of the infiltration must have been carried out by rainwater, since the only evidence of any surface drainage within the formation is the single green clayey silt layer at Eyton (SJ 570059), and it seems unlikely that the influx of water that must have accompanied the Pebble Beds could have infiltrated clay into sand extending to depths of 300 m. It is difficult to ascertain the proportions of the pigment due individually to the intractable red staining of all the grains, infiltration, and intrastratal modification. Probably the first and last mechanisms are most important, uniequivocal infiltration textures being present in only one quarter of the thin sections examined.

There is a certain amount of uncertainty about the explanation of the remainder of the diagenesis of

the formation. Waugh (1978) links the development
of illite and feldspar overgrowths to the release
of aluminium, silica, potassium and bicarbonate ions
into the groundwater by the hydrolysis and carbonation
of K-feldspar. The potassium, silica and aluminium
is then used in the precipitation of illite and feld-
spar overgrowths. Such processes are presently active
in Tertiary fluvial arkoses in the S.W. .U.S.A., the
authigenesis being an early diagenetic process occuring
both above and below the water table (Waugh, op. cit.).

This mechanism seems valid but cannot be extended
to account for the authigenic kaolinite. The
precipitation of kaolinite is favoured by "slightly
acid conditions where there is an excess of silica
but few K and Mg ions in solutions. It forms during
early diagenesis by the invasion of groundwater with the
above composition, or during moderate burial depths
(4,000-13,000 feet) in the presence of formation
water of similar composition" (Jonas and McBride,
1977, p.81). The early diagenetic groundwaters in
the Bridgnorth Sandstone, being desert groundwaters,
are likely to have been alkaline, developing the
feldspar overgrowths and illite. There therefore
seems no alternative to attributing the kaolinite to
a later stage process, perhaps to waters rising from
the underlying Carboniferous. There is, however, no
obvious intraformational source for the necessary
alumina and silica- the feldspar overgrowths are
inviolate.

Similarly vague origins must be given for the tiny quantities of barytes, calcite and gypsum present in the formation. The barytes in Preston Brockhurst roadcut is thought to be related to the local Cu-Ba mineralisation, regarded by Poole and Whiteman (1966, pp.56-58) as of Tertiary age. The calcite can only be said to be post pigment and showing signs of dissolution. The gypsum was never seen in thin section.

SECTION 8.5   CONTRASTS WITH THE YELLOW SANDS

The Bridgnorth Sandstone and Yellow Sands were laid down in very different ergs. The Yellow Sands were deposited under a metasaturated sand-drift on the margin of a large sedimentary basin. The bedform pattern (sinuous transverse dunes on longitudinal draa) was very sensitive to wind regime, recording a bimodality. The longitudinal draa probably grew to equilibrium relatively quickly, thereafter remaining stable and retaining very little of the passing sand-drift.

The Bridgnorth Sandstone was deposited under a saturated sand-drift in a relatively small sedimentary basin. The erg had complete sand cover and this probably constrained bedform shape, obscuring the true nature of the original wind regime. The bedforms (transverse dunes on transverse draa) were efficient sand-trappers. They probably grew quite quickly at the upwind margin of the erg, subsequently migrating steadily across it in equilibrium. A significant

proportion of the sand-drift was retained within the erg. Intra-bedform sand transport was much more important than in the Yellow Sands.

Being a sequence of piled up draa, the Bridgnorth Sandstone contains all 4 orders of bounding surface. The Yellow Sands are a single layer of draa and the only first order surface is therefore the base of the formation.

The contrast in bedform shape and erg type has led to a grain size contrast between the formations as mentioned previously. The laminations of the formations also contrast, the Bridgnorth containing more sandflow, less wind-ripple and almost no sand-sheet. This is caused by the lower sinuousity of the Bridgnorth dunes.

Summary

The Bridgnorth Sandstone is mostly of Early Permian age and, as exposed, entirely aeolian in origin. Two scales of cross-bedding are evident, one with trough-shaped sets 2-4 m thick and 20-40 m wide, the other with much more tabular sets 6-10 m thick (up to 30 m) and with a lateral extent of probably 100 m; larger than most exposures. The former is believed to have been deposited by dunes on the leeside of slipfaceless draa, the latter by slipfaced draa. The paucity of exposure and the lack of any stratigraphic markers in the formation precludes the detection of any systematic ordering of the two styles of cross-bedding.

Net sand drift is shown by cross-bedding dip azimuth data to have been unimodal and from east to west. This does not uniquely define the wind regime.

55% of the formation consists of wind-ripple laminae, 44% of sandflows and 1% sand-sheet. Grainfall lamination may have been overlooked. The contrast with the Yellow Sands (figures 71%, 24% and 4% respectively) is attributed to the lower sinuosity of the Bridgnorth Sandstone dunes.

It is possible that there may be a significant difference in maximum thickness between sandflows on draa and dune slipfaces. The limit of thickness of sandflows deposited on dunes in the formation is 80 mm, whereas sets interpreted as deposited by draa slip-faces contain sandflows up to 120 mm thick.

Grain size analysis of the Bridgnorth Sandstone shows it to be somewhat finer than the Yellow Sands. This is believed to be related to the contrasting abilities of transverse and longitudinal draa at trapping sand, the transverse bedforms in complete sand cover holding back coarse grains more effectively. The relationships of grain size characteristics and lamination observed in the Yellow Sands are confirmed in the Bridgnorth Sandstone.

The principal cement of the formation is the pigmenting clay which accumulated during early diagenesis by infiltration and the intrastratal alteration of labile rock fragments. The effects of later diagenesis are minimal.

CHAPTER 9

THE PENRITH SANDSTONE

Introduction

The most complete previous study of the Penrith
Sandstone is that of Waugh (1967) who dwelt particularly
on the petrography and diagenesis of the formation.
Through lack of time the present work is not as detailed
as that related in the previous 4 chapters. It neverthe-
less covers cross-bedding, lamination, petrography and
diagenesis, and serves to supplement, modernise and
dispute some of Waugh's conclusions.

SECTION 9.1 STRATIGRAPHY AND LOCATION

General

The outcrop of the Penrith Sandstone occupies about
300km^2 of the county of Cumbria and is almost entirely
confined to a single river valley, the Vale of Eden.
The outcrop extends from Kirkby Stephen in the south
almost to Carlisle in the north (fig. 9.1). The formation
reaches a maximum thickness of 300-400m in the central
part of the outcrop, from Penrith to Appleby. It is
known to thin markedly to the east, north, and south
(Burgess & Holliday, 1979; Arthurton & Wadge, 1981).
It rests unconformably on Carboniferous and is overlain
by the Late Permian Eden Shales.

Around Appleby and further south the Penrith Sandstone
is interbedded with rudites known locally as "brockrams".
These comprise coarse alluvial deposits generally
divided into a Lower and Upper Brockram.

Waugh (1967) defined the Lower Brockram as a single massive unit up to 150m thick without thick interbedded sandstones, confined to an area within a few kilometres north and south of Appleby. The Lower Brockram, on the basis of its lateral thickness variations and internal structure is considered by Waugh to comprise a single alluvial fan with the sediment source to the S or SW. The Upper Brockram occurs above the Penrith Sandstone on the eastern side of the Vale and comprises wedges of rudite up to 3m thick separated by water-lain sandstones up to 9m thick. It is altogether about 75m thick. Waugh also distinguished the Penrith Brockrams into which the Penrith Sandstone passes laterally around Kirkby Stephen. In this area brockram deposition continued into Eden Shales times.

Brockrams (using the word as a facies term) also occur in the northernmost parts of the outcrop, between Armathwaite and Carlisle (Arthurton & Wadge, 1981).

The information given in Waugh (1967), his Ph.D. thesis, concerning the Brockrams and many other aspects of the Permo-Triassic sequence of the Vale of Eden has never been published elsewhere. Waugh did however publish material concerning the diagenesis of the Penrith Sandstone (Waugh, 1965, 1970a, 1970b). This will be discussed further in Sections 9.3 and 9.4.

The Penrith Sandstone is fairly well exposed by British standards. There are many old quarries in the area north of Cliburn where the formation is well cemented. The only presently active quarry is at

NY 533352 where the formation is worked for rough
cladding, selling at £14.75 per square yard in October
1980. North of Kirkoswald (see fig. 9.1) the River
Eden has cut a gorge up to 100m deep through the
formation, producing numerous large (though generally
wooded) exposures.

Palaeogeography

The palaeogeography of the Vale of Eden during the
Permian to a large extent hinges on the interpretation
of the structural history of the eastern margin of the
Vale, where the Carboniferous rocks of the Pennine
escarpment are faulted against Permo-Triassic rocks.
Waugh's work on the sediments and two recent Memoirs
of the I.G.S. (Burgess & Holliday, 1979, Brough-under-
Stainmore; and Arthurton & Wadge, 1981, Penrith) now
admit firmer conclusions than have ever previously
been possible.

The presence of alluvial facies around the edges
of the present outcrop demonstrates that the Penrith
Sandstone erg probably did not extend much beyond the
present Vale of Eden. The poorest definition of this
is along the western limit of the outcrop; here the
erg may have extended 10 or 20km further. To the south
and west lay the Lake District massif sourcing the Lower
Brockram alluvial fan at Appleby. To the south and
south east both the Penrith Sandstone and Eden Shales
thin and pass into marginal brockrams. To the east
the Upper Brockrams are encountered and the Penrith
Sandstone thins. This last fact - the lack of thick

sediments along the line of the Pennine Fault system - suggests that there was no active faulting on these fractures during the Permian (Arthurton & Wadge, 1981, p.107). The Pennines were however an elevated area and the Penrith Sandstone accumulated in a topographic hollow.

SECTION 9.2 SEDIMENTARY STRUCTURES

Cross-Bedding

The formation is cross-bedded throughout. The maximum thickness of sets generally varies from 2-6m. Where suitable sections transverse to palaeocurrent are available the sets may be seen to be trough-shaped and usually 20-60m wide. Such sections may be seen at Appleby (NY 688200), in Trough Gill (NY 588240), by the River Eden at Baronwood (NY 520435) and in Halfwaywell Quarry (NY 533352). This last locality is shown in fig. 9.2. The whole quarry, some 40m square and 8m deep, is excavated in a single set with its axis directed towards 310° and the limbs dipping to 350° on the SW side and 290° on the NE. The base of the set occurs in the floor of the quarry but the top is not seen. At Cowraik Quarry (NY 542310) near Penrith one set attains a thickness of 10m in a section approximately parallel to palaeocurrent (fig. 9.3).

The largest and least obscured exposure of the formation is in an intermittent cliff section 300m long and 25-30m high on the west bank of the Eden at NY 518442. The section is more or less parallel to palaeocurrent and

a suggestion of a bounding surface hierarchy is evident from simple inspection. Laterally extensive horizontal surfaces are present, apparently truncating others and spaced 6-10 m apart. Between these are sets of cross-bedding 1-4 m thick bounded by surfaces with a general westwards inclination. These sets have internal modification surfaces.

At Bowscar Quarry (NY 519343), Waugh (1967) describes a 140 m long E-W face of the quarry as being occupied by a single set with a sweeping, concave shape in plan. The face is nowhere more than 5 m high and neither top nor bottom of the set are exposed.

Palaeocurrent data for the formation, derived from cross-bedding dip azimuths, is given by Waugh (1967, 1970a) and Arthurton and Wadge (1981). A resultant to WNW is indicated, from unimodal distributions. Waugh found that to the south of Appleby his resultants were to 310°, further north between 270° and 290°. He attributed this divergence to the influence of topography in the south, the hills that provided the source of the brockrams altering the flow of the wind from its regional easterly.

The cross-bedding pattern conforms to the predictions set out in Chapter 2. Its form as broad, sweeping troughs indicates deposition by slightly sinuous transverse bedforms. This is confirmed by the unimodal cross-bedding dip azimuth results. There is no marked bimodality of set size evident as was discovered in the Bridgnorth Sandstone, though careful

surveying and sketching of all the available outcrops might reveal something of this nature. The size of most sets is suggestive of a dune origin. For the thicker and more extensive sets, as for example at Cowraik and Bowscar, deposition by slipfaced draa cannot be ruled out, but they could easily represent the fortuitous perservation of thick dune sets.

Lamination

In all the exposures of Penrith Sandstone visited the observed sandflow thickness never exceeded 80 mm, the maximum usually being between 60 mm and 80 mm.

Despite diagenesis and weathering which, in the cemented horizons, commonly do not enhance the lamination, the distribution and variety of lamination within the formation seem normal. Sandflows dominate at high angles of dip, wind-ripple laminae at low. No sand-sheet material was seen. Once again grainfall lamination may have been missed.

The lamination therefore confirms the suggestion made from the cross-bedding pattern: that only dune slipfaces are preserved and that slipfaced draa probably did not contribute to the Penrith Sandstone.

Other Structures

Contortions of the cross-bedding occur at several localities in the formation. It may be seen by the River Eden at NY 520439, and in two exposures in Hilton Beck (NY 712201). Arthurton & Wadge (1981)

mentioned such features as being present locally at Nunnery Walks (NY 536422) near Kirkoswald. Those examples inspected are the subsurface deformation features of Doe & Dott (1980), though never involving more than a few square metres of the exposure. They pass gradationally into undisturbed sand. All the examples lie fairly close to the top of the formation.

Where the rock is cemented well enough to give bedding-plane exposures, parting lineation is very common. This is not primary current lineation as described by Allen (1968, pp. 29-34) - there is no visible streaking of the sediment. The feature is manifest as a preferential parting direction when the rock is split along thin laminae (fig. 9.5). The parting steps generally face both ways but occasionally are one-sided. Where one-sided (i.e. all the steps face the same way, in staircase fashion) they have a consistent wavelength, generally of the order of 0.1 m. Where they are two-sided there is ususally no consistent wavelength. The parting steps are in most examples oriented parallel to the slope of the cross bedding, but are oblique in some areas. The feature is only developed on wind-ripple laminae. It was also noted (see Appendix) in the Locharbriggs Sandstone (fig. 9.4). The asymmetrical variety was termed pseudoripples by Kocurek and Dott (1981) and envisaged as forming by stair-step fracturing across a succession of wind-ripple or climbing translatent laminae.

The persistent orientation of the parting steps could be due either to a preferred orientation of the grains, with the steps parallel to long axes, or to some other inhomogeneity in grain packing or size. The latter would be envisaged as being parallel to the original ripple crests and might involve pauses in migration or   position, or fluctuations in wind speed. The stair-step, one-sided parting pattern or pseudoripples might closely approximate the original ripple forms; their wavelengths are compatible (0.05 to 0.15 m, typically) and the regularity suggestive.

Waugh (1967) studied grain fabric in the Penrith Sandstone, finding a consistent long axis orientation parallel to cross-bedding dip and also an upwind imbrication of 0.5-12.5°. Unfortunately these results were not related to lamination type.

Two bedding plane sections of wind-ripple laminated Locharbriggs Sandstone measured for the present work showed no well defined grain orientation. A sandflow sample had a well defined preferred long axis orientation parallel to dip, with no imbrication detectable.

These results, though sparse, suggest that Waugh's measurements (7 thin sections of each orientation) were made on sandflow samples. The apparent lack of grain orientation in wind-ripple laminae must indicate that the parting lineation develops along inhomogeneities induced by fluctuations of ripple migration. The lineation is therefore probably a good marker of ripple orientation. How closely the stair-step parting spacing corresponds to original ripple wavelength will be extremely difficult to determine.

SECTION 9.3 PETROGRAPHY

Previous Work

The major features of the diagenesis of the formation are its red colouration, and cementation by syntaxial quartz overgrowths. Other diagenetic phases include feldspar overgrowths, authigenic dolomite, authigenic illite in a pore-lining habit, and very rare gypsum cement and pore-filling authigenic kaolinite.

Waugh (1967) devoted considerable attention to the colouration and quartz overgrowths, following up the latter in his 1970a and 1970b papers. Waugh attributed the red colour of the formation as being the product of red-stained clays derived from humid upland soils developed in the source areas of the Penrith Sandstone during the Coal Measures to Zechstein interval. This is essentially the view put forward by Anderson and Dunham (1953) and is now outdated, having been proposed at a time when the formation of red beds was poorly and incompletely understood. The supposed humid upland soils are most incompatible with the unremitting aridity of the time attested by the preserved sediments. That such soils might have developed before the aridity was suggested by Waugh, but there is then the problem of persuading these soil profiles (which tend to be one-off jobs) to continuously provide red clay throughout the entire period of deposition of the Penrith Sandstone, a period which might amount to 20Ma.

Waugh (1965, 1967, 1970a, 1970b) gave painstaking and detailed descriptions of the distribution, quantity,

textures and development of the quartz overgrowths
in the formation. He showed that in the southern part
of the outcrop, from Appleby to Kirkby Stephen, quartz
overgrowths are completely absent. It is in this area
that brockrams are developed. North of Cliburn and
Kirkby Thore the rock is silicified and "there is a
general uniformity in the degree of cementation" (1970a,
p.1232), though "even within the area of dominant
silification there is an irregular distribution of
uncemented or weakly cemented beds" (ibid., p.1231).
Near Holmwrangle (NY 515489) brockrams appear again in
the formation and in this area the surrounding aeolian
sandstones are unsilicified.

Arthurton & Wadge (1981, pp.72-73) considered that
although the distribution of silicification in the
Penrith district is variable, it is generally concentrated
in the upper part of the formation. The division is
not clear cut though; silicified material being present
but very subordinate at lower levels while unsilicified
rock are found in the top part.

Waugh describes the source of the silica as being
the dissolution by desert groundwaters and dew of
sub-50μ quartz dust particles derived from the natural
abrasion of the sand during transport. This occurred
penecontemporaneously with the deposition of the
formation. The very small size of the particles
invoked as the source of the silica promoted relatively
high concentrations of silica in the dew and groundwater.

According to Waugh, the inverse relationship of brockrams and silicification arises because in the areas of brockram deposition, groundwaters would probably have been enriched in calcium bicarbonate produced by the reaction of atmospheric CO_2 and rain or groundwater with the abundant limestone clasts. This solution, it is stated, "is chemically aggressive and would preferentially dissolve, rather than precipitate, quartz". The upward movement of these groundwaters and subsequent evaporation would give the $CaCo_3$ cement found in the brockrams, but not quartz for some unexplained reason. Away from the brockrams the alkaline ground waters were probably enriched in dissolved silica, as described above, rather than carbonate. Evaporation therefore gave a quartz cement (Waugh, 1970a, p. 1239).

Composition

In this study, the average composition of the Penrith Sandstone was found to be $88.2 \pm 3.3\%$ quartz (range 82-96%), $5.6 \pm 2.69\%$ feldspar (range 1-11%) and $6.1 \pm 2.1\%$ rock fragments (range 2-10%). Most specimens would therefore be classified as sublitharenites or subarkoses according to the scheme used for the Yellow Sands and Bridgnorth Sandstone.

One rock fragment type may be worthy of note. These are grains comprising a network of altered feldspar laths, with interstitial opaque iron oxide probably representing a former ferromagnesian mineral. These may be fragments of dolerite, and if so, derivation from the Whin Sill cannot be ruled out. They were especially abundant in samples from Hilton Beck, half a dozen or so being seen in one slide.

Pigment and Pore-Lining Clays

In repect of these components the Penrith Sandstone is almost identical to the Bridgnorth Sandstone. The visible result of this, the colour of the formation, is also similar to the Bridgnorth though somewhat more variable (Table 9.1; mostly red, reddish brown or yellowish red). This is mostly due to the effect of the quartz overgrowths smothering the pigmenting agents and hence subduing the colour.

All grains in the formation bear an extremely thin red stain which cannot be removed with a needle, like that in the Bridgnorth Sandstone (pp. 228-229). It is at most only 1 or 2μ thick and is present at many grain contacts though absent from some. Where part of a grain surface lies in the plane of a thin section it appears speckled colourless and orange (fig. 9.6), the latter being the staining in question.

The grains also bear a sporadic thicker (up to
10 μ) coat of pigment with accumulations in hollows
in grain surfaces and in association with the
breakdown of labile rock fragments. In a few slides
this shows good menisci and geopetal textures. It may
be present or absent at grain contacts. Where its
birefringence is visible through the colour it is
high, suggesting illite or montmorillonite, with
extinction parallel to grain surface. SEM examination
reveals the presence of boxwork/textured authigenic
illite coating many grains (fig. 9.7). It is stained
red and is absent from the scars of grain contacts.
Suggestions of its presence may be seen in many thin
sections.

Waugh (1967, p. 131) found the various grain
coating phases (concentrated by ultrasonic cleaning)
to be composed of illite and hematite with subordinate
kaolinite. This finding is endorsed here. This
probably represents the grain coating phases though
may include the surface stain. Attempts were made to
concentrate the staining phase by ultrasonic treatment
of samples which in thin section showed the smallest
quantities of the other pigment types. XRD results
from these showed the presence of illite, mixed layer
clays, disordered kaolinite, and geothite.

Quartz Overgrowths

The description Arthurton & Wadge (1981) give of
the distribution of silicification within the
formation is here held to be the most accurate. They
differ from Waugh in laying more stress on the
variability of the cementation, having no illusions
about its uniformity or completeness in any particular
area. However, its confinement to the area north of
Cliburn is undeniable.

Little can be added here to Waugh's description of
the petrography of the overgrowths. Where best
developed they nucleate on every available clastic
quartz grain and plug much of the available pore
space, though residual intergranular porosity usually
amounts to about 10%.

A feature mentioned by Waugh in his thesis (1967,
pp. 132-136), but not in any subsequent publication,
is the inverse relationship of the overgrowths and
pigmenting clays. This may be seen on one scale in
slides where the overgrowths are not well developed;
the inchoate growths avoid areas of thick pigment, and
more widely by the graph of pigment quantity v. per-
cent authigenic quartz shown in fig. 9.9. This shows
that south of Cliburn, where overgrowths are absent
the pigment quantity (including the surface stain,
illite and material from degrading rock fragments)
varies from 5 to 9%. North of Cliburn, where
silicification has occurred, the inverse relationship is
very well marked, pigment quantity varying from 1-8%
and secondary quartz from 0-26%.

Any petrographic evidence for the place of the
quartz overgrowths in the diagenetic sequence must be
crucial in formulating and evaluating hypotheses for
their origin. Thin sections show that the quartz
overgrowths post-date almost all of each pigment type,
the authigenic feldspar and kaolinite, and the pre-
cipitation of dolomite. They may pre-date the
dissolution of the feldspar overgrowths, but the
direct evidence on this point is not clear.

The overgrowths are a post-compaction feature, minus-cement porosities indicating 20% compaction before silicification. Specimens with well developed overgrowths generally do not show pressure solution. This may be because the overgrowths pre-date the pressure solution and prevented its development, or merely a reflection of the fact that pressure solution is most common low in the formation, away from the cemented areas. Also, pressure solution is known to be promoted at grain contacts bearing a certain amount of clay, the clay appearing to act as some kind of catalyst.

Other Minerals

Syntaxial overgrowths on feldspar grains are present throughout the formation, constituting up to 1.3% (average 0.25%) of the rock in the slides studied. The overgrowths, like their hosts, are always of alkali feldspar and show all stages of growth from being absent to complete euhedral rims, exactly as is the case in the Bridgnorth Sandstone. Waugh (1978) found the overgrowths to be composed of stoichiometric $KAlSi_2O_8$, structurally arranged as potassian intermediate sanidine.

Unlike the Bridgnorth Sandstone, all the overgrowths show much the same degree of dissolution as their host grains and also in some cases bear pigment. In some examples this appears to be an influent coat of illitic material oriented parallel to the grain surfaces, in other cases it may be authigenic boxwork illite; the distinction is often difficult in thin section.

Both the overgrowths and some of the subsequent dissolution developed before compaction. This is shown where partially dissolved grains have collapsed under pressure, and rare examples of feldspars with overgrowths pressolving into other grains.

Only one surviving dolomite crystal was seen in all the thin sections studied. However, there is evidence that it was formerly much more abundant. This is manifest as moulds of euhedral $10-20\mu$ rhomb-shaped

crystals within quartz, and very rarely feldspar,
overgrowths. In specimens where overgrowths are
present these moulds average 0.3% of the rock, ranging
from zero to 2%. They are visible in fig. 9.8. The
moulds often have a lining of pigment which appears to
be of the grain coating type, though as in the case of
the feldspar overgrowths distinction between this and
the boxwork illite is not easy in thin section. At
only one locality were the moulds found to be absent
from all specimens of silicified sandstone. This was a
quarry at NY 583264, 2 km north of Cliburn.

Porosity and Compaction

Present porosities range from 15-25% for
unsilicified parts of the formation, getting down to
10% and less where silicification has occurred.
Secondary porosity has developed by the dissolution of
feldspars and rock fragments. This is found in all
specimens and amounts to an average additional porosity
of 1.4%, with a maximum of 4.3% in one specimen.
Moulds of dolomite crystals are found in quartz
overgrowths, these account for a further 0.3% secondary
porosity on average.

The minus-cement porosities of wind-ripple
laminated samples averaged 18%, those of sandflow
samples 25%. The porosities were measured by point-
counting, classing quartz and feldspar overgrowths as
cement. Following the exercise of Chapters 7 and 8 and
comparing these with Hunter's (1977) figures for modern

dunes indicates that the Penrith Sandstone has been compacted by about 20%, somewhat more than the Yellow Sands and Bridgnorth Sandstone.

Commensurate with this compaction there is good evidence for pressure solution in the formation. Waugh (1967, pp. 125-131) documented this point with detailed measurements of grain contact numbers and types. As he states, pressure solution is best (but not exclusively) developed in the lower levels of the formation, though even there the effects of the process are slight.

SECTION 9.4 INTERPRETATION OF THE DIAGENESIS

The earliest diagenetic modification of the formation was its acquisition of the pigment, by exactly the same methods as identified for the Bridgnorth Sandstone: the infiltration of clays in rain and floodwater, and the intrastratal breakdown of labile grains. Some of the pigment was probably acquired before final deposition, how much remains uncertain.

While the pigment was still forming, the dolomite was precipitated, probably by desert groundwaters.

The feldspar overgrowths in the Penrith Sandstone do not fit into Waugh's (1978) hypothesis, which was used to explain the origin of such features in the Bridgnorth Sandstone. Waugh envisaged part of the constituents necessary for the precipitation of authigenic feldspar as being derived from the previous dissolution of clastic feldspar grains. This would lead to fresh overgrowths on corroded hosts, as is found in the Bridgnorth Sandstone. In the Penrith

Sandstone the overgrowths are just as fresh or corroded as their host grains - evidently the dissolution occurred after the overgrowths had developed. The necessary material could, however, have been derived from the dissolution of other labile mineral and rock fragments. The authigenic illite could be derived from similar sources, and both probably formed relatively soon after deposition.

Waugh's hypotheses for the origin of the authigenic quartz - dissolution and reprecipitation of quartz dust by dew or groundwater, and differing groundwater composition in brockram and brockram-free areas - have a number of weaknesses. Firstly, note that the processes of solution and precipitation of abrasion-derived quartz dust are not specific to the Penrith Sandstone. If these events can occur in the Penrith Sandstone reasons need to be offered to explain why they are not evident in every other aeolian sand under the sun, ancient and modern. Yet the degree of silicification suffered by the Penrith Sandstone is unique among British aeolian sandstones. Surveying the literature, records of modern silicified aeolian sands are conspicuous only by their absence. Waugh offers the example of Williamson (1957), whose silicified "aeolian" sands were apparently only identified as such on the basis of the rounding and sorting of the grains; no reference is made to field relations.

Waugh (1970a, p. 1236) suggests that cementation was penecontemporaneous with deposition, with silicified crusts forming at the surface. If this were

the case, there ought to be some evidence in the sediments - the formation of such crusts could freeze the whole erg. There is nothing within the formation to suggest that this took place. The examples of contorted bedding positively refute such a possibility. Modern silcretes develop very slowly in areas of minimal erosion and deposition. Active ergs do not fit into this category.

If, as Waugh suggests, siliceous dust was being completely dissolved by desert dews, one might expect to find evidence of at least some corrosion of the framework grains, but there is none. If the cement were precipitated early and from evaporating dew, commensurate textures should be evident; for example, the concentration of cement at grain contacts through precipitation from menisci, and the preservation of a very high minus-cement porosity, but this is not the case. Compaction of some 20% had occurred before cementation. This is hard evidence that the silicification took place at depth - the Penrith Sandstone is more compacted than either the Yellow Sands or Bridgnorth Sandstone.

Waugh's hypotheses can therefore no longer be retained. The siliceous dust invoked as the source of the silica probably never landed on the desert floor - it was just picked up into suspension and blown away.

The quartz overgrowths are a product of diagenesis at depth. The variability of the cement within the silicified area is attributable to the inhibiting effect clay pellicles around grains have on cementation

(e.g. Heald & Larese, 1974), clearly shown by fig. 9.9. Since the quantities of pigment found in the silicified and unsilicified areas (north and south of Cliburn, respectively) overlap considerably it is surprising that there is no secondary quartz at all in the unsilicified area. This might suggest that the conditions which brought about the cementation did not pertain south of Cliburn. It is possible that this is due to the influence on the pore water chemistry of the brockrams with their abundant dolomitised and undolomitised limestone clasts and calcite cement.

The brockrams may account for any increase in the quantity of pigment present in the southern half of the outcrop. The presence of the brockrams implies sporadic surface drainage with subsequent influent seepage and this must have carried clays into the underlying sediment. Also the brockrams are more likely to contain mechanically and chemically fragile rock and mineral fragments which might then break down subsurface, releasing pigmenting material. Thus both main processes of pigmentation are aided by the presence of water-lain sediments.

There is no credible intraformational source for the silica which is now incorporated in the quartz overgrowths. Nor is there any compelling extraformational source.

Bleached and silicified veins are present in most outcrops and locally abundant, being most apparent in unsilicified rocks where their presence is accentuated by weathering. Sections cut of these showed them to

consist of clastic grains (without overgrowths), some
possibly fractured, set in a groundmass of crystalline
to microcrystalline silica, the finer clastic grains
merging into the coarser groundmass crystals. Whether
these veins have anything to do with the cementation is
uncertain - not enough is known of their field and
textural relations to the Penrith Sandstone and
overlying rocks. Nevertheless, they merit a mention,
if only for completeness.

Other straws that might be clutched at are clay
mineral transformations in the overlying Eden Shales or
silica internally derived from pressure solution.
However, the Eden Shales are only ~ 150 m thick at
maximum and the effects of pressure solution in the
Penrith Sandstone are slight. Neither seems capable of
producing sufficient quartz to accomplish the
cementation observed.

Summary

The Penrith Sandstone is a red aeolian sandstone of Early Permian age. Facies and thickness patterns suggest that the original erg may not have been much larger than the present outcrop area of some $300km^2$. The formation is cross-bedded throughout, the sets generally taking the form of troughs 2-6m thick and 20-60m wide. Sandflows reach but do not exceed 80mm in thickness. The cross-bedding pattern and sandflow thickness data suggest that the formation was deposited by slightly sinuous transverse dunes on slipfaceless draa.

The major features of the diagenesis of the formation are its red colouration and extensive cementation by authigenic quartz. The red colouration developed by the infiltration of clays bearing iron oxides and the intrastratal breakdown of labile grains. Textures indicative of accumulation above the water table are evident.

The quartz cementation is present only in the northern half of the outcrop and is there concentrated in the higher parts of the formation, though its distribution is everywhere patchy. This unevenness of silicification is determined by the quantity of pigmentating clays in the rock, pigment and cement having a clear inverse relationship. The quantity of pigment is enhanced in areas where contemporaneous water-laid sediments are present. The silicification took place at depth, after 20% compaction had been achieved. Previous ideas of attributing the cement to a syndep-

ositional process of dissolution and reprecipitation of quartz dust are rejected. There are no obvious sources for the silica present in the quartz overgrowths.

CHAPTER 10

THE SIGNIFICANCE, CHARACTERISTICS AND IDENTIFICATION

OF AEOLIAN SANDS

Introduction

The aim of this chapter is threefold: firstly,
to summarise and assess the palaeogeographical and
stratigraphical implications of the previous
chapters; secondly to describe and review the
sedimentology of aeolian sandstones, as presented
in the British geological record; and thirdly, to
set out criteria for the identification of aeolian
sandstones.

SECTION 10.1   THE PALEOGEOGRAPHIC AND STRATIGRAPHIC
SIGNIFICANCE OF AEOLIAN SANDSTONES IN
THE U.K.

Stratigraphy

Much of the British Permian record consists of
unfossiliferous continental sediments, and consequently
the existing correlation is largely lithostratigraphic.
However, the sedimentary sequences, viewed on a large
scale, are broadly similar in many areas and an
important aspect of this similarity is the widespread
occurrence of thick aeolian sandstones in the lower
part of the Permian.   These formations are regarded as
an important stratigraphic marker (Smith et al., 1974).
There are 6 such formations in S.W. Scotland, the

Penrith Sandstone in Cumbria, the Yellow Sands in
N.E. England, the Collyhurst/Kinnerton/Bridgnorth
Sandstone unit in the West Midlands and N.W. England,
and the Dawlish Sands in S.W. England. All these
formations seem to represent the onshore margins of
four major sedimentary basins: the Northern and
Southern Permian Basins in the North Sea, the East
Irish Sea Basin, and a basin under the present
English Channel (palaeogeography on fig.5.2b). The
Yellow Sands, the Penrith Sandstone, the Collyhurst
Sandstone and parts of the Bridgnorth Sandstone are
effectively bracketed as Lower Permian by fossils
in conformably overlying strata.

Most of the formations are thick, exceeding
200 m, though the contrast between the Locharbriggs
Sandstone (1000 m) and the Yellow Sands (mean spread
out sand depth ~ 15 m) should be borne in mind.

Smith et al., (1974) suggest that aeolian
deposition took place at roughly the same time in
each area, the gross aspects of sedimentary record
being determined by climate and the overall evolution
of the landscape. The incoming of the Zechstein and
Bakevellia Seas east and west of the Pennines is
thought to have been synchronous and glacio-
eustatically prompted. This transgression terminated
aeolian deposition in the North and Irish Sea areas
and also in the Vale of Eden. The climate over the
British Isles is then said to have become somewhat
wetter, inducing fluvial action wich replaced aeolian
deposition in the remaining land areas.

These assumptions seem reasonable and produce a correlation which is satisfying and on the whole consistent. The only problem would seem to be the possible persistence of aeolian sedimentation into Late Permian times in the Cheshire Basin, where the marine, early Late Permian Manchester Marls are thought to pass laterally into aeolian sandstone of the Collyhurst/Kinnerton/Bridgnorth unit. This persists up to the erosive base of the Bunter Pebble Beds (= Kidderminster Conglomerate Formation = Chester Pebble Beds), regarded as Lower Triassic. This boundary represents an exceedingly abrupt and complete change in sedimentary environment, and one that occurs over a very wide area: from Worcestershire to Merseyside and probably beyond. Any time gap represented by the boundary may never be quantified, unless by some fortuitous remittance of the problems surrounding the measurement of palaeomagnetism in sediments.

A further 4 units of aeolian sand are found in the Lower Triassic: the Hopeman Sandstone, Kirklinton Sandstone, Frodsham Member and Otter Sandstone. These are considerably thinner than their Lower Permian counterparts, ranging from only 45 m to 120 m. The Hopeman Sandstone is part of only a very thin known sequence; too thin to define any overall trends in the record, but seems to be earliest Trias by the reptilian fauna (Peacock _et al._, 1968). The Kirklinton Sandstone occurs above a thick fluvial sequence (the St. Bees Sandstone) and below thick

mudstones (the Stanwix Shales). The Frodsham Member also occurs above a thick fluvial sequence (underlain successively by the rest of the Helsby Sandstone, the Wilmslow Sandstone and the Chester Pebble Beds), but lies below the supposedly marine Tarporley Siltstone (= Waterstones), which in turn are succeeded by thick mudstones (the former Keuper Marl). The Otter Sandstone rests on the fluvial Budleigh Salterton Pebble Beds and underlies mudstones.

Thus, all but the Hopeman Sandstone lie in broadly comparable sequences and it may well be that the Kirklinton Sandstone, Frodsham Member and Otter Sandstone were being laid down at the same time. They are tentatively shown so on the correlation charts of Warrington et al. (1980), at around the Scythian-Anisian boundary, though no explicit mention is made of reasoning similar to that of Smith et al. (1974).

Wind Regime and Climate

Units of aeolian sand are rare in the U.K. Old Red Sandstone, strikingly so when compared to their abundance in the New Red Sandstone. Presumably this must indicate that the climate under which the O.R.S. was deposited was not dry enough for fluvial deposition to relent for sufficiently long periods for aeolian sands to accumulate.

Fig. 4.3 shows 8 cross-bedding dip azimuth resultants from various Lower Permian aeolian formations in Britain. Although there is an overall uniformity, in that all the resultants have a dominant westerly component, there is considerable divergence from

one formation to the next. The Corrie Sandstone,
Locharbriggs Sandstone and Yellow Sands directions
lie between SW and WSW, whereas the Penrith Sand-
stone, situated midway between Dumfriesshire and
County Durham, gives a result to the WNW. This is
a divergence of 60° from its nearest neighbours.
The sandstones in the Vale of Clwyd, in southern North
Sea gas wells, and the Bridgnorth Sandstone give
results to due west. The Dawlish Sands contrast
strongly, with a resultant to the NNW.

Apart from the Dawlish Sands, the main oddball
is the Penrith Sandstone result. The peculiarities
of the Yellow Sands results must not be forgotten -
its bimodality (to the WNW and S), and the possibility
that the incomplete sand cover may make the formation
more sensitive to the real wind regime. In discussion
of the Yellow Sands wind regime in Chapter b, it was
suggested that the bimodality reflected real winds,
either a summer N'ly and winter ESE'ly or a regional
ESE'ly modified by a katabatic N'ly. The Yellow
Sands WNW mode parallels the Penrith Sandstone
resultant.

Three questions can be asked :

(i) in northern Britain, was the regional wind
unimodal or bimodal, and is the true resultant
to the WNW or WSW?

(ii) Are the northern (1-4 on fig. 4.3) and Midlands
(5-7 on fig. 4.3) results compatible?

(iii) Why are the Dawlish sands so different?

In answer to the first question, it could be argued that the true resultant wind was ENE'ly (parallel to the Yellow Sands, Locharbriggs and Corrie results) and that the Penrith anomaly is due to local orographic flow; deflection off the Lake District Massif for instance. Following this argument necessitates consideration of whether the bimodality of the Yellow Sands is valid, whether it should also apply to the Locharbriggs and Corrie results, why only one mode should apply to the Penrith Sandstone, and whether any light is consequently shed on the origin of the bimodality.

The alternative, that the true regional wind was WNW'ly implies that such local forces also prevailed at Locharbriggs and Corrie, but not in the Penrith Sandstone.

In the author's opinion the issue cannot be decided - the indeterminacy and plausibility of local orographic effects render every option credible.

To the second question - the compatibility of the northern and midlands results - a similarly nebulous (but honest) answer must be given. All the available evidence points to the concurrence of aeolian deposition in the two areas; therefore their respective wind regimes must have been compatible. Presumably the 170 km distance between the two areas was sufficient to allow a divergence of net sand-drift and airflow of 25°-30°, whatever the true resultant in the north is said to be. Orographic effects cannot be ruled out in the midlands area - neither the Vale of Clwyd, Bridgnorth or the southern North Sea gas fields are far removed from the presumed basin margins and possible areas of high ground.

The winds in Devon must also be put down to orographic effects - there seems little option when the result diverges from those of the remainder of the formations by 45° to 105°.

SECTION 10.2   THE CHARACTERISTICS OF AEOLIAN SANDSTONES

Cross-Bedding

The vast majority of modern ergs contain both dunes and draa, draa always bearing superimposed dunes. Draa are seldom absent, typically in areas of low sand cover or around the periphery of an erg, in the transition from an environment of aeolian deposition to one where water-lain sediments are slowly accumulating.

The climbing of these bedforms during migration should lead to cross-bedding showing a hierarchy of bounding surfaces comprising up to 4 orders. First and 3rd order bounding surfaces form by the migration of draa and dunes respectively, and are consequently termed migration surfaces. Second and 4th order bounding surfaces develop by the erosive modification of draa and dune shape caused by changes in the air-flow pattern. These are termed modification surfaces.

Two styles of cross-bedding should be found. The first, showing only 1st and 2nd order bounding surfaces, consists of thick and laterally extensive sets deposited by slipfaced draa. The second type displays all 4 orders of bounding surface, though 2nd order may be rare. Third order surfaces are strung

between 1st order surfaces, dipping gently down-current. The third order surfaces bound sets of cross-bedding, whilst first order surfaces define cosets. This type of cross-bedding is deposited by dunes migrating down the leeside of a migrating and climbing slipfaceless draa. The cross-bedding cannot be fully understood unless allowance is made for the random component of bedform behaviour.

Mechanical and topological considerations demand that where an erg is aggrading (i.e. virtually all ancient U.K. examples), the cross-bedding should be formed only by transverse dunes and draa, though the sinuousity of the bedforms may vary. Longitudinal and star-shaped bedforms cannot form thick, continuous sequences of aeolian sand because they lack continuity in the direction normal to migration.

The thickness of individual sets depends on the migration rates of the depositing bedforms, the angle of climb of those bedforms, and for dunes, the angle of slope of the draa leeside. Draa sets should be thicker and more laterally extensive than dune sets. These hypotheses are set out in detail in Chapter 2. Quintessentially, the theory is merely a reasoned geological judgement of the demands and effects of preservation potential on a realistic perception of the modern situation.

All of the formations studied are cross-bedded throughout, and this cross-bedding conforms to the pattern set out above. When cut transverse to palaeo-current the cross-bedding is invariably seen to consist of trough-shaped sets of varying width and

depth. The lower bounding surface cuts erosively into material below and is concave upwards, as are the laminae within the set, these approaching the lower surface asymptotically. The troughs may be so broad and the lower surfaces so flat as to merit description as tabular-planar cross-bedding in some cases.

When cut parallel to palaeocurrent the bounding surfaces are generally subhorizontal and the laminae tangentially based. The sets have a finite length, either lensing out or apparently erosively removed by a successor.

This pattern is indicative of sinuous transverse bedforms, confirming the postulated restriction on the shape of perservable bedforms.

Analysis of the patterns of set thickness and bounding surfaces requires large exposures. Typically the cross-bedding is disposed in sets 2-6 m thick and 20-60 m wide. These are believed to be dune sets, and this is confirmed where bounding surface analysis is possible. Reliable confirmation that the cross-bedding pattern is due to dunes on slipfaceless draa, and not simply to dunes in a draa-less erg, is possible only in a few large exposures cut parallel to palaeocurrent (in the Penrith Sandstone by the River Eden at NY 518442, in the Locharbriggs Sandstone at the Locharbriggs Quarries, and in the Yesnaby Sandstone in sea cliffs at HY 219153).

Sets of a size and with a bounding surface system compatible with deposition on slipfaced draa are

relatively rare in the British record. The Bridgnorth
Sandstone is most instructive in this respect, since
this is interpreted as containing the deposits of
both slipfaced and slipfaceless draa (Chapter 8).
Dune sets in the formation are 2-4 m thick and 20-40 m
wide. Sets interpreted as deposited by slipfaced draa
are consistently 6-10 m thick and extend laterally for
at least 100 m. Occasionally, examples exceed 20 m in
thickness. The bounding surface pattern is consistent
with this interpretation, but exposure is not good
enough for it to be positively confirmed.

Slipfaced draa sets are also present in the
Corncockle Sandstone (Appendix   ) where a set 30-40 m
thick is visible in one locality, and in the Hopeman
Sandstone (Appendix   ), where similar dimensions are
attained.

One of the major lessons gained from the research
recounted in this thesis is the realization of the
difficulty of confidently labelling bounding surface
types. The concept and enactment of the hierarchy is
complicated by the sequential development of the
surfaces. When one bounding surface truncates another
below, the erosion recorded does not necessarily
reflect a hierarchical relationship. For instance, it
is possible to envisage dunes migrating down a draa
leeside, hence generating 3rd order surfaces, cutting
across and truncating a previously developed 2nd order
surface. Only where one surface approaches another
from above and is truncated is a hierarchical relation-
ship proven.

Where the base of a draa leeside is occupied by dunes, dune-sized scours are probably cut into the first order surface being created by the migration of the draa. Thus, when cut transverse to palaeocurrent, the resulting cross-bedding pattern will show no evidence of relatively flat, laterally extensive first order surfaces, appearing to record just a succession of stacked dune sets. The first order surfaces should however, be detectable in suitably long sections parallel to palaeocurrent (see fig.2.16).

It is relatively easy to distinguish migration surfaces from modification surfaces, but to separate these types into their components can be exceedingly difficult. In general, migration surfaces are sub-horizontal and truncate the laminae below with considerable angular discordance, whilst the laminae above meet the surface tangentially or asymptotically, increasing in dip away from the surface (these relationships hold only in sections parallel to palaeocurrent). Modification surfaces often dip at an angle approaching that of the lamination with only a gentle discordance below. The laminae above tend to approach tangentially and may show only a slight increase in dip away from the surface. The modification type is also manifest as 'hanging' surfaces, that is, surfaces present in the top part of a set, but which do not extend to the bottom, apparently being replaced by concordant laminae.

The cross-bedding displayed by the formations inspected is dominated by dune migration surfaces

(3rd order). First order surfaces are recognizable in suitably large and appropriately oriented sections. Modification surfaces are present in many sets, but are not nearly so densely developed as in the active dunes trenched by McKee (1966). This is because only the lower part of bedforms is preserved, and these are areas last and least affected by the fluctuation which cause 2nd and 4th order surfaces - preservation potential filters out a proportion of the randomness of bedform behaviour.

The rule that only transverse bedforms are preserved has one notable exception. This is the Yellow Sands of N.E. England, where longitudinal draa are preserved beneath a suspected 'instant' transgression (Chapters 5, 6, 7). The Yellow Sands exist because of this unusual preservation - they had little chance of surviving otherwise. Fig. 5.4, the thickness map of the Yellow Sands, constitutes the most explicit evidence yet recorded for the existence of draa in the geological record. All other examples, though often fairly unambiguous, rest to a certain extent on speculations and theories about bedform behaviour. Little sand has been removed from the Yellow Sands draa in preservation and the quarries dug in the formation probably represent the best sections through complete aeolian bedforms available anywhere in the world.

An account of the cross-bedding of aeolian sandstones must take note of the many aeolian sandstones magnificently exposed in the Colorado Plateau

area (e.g. lists in Walker and Middleton, 1979, and McKee, 1979b, p. 190). Especially notable is recent excellent work by Kocurek (1981*a,b*   ) on the Jurassic Entrada Sandstone. The quality of exposure available in the U.S. formations is amply illustrated by the two sections featured in Kocurek (*1981a*   ). Combined, these two expose 10 times more rock than is drawn on the Yellow Sands sections included in this thesis (Enclosures 1-8). Brookfield (pers. comm.) mentions an exposure of the Navajo Sandstone 300 m high and 240 km long!

Stokes (1968) described multiple-parallel truncation planes bounding very thick sets of cross-bedding from several U.S. formations. Given the lateral extent of these planes, the size of the sets between them (actual typical figures are hard to extract from the literature, but 15-30 m does not seem exceptional), and the difficulties inherent in Stokes' interpretation (that they mark deflation of the whole erg to the water table) these planes are now regarded as 1st order bounding surfaces and the cross-bedding as deposited by slipfaced draa (Brookfield, 1977; Kocurek, *1981b*   ). These sets are generally tabular-planar in form and it is this that probably drove McKee (1979b, p. 192) to state that cross-bedding in aeolian deposits is mostly tabular-planar or wedge-planar, with trough-shaped types rare. Such a conclusion could not possibly be reached from the British record.

A further contrast with the U.K. is documented by
Kocurek (1981a, b). In the Entrada Sandstone of
Colorado and Utah, 'interdune' deposits are preserved
along most of all the 1st order bounding surfaces
present in the sections presented. These 'interdune'
deposits are horizontally bedded, up to 4 m thick and
contain such features as water-lain current ripples,
small deltas, algal structures, adhesion ripples,
deflation scours, lag surfaces, sand-sheet deposits and
thin aeolian dune (s.s.) sets (Kocurek, 1981b). These
surfaces are deduced by Kocurek (1981a, fig. 10) to
have been interdraa planes some 400 m long (parallel to
palaeocurrent) and 1500 m wide. Interbedform deposits
occur in several other U.S. formations (Kocurek, 1981b)
and in the Southern North Sea Rotliegendes (Glennie,
1972). The significance of these features is that, in
the Entrada at least, the interdraa areas migrate and
climb <u>with</u> the draa. These, therefore are examples of
ergs aggrading with incomplete sand cover. These are
accommodated in the erg models derived in Section 1.3
(the accommodation being prompted by Kocurek's work).

The only interdraa deposits known to the author in
the onshore U.K. are those exposed in the Preston
Brockhurst road cut (pp. 217-218 ; figs. 8.4, 8.5;
Enclosure 14), and the "playa" sediments between draa
sets in the Hopeman Sandstone described by Williams
(1973; see also the Appendix to this thesis).

Many different approaches can be used in the
reconstruction of bedform types in aeolian sandstones.
If the sandstone is more than a few metres thick draa
should be recorded; this may be verified and identifi-
cation of slipfaced and slipfaceless types made by
inspection of the bounding surface pattern, set and
sandflow thicknesses, and the lateral extent of sets.

The sinuousity of the bedforms may be observed in horizontal exposures or inferred from foreset dip azimuth patterns. Kocurek (1981) estimated original dune height in the Entrada Sandstone by utilizing Finkel's (1959) data of barchan dune morphology. Finkel gives a height to width ratio of about $\frac{1}{10}$ for barchans, and Kocurek therefore suggested that 8 m high dunes laid down his trough-shaped sets, which reach a maximum width of some 80 m. However, Finkel's figures imply that simple transverse dune should have an infinite height and use of this data on sinuous transverse bedforms can barely be justified. For the Entrada draa, Kocurek (1981) estimates bedform width from the extent of interdraa deposits and calculates draa length by projecting 2 successive 1st order bounding surfaces onto a horizontal plane and measuring the distance between them. Both these techniques require U.S.-style exposure. By draa "length" is meant the distance from the foot of the windward side to the foot of the leeward side. Draa wavelength equals this plus the length of an inter-draa area. Kocurek seems to be mistaken in describing and treating the figure for draa length as draa wave-length - he resorts to the data of Breed and Grow (1979) to derive a false draa length from his "wave-length".

The upshot is that he cites Wilson (1972, 1973) as giving a ratio of draa height to wavelength of about $\frac{1}{15}$ and deduces an Entrada draa hight of about 110 m. Given his mistake, this

must be an underestimate.

The relationship of lamination distribution and sandflow thickness to bedform size and type is discussed in the next subsection.

Ancient ergs typically are lenticular or half-lenticular bodies of sand showing considerable thickness variations (e.g. the Penrith Sandstone, Chapter 9, the Bridgnorth Sandstone, fig. 8.1; the southern North Sea Rotliegendes, Glennie, 1972). Depending on the manner in which ergs aggrade, this may have implications for the average set thickness of cross-bedding deposited.

An erg in an endoreic basin with a central playa is shown in fig. 10.2a. Deposition in this situation may be controlled by either subsidence or sand supply. However, if deposition outpaces subsidence, the drainage may eventually be reversed and the central playa destroyed. On the left hand side of fig. 10.2a the erg margin remains more or less fixed and draa nucleate and grow at this margin before migrating inwards (the draa could equally well form at the centre and migrate outwards). In this model every part of the erg is traversed by the same number of draa and therefore the erg body is potentially composed of the same number of 1st order sets. Since the erg thickness decreases towards the margin the average set thickness should also. On the right hand side of fig. 10.2a a different pattern is drawn. Here the erg has expanded laterally as it has grown in thickness. Thinner areas

are therefore made up of fewer sets and set thickness
need not vary. This implies that the area where sand-
drift decreases with distance has increased gradually
with time and possibly more likely indicates control
by subsidence.

Analogous situations are drawn in fig. 10.2b for
the case of an erg accumulating as a prism above the
surrounding ground surface. On the left hand side the
erg margin is assumed *to be* fixed and *to be* the site of bedform
generation (or dissipation). Again set thickness
decreases toward the margin. On the right hand side
the erg is drawn as expanding with aggradation,
producing sets that do not necessarily vary in thick-
ness. Again this may more likely indicate subsidence,
though on a scale much larger than that of the erg.

Unfortunately none of the U.K. formations
studied are well enough exposed to test these
possibilities,   nor is there sufficient data on
other areas in the literature to make any constructive
comments. The validity of the suggestions therefore
remains unconfirmed.

Lamination

Two types of lamination make up most of the
(U.K.) formations studied. These are sandflow
cross-stratification and wind-ripple lamination. An
additional, rare type is sand-sheet lamination. These
are described and illustrated in Chapter 3. The
grainfall and planebed lamination types described by
Hunter (1977, 1981) have not been recognised; the

former possibly by remission, the latter because of
non-preservation. If present, grainfall lamination
is probably rare and may have been passed over as a
fine and faint wind-ripple lamination. Kocurek
(pers. comm.) reports having a "devil of a time"
distinguishing them in the field.

Sandflows are found only at angles of dip
greater than about 20°. They are formed by the
cohesionless avalanching of grains initiated on
oversteepened leesides. The relatively low angles
of dip at which they may be found are partly an
effect of compaction and partly may be an illus-
tration of the momentum such flows can build up on
large bedforms.

Wind-ripple laminae are formed by migrating
and climbing wind ripples. They may be found at any
angle of dip and are the dominant stratification
type at dips lower than 20°.

Sand-sheet material is an interbedform or
marginal erg deposit. It is found at low angles
of dip and is distinguished from wind-ripple
lamination by the presence of small scale cross-
lamination believed to be deposited by megaripples,
the greater irregularity of the lamination, the
presence of discrete lenses of coarse sand, and its
strikingly bimodal grain size distribution.

The proportions of the various lamination types
present within a formation are governed by the types
of bedform present in the original erg. Sand-sheet
material is rare or absent in formations deposited

from transverse bedforms in an erg with complete
sand cover. It is however present in the Yellow
Sands (longitudinal draa, incomplete sand cover) and
in formations of transverse draa with incomplete
cover (e.g. Kocurek, 1981). In the former instance
its presence is attributed to the ease of penetration
of a longitudinal draa erg by coarse sand and the
proximity of potential sand sources. In the latter
case it may record prolonged and repeated deflation
of the interbedform areas removing the medium sand
to leave a bimodal lag.

The relative proportions of wind-ripple and
sandflow lamination depends to a large extent on
the sinuousity of the formative (transverse) bedform.
More sinuous bedforms tend to have longer toesets
(Allen, 1968; Hunter, 1981), which are made up of
wind-ripple laminae. Thus in the Yellow Sands,
deposited from strongly sinuous bedforms, the
proportions of wind-ripple to sandflow to sand-sheet are
71:25:4 (Chapter 7). In the Bridgnorth Sandstone,
deposited by much less sinuous bedforms these ratios
are 55:44:1 (Chapter 8). Slipfaced draa sets
probably contain more sandflow stratification than an
equal thickness of dune sets: thus bedform order may
affect the lamination proportion. Also, in a slowly
aggrading erg where only relatively thin sets are
preserved, the proportion of toesets, and hence wind-
ripple laminae, will be enhanced.

The maximum sandflow thickness found on any
slipface is found to increase with the height of that
slipface (Kocurek and Dott, 1981). In the formations

studied, sets interpreted as deposited by dunes
contain sandflows up to 80 mm thick. Those inferred
to be the deposits of slipfaced draa contain sand-
flows up to 120 mm thick. If confirmed this may be
a useful method of distinguishing draa and dune sets
in restricted exposures.

Fig. 10.3 is a plot of the maximum exposed set
thickness v. maximum recorded sandflow thickness for
the 9 formations for which   data are available. The
plot shows a bimodal grouping into those where only
dune sets (i.e. dunes on slipfaceless draa) are
preserved and those believed to preserve slipfaceless
draa. The Entrada Sandstone, though containing
slipfaced draa deposits and having a maximum avalanche
thickness of 90 mm (Kocurek and Dott, 1981) is some-
what removed from the bimodality since it has a
maximum set thickness of only 8 m. The general
increase of sandflow thickness with set thickness
among the formations preserving dunes may be a
spurious effect of the small data base. If not, 2
factors contributing to its occurrence can be
envisaged. Since sandflows lens out downslope,
being replaced by wind-ripple laminae, the thickest
parts of individual flows are found part way up the
slipface. Increasing truncation below this point
will lead to lower preserved sandflow thicknesses.
Thus if a 15 m high dune were truncated to form 7 m
and 2 m thick sets, the 7 m set would most probably
give the greatest sandflow thickness. The second
factor arises because of the broad correlation which
must exist between bedform height and preserved set

thickness. Taller bedforms contain greater volumes
of sand and therefore in a given sand-drift must
migrate more slowly than small bedforms - this is
why a draa can never overtake a ripple. The slower
the bedform, the thicker the preserved set, whether
it be a dune or a draa (see Chapter 2 for details of
the theory).

Hunter (1977) reports the initial porosities of
sandflows and wind-ripple laminae as averaging 45%
and 39% respectively. This knowledge can be used to
gauge the amount of compaction ancient aeolian
formations have undergone - simply by measuring minus-
cement porosities in thin sections and subtracting
this figure from the initial porosity of the particular
lamination type. Such exercises yielded compaction
figures of about 13% for the Yellow Sands, 17% for
the Bridgnorth Sandstone, 20% for the Penrith Sandstone
and 25% for the Locharbriggs Sandstone.

Compaction figures can also be derived from the
dip frequency distribution of foreset laminae.
This invariably peaks at a value some way below 34°,
the angle of repose of dry sand, or 32°, a typical
initial sandflow dip (Hunter, 1981); typically at a
value between 25° and 28°. This can be attributed
to a mixture of compaction and the selective
preservation of lower, more gently dipping slopes.
Comparison of compaction figures derived from dip
angles with those from porosity was made for the
Yellow Sands and Bridgnorth Sandstone. These
indicated that the selective preservation of more
gently dipping slopes does indeed occur.

Grain Size Characteristics

Each type of lamination results from a particular depositional process and each type therefore is the result of a particular size-sorting process. Grain size analysis of aeolian sands is considerably devalued if it is not related to the lamination.

The grain size of sandflows is found to be unimodal, very well to moderately well sorted, positively skewed, and meso- or leptokurtic. Wind-ripple samples are unimodal or bimodal, moderately or moderately well sorted, tend to be positively skewed, and are platykurtic. Sand-sheet samples are markedly bimodal (the modes being separated by 2-3 ϕ), moderately sorted, variably skewed and platykurtic.

The range of grain sizes present falls almost entirely into the sand category (-1 to 4ϕ) with only minimal overlap into very fine pebbles or coarse silt. Mean sizes typically range from 1.5 to 2.5 ϕ (350μ to 175μ); the vast majority of grains are coarse, medium or fine sand grade.

Sorting process particular to each lamination type may be deduced from these data. The action of deposition from climbing wind-ripples tends to develop a bimodal grain size distribution, though the exact reasons and means by which this influence acts are not understood. It occurs during the reworking of grainfall and sandflow deposits on the bedform leeside and during transport from the leeside of one bedform to the brink of the next. At this

point any further movement is by grainfall. This
is a potentially very efficient sorting process but
the thinness of the lamination it produces will
probably obscure this feature in analysis by sieving.
No grainfall deposits were analysed. Material
deposited by grainfall is likely to be reworked in
sandflows. These sort the sand further, coarse
grains tending to outpace and travel a greater
distance than fines in any flow. Thus preserved
lower leesides may selectively display only the
coarsest parts of individual sandflows. Some
confirmation of this influence was obtained; sand-
flows tending to be coarser than the average in the
two formations analysed (the Yellow Sands and the
Bridgnorth Sandstone).

Sand sheets may represent the ultimate result
of wind-ripple sorting, whereby most medium grains
have been removed, leaving a coarse plus fine
bimodality. The gap in the sand-sheet distribution
corresponds to the mode of the 'bedform distribution',
supporting the hypothesis of such a deflationary
origin. An alternative explanation is that sand
sheets are the result of interpenetration sorting,
that is the filtering of fine grains into the gaps
between the grains in a coarse surface carpet.

An overall contrast in grain size between the
Yellow Sands and Bridgnorth Sandstone was found, the
latter averaging 0.4 ϕ finer. This may be a feature
of provenance, but is also believed to be due to the
contrasting bedform shape and erg type. The Yellow
Sands are longitudinal draa on a bedrock surface; the

Bridgnorth Sandstone was an erg of transverse draa with complete and deep sand cover. The Yellow Sands situation expedited the penetration of coarse grains into the formation.

No systematic lateral and vertical variations in grain size were detected, either by eye or by sieving, in any of the formations studied, except that the lowest few metres of the Yellow Sands contained fine pebbles and more coarse sand than normal.

The Penrith Sandstone at some localities contains a considerable quantity of coarse or very coarse and very well rounded sand (e.g. at Cowraik Quarry, NY 541310; and Bowscar Quarry, NY 519343). This was initially regarded as an effect of the small size of the Penrith erg and the proximity of the area to potential sand sources. However, such prominent quantities of coarse sand were not seen in the Locharbriggs or Thornhill Sandstones, which were both once small ergs close to potential sources. Thus the coarse sand in the Penrith Sandstone may also reflect the availability of such material - it may be an illustration of the effect of provenance.

The grain size analyses carried out in the present research tend to confirm the hypothesis of Bagnold and Barndorff-Nielsen (1980) that natural grain size distributions conform to a log-hyperbolic, rather than a log-normal, or Gaussian form. This is manifest as a hyperbolic rather than parabolic shape to the log-frequency curve. Whatever the form of the distribution, it is felt that log-frequency curves

and frequency histograms are by far the best methods of portraying grain size data. Cumulative curves conceal significant trends and tendencies, and can be completely misleading.

Grain Shape and Surface Texture

In the past much has been made of the well rounded shape of aeolian sand grains. Thin sections of the formations examined in this thesis showed grain shape to vary from sub-angular or angular in very fine sand grains to well rounded in coarse and very coarse grains. The modal sizes, medium and fine grains, tend to be subrounded to rounded. These conclusions are based on visual estimates of shape; no measurements were made.

There seem to be a number of different causes of the famed frosting of aeolian sand grains. When an aeolian sandstone is examined under the microscrope in hand specimen the entire surface of every grain may be seen to be frosted. The frosting is visible even on authigenic and infiltrated clays coating the grains. Grains in water-laid sandstones associated with the aeolian formations are frosted to the same degree. This being the case, the frosting cannot be attributed to any specific aeolian grain surface texture. It seems to occur wherever a surface is minutely irregular, reflecting and refracting incident light in all directions. The frosting of modern aeolian sand grains cannot be denied (i.g. illustrations in Glennie, 1970, pp. 164-171) and it certainly seems to be a distinctive feature of such

grains. However, in the ancient so many textures and features appear to be able to produce frosting that its presence in untreated specimens should not be regarded as remarkable.

Under the S.E.M. upturned or cleavage plates have been identified as an aeolian abrasion texture and these features have been found in the formations studied. Their frequency decreases as the severity of diagenesis increases.

Diagenesis

In the U.K. the only non-red aeolian sandstone formations are the Yesnaby Sandstone, much of the Hopeman Sandstone, the Yellow Sands, and parts of the Frodsham Member.

The red colour of most of the formations is produced by coatings on most grains of red-stained clays and minor iron-oxides. X-ray diffraction analysis of this material shows it to consist of illite, mixed layer illite-montmorillonite, possibly kaolinite, and hematite. It is present in several distinct habits: as very thin layers or stains covering most of the surface of every grain; as concentrations in hollows on grain surfaces; as layers and partial coats on grain surfaces, often geopetally arranged or showing meniscus-bridge textures; in accumulations associated with, or obviously derived from, a degraded rock fragment; and as pore-lining authigenic clays, generally illite. This material was emplaced by the processes of clay infiltration by rain or floodwaters and intrastatal

alteration of labile rock fragments described by
Walker (1975, 1979). These are early diagenetic
process, occurring in the desert, and probably are
completed within one or a few million years after
deposition. Most of the sand had probably acquired
some pigment and was red before final deposition.

Intrastratal alteration and dissolution has
also affected the feldspars in most formations, and
a further common feature is the presence of authigenic
K-feldspar overgrowths. These also are believed to
have formed during early diagenesis.

The end result of the pigmentation is to produce
different formations of remarkably similar colour; red
or reddish brown in Munsell terms, with 2.5 YR 4/6
a typical figure.

Fluvial sandstones interbedded with the aeolian
formations studied commonly contain more pigment
and may be more strongly coloured (e.g. Piper, 1970).
This is believed to be because such sediments may
have originally contained more labile, iron-rich
mineral and rock fragments than their aeolian
neighbours. Such grains are often mechanically weak
as well as chemically unstable and are more likely
to be destroyed by the rigours of aeolian transport
than in subaqueous movement. The presence of flowing
water also signifies influent seepage and is thus
likely to enhance the deposition of infiltrated clays
as a further source and medium of pigmentation.

There is no evidence of any early diagenetic hard
cement in any onshore formation, though this is a

common feature of interbedded fluvial material. The cement in these is usually sparry or poikilotopic calcite (e.g. in the Brockrams (Chapter 9), and the Rotliegendes of the southern North Sea (Glennie et al., 1978)) and is common in modern deserts where limestone bedrock occurs (Glennie, 1970, pp. 33-36). Glennie et al. (1978) report gypsum (converting to anhydrite with burial) as an early cement in some aeolian sandstones under the southern North Sea.

This general lack of any early hard cement must come about partly because of the lack of any intrastratal source for such a cement in most horizons and partly because of a lack of groundwater to transport and precipitate the material. Only at the centres of the Irish Sea and North Sea Basins is there any evidence that the water table was at or near the surface during the Early Permian. There is little chance of achieving upward evaporative pumping of groundwater, with possible consequent mineral precipitation, through anything more than a few metres of blown sand. It can be no accident that the only recorded early hard cement in British aeolian sand-stones occurs in the southern North Sea, where inter-bedform sabkha deposits are common and the erg evidently lacked complete sand cover.

Unfortunately, little information is available on the diagenesis of the Colorado Plateau aeolian sandstones for an extension of any of these processes and tendencies to be illustrated. However, in those formations where early post-depositional deformation

by liquefaction has occurred the presence of an early cement would seem to be ruled out.

Most of the formations studied also show a lack of any cement due to later diagenesis. Most of the aeolian sandstone in the U.K. is held together only by the pigmenting clays. This must again be partly due to the lack of any intrastratal sources, other than pressure solution, for such a cement. The compositional maturity of aeolian sandstones induced by the mechanical destructiveness of transport by the wind is further enhanced by the early dissolution of labile grains inherent in desert diagenesis. Once this has occurred, there is little left to dissolve and re-precipitate as cement. The early grain-coating phases must inhibit cementation by quartz by partially or completely isolating grain surfaces from the pore water. With the pigment in place there are few sites, or seed crystals, on which silica in solution can nucleate.

Explanations can be found for the 4 exceptions to the rule of redness. The Yesnaby Sandstone and the Yellow Sands are both thin and trapped by one impermeable boundary against thick sequences rich in organic material and likely to bear reducing groundwaters. The Yesnaby Sandstone rests on crystalline basement and is overlain by several kilometres of lacustrine sediments. The Yellow Sands are overlain by tight carbonates and evaporites and rest on Coal Measures. The bleaching of parts of the Frodsham Member is believed to be due to mineralisation along

nearby faults (Appendix), and it is interesting

to note that the Hopeman Sandstone is also locally

mineralised (Peacock, et al. 1968). It must be stressed

that this last sentence is rampant speculation - no

systematic study of the diagenesis of either

formation has been made.

Facies Relations

The external facies associations of ergs were

discussed in Chapter 1 and there is little to add

here. However, a little can be said about facies

differentiations within ergs.

In the onshore U.K. the only conceivable facies

distinction that could be made in aeolian sandstones

is between sand-sheet deposits and large-scale

cross-bedded deposits. There may be some justification

in this, in that sand sheets do represent distinct

sub-environments within ergs. They are most likely

to be found on longitudinal draa (to judge by the

Yellow Sands), around the margins of ergs with

complete sand cover, or in interbedform areas. In

the latter case, sand sheet deposits would be a

sub-facies of an interbedform facies which could

include such things as freshwater to hypersaline

carbonates, adhesion-rippled sand, algal mats,

small deltas and channel sands, evaporites and sabkhas,

playa sediments, small dune sets, etc., etc. (e.g.

Kocurek, 1981b).

Kocurek notes that the interbedform sediments in

the Entrada Sandstone become 'wetter' (i.e. show more

influence of standing or flowing water) towards the
palaeo-shoreline. This could be a useful palaeo-
geographical indicator in ancient formations and
need not be restricted to coastal ergs. In the
Bridgnorth/Kinnerton/Collyhurst Sandstone such
sediments seem to develop and increase towards the
centre of the original drainage basin (Chapter 8),
likewise in the Rotliegendes under the southern North
Sea (Glennie, 1972).

Further illustration of facies differentiations
possible in suitably well exposed aeolian sandstones
is provided by Kocurek (1981a). He divides the
Entrada Sandstone into a central erg facies, a
coastal erg facies, and inland margin erg facies and
a local sheet-sand facies. The central erg facies
is characterised by slipfaceless draa sets with well
developed interbedform areas and deposits. In the
coastal erg facies,dunes seem to have been smaller
(it is not clear whether they were on slipfaceless
draa or not), the palaeocurrent pattern more variable
(attributed to sea breezes) and marine sediments are
interbedded with the aeolian. The inland margin erg
facies comprises small sets of aeolian cross-bedding
and beds of wind-ripple deposits interbedded with sandy
wadi deposits. Draa are believed to have been absent.
The sheet-sand facies consists of what are termed
sand-sheet deposits in this thesis and make up as much
as 40 to 50% of the sequence at 3 localities within
the central erg area.

Reservoir Properties

As shown by the formations described in this thesis, aeolian sandstones are typically highly porous and permeable, porosities commonly exceeding 20%. This is due to an inherent resistance to extensive diagenetic modification and cementation. Many aeolian formations are also homogeneous bodies of sand, uninterrupted by interbedding with the deposits of other sedimentary environments, and are thus free of large-scale permeability barriers. However, other inherent features of aeolian sandstones are a remoteness from~contemporaneous~potential hydrocarbon source rocks and a tendency to be overlain by porous arenaceous or rudaceous rocks.

Permeability in aeolian sandstones must be strongly anisotropic, favoured in the direction parallel to palaeowind. In individual cross-bedded sets poroperms should be higher in the sandflow laminated foresets than in the wind-ripple laminated toesets. Since sinuous bedforms tend to have more toesets, straighter-crested beforms should produce better reservoir properties. Similarly, the proportion of sandflow laminae is enhanced in the thick sets deposited by slipfaced draa.

If original topographic inclinations are preserved, subsurface aeolian sandstones may be found to finger out up dip into fringing alluvial fans. If these are suitably tight a potential stratigraphic trap may result. Such traps are found in the Rangely oilfield, Colorado, (Fryberger, 1979b). Basin-centre

interfingering with playa sediments offers similar opportunities, though the dip is less likely to be favourable and cementation by evaporite minerals probable.

The interbedform deposits of some aeolian formations constitute potential barriers to fluid migration. If such horizons are laterally extensive, and merge and separate across a formation, potential stratigraphic traps may be generated.

Aeolian sandstones of coastal ergs are probably more favoured for the entrapment of hydrocarbons than their inland counterparts. They are likely to interfinger basinwards with marine sediments and thus may be in communication with hydrocarbon source rocks. If overlain by transgressive marine shales or other tight formations, stratigraphic or structural traps with very great potential may be formed.

SECTION 10.3   THE IDENTIFICATION OF AEOLIAN SANDSTONES

General

Seen as one of a vast array of natural sedimentary environments, aeolian sandstones are very easy to distinguish. Seen as the results of one of 3 sediment depositing media, their identification is more difficult.

In the past aeolian sandstone formations have generally been identified as such by the large scale of the cross-bedding, the excellence of sorting and absence of pebbles, the well rounded and frosted nature of the sand grains, and the absence of any

fossils other than those of land plants and animals.

Aeolian v. Shallow Marine Sands

In the 1960s and 70s a number of formations, previously accepted as being aeolian, were re-interpreted, generally as *deposits of* some shallow marine environment. The formations receiving this treatment were the Cedar Mesa Sandstone (Permian, western U.S.A.), from Baars (1962), the White Rim Sandstone (Permian, western U.S.A.) from Baars and Seager (1970), the Yellow Sands and their correlates in Germany, the Weissliegendes (Pryor, 1971b, 1971a respectively), and the Navajo Sandstone (Jurassic, western U.S.A.) from Freeman and Visher (1975).

The root of these re-interpretations was a suspicion that the presence of large-scale cross-bedding was not unique to aeolian sands. Pryor (1971b) mentions 8 m thick planar cross-bedding in the Brazos River in Texas. The discovery of large, submarine bedforms known as sand waves or tidal sand ridges on the continental shelf of many oceans also seemed to offer an environment where large-scale cross-bedding might form. Tidal sand ridges, described for instance by Houboult (1968) from the southern North Sea, occur in water depths of 40-60 m and may be 30-40 m high, 1-3 km wide and 20-60 km long. Houboult's sparker profiles through two of these ridges appear to show high-angle cross-bedding in sets up to more than 40 m thick. This feature was

used by Pryor (1971 a, b) as justification for his
interpretation of the Yellow Sands and Weissliegendes
as such bodies. However, as Walker and Middleton
(1979) have pointed out, Houboult's sparker profiles
are considerably vertically exaggerated (about 13
times). The "steep" faces and internal cross-bedding
in fact dip at only 5°.

Sand waves are large, asymmetrical, transverse
bedforms, typically having wavelengths of a few hundred
metres and heights up to 10 m, exceptionally 20 m.
They have been described from the North Sea by
Terwindt (1971) and McCave (1971), from the
continental shelf of the north-eastern U.S.A. (Jordan,
1962), and from off the east coast of South Africa
by Flemming (1980). Allen (1980, p. 285) gives an
exhaustive list of references. Walker and Middleton
(1979) show that many of these features have "steep"
faces with maximum slopes ranging from $11^{O}-20^{O}$,
though some of those depicted by Flemming (1980) may
have lee slopes at the angle of repose.

It is also evident from published sparker profiles
that most, if not all, these bedforms bear super-
imposed megaripples on both stoss and lee slopes.
Quoting Walker and Middleton (1979), "their migration
would produce cross-bedding of a scale up to about a
metre and the probable internal structure of the sand
waves and tidal ridges would be complex, medium scale
cross-bedding (with sets less than 1 m thick)" This
situation is analogous to that of dunes on slipfaceless
draa. Similar styles of cross-bedding might be

produced, though with the scale larger by at least a factor of 5 in the aeolian case.

Allen (1980) gives an elegant and comprehensive account of the theory, origin, development and internal structures of sand waves. He restricted the use of the term sand waves to those bodies formed in reversing (i.e. tidal) flows. Bodies of similar size formed in one-way flows may be dunes. Allen drew up models of the postulated internal structure for a variety of flow velocity strengths and symmetries. Where the unsteady component of the velocity-time regime is such that periods of slack water do not occur, asymmetrical bedforms with leeside separation bubbles develop, giving slopes at the angle of repose and with deposition dominated by avalanching. High flow strengths lead to long bottomsets with backflow ripples, thus depositing small-scale cross-lamination directed up slope. Lower flow strengths result in the avalanche laminae meeting the basal bounding surface abruptly and acutely. As the influence of the unsteady component increases, periods of slack water develop. Mud drapes, mud-pellet breccias, and bioturbation of the foresets are generated. Super-imposed smaller dunes (megaripples) are common. As velocity symmetry increases, the sand wave becomes more symmetrical and large-scale flow separation is lost. Internal structures are then dominated by medium-scale cross-bedding with abundant mud drapes and mud-pellet breccias. If the velocity regime is still more symmetrical, herringbone cross-bedding is

formed. Allen (1980) shows that the validity of his model is confirmed in the stratigraphic record.

The abundant presence of ripple cross-lamination, bioturbation of the foresets, medium-scale cross-bedding, herringbone cross-bedding, mud drapes and mud-pellet breccias in many of these scenarios instantly precludes their confusion with aeolian sands.

Allen's (1980) classes IB and IIB show the most aeolian-like features. These are dunes rather than sand waves, being formed in unidirectional flows. Flow strength is not sufficient to develop toesets and ripple cross-lamination is absent. This immediately generates a distinguishing feature - only very rarely does aeolian cross-bedding lack toesets.

Numerous ancient shallow marine sand wave and dune bodies have been identified. R.G. Walker (1979) gives a list of those in the Mesozoic of the western U.S.A. The Eocene Roda Sandstone of the Pyrenees is also believed to be of this type (Nio, 1976). This formation is notable for containing a set of cross-bedding 20 m thick, "the largest demonstrably marine angle-of-repose set of cross-bedding yet described," (R.G. Walker, 1979, p. 81). In the U.K. Lower Greensand Group (Aptian-Albian), the Folkestone Beds in Surrey and West Sussex and the Woburn Sands in Bedfordshire are also of this type (Allen and Narayan, 1964; Nio, 1976; De Raaf and Beorsma, 1971).

These sediments are well exposed in numerous large sand pits around the town of Leighton Buzzard. The sands are completely unconsolidated and are dug for glass sand, aggregates and building sand, e.g. at

SP 940287, SP 929290, SP 928287, SP 929241 and

SP 920233. Quarries in the Surrey-Sussex outcrop

may be found for instance at SU 964188, TQ 125237,

SU 889202, SU 864473 and SU 785228.

A number of features are apparent which preclude
the confusion of sand wave deposits (even those of
Allen's (1980) Classes IB and IIB) with aeolian
sandstones.

Grain size, shape and frosting are somewhat
maligned, or at least unfashionable, clues to a
sedimentary environment. However, the Lower
Greensand grains are distinctive in lacking frosting
and are also often quite angular. Very fine to medium
pebbles are common, especially at the base of, but
also within sets. Phosphate, or phosphate-coated
grains are abundant for example in Arnold's Quarry
south of Leighton Buzzard.

Avalanche laminae in the Lower Greensand are
similar to aeolian sandflows, being laterally
extensive, tabular, homogeneous (lacking any
perceptible grading), typically a few centimetres
thick and bounded by thin finer-grained laminae (figs.
10.5-10.7). Purely in terms of the lamination the
only difference between these subaqueous avalanches
and subaerial sandflows seems to be that the upper
and lower surfaces of the former are slightly more
diffuse. However, subaqueous avalanches do not pass
into or are interbedded with wind-ripple laminae,
and the constituent grains of the two types of flow
are easily distinguishable. The subaqueous laminae

are also often bioturbated (e.g. fig. 10.4).

Further distinctions between shallow marine and aeolian sand bodies can be made. Any preserved fauna are likely to be instantly decisive. The stratigraphic context and overall sand body shape of the two environments is entirely different. None of the shallow marine formations listed by R.G. Walker (1979, p. 84) exceeds 40 m in thickness, and only one is an extensive sheet sand (more than 100 km by 100 km). Shallow marine sands tend to occur enclosed in marine shales, or at the base of a transgressive sequence. Shallow marine sands are not characterised by thick, unbroken sequences of large-scale cross-bedding consisting of nothing but sand. Nor, indeed, are the deposits of any other sedimentary environment.

Identifying Criteria and their Pitfalls

There should be no problem in identifying formations of aeolian sand. Difficulties should only arise on the scale of beds - for instance in picking out thin aeolian intervals in desert fluvial sequences, (including arctic deserts) or in identifying aeolian dune deposits in barrier island and other coastal sequences. No competent sedimentologist should be able to walk past or drill through a whole fossilized erg without knowing it.

When considering a whole formation, probably the

most obvious feature of aeolian sandstones is the ubiquitous large-scale cross-bedding, dune sets being 2-6 m thick, slipfaced draa sets 8-40 m. This must also be one of the safest distinguishing criteria.

The lamination of aeolian sandstones is very characteristic. Its recognition is of enormous significance where a single bed or outcrop, or a borehole core is under examination. Where well shown, the lamination can confirm or deny an aeolian origin on its own. The lamination should be dominated by wind-ripple (fig. 3.1) and sandflow (fig. 3.2) types, with grainfall and sand-sheet (fig. 3.3) rare. However, avalanches on the leesides of large subaqueous bedforms can make a very passable imitation of aeolian sandflows. Wind-ripple lamination is similar to beach lamination (e.g. Clifton, 1969) and to various types of subaqueous planebed laminae (Picard and High, 1973; N.D. Smith, 1971; McBride et al., 1975). However, beach lamination commonly contains heavy mineral stratification and low angle truncations of laminae, while the water-lain planebed laminae are almost always horizontal and individual units seldom exceed a few decimetres in thickness. None of these features are characteristic of wind-ripple laminae. Also, wind-ripple laminae show only parting lineation, whilst water-lain planebed laminae commonly show both parting and streaming lineation. One-sided, stairstep parting lineation may be exclusive to wind-ripple lamination.

Wind-ripple form-sets may be preserved beneath non-erosive sandflows or grainfall laminae. These - straight crested, very low amplitude (typically 1-3 mm) and with wavelengths of the order of 0.1 m - are highly distinctive.

The only divergences from these basic types of lamination in aeolian formations should be found in interbedform deposits. These should be relatively thin (a few metres), laterally extensive and contain a suite of distinctive features indicative of sedimentation in wet, damp and dry conditions. Their presence reinforces rather than weakens the probability that any surrounding sands are aeolian (c.f. Freeman and Visher, 1975).

Aeolian sandstones are free of any grains finer than coarse silt and coarser than very fine pebbles. Silts and clays may be found in interbedform areas, pebbles only near the margins or base of a formation. No other sedimentary environment is so exclusive in containing only sand.

Any grains coarser than fine sand in an aeolian formation tend to be subrounded, rounded or well rounded. Coarse and very coarse sand grains are almost exclusively well rounded, even near the margins of a formation. The grains are also always frosted, though this does not necessarily indicate an aeolian abrasion texture : the presence of this feature can only be ascertained by electron microscopy. The drawback of grain-related parameters is that grains can be reworked.

The early diagenesis of desert sediments is probably distinctive enough for it to be used as a criterion for the identification of such rocks. Its characteristics are red grain coats, concentrations of pigment in hollows on grains, the concentration of pigment around degrading rock fragments, geopetal arrangement of the pigment, meniscus bridging textures and infiltration textures.

Facies association is important - always remember Walther's Law. For instance 10 m of large-scale cross-bedded sand is more likely to be aeolian if included in a fluvial red-bed sequence, than if conformably under- and overlain by marine shales.

Other rare features can contribute important evidence. These include fulgurites, raindrop impressions, granule-lag monolayers of evenly spaced fine pebbles, dinosaur footprints and other terrestrial traces, dead dinosaurs and such, and small-scale surface deformation and fracturing of laminae.

Negative criteria must be listed:- the lack of mud (except in interbedform deposits), the near absence of pebbles, no marine or other aqueous fossils (though Goudie and Sperling (1977) report foraminifera being blown 800 km inland into the dunes of the Thar Desert in India), no extensive bioturbation except possibly in interbedform areas, no glauconite, no mica, etc., etc.

Summary

There appear to be two distinct periods during
the Permo-Triassic when conditions of climate and
geomorphology favoured the widespread deposition of
aeolian sand in the British area. These are in
the Early Permian and Early Triassic. The presence
of aeolian sandstone in any given area may be used
as an aid in correlating rocks in that area with
those elsewhere. Palaeowind resultants for British
Lower Permian aeolian sandstones show some divergence.
It is not altogether clear to exactly which direction
the true wind resultant lay. Much of the contra-
diction must be attributed to local orographic
effects.

The cross-bedding model presented in Chapter 2
is endorsed by onshore U.K. aeolian formations.
Evidence from the U.S.A. shows that ergs do not need
complete sand cover to aggrade: interbedform areas can
also migrate and climb. This can be accommodated by
the cross-bedding model and explained in terms of the
erg model presented in Chapter 1.

The identification of any sedimentary environment
must always rest on a balance of probabilities - there
is no black and white in geology, just lots of grey
areas. The philosophy must be to obtain all the
information possible and then go back for some more.

Aeolian sandstones can now be more confidently
identified than ever before, even to the extent of
distinguishing different facies within aeolian
formations. This progress is largely a result of

Hunter's (1977, 1981) elucidation of the lamination of aeolian sandstones. Ambiguity only arises where the exposure is poor, the rock badly weathered, or the lamination not well shown.

APPENDIX

OTHER AEOLIAN SANDSTONES IN THE U.K.

Introduction

This appendix comprises a list of the remaining aeolian sandstones in the U.K., with brief descriptive notes. These latter are compiled from varying amounts of field work and petrography undertaken during the present study, and also from the literature. Outcrop areas of these formations or members is shown on fig. A.1. For want of a better system, the formations are listed in progression from north to south.

1.   Yesnaby Sandstone Formation

Exposures: Spectacular sea cliffs on W. side of Orkney Mainland.

Stratigraphy: ? Lower Old Red Sandstone (House et al., 1977). Rests on basement; overlain by water-lain sandstones and thick, lacustrine, middle O.R.S. (Mykura, 1976). c.100 m thick around HY 219153 (figs. A.2, A.3).

Structures: Cross-bedding in sets up to 3 m thick (avge. 1.5 m) marked by major bedding planes dipping N.W. (structural) at 5 - 6 m intervals. 140 spread of dip directions (data from Fannin, 1970). Therefore very likely sinuous transverse dunes on slipfaceless draa.

Diagenesis: Formation is grey at outcrop.   No clay pellicles, cement by quartz overgrowths plus minor ferroan calcite and dolomite.   Carbonate replaces feldspars and rock fragments and is itself corroded. This secondary porosity filled or lined by tarry hydrocarbon.

2.   Hopeman Sandstone

Exposures: Excellent cliff sections for several km around Hopeman (NJ 1469) on S. shore of Moray Firth. Also small inland exposures.

Stratigraphy: Probably straddles Permian-Triassic boundary (Warrington et al., 1980).   Rests on O.R.S. and overlain by later Triassic strata. c.60 m thick.

Structures: Notable for enormous sets and spectacular contortions (figs. A.4 - A.7; Williams, 1973; Peacock, 1966).   Large sets must record slipfaced transverse draa.   Also has water-lain interdraa deposits (Williams, 1973).   Deformation due to liquefaction below water table, in manner described by Doe & Dott (1980).   Contains reptilian footprints and other possible trace fossils (figs. A.8, A.9).

3.   Lossiemouth Sandstone

Exposures: Very poor: rare old quarries and coastal exposures around Lossiemouth, Grampian (NJ 2370).

Stratigraphy: Probably late Triassic (reptilian fauna). Part of same sequence as Hopeman Sandstone. c.30 m thick.

Structures: "Barchan" - type cross-bedding, sets to 15 m thick. Palaeocurrent to N.E.

Data from Williams (1973).

4.   Old Red Sandstone, Scottish Midland Valley

Details: 2 possible aeolian sandstones in Upper O.R.S. mentioned by Bluck, (1980), one in lowest part of sequence, one in highest.   Both within fluvial sequences.

5.   New Red Sandstone, South-West Scotland

Exposures: 6 out of 7 areas of outcrop contain aeolian sandstones; Arran (Corrie Sst.), Mauchline (Mauchline Sst.), Thornhill (Thornhill Sst.), Moffat (Corehead Sst.), Dumfries (Locharbriggs Sst.), Lochmaben (Corncockle Sst.).   Exposures generally poor, though some good quarries (e.g. Corncockle, Lochmaben (NY 086870 ; fig. A.10); Locharbriggs area, Dumfries (NX 9980, 2 still working)), and good coastal exposures in Arran.

Stratigraphy: All aeolian sandstones believed Lower Permian (Smith et al., 1974).   Outcrop areas mark sites of former intermontane basins, probably linked originally. May have been contiguous with Irish Sea Basin and thus also the Penrith and Bridgnorth Sandstones. Some of the basins had penecontemporaneously faulting margins. Aeolian sands generally at base of, or low in sequence, interfingering with, and passing into marginal alluvial fan rudites.

Thickness: Locharbriggs Sandstone, 1000 m;
Corncockle Sandstone, 900 m; Mauchline Sandstone,
450 m; Corrie Sandstone, 300 m; Thornhill Sandstone, c.200m;
Corehead Sandstone, at least 30 m. Most of above data
from excellent documentation by Brookfield (1977, 1978,
1979, 1980) and Piper (1970).

Structures: Mostly trough cross-bedding in sets
a few metres thick deposited by sinuous transverse
dunes on slipfaceless draa. Slipfaced draa at type
locality of Corncockle Sandstone (NY 086870).
Hierarchy of aeolian bounding surfaces was first
described from the Locharbriggs Sandstone (Brookfield,
1977). Lamination dominated in most cases by
wind-ripple and sandflow types with grainfall and sand-
sheet subordinate. Sandflow thicknesses observed did
not exceed 60 mm except in rocks thought to be from the
slipfaced draa sets at Corncockle (these up to
100 mm).

Diagenesis: Colour and pigmenting phases as for
Bridgnorth and Penrith Sandstones; colour of all
formations remarkably similar. Principal lithifying
cement is quartz. 25% compaction deduced for
Locharbriggs Sandstone (only that, Corncockle and
Thornhill Sands examined petrographically).

6.   Kirklinton Sandstone

Exposures: Very few. Best at NY 492603,
NY 413662.

Stratigraphy: Believed Lower Triassic, part of same sequence as Penrith Sandstone. Passes by interbedding to fluvial St. Bees Sandstone below and Stanwix Shales above. Mentioned in Warrington et al., (1980), Trotter and Hollingworth (1932), Dixon et al., (1926). Maximum thickness, c.100 m around Carlisle. Colour and cementation: Red (very similar to Bridgnorth Sst). Mostly uncemented at outcrop except for pigmenting clays.

7. Frodsham Member

Exposures: Good in railway cuttings at Frodsham, Cheshire (SJ 5177). Many old quarries, one working quarry, road and natural sections near villages of Nesscliffe, Ruyton-XI - Towns, Myddle, Grinshill and Weston, northern Shropshire. The two outcrops are probably correlatable, but are separated by an area of thick drift with no exposure: continuity cannot be confirmed.

Stratigraphy: Lower Triassic; forms part of the Helsby Sandstone Formation in the Frodsham area. Overlies the Wilmslow Sandstone (ex-Lower Mottled Sandstone); overlain by the Tarporley Siltstone (ex-Keuper Waterstones; Warrington et al., 1980). Maximum thickness c.45 m.

Structures: Mostly structureless in the southern area with only sporadic evidence of large-scale cross-bedding (fig. A.11) and aeolian lamination. Crossbedding well preserved in northern area;

interpreted as dome-shaped dunes by Thompson (1969) but probably records both slipfaced and slipfaceless draa.

Colour and cement: Red (very similar to Bridgnorth Sst) in northern and part of southern outcrop. Upper parts yellow to buff in southern outcrop. Cement generally only pigmenting clays, but also barytes in south; this and the yellow colour linked to Tertiary mineralization (Thompson, 1970; Poole and Whiteman, 1966).

8. **South-West England**

(a) 6 m thick unit within Tor Bay Breccias (Early Permian), visible in coastal section at Paignton. Tentative palaeocurrent result to S.S.W. (Laming, 1966).

(b) Dawlish Sands (350 m thick; lowest 230 m aeolian). Probably Early Permian. Cross-bedding in sets up to 9 m thick and 60 m wide, deposited by "barchans". Palaeocurrents to N.N.W. (Laming, 1966; Smith et al., 1974). Well exposed in coastal section at Dawlish.

(c) Otter Sandstone (c.120 m thick), Early Triassic (Warrington et al., 1980). Exposed in coastal section at Budleigh Salterton overlying the Budleigh Salterton Pebble Beds.

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