DESERT SEDIMENTS - PRESENT AND PAST

by

Kenneth W. Glennie

Published and unpublished works presented for the degree of Doctor of Science at the University of Edinburgh, 1984
work published by Kenneth W. Glennie in order of publication date.

Abstract

1) 1968 Dikaka: plants and plant-root structures associated with aeolian sand.
Palaeogeography, Palaeoclimatol., Palaeoecol., 4: 77-87.
Co-author B.D. Evamy (Shell).

This paper has been reprinted in:
Sarjeant, W.A.S. (Editor). Terrestrial Trace-Fossils
Benchmark Papers in Geology, volume 76,

2) 1970 Desert Sedimentary Environments.
Developments in Sedimentology 14.

3) 1972 Permian Rotliegendes of Northwest Europe interpreted in light of modern desert sedimentation studies.

This paper has been reprinted in:
Weimer, R.J. (Editor)
Sandstone Reservoirs and Stratigraphic Concepts II.

4) 1976 A reconnaissance of the Recent sediments of the Ranns of Kutch, India.
Sedimentology 23: 625-647.
Co-author: G. Evans (Imperial College).

5) 1978 Depositional environment and diagenesis of Permian Rotliegendes sandstones in Leman Bank and Sole Pit areas of the UK southern North Sea.
JL. geol. Soc. Lond. 135: 25-34.
Co-authors: G.C. Mudd (Esso) and P.J.C. Nagtegaal (KSEPL, Rijswijk).

6) 1978 (i) Desert Sedimentary Environments
(ii) Eolian Sands
Contributions to: Fairbridge, R.W. and Bourgeois, J. (Editors).
The Encyclopedia of Sedimentology. Dowden, Hutchinson & Ross,
Eolian Sands co-authored by David Krinsley.
7) 1981 Sole Pit Inversion Tectonics. 
   In: Illing, L.V. and Hobson, G.D. (Editors). 
   Petroleum Geology of the Continental Shelf of North-West Europe. 
   Co-author: P.L.E. Boegner (Shell).

8) 1983 Early Permian (Rotliegendes) palaeowinds of the North Sea. 
   Sedimentary Geology 34: 245-265.

9) 1983 The Permian Weissliegend of N.W. Europe: the partial deformation of 
   dune sands caused by the Zechstein transgression. 
   Sedimentary Geology 35: 41-81. 
   Co-author: A.T. Buller (formerly Shell Expro, now University of 
   Trondheim).

10) 1983 Lower Permian Rotliegend desert sedimentation in the North Sea area 
    In: Brookfield, M.E. and Ahlbrandt, T.S. (Editors). 
    Eolian Sediments and Processes. 
    Developments in Sedimentology 38. 

11) 1984 Early Permian - Rotliegend. 
    In: Glennie, K.W. (Editor). 
    Introduction to the Petroleum Geology of the North Sea . 

12) in press Experimental soft-sediment deformation in artificial dune sands 
    Submitted to Sedimentology for possible publication.
The collection of papers presented in this thesis comprises the results of studies undertaken both in Company and in 'free' time by an exploration geologist of a major oil company. Their origins had a strong element of chance - being in the right place at the right time. Their progress depended on a happy marriage between Company directive and personal choice, and followed a natural development in which studies of the present became the natural key to the past; and in a few cases it was the past that stimulated interest in the Present.

In the Autumn of 1963, the writer was told that he was to be transferred to Shell's Exploration and Production Research Laboratory (KSEPL) in The Netherlands to take charge of the existing research programme on turbidites (previous experience of the subject in New Zealand and the Canadian Arctic). It was just at this time, however, that Shell realised that there was sufficient gas in the giant Groningen field in northern Netherlands to alter the whole fuel economy of N.W. Europe, and that there was the possibility of finding additional gas in the southern North Sea. The Early Permian Rotliegend reservoir sandstone in the Groningen field was thought to be of desert origin, but as no one in Shell had a personal knowledge of desert sediments and there was an almost complete lack of modern literature on the subject, it was decided that a programme of research into desert sediments should be undertaken; the 1964 budget for turbidite research was cancelled and forthwith transferred to desert research - which Glennie was appointed to carry out.

There was now a strong need to visit a desert, if only to be in a position to assess the value of the limited literature on the subject. Glennie already had a little experience of the desert in northern Libya, acquired during army service in 1947-1948, prior to his university studies. For logistical reasons, the choice of deserts was limited initially to areas where Shell had an operating company.
Thus in February 1964, the writer, accompanied by his colleague E. Oomkens (who made the initial interpretation that the Groningen Reservoir was of desert origin) began a three-month reconnaissance of the deserts of Libya, both inland and coastal. It was here that Oomkens recognised the modern equivalents of both sand dykes and adhesion ripples, which he had first seen in Rotliegend cores.

During the Summer of 1964, the writer visited some areas of Rotliegend in Germany, and made an extensive reconnaissance survey of Permo-Triassic outcrop areas in Britain from South Devon to the Midland Valley of Scotland.

In the Autumn of 1964, he was able to combine attendance at the International Geological Congress in New Delhi with a reconnaissance of the Rajasthan Desert and that almost unknown supra-tidal salt-covered coastal embayment, the Ranns of Kutch. Because of the existing co-operation between KSEPL and Imperial College, London, on carbonate research in the Persian Gulf, Dr. Graham Evans, of Imperial College, was invited to join the writer on this trip, to their mutual benefit. Publication No. 4 was the outcome of this co-operation, in which geochemical analyses were undertaken by colleagues at both KSEPL and Imperial College.

Field work recommenced in January 1965 with a 3-month survey of desert sediments in eastern Arabia (Qatar, Trucial States, Oman) on which the writer was accompanied by B.D. Evamy, who had recently joined KSEPL's carbonate research team. Publication No. 1 was the direct outcome of this co-operation, in which the petrological work was undertaken by Evamy.

Further limited field work was carried out during 1965 on Permo-Triassic outcrop areas in Britain. Much of the year, however, and extending into 1966, was taken up with report writing, which eventually led to completion of a major internal company report on Desert Sedimentary Environments.

Permission to publish the bulk of this report as a book was granted late in 1967, at which time the writer was leading a team mapping the
geology of the Oman Mountains. Preparation of a version for publication was made in the Summer of 1968 prior to field work in Iran. After an interval of 6 months or more, the publisher's reviewers suggested that certain sections should be expanded - this was mostly a spare-time activity during the Spring and Summer of 1969. The book (Publication No. 2) finally appeared late in 1970.

In the Summer of 1970, Shell asked the writer to prepare a paper on the Rotliegend of N.W. Europe suitable for presentation on an AAPG 'Distinguished Lecturer' tour of North America. The lecture tour was undertaken early in 1971 and the ensuing publication (No. 3) appeared in June 1972. Preparation of this paper was helped considerably by reference to unpublished internal reports by Shell colleagues, in which it was gratifying to learn that many of their environmental interpretations were based on Glennie's original KSEPL report on Desert Sedimentary Environments. Despite this help, compilation of the paper, selection of the illustrations and drawing the figures was entirely the responsibility of the writer.

Glennie was transferred to Shell UK's North Sea exploration team in the Summer of 1972, to work first on the NW Shelf and then, from 1975 to 1981, on the Southern Permian Basin. During an orientation field trip in 1972, the writer was shown the deformation structures in the Hopeman Sandstones north of Elgin by Dr. D. Peacock of the UK Geological Survey. This experience helped in an understanding of the origin of the deformation structures seen in Weissliegend dune sands, and led to Publications 10 and 12.

A thematic meeting of the Geological Society of London in March 1977 on the topic of Sandstone Diagenesis led directly to Publication No. 5. Shell asked Glennie to prepare a paper on Rotliegend diagenesis; of his co-authors, G.C. Mudd (Esso) had earlier experience only of Middle East carbonates, but P.J.C. Nagtegaal (KSEPL) had already published on desert sedimentology and is an excellent petrologist. As with so many oil-company publications of this type, much of the factual data was based on the work of colleagues that was spread over the previous 12 or 13 years of North Sea exploration; the current understanding of Rotliegend diagenesis, and thus the way in which the
paper was written was entirely the senior author's responsibility; Nagtegaal's main contribution was in ensuring that the correct emphasis was given to the different aspects of burial diagenesis, and in selecting the electron photomicrographs to illustrate the habit of authigenic illite.

The title of the 1981 publication (No. 7) was imposed by Shell Expro's Exploration Manager when deciding what contribution should be made by the Company to the 1980 Lancaster Gate Conference on the N.W. European Continental Shelf. Glennie was invited to prepare the article together with P.L.E. Boegner, whose overall contribution was largely confined to discussion of some of the incorporated ideas. Our main interest was the history of Rotliegend burial during the past 250 Ma. The text and all the figures were prepared by Glennie.

Publication No. 8 resulted from an invitation to prepare a special volume of Sedimentary Geology in honour of Prof. J.F.M. de Raaf of Utrecht University, from whom Glennie was to have taken over responsibility for turbidite research at KSEPL in 1964. Much new palaeowind data could be deduced from the dip-meter logs of recently released North Sea wells; it just waited analysis.

A request from Polish geologists to help interpret the origin of deformation structures and structureless sands seen in photos of Weissliegend sequences, led to dissatisfaction with the 1972 interpretation of similar structures seen in North Sea wells (Publication No. 3). Glennie's developing ideas on the subject needed a 'sounding board'; this was found in A.T. Buller, initially Shell Expro's senior sedimentologist, and latterly Head of the Petroleum Geology Dept. at Trondheim University. Apart from being a sound sedimentologist, Tony Buller would not readily accept an interpretation that he did not fully understand. The bulk of the text of Publication No. 9 was written by Glennie but with many minor modifications by Buller. The inspiration that led to the understanding of the pressures associated with air pockets in a submerged dune came from a chance remark during discussion with a young Swiss colleague, Gerard Bloch. Publication 9 led directly to the need to test its plausibility with a small-scale tank experiment - with the very satisfying results recorded in the unpublished paper No. 12.
An invitation to speak at the meeting of International Sedimentologists held in Hamilton, Canada, in 1982, gave rise to Publication No. 10.

Publication 11 forms chapter 3 of the Course Notes given to delegates to a 2-day presentation 'Introduction to the Petroleum Geology of the North Sea' held in Burlington House, London. In addition to the environmental and diagenetic aspects of the Rotliegend, the chapter also summarises the economic importance of the formation as a gas reservoir. The course was organised and the Notes were edited by Glennie. The Course Notes were formally published in May 1984 in time for the fourth verbal presentation of the course.

In addition to the above, acknowledgements have been made in the different publications to those others who have assisted in the development of ideas or in the presentation of data on Desert Sediments - Past and Present, during the past 20 years.

Apart from the cases listed above, all the publications presented in this thesis are entirely the work of the undersigned.
NOTE: Because this submission comprises 9 separate publications, including a book, and an article 'in press' on the broad subject of deserts, past and present, the writer has taken the unusual course of presenting a fairly long abstract with sub-titles. The abstract is an update of a summary that has been handed out on various occasions to students and others to whom the writer has lectured.

ABSTRACT

Introduction

Deserts cover about one fifth of the world's land surface, but only one fifth of the desert surface is covered by dune sand. The remainder comprises areas of outcrop, fluvial (wadi) sediments, sediments of temporary desert lakes and inland sabkhas and coastal sabkhas.

Definition (Publications 2, 6)

A desert is an almost barren tract of land where rainfall is so limited and spasmodic that it will not adequately support vegetation, and where the potential rate of evaporation far exceeds precipitation (e.g. 2500-6000 mm/annum in Sahara).

Erosion and Deflation (Publications 2, 6)

Chemical corrosion - breaks down carbonates - moisture from dew.

Exfoliation - argillaceous - rounded boulders.
- siliceous - (insolation) splits boulders - ventifacts?

Deflation - process of sand removal by wind action.
Deflation surfaces - fine products of erosion or sedimentation removed by wind to leave a lag of coarse grains.

**Wadis (Publications 2, 6)**

Desert watercourses, dry except after rain. Rain may vary from seasonal to sporadic thunderstorms 1-30 years or so apart.

Products of erosion carried in wadis by water following rain.

Terminal point of water in a wadi depends on volume of water (rainfall), and may be the sea, temporary or semi-permanent desert lakes (e.g. Lake Chad), inland or coastal sabkhas or where it soaks into ground somewhere along its length.

Sediment carried by stream flow (saltation and traction) or mud flow (suspension) depending on sediment/water ratio; forms major source for aeolian sands.

Braiding - when the fluvial transporting medium becomes overloaded with sediment - change in slope or loss of water - alluvial fans.

Cementation - wadi water commonly rich in \( \text{CaCO}_3 \) - precipitated as cement following near-surface evaporation - finer surface sediments commonly not cemented - deflated - source of dune sand or loess. Ancient better-cemented wadi channels exhumed by differential deflation and now stand out in relief.

Aeolian sands preferentially deposited against wadi banks - preserves curled clay flakes - diverts water to increase braiding action.

Aeolian sands remarkably resistant to action of flowing water.

Areal balance between sediments of wind and water at any one time.
Desert Lakes (Publications 2, 6)

Groundwater supply - semi-permanent - evaporation - salt concentration.

Wadi supply - temporary - clay lags plus gypsum crystals preserved by wind-blown sands. Older dune relief remarkably well preserved.

Inland Sabkha (Publications 2, 6)

Flat area of clay, silt or sand, often with salt crust.

Sand dykes - fracturing of clay layer over hydrostatically pressured sandy slurry - often preserved by overlying aeolian sand.

Adhesion ripples - adhesion of wind-blown sand and silt to a surface kept moist by capillary supply of water from water table. Source of aeolian sediment when water table is low during dry season.

Near-coast inland sabkhas locally supplied by groundwater of marine origin but not flooded by storm tides, therefore no living marine fauna.

Aeolian Sands (Publications 2, 6)

Origin of sands - outcrops, but mainly via older dune sands, non-cemented wadi and sabkha sediments, coastal beaches and sabkhas.

Movement of sand: Proportional to: (Excess over 'Threshold Velocity')³

Movement by suspension up to 100 μ
- saltation 100 to 2000 μ
- surface creep 6 x diameter of saltating grains.

Structures: Sand ridges - part surface creep - part deflation - transverse to wind.
Sand ripples: 'Wind forms whose wavelength depends on wind strength and is constant with time'. Coarsest grains on top. Height depends on range of grain sizes and wind strengths. Transverse to wind.

Dunes: Sand accumulations formed by the wind in which finest grains are found at the top. Unlimited extension down wind. Size dependent on supply of sand and ability of wind to carry grains to top without blowing them off again, which may be controlled by the depth of the unstable atmospheric layer above the ground.

Barchans: Crescentic, formed by moderate winds with a limited supply of sand; migratory 34° slip-face. Maximum wind velocity along flanks - sand transport - horns. With an unlimited sand supply - transverse dunes are the normal form provided that wind velocities are not too high for them to be stable.

Sifs: Longitudinal, with long axis parallel to dominant wind. Origin possibly dependent on following three statements:

a) "Strong sand-laden wind causes transverse instability so that sand tends to be deposited in longitudinal strips parallel to wind" (Bagnold, 1941, p. 178).

b) "Strong wind causes accretion of sand on an existing sand patch and an extension up wind - lasts only as long as supply of sand is plentiful" (Bagnold, 1941, p. 171).

c) "A given wind can drive sand over a hard immobile surface at a considerably greater rate than is allowed by the loose sand-covered surface" (Bagnold, 1941, p. 72).

Results in air-pressure gradients between interdune and dune areas. With strong winds, this results in development of spiral (helical) wind vortices that transport sand obliquely from the middle of the interdune area towards the dune's crest.

34° slip-faces form perpendicular to axis of dune, which is parallel to the axes of the helical wind vortices.
Dikaka: accumulations of dune sand covered by scrub or grass vegetation; extended to include plant-root burrows in ancient dune sands. (Publications 1, 2).

**Effect of Global Wind Patterns on Desert Conditions**

/Publications 2, 10, 11/

(Southern Hemisphere in brackets)

<table>
<thead>
<tr>
<th>Wind System</th>
<th>Direction</th>
</tr>
</thead>
<tbody>
<tr>
<td>Polar</td>
<td>from NE (SE)</td>
</tr>
<tr>
<td>Westerlies</td>
<td>from SW (NW)</td>
</tr>
<tr>
<td>High pressure horse latitudes</td>
<td>diverging</td>
</tr>
<tr>
<td>Trade winds</td>
<td>from NE (SE)</td>
</tr>
<tr>
<td>Equatorial doldrums</td>
<td>convection</td>
</tr>
</tbody>
</table>

Increase Laurentian, Scandinavian and Antarctic ice caps - all climatic zones concentrated towards Equator. Increased wind velocities - higher rates of evaporation but colder climate - seif dunes predominate - desert dune formation extends to below present sea level. Extensive deflation of unconsolidated shallow-marine faunas - Miliolite. Creation of major dune systems of Afro-Arabia and Australia.


Variations in aridity of tropical deserts seem to follow changes in polar glaciations, with the most arid conditions coinciding roughly with glaciations and desert 'pluvials' with interglacial periods.

**Desert Coasts** (Publications 2, 4, 6)

No continuous flow of water from wadis - no fluvial deltas. Warm seas, clear water - manufacture of carbonate sediment.
Longshore transport of carbonate sediment - spits, bars.
Formation of lagoons and coastal sabkhas.
Tidal-flow transport of sediment in lagoon - tidal delta.
Inward and outward-facing oolite tidal deltas.
Daily onshore winds - beach sands transported inland.
Forams and shell fragments carried into inland dunes.
Nightly offshore winds much milder than daytime winds.
Storm tides over low flat coastal areas - coastal sabkhas.
Salt encrusted, algal mats, diagenetic alteration of sediment - dolomite, gypsum.

The interstitial waters of coastal sabkha sediments become concentrated as result of evaporation of flooded sea water. Precipitation of gypsum increases Mg$^{2+}$/Ca$^{2+}$ ratios, thereby creating conditions favourable for dolomitisation of carbonate sediments. In the Ranns of Kutch, however, there is virtually no carbonate to dolomitise and Mg$^{2+}$/Ca$^{2+}$ ratios increase from 3 on the coast to 240 on the shoreline of Pachham Island. (Publication 4).

Ancient Desert Sediments

North America - Late Carboniferous to Jurassic
South America - Jurassic-Cretaceous
N.W. Europe - Late Carboniferous to end Triassic (Rotliegend, etc.)
Devonian dune sands (S.W. Ireland, Orkneys, Caithness)
Early Permian Rotliegend of N.W. Europe (Publications 3, 5-11).

Southern and Northern Permian Basins, E. Devon, Worcester Graben, Cheshire, Irish Sea, Vale of Eden - Solway Firth area, Ayrshire, Arran, Moray Firth Basin, ? West of Shetlands. Southern Permian Basin stretches from Durham to Russo-Polish border. Four major desert facies recognised: Fluvial (wadi) along southern margin; grades north into aeolian dune facies. This facies grades into that of a sabkha marginal to the site of a more or less permanent desert lake, which extended over the sabkha area at times of reduced evaporation. During windier periods of higher rates of evaporation, bedded halite was deposited on the lake bottom. This Rotliegendes desert lake measured.
some 1200 km x 200 km, and its surface may have been up to 300 m or so below the level of the Permian Oceans at the time of the Zechstein transgression.

Early Permian (Saxonian) winds blew over Southern Permian Basin from about NE to SW, indicating deposition in a Northern Hemisphere 'Trade-Wind' desert. Over the Northern Permian Basin winds mostly from W or NW. Consistency of palaeowind directions with time indicates that a semi-permanent barometric high must have existed over the Mid North Sea High throughout the Saxonian.

Desert conditions were brought to a close by the very rapid flooding of the Rotliegend desert basins by the waters of the Zechstein Sea. Although very fast, the flooding was not destructive and dune relief was largely preserved beneath the Kupferschiefer. In areas of transverse dunes, however, part of the dune-sand sequence above the level of the former water table was partly deformed and homogenised by the escape of air that had been trapped beneath sands wetted by the rapid capillary rise of water along fine-grained laminae. (Publication 10, submission 12).

Rotliegend dune sandstones and, to a limited extent wadi sandstones, form major reservoirs for gas in the Southern Permian Basin with porosities averaging 15-25% and permeabilities in the order of several to hundreds of mD. In areas where subsiding basins brought the Rotliegend to depths of 4-5 km or more, burial diagenesis brought about loss of original 45% or so primary porosity by compaction, pressure solution and growth of authigenic minerals to around 10%, and primary permeabilities of several Darcies to the order of 0.3 mD or less.

The source of this gas is coal and carbonaceous shales of the underlying Carboniferous. The top seal is mainly the Zechstein Cycle II Stassfurt Halite.

In the Northern Permian Basin, oil occurs in Rotliegend reservoirs in the Auk and Argyll fields, the source rock being the Kimmeridge Clay, which matured deep in the adjacent Central Graben.
DIKAKA: PLANTS AND PLANT-ROOT STRUCTURES ASSOCIATED WITH AEOLIAN SAND

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(Received April 25, 1967)

SUMMARY

Plant-root structures, which are normally associated with low-lying swamp-like environments, have locally been found in the dune sands of arid deserts. The Arabic word "dikaka", meaning scrub-covered dune sand, has been chosen to designate the latter type of plant structure. Such structures are common only in Tertiary and younger dune sands and represent desert environments having at least a sporadic supply of water, for example from the sea or an occasional rain storm. Dikaka, preferentially cemented by gypsum, is found where this water has been particularly saline. An abundance of Recent dikaka can stabilise formerly mobile dune sand but, in so doing, can destroy the dune bedding so that the aeolian origin of the sand becomes difficult to recognise.

INTRODUCTION

Most fossil root structures so far described in the literature are associated with sediments deposited in low-lying swamp-like environments where the necessary water for plant growth was plentiful. An adequate local supply of water is also indicated for certain arid desert environments, however, since it has been found that plant-root moulds are quite common sedimentary structures in some Tertiary and younger desert dune sands. Root structures in desert sediments are therefore considered to be important environmental indicators of nearby water—present either more or less permanently in wadi gravels at shallow depth or temporarily in dune sand following rain. The Arabic word "dikaka", meaning scrub-covered dune sand, has been chosen to designate these structures and their environment, which form the subject of this paper.
AGE RANGE AND DESCRIPTION OF DIKAKA

Although some plants had already adapted themselves to desert conditions by the Early Permian (e.g., *Walchia piniformis* and *Walchia filiciformis*, Mägdefrau, 1956, pp.177–179), their root systems, if any, seem to have been only lateral and not to have penetrated deeply into the underlying sediment. Such plants apparently grew in water-transported sediment. The earliest dune (but not necessarily desert) plants with sediment-penetrating roots (*Nathorstiana, Weichselia, Hausmannia*) are known from Lower Cretaceous dune sands of possible coastal origin from Germany (Mägdefrau, 1956, pp.279–286). By Tertiary time, however, several root-forming plants had adapted themselves to life in dune sands of arid deserts. From our own observations of the Late Tertiary, their root moulds (dikaka) are preferentially preserved in dune sand adjacent to fossil wadi channels where formerly there was a regular supply of water. A plentiful supply of water, in this case of marine origin, would also have been available for the abundant dikaka seen in the Quaternary coastal dunes of the Trucial Coast of the Persian Gulf and Qatar (Plate IIA). In both Tertiary and Quaternary examples root penetration can be measured in lengths of at least tens of centimetres.

The Late Tertiary aeolian sand of Jebel Baraka on the Trucial Coast (see Fig.1 and Plate IA) shows a considerable development of dikaka, but the original dune bedding of the sediment is still clearly visible. In other dune sands from the same locality (Fig.1 and Plate IB and C) the dikaka root moulds are so numerous that the original dune bedding is only discerned with difficulty. Many of the moulds are oriented parallel to the sand laminations (almost horizontal in Plate IC), since this presumably was the direction of greatest permeability, and therefore the direction of flow of the intergranular water, so necessary for root-forming plants.

PREFERENTIAL CEMENTATION OF DIKAKA

The roots themselves, apart from living and not-long-dead examples, are not preserved since, in arid environments, oxidation soon brings about their almost complete conversion to carbon dioxide. The former distribution of roots in sediment deposited in an arid climate is, however, commonly exhibited by preferential cementation of the sand grains which had encased the roots (Plate I). The typical cemented envelopes which result are particularly apparent in Plate IIA. If, after such preferential cementation, wind deflation removes the loose or less well-cemented sand between the former roots, the root pattern becomes exhumed (Plate IIB) and stands out in relief above the surface of deflation. Hoffmeister and Multer (1965) have described, from Biscayne Bay (Florida), a similar phenomenon involving firstly preferential cementation of mangrove roots and sub-
Modern plants living along the Trucial Coast are generally halophytes, i.e., xerophytes \(^1\) which can live in soils having ground waters of high salinity. The leaves of these cactus-like plants contain an abundance of liquid so that it is relatively easy to obtain sufficient material to analyse their water content for inorganic ions. Such an analysis shows the leaves to contain water with 13.95 g/l \(\text{SO}_4^{2-}\). This is five times the concentration of \(\text{SO}_4^{2-}\) in normal sea water. Even if the liquid in the leaves only approximately represents that in the rest of the plant, it would seem that, on decay, the plant would be a likely site of calcium sulphate precipitation. It is significant, therefore, that the dikaka of the Trucial Coast is typically cemented by gypsum euhedra (Plate IIIC), arranged in vaguely concentric patterns around former roots. The cement of the host rock, however, need not be gypsum. Along the Trucial Coast, for example, the Quaternary coastal dunes, away from localised gypsum-cemented dikaka, are cemented by calcite.

\(^1\) Plants adapted for growth with a limited water supply.
Marine limestone.
Not exposed.

Aeolian sandstone riddled with cemented plant-root moulds (DIKAKA), Plate IB, C.

Alternating wind-blown sand and water-transported conglomerate with pebbles up to 2-3 cm diameter.

Pebble conglomerate.
Aeolian sand.

Pebble conglomerate with MASTODON bone fragments and TOOTH partly fixed to jaw bone.

Aeolian sand.

Pebble conglomerate.

Foresetted dune sand with cemented plant-root structures occurring throughout but most commonly in upper half, Plate I A.

Sandstone changing upwards from having horizontal to ripple lamination, the ripples being filled with clay lenses.

Red argillaceous sandstone having horizontal to ripple lamination and rootlet moulds.

Grey sandy and silty clay with iron-stained rootlet moulds.

Sandstone with clay-pebble conglomerate, clay lenses in ripple hollows and SEA LEVEL.

Grey argillaceous sandstone with rootlet moulds.

Not exposed.

Fig. 1. Part of the Pliocene continental sequence, Jebel Baraka, coast of the Trucial State of Abu Dhabi.

sequent marine erosion to expose “a lattice of roots in their original position”.

Modern plants living along the Trucial Coast are generally halophytes, i.e., xerophytes1 which can live in soils having ground waters of high salinity. The leaves of these cactus-like plants contain an abundance of liquid so that it is relatively easy to obtain sufficient material to analyse their water content for inorganic ions. Such an analysis shows the leaves to contain water with 13.95 g/l SO₄²⁻. This is five times the concentration of SO₄²⁻ in normal sea water. Even if the liquid in the leaves only approximately represents that in the rest of the plant, it would seem that, on decay, the plant would be a likely site of calcium sulphate precipitation. It is significant, therefore, that the dikaka of the Trucial Coast is typically cemented by gypsum euhedra (Plate IIC), arranged in vaguely concentric patterns around former roots. The cement of the host rock, however, need not be gypsum. Along the Trucial Coast, for example, the Quaternary coastal dunes, away from localised gypsum-cemented dikaka, are cemented by calcite.

1 Plants adapted for growth with a limited water supply.

*Palaeogeography, Palaeoclimatol., Palaeoecol.*, 4 (1968) 77-87
The fossilised mangrove roots from Florida are reported by Hoffmeister and Multer (1965) to be preferentially cemented by cryptocrystalline calcite. These authors also relate the selective cementation to the decay of the plant matter. They suggest that the decay provided CO₂ which formed carbonic acid in water. This, in turn, dissolved calcium carbonate from the surrounding sediment. The highly calcareous solutions so produced are believed by them to have then re-precipitated their calcium carbonate as cryptocrystalline calcite cement. The decay of plant material has furthermore been suggested by Siever (1962) to be the cause of the common replacement of plant tissue by silica.

An alternative explanation for the preferential cementation of plant roots may be that ions that combine to form such salts as calcite, gypsum and halite are only incorporated in the plant tissue in limited quantities. The remainder are selectively rejected as the plants absorb moisture from the sediment. There would then be a concentration of ions capable of cementing sediment in the immediate vicinity of the plant roots. No living plants, however, were found to have their roots encased in cemented sediment along the Trucial Coast. It is interesting to note, in the context of this mechanism of cementation, that in Florida, where the climate is relatively humid, the cementing mineral is calcite, whereas along the arid Trucial Coast of the Persian Gulf the cement is gypsum. The two different cement types may represent successive stages in the normal evaporite cycle.

Either mechanism for the preferential cementation of plant roots by gypsum requires mineral-rich ground water, for example that found in a wadi or that derived from the sea. Plants that grew on dune sands away from mineral-rich ground water are unlikely to be recorded in fossil sediments by the preferential cementation of their root moulds.

ASSOCIATION OF WADI SEDIMENTS WITH FOSSIL DIKAKA

The following is an example of how the environmental significance of dikaka has been of use. At Jebel Baraka on the coast of the Trucial State of Abu Dhabi, a large mammalian tooth and part of the attached jawbone were collected.

PLATE I

Fossil dikaka from the Tertiary desert sediments of Jebel Baraka on the coast of the Trucial State of Abu Dhabi.
A. Preferentially cemented plant-root moulds (dikaka) in dune sands in which the dune bedding is still clearly visible.
B. Dune bedding almost destroyed by fossil dikaka.
C. Close-up of B showing the almost complete destruction of dune bedding (sub-horizontal) by dikaka.
by the writers. The tooth was found in one of a series of water-borne pebble conglomerates interbedded with aeolian sand (Fig. 1), and belongs to a large *Mastodon* (*Tetralophodon*) sp., which is confined to the Pontian stage (upper half of the Lower Pliocene). The thought of a large *Mastodon* roaming an arid desert is at first a little disconcerting, since such mammals are usually associated with more temperate regions. After consideration of the section in which the tooth was collected, however, a plausible explanation can be found for its presence.

The section shows not only alternations of wind-blown sand and water-borne pebble conglomerates (Fig. 1), but also abundant dikaka (Plate I). The sequence is interpreted as having been deposited in a wadi environment within a desert. The prolonged presence of a wadi in this area apparently permitted sufficient moisture to be retained in the associated aeolian sediment to allow considerable plant growth there. This situation is also found in modern deserts where wadi tracts can be recognised at considerable distance because of the relatively abundant vegetation associated with them (see Plate IIIA, an oblique aerial photograph of the wadis of an alluvial fan crossing the desert west of the Oman Mountains). The occurrence of the *Mastodon* tooth in the fossil wadi gravel may be explained by the presence of such vegetation, which may have attracted the animal along the wadi tract.

According to Osborne (1942), the *Tetralophodontinae*, which form a subfamily of the *Mastodontoidea*, range from Mio-Pliocene to Middle Pleistocene in age. They migrated during that time from Africa to Europe, and via Asia to North America. They became rare in Eurasia early in the Pliocene, but lingered on until Middle Pleistocene time in North America. It is interesting to note that the *Mastodon* (*Tetralophodon*) considered here appears to fit into this migration pattern, ending its life approximately on the migration path in the late Tertiary desert environment of Arabia. To judge from Osborne's (1942, pl.16) map of the migration and evolution of the *Tetralophodontinae*, this *Tetralophodon* must have been one of the last to have lived in Arabia.

---

**PLATE II**

Fossil dikaka.

A. Preferentially cemented envelopes of former roots in dune sand. Quaternary, Jebel Fuweirat, Qatar.


PLATE II

Palaeogeography, Palaeoclimatol., Palaeoecol., 4 (1968) 77–87
DIKAKA AND STABILISATION OF DUNE SAND

There are two methods by which plant roots may become incorporated in aeolian sands and thereby stabilise dunes. The first (Plate IIIB) involves a situation in which pre-existing vegetation is slowly enveloped by drifting sand. Since this sand is trapped among the roots and branches of the plants, well-defined laminae are unlikely to form except in the lee of scrub bushes. The aeolian origin of such a sand would be more than usually difficult to interpret in the geological record. The second method is by the normal growth of a plant from seed on a pre-existing sand dune (Plate IVA). It is, however, extremely difficult for plants to establish themselves on active windblown sand. They are usually either smothered by drifting sand, or their soil, the dune itself, is removed from their roots by wind erosion.

Unlike the cemented dikaka seen at Jebel Baraka (Fig.1, Plate I), the aeolian sand stabilised by dikaka of the Rajasthan Desert, India, (Plate IIIC), has no wadis associated with it. From observations in the latter area, it is suspected that the process of stabilisation of these dunes by plants was assisted by the formation of a thin crust cemented by carbonate, halite or gypsum. Halite and gypsum particles, blown inland from the Rann of Kutch or other salt-rich depressions by the Southwest Monsoon, dissolve slightly during periods of rain or heavy dew. During subsequent periods of evaporation they can precipitate and cement the surface sand. Such lightly cemented crusts would prevent young plants from being overwhelmed by shifting sand and the effect of both crust and vegetation would be to stabilise the dune.

Destruction of a protective cover of vegetation by man or grazing camels or goats can result in stabilised dunes reverting to mobile dune sand (Plate IIID). The sand eventually swamps and kills other plants, as seen in the interdune area in the foreground of Plate IIID. Although no exhumed cemented dikaka was found in this area, occasional fragments of calcite-cemented dikaka are incorporated in the mobile dune sands seen in Plate IIID.

It appears that seeds (Plate IVA) may become incorporated into a sand dune and remain dormant until the sand is moistened by rain. If there is insufficient moisture for the plant to break through to the surface and grow to maturity, it may only sprout, make a boring in the sand, and die (Plate IVA). In time, the carbonaceous matter will probably be completely removed as a result of oxidation, but the evidence of its existence may be preserved by the borings, infilled from above with sand of different colour or texture (Plate IVB).

Not all borings in dune sands are made by plants. Some may be made by insects such as the large black ant, seen in Plate IVC, which throws out pellets of sand possibly bound by mucus. Borings may also be produced by marine organisms if the sea transgresses a desert.
Recent dikaka.

A. Oblique aerial photograph showing the preferential distribution of vegetation along the wadi tracts of an alluvial fan. West of the Oman mountains, Arabia.


C. Dikaka-covered sand dunes. Rajasthan Desert, India.

D. Grazing animals may destroy dikaka, which has previously stabilised dunes, so that mobile dune sand results. Shahgarh, Rajasthan Desert, India.
PLATE IV
CONCLUSIONS

It is axiomatic that dikaka, as represented by fossil plant-root moulds in dune sand, indicates that vegetation grew in a desert environment because of a sufficient supply of water. Cementation, associated with the dikaka, may indicate proximity to a former wadi or marine environment. The absence of cemented dikaka suggests the lack of such mineral-rich ground water and instead, plant-growth in dune sands moistened mainly by rain water.

Dikaka assists the stabilisation of sand dunes. In so doing, however, it may destroy most of the original dune bedding, an important sedimentary feature from which aeolian deposition may normally be inferred. Consequently where plant-root structures are found, evidence should be sought in the form of locally preserved dune bedding, to determine whether or not the host sediment is of aeolian origin.

ACKNOWLEDGEMENTS

The authors are indebted to Professor Dehm of the Institut für Paläontologie und Historische Geologie, Munich, for identifying the tooth, found in the Tertiary desert deposits of the Trucial Coast, as belonging to a *Mastodon (Tetralophodon)* sp. of Pontian age. They are grateful to the Management of Shell Research N.V. for permission to publish this paper and to several colleagues at the Koninklijke/Shell Exploratie en Produktie Laboratorium for assistance and guidance in its preparation.

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PLATE IV

Recent dikaka and associated structures.
A. Plant seeds, buried in recently rain-moistened dune sand, in the process of sprouting and burrowing. Zanzur, Libya.
B. Quartz/carbonate dune sand riddled with burrows made by plant roots. Zanzur, Libya.
C. Large black an (arrow) in the process of making the burrows of its nest in dune sand. It throws up a pile of sand pellets bound possibly by mucus. Near Zliten, Libya.

*Palaeogeography, Palaeoclimatol., Palaeoecol.*, 4 (1968) 77–87
DEVELOPMENTS IN SEDIMENTOLOGY 14

DESSERT SEDIMENTARY ENVIRONMENTS

By

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ELSEVIER PUBLISHING COMPANY Amsterdam London New York, 1970
Frontispiece: Space photograph of the eastern part of the Arabian Peninsula, showing the eastern Oman Mountains, Wahiba Sands, wadi gravel fans and the Arabian Sea. (Taken from the Gemini IV spacecraft during its June 3-7, 1965 orbital mission. Photo credit: National Aeronautics and Space Administration.)
About one fifth of the world’s land surface is desert but, contrary to the popular “Lawrence of Arabia” concept, only one fifth of the desert areas is covered with sand (Holmes, 1965, p.760; see also Fig.1). Fossil desert sediments (mostly dune sands) have been recognised in various periods from the Cambrian to the present. The desert origin of many of such formations has been in dispute, however, for various reasons; sometimes it was because authors were discussing sediments of the same age but of different facies and locations, the one in a desert facies, and the other closely associated with marine fossils (McKee, 1962, p.558). In other cases it is apparent that the authors did not know what factors were significant in deciding between desert and non-desert sediments, or the differences between aeolian and water-laid sands.

Widespread desert conditions are known to have occurred during the Permo-Trias and continental formations of this age, often referred to in Europe as the New Red Sandstone, contain important oil and gas reservoirs in North Africa (Algeria), Europe (Germany and The Netherlands), and in the United States. It follows that, since little is known about desert sediments, an increased knowledge of them and their inter-relationship with sediments deposited in other environments is of prime importance to the oil industry. Geologists in other walks of life, whether they be concerned with palaeontology or palaeobotany, sedimentology or stratigraphy, will wish to relate their observations to a particular environment within a palaeogeographic setting, and in all these branches of geology, the environment in question could be that of a desert. To them also, then, a knowledge of deserts and their sediments could be of importance.

Deserts were first described for the specific benefit of the geologist by Walther in his paper of 1888. He subsequently published several more papers and a textbook, Das Gesetz der Wüstenbildung, which ran to four editions and is still the most complete authority on the subject for German students.

Since Walther’s classic papers and book, much has been published on desert travel and life in the desert by such authors as Lawrence (1935), Thesiger (1948, 1959), Gautier (1935) and Gerster (1960). While of geographical and ethnological interest, their works do not help greatly in the geological interpretation of deserts.
sediments. Bagnold (1941) in his classic work *The Physics of Blown Sand and Desert Dunes* tells how sand is transported in air and suggests the process by which dunes are formed. Several other authors have written on deserts, but their theme has usually been concerned with the morphology of dunes (Capot-Rey, 1949; Holm, 1953, 1960; Smith, 1965), or with granulometric detail (Alimen and Fenet, 1954; Tricart and Mainguet, 1965).

Perhaps of greater use to the field geologist are the interpretations made of fossil desert environments by Sherlock (1947), Shotton (1937, 1956) and Laming (1966) in Europe, and in North America by McKee (1934, 1945, 1954). These authors, however, tend to confine their environmental analyses to the more obvious dune, coastal and wadi sediments. The shortage of literature from outside North America prior to about 1960 that covers all geological aspects of deserts and their sediments can be attributed to the past difficulties and cost of travelling over many of these inhospitable areas. A few years ago, however, the search for oil was intensified in desert areas. This led to an invasion by nomadic technicians, and the difficulties of travel were eased by the use of mechanical transport that could cover in a day distances that still take a week by camel. Those geologists who entered the desert were, however, usually too busy studying older non-desert rocks to have much time to spare for the recent desert environment. The subject remained more or less neglected outside North America and the "French" Sahara until the need for agricultural development of modern deserts and the finding of hydrocarbons in ancient desert rocks led to the economic necessity of understanding them. This need also resulted in the recent surveys by the writer. The recent geological interest on deserts and their sediments can also be seen from the great increase in literature on deserts during the 1960's.

By presenting pictorially a wide range of recent desert sedimentary structures and environments in relation to their geographic setting, the writer hopes to add to the knowledge of the geological aspects of present-day deserts. The examples are based on a personal knowledge of modern deserts in Libya, southeast Arabia and India. The geographic distribution of desert sediments of parts of Libya and southeast Arabia are given in Enclosures 1-4. Comparative fossil occurrences from the Permo-Triassic sediments of northwest Europe and the Tertiary to sub-Recent deserts of Arabia and Libya are used to indicate the appearance of the sedimentary structure when seen in outcrop or core. Although proof of the mechanism is normally lacking, tentative explanations are given to account for the process by which many of the sedimentary structures are formed.

Although the principles outlined in this book are thought to apply to deserts in general, it is realised that other, so far unrecognised, sedimentary processes may occur—both in deserts of which the writer has a personal knowledge and those of which he has not. Others, no doubt, will remedy this deficiency.

Deserts occurring in other environments such as the high altitude and moderate latitude (45° N) Gobi Desert, which is a very long way from the sea, or the polar areas of low precipitation, have no place in this book except, perhaps occasionally, for purposes of contrast with the hot deserts of more tropical latitudes. Explanations for many of the more unusual terms used in describing deserts and their sediments are given in the glossary at the end of the book.

Provided that the reader can acquire sufficient data from the book to enable him to interpret a group of ancient rocks as having been deposited in a desert environment, and can interpret particular beds as "probably water-laid" or "probably aeolian" from cores and drill cuttings as well as in outcrop, then the book will have served its initial purpose. If it has assisted in making more precise palaeogeographic interpretations, so much the better.

This book is the outcome of research undertaken at the Koninklijke/Shell Exploratie en Productie Laboratorium (K.S.E.P.L.), Rijswijk, The Netherlands, with the addition of facilities for field work provided by the Royal Dutch/Shell Group of companies in Europe, Libya, eastern Arabia and India. In the latter country, Mr. S. Biswas and Mr. R. Mahotra, geologists of India's Oil and Natural Gas Commission, acted as guides in, respectively, the Ranns of Kutch and the Rajasthan Desert; without their help, much less of this interesting area would have been seen. In Libya and on three visits to Germany, the writer was accompanied by E. Oomkens. In Arabia, his companion was B. D. Evamy whose particular interest on the trip was the diagenesis of carbonate rocks. In India, G. Evans of Imperial College, London, accompanied the writer and was able to make valuable comparisons between the sediments of the Ranns of Kutch and those of the Persian Gulf. Many of the writer's ideas have been discussed with his field companions and other colleagues both in The Netherlands and elsewhere; this has proved invaluable as an aid to clear presentation. In addition, the writer was able to draw upon his colleagues' wide range of experience in many branches of geology. The field study was completed in under three years. It is natural, therefore, that gaps should remain in our knowledge and questions be left unanswered. It is hoped that there are no errors in interpretation.

The bulk of this book was prepared as a K.S.E.P.L. report in 1966. Since then, the writer has spent two winters in Oman and the Trucial States studying an entirely different geological problem. The opportunity to revisit this desert region has enabled him to recheck some of his ideas. Two short visits to Iran have added to his understanding of deserts in basins of inland drainage.

This book is published by permission of Shell Research N.V., The Hague, The Netherlands, to whom thanks are given. The maps and figures were prepared under the direction of Mr. R. R. J. Davilar, Head of the K.S.E.P.L. Exploration Draughting Department, whose assistance is gratefully acknowledged. Thanks are extended to Professor K. H. Wolf of Oregon State University and Dr. E. D. McKe of the United States Department of the Interior for advice and constructive criticism after reading a draft of the book. The writer remains, however, solely responsible for any errors or defects in style.
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Chapter 1

SOME GENERALISATIONS CONCERNING DESERTS

ARIDITY LEADS TO DESERT FORMATION

Essentially, a desert, in the sense that we are concerned with of a hot tropical desert, is a land area on which there is little or no plant cover because of an insufficiency of regular rainfall; a place where the potential rate of evaporation exceeds the annual precipitation. This may be the result of several factors such as the presence of a mountain barrier in the path of the prevailing wind and the creation of a rain shadow beyond, prevailing off-shore winds, insufficient relief to induce rainfall from temporary onshore winds or wind systems of natural low humidity (Fig.1).

The upper limit for average annual rainfall in a desert is 10 inches (25 cm) according to Holmes (1965, p.769) and this is reached only near the desert margins, with the adjoining semi-arid regions averaging between 10 and 20 inches (25–50 cm) annually. This obviously applies only to “hot deserts” where the potential rate of evaporation is much greater than the rainfall because of the high average temperature and low humidity.

In some of the arctic regions of North America and Eurasia, the annual precipitation (including the snow equivalent) is less than 10 inches, but the annual rate of evaporation does not exceed the rainfall. Plants have difficulty in maintaining an existence in these polar areas because of the problems of absorbing water into the plant tissues at sub-freezing temperatures. In the hot deserts, on the other hand, there is often little moisture to absorb and the xerophytic plants that inhabit these regions are adapted to reduce evaporation from their surfaces to a minimum.

Whether near a coastline or in the interior of a sand sea, it is rarely so arid in a desert that no plant life can exist. In the few areas where no plants occur, their absence is usually the result of continually shifting sands that prevent the plants from obtaining a firm hold with their roots, or of a salt concentration that is too great to permit plants to exist.

1 For example, Kendall and Skipwith (1969) quote Privett (1959) as estimating that evaporation rates in the southern Persian Gulf can be as high as 124 cm per annum. This is more than thirty times the annual precipitation recorded at Tarif on the coast of Abu Dhabi (average of 3.7 cm per annum) from 1958 to 1964, with a maximum of 6.73 cm and a minimum of 0.33 cm (Kendall and Skipwith, 1969, table 1).
METEOROLOGICAL REASONS FOR DESERTS

The atmospheric conditions that aid the initiation, growth and maintenance of deserts result from the planetary system of high and low air-pressure belts (Fig. 2). These air-pressure belts are caused by the circulation of air between the hot equatorial belt and the cold polar regions. If the earth were stationary with the sun revolving around it, a simple convection system would exist with hot air rising over the equator and cold air descending over the poles so that there would be a surface wind blowing towards the equator. The earth, however, rotates, and the winds of the convection system are consequently deflected by the Coriolis force, which is associated with the difference in the earth's rotational velocity at the equator and at the poles. In the equatorial regions, winds tend to be left behind by the earth, so that the convection pattern is deflected towards the west. A further complication is due to the build-up of other high-pressure belts between the equator and the poles. They are formed by hot equatorial air passing into latitudes that are shorter than the equator. These high-pressure belts, known as the Horse Latitudes,
occur roughly 30° N and 30° S of the equator although, because of the unequal distribution of land and sea, the northern one is slightly further from the equator than the other. The descending high-pressure air of the Horse Latitudes gives rise to clear skies that are not conducive to rainfall. The outflowing air that is directed towards the equator—the “trade winds”—is deflected to the west. The winds that blow towards the poles from the Horse Latitudes are deflected to the east and are known as the “westerlies”.

The sun appears directly overhead at the equator only at the two equinoxes. At the solstices, the sun is vertically overhead at either the Tropic of Cancer or the Tropic of Capricorn. As a result of these seasonal changes in the declination of the sun, the wind systems move about 5° north of their mean positions in the Northern Hemisphere summer and an equivalent distance south in its winter.

TRADE-WIND DESERTS

When the trade winds blow across continental areas such as North Africa and Arabia, they usually do so under an almost clear blue sky and no rain falls on the desert landscape across which they pass. In fact the air heats up over the barren desert surface as it moves towards the equator and tends to desiccate still further the land beneath because of its ability to absorb more moisture. This is the reason why the hot and very dry Harmattan, blowing from east or northeast from the Sahara Desert, brings such welcome relief to the inhabitants of the hot and very humid Guinea Coast of West Africa. It is also, no doubt, a factor that gives rise to occasional reports that certain deserts are extending downwind.

Most of the major deserts of the world lie in areas where the trade winds originate in, or pass over, extensive land masses (Fig.1). This is especially true of the Arabian and North African deserts.

Over the sea, the trade winds have a shallow moist layer capped by a stable inversion layer above which the air is very dry (HARE, 1961). When, however, there is a mountain barrier in their path, rain can be precipitated as the surface air is forced to rise over the mountains and a desert can occur in the rainshadow region beyond. This appears to be the reason for the existence of the deserts of central and western Australia. The Southeast Trades lose their moisture on the mountains of the Great Divide in eastern Australia, and although there are other ranges in central Australia, the air is already too hot and relatively dry for them to induce further rainfall (Fig.1). Similarly in South Africa, the trades lose much of their moisture over the Drakensberg and Cape ranges; the Kalahari Desert is beyond.

In the region of South America, the Horse Latitudes are virtually split into two by the Andes, with one high-pressure system centred over the southern Pacific and a second high-pressure system centred over the southern Atlantic. The anticlockwise winds associated with the South Pacific high-pressure system, blow across the Atacama Desert from the southwest and swing north parallel to the Andes without normally precipitating rain.

In the very narrow southern part of the Atacama Desert, the effect of strong onshore winds—possibly accentuated by convection currents over the Andes foothills—causes sand transport towards the east, that is towards the Andes. Further north, sand is transported to the north by trade winds that here blow roughly parallel to the Andes (Fig.1). The stabilising effect on the atmosphere of the cold Humboldt current that rises along this coastline, is an added factor that prevents rain (WALLÉN, 1966, p.34). Similar conditions exist along the west coast of South Africa where the Benguela current in the South Atlantic flows north. As a consequence, much of the coastal area of southwestern Africa (Namib Desert) ranges from semi-arid to intensely arid (FLINT, 1959).

Drought-inducing rain shadows exist also in the lee of the westerlies for both the arid plains of Patagonia and the intermontane Mohave and Sonoran deserts of the U.S.A. and Mexico. They are to the east of the Andes and Coast Ranges, respectively. The Gobi Desert represents an extreme case of a rain-shadow desert. It lies far from the sea in a great continental basin that is ringed by high mountain ranges.

MONSOON DESERTS

The existence of large land masses and the presence of high mountain ranges can strongly modify the ideal wind patterns shown in Fig.2. The Southwest Monsoon has its origins in the southern part of the Indian Ocean as the Southeast Trade Winds. It is drawn north of the equator by a low-pressure area over northwest India. As it crosses peninsular India, part of it swings north and then west over the Ganges Plain, to lose the last of its moisture on the eastern slopes of the Aravalli Range. The Rajasthan Desert lies beyond, to the west. The moist air that forms the major part of the airstream that crosses Rajasthan during the monsoon blows over the desert from the southwest and is confined to the near-surface layers.

In order to produce rain, moist air must be cooled. This is often achieved by an increase in altitude. PRAMANIK (1952) and RAMANATHAN (1952) have both pointed out, however, that over Rajasthan the near-surface air of the Southwest Monsoon is overlain by relatively warm, dry, anticyclonic air centred over Baluchistan. This will absorb any moist air that may rise and results in the dissipation of any cloud and decreases the likelihood of rain over the desert. The Southwest Monsoon is replaced in the autumn by light winds of the Northeast Monsoon. These, having had a long continental route before reaching Rajasthan, have little chance of bringing rain to the area.

1 Although very arid, mist keeps the surface air over coastal parts of the desert saturated with moisture during much of the year; it rarely rains, however (LOGAN, 1960).
POLAR DESERTS

Like the Horse Latitudes, the poles also are regions of high atmospheric pressure. This is because of the weight of cold, relatively dry, descending air. As the air flows away from the poles it is deflected to the west in response to the higher rotational velocity of lower latitudes. The high-pressure polar air system is characterised by clear skies and low precipitation, whether it be in the form of rain or snow. Because of this relative aridity, late Pleistocene winds were able to pick up the dry, unconsolidated sands deposited by glacial rivers during spring floods or left behind by retreating ice sheets, and build them into sand dunes such as can now be seen on the northern plains of the United States (Smith, 1965; Germany and The Netherlands. The morphology of these dunes appears to have much in common with that of dunes in hot deserts.

It is apparent that deserts occur in continental areas of clear skies associated with high atmospheric pressure. The presence of a desert may modify the distribution of the high-pressure belt, but it is not the reason for its occurrence. As many a torpedoed seaman knew during the early 1940's, the trade winds might help him to sail to safety, but they did not always provide him with drinking water. “Water, water everywhere, nor any drop to drink.” (The Rime of the Ancient Mariner, Coleridge.)

DESERT RAIN

As a result of temporary meteorological disturbances—often initiated outside the desert—rain does occur in desert areas. The frequency of rainfall may vary from several times annually to once in ten or even fifty or more years. Areas such as mountain ranges bordering deserts, or deserts bordering the sea usually receive annual though sparse rainfall. Away from the coast, it becomes increasingly insufficient to overcome the evaporation which takes place during the rest of the year.

Torrential rain can fall in a desert for a matter of hours, or even days. Since there is little soil or vegetation to hold the water in hilly areas, this results in rapid run-off; floods ensue on the plains with little warning and transport sediment over great distances. With no more rain to maintain the flow, the waters may subside almost as rapidly (McGee, 1897). The sporadically flowing watercourses that form as a result of desert rain are known to the Arabs as wadis. Deposits of thin clay form films, which dry out, crack, curl up, and may be preserved in the geological record if covered by a protective layer of wind-blown sand. If not protected in this manner, then they are either fragmented to dust and blown away, or washed away by a succeeding flood and again deposited as a layer of clay, or as clay pebbles.

SOURCES OF DUNE SAND

The sand that wind transports comes, in part, from chemical weathering and from the fracturing of rock surfaces by rapid temperature changes and the abrasive action of wind-blown sand. This desert weathering and erosion occurs not only
over the broad expanses of outcrop, but applies also to areas where aquatic sedimentation is predominant. Any breakdown of older rock or recently cemented sediment gives rise to a supply of sand that the wind is capable of transporting.

A second and possibly equally important source of sand is the sediments that are deposited in alluvial fans by the wadis. The loose sand is picked up and removed by the wind, leaving a lag of coarser pebbles and boulders.

A third and more restricted source of sand is the coastal beaches bordering the deserts. With a tropical sea, this sand often consists almost entirely of shell fragments, ooliths and Foraminifera. Under near-coastal conditions, this carbonate-rich sand can cement rapidly and form a protective barrier against the incursion of the sea. The shell fragments tend to abrade more rapidly than quartz, however, and are not usually recognisable in quantity more than 50 or 100 km from the coast.

**AEOLIAN SEDIMENTATION**

Whatever their derivation, sand grains are transported over the desert by the wind, until the surface wind velocity is reduced sufficiently for them to come to rest. On the large scale, this occurs in sinking continental basins or topographic traps formed by the uplift of a range of hills of mountains. On the small scale, the sand grains may come to rest in the comparative protection of the lee side of boulders or vegetation where sand drifts form, or in the shelter of a wadi bank where the aeolian sands cover fluviatile sediments. With greater quantities of sand, sand dunes develop, with deposition taking place, in the case of the simple crescent or barchan dune, on the lee slope. Because loose sand is readily transported by the wind, most major accumulations of sand dunes—the sand seas of desert regions—occur in depressions. These depressions may be structural, in the form of grabens or basins, or they may be lowland areas, surrounded by hills on two or more sides, that were eroded or peneplaned before the onset of desert conditions.

**BASINS OF INLAND DRAINAGE**

Many basins of inland drainage form the sites of deserts (Gobi Desert in Central Asia, Great Kavir of central Iran, Mohave Desert of California). This is very much the case with some of those basins that are situated within or near to the Tropics and have formed as the result of Tertiary mountain building. Since the river water of such regions never reaches the sea, evaporation from the surface of the terminal lake or swamp results in a steady increase in the concentration of salts that were carried in solution. Most plants are unable to live in saline soils; a decreasing cover of vegetation encourages more rapid evaporation from the surface of the soil and desert conditions ensue.

![Fig. 1: Schematic cross-sections to illustrate how unconsolidated dune sands may be preserved beneath the wave base of a transgressing sea.](image-url)
Within a basin of inland drainage, the type of sediment fill will partly depend upon relief and climate. Provided there is sufficient rainfall, a lake will form at the lowest point in the basin and fluvial sediments will be deposited in channels leading to the lake. As the central part of the basin fills with sediment, the local base level will slowly rise (Davis, 1936). With a decrease in rainfall, rivers may flow only seasonally. Much sediment will be deposited on alluvial fans at the foot of the surrounding hills, but large amounts of gravel, sand, silt and clay can still be carried into local terminal basins by floodwater. After drying out, the wind can remove from the area the finer fractions of this fluvial sediment with sand-sized particles forming migrating dunes. Under increasingly more arid conditions, the base level in the central part of the basin may be lowered as sediment is removed by the wind more rapidly than it is replaced by stream flow. Occasional fluvial transport of sediment into these terminal basins results in sand, silt and clay settling between dunes. Interstratification of aeolian sand and water-laid sediment is characteristic of many areas of Central Asia (Horner, 1936).

**Preservation of Fossil Dune Sands**

The question is sometimes posed: "How can unconsolidated dune sands survive the erosive action of a transgressing sea?" Fig. 3 goes far to answer this. Dune sands of the coastal plain are protected behind cemented carbonate dunes and even lightly cemented quartz dunes of the coastal dune barrier. These coastal dunes can themselves be protected from strong wave action by the upward and seaward development of limestone reefs and coastal lagoons. Even in the case of completely unconsolidated dune sands, although the surface sand will be reworked as the relative sea level rises, it thereby forms a "buffer zone" preventing erosion of the aeolian sand beneath.

Inland, great thicknesses of unconsolidated dune sand may accumulate in subsiding continental basins. A transgressing sea will only be capable of reworking the dune sands down to the wave base. For example, in Libya, 120 m of unconsolidated sands overlie the Cretaceous in one basin, and 1,200 m overlie Palaeozoic formations in another (R. Jordi, personal communication, 1964; see also Enclosure 4). McKee and Tibbitts (1964) refer to the presence of about 300 m of aeolian sand south of Bir Zelten in Cyrenaica. This places the base of the sand at this locality some 100 m below the present level of the Mediterranean Sea.

**Chapter 2**

**Some Criteria for Identifying Ancient Desert Sediments**

The following criteria are given as an aid to identifying fossil desert sedimentary environments by recognition of: (a) wind-deposited sands; (b) water-laid sediments; (c) sub-aerial exposure of water-laid sediments.

They are presented here as unsupported statements so that the reader may appreciate the significance of the various sedimentary patterns shown in the following chapters. That these statements are essentially correct will become apparent from the examples given. The numbers in brackets refer to figures in which these features are well shown. It is emphasised that no one criterion is sufficient to permit a positive identification of a desert environment to be made, but it may be enough to suggest it. Individually, many of the criteria can also be found in the sediments of non-desert environments. Criteria for the subsurface recognition of desert sedimentary rocks from cores and drill cuttings are given in Table I (p. 144).

**Identification of Wind-deposited Sands in a Desert Environment**

1. Sand laminae horizontal (Fig. 88–90, 115, 122) or usually with either fine (centimetre) or large-scale (several metres) foresets having either: (a) constant orientation (crecent or barchan dune type) (Fig. 84, 87, 102, 115, 117, 120, 121); (b) multiple orientation (linear or seif dune type) (Fig. 79, 103, 122).
2. Individual laminae well sorted especially in the finer grain sizes; sharp grain-size differences between laminae common (Fig. 88, 90, 120–122, 124–128).
3. Grain size commonly ranges from silt (60 μ) to coarse sand (2,000 μ). Maximum size for grains transported under the action of wind is in the order of 1 cm but grains over 5 mm are rare (Fig. 6, 30–36, 43–48, 64–66, 124–129, 131–136, 138, 139).
4. The larger sand grains (500–1,000 μ) tend to be well rounded (Fig. 124–129, 131–139).
5. Clay drapes very rare (Fig. 62, 79, 87, 88, 90, 91, 102, 103).
6. Sands free of clay1 (Fig. 66, 125–128).

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1 This criterion implies that the sand was free of clay at the time of deposition and does not preclude the possibility that authigenic clay may be present in ancient dune sands as a result of later diagenesis.
Quartz grains that have not been cemented by calcite usually exhibit a frosted (pitted) surface under the microscope (Fig. 124, 131–139).

Mica generally absent.

IDENTIFICATION OF WATER-LAIÉD SEDIMENTS IN A DESERT ENVIRONMENT

1. The bedding may be that of normal stream-flow sediments of both the upper and lower flow regimes (foresetted gravels, sand ripples, dunes, plane-bed, antidunes etc.) (Fig. 18, 25, 27, 28, 33, 36, 42, 43).
2. The bedding may also appear as unsorted mudflow conglomerates (Fig. 118), sheet-flood sediments, horizontally laminated and steeply foresetted flood-plain deposits, and as contorted (recumbent) folds (Fig. 94, 95).
3. Channeling (Fig. 27, 28) and accretion foresets common in stream-flow sediments.
4. Clay laminae or clay drapes common (Fig. 37, 38, 43, 56–60, 62).
5. Many sands are not well sorted, being sometimes argillaceous and pebbly (Fig. 31, 34, 45).
7. Sand and silt layers commonly missing from graded conglomerates (Fig. 18, 27, 28).
8. Majority of quartz grains that are not cemented by calcite exhibit frosted (pitted) surfaces under the microscope.
9. Water-laid sediments that are deposited in a desert environment may also possess any of the features mentioned in the succeeding set of criteria.

RECOGNITION OF SUB-AERIAL EXPOSURE OF WATER-LAIÉD SEDIMENTS IN A DESERT ENVIRONMENT

1. Presence of curled (generally with concave side up) clay flakes (Fig. 36, 41–48, 110).
2. Common presence of clay pebbles (Fig. 36, 43–48, 56).
3. Presence of mud cracks with sandy infill (Fig. 37–42, 110–113).
4. Presence of sandstone dykes (Fig. 42, 56–58, 110–113).
5. Aeolian sand interbedded with beds of aquatic origin. The contacts may show evidence of fluvial erosion or deflation (Fig. 24–27, 30–36, 43–48, 56–58, 61–62).

As a first approximation, the presence of red beds in an ancient sequence can be taken as an indication that they are of continental (although not necessarily desert) origin, were derived from nearby continental sediments or were secondarily reddened as the result, perhaps, of later desert conditions. Many ancient desert sands are often red, but many instances are known where they are grey, brown, yellow or even white. This subject will be dealt with in some detail in chapter 9, where additional criteria for the recognition of fossil desert sediments are given in Table IV (p. 192).

The above criteria have been found useful in the identification of different desert environments of deposition in the Permo-Triassic sedimentary rocks of northwest Europe.

1 So far, mica is only known from coastal dunes adjoining micaceous beach sand and in other wind-blown sands derived directly from mica-rich water-laid sediments, or nearby micaceous outcrops where the distance of aeolian transport from its source is assumed to be small. The absence of mica is not, by itself, indicative of dune sands but may rather be the result of provenance in water-laid sediments.
Chapter 3
DESERt EROSION AND DEFLATION

GENERAL

In the arid environment of a desert, flowing water plays only a small part in processes of erosion. More important are the effects of changes in temperature, which help to bring about exfoliation of the rock surface and the breaking up of rock and rock fragments. Coupled with these effects of insolation, are those of crystal growth within the rock itself (Wellman and Wilson, 1965), micro-chemical corrosion (Kuenen and Perdok, 1962) aided by wetting effects from rare rainfall and the frequent early-morning dews, and the abrasive action of wind-blown sand and dust.

EXFOLIATION AND INSOLATION

Rapid temperature changes in a desert environment as the means of splitting rocks was brought into disrepute by Blackwelder (1933, 1936). He pointed out that experimental work indicated that most rocks did not break at four times the temperature range of solar heating. Hobbs — in a discussion on Blackwelder's (1936) paper — opposed these arguments by pointing out that the presence in deserts of split rocks, large and small, called for an explanation. Morris (in the same discussion) pointed out that the rocks themselves had many inhomogeneities that predisposed the line along which it would fracture irrespective of the final "trigger" of splitting.

In another discussion on Blackwelder's paper, De Terra (1933) states "Insolation as a geologic process cannot be denied...". He supports this by pointing out that black surface gravel shows the effects of spalling more frequently than lighter rock fragments. He also quotes descriptions by Kaiser (1925) of angular fragments from disintegrated quartz veins that fitted each other when pieced together.

More recently, Sugden (1964) suggested that the facets found on ventifacts originate by fracturing, "partly attributable to diurnal heating and cooling, but ordinarily following flaws initiated otherwise". Bruckner (1966) attributes much...
DESERT EROSION AND DEFLATION

DEFLATION SURFACES

Over areas of flat desert, the processes of deflation cause the removal of the finer fragments by the wind, leaving a lag of coarser pebbles and boulders composed of more resistant rock (Fig.8). Before sand grains can be removed by the wind, they must first, of course, be deposited there by another means such as flowing water, or be derived locally from larger rocks. Chemical corrosion and insolation bring about slow fragmentation of the solid rock beneath the desert surface lag. Hörner (1936) and Beaumont (1968) believe that in areas of alternating wet and dry conditions of the ground, a high salt concentration in gravels can also play an important role in the breakdown of pebbles into finer fragments. Cooke and Smalley (1968) consider that thermal expansion of salts as well as growth of crystals from solution and hydration of anhydrous salts are also important factors in weathering of desert areas. If these hypotheses are correct, then this form of weathering may

DEFLATION SURFACES

of the reason for breakdown of the rock surface to “exfoliation in the widest sense, that is to say, the result of release of internal stress in the bedrock at or near its surface, ascribed solely to removal of load by erosion, or to temperature variations, or to differential hydration, or a combination of these and other causes”.

The writer has observed many boulders in the desert that can be seen to have split in situ, the pieces lying fractionally apart. Some have been seen on hillsides (where tumbling boulders could, of course, have caused them to crack by impact), but others have been noted on hilltops and open plains where no mechanical fracturing can be expected. Insolation seems the most likely reason for the splitting of larger boulders, and is also thought to play a part in the formation of much finer fragments (see p.167).

Fig.4 shows the south flank of the Fahud anticline in Oman. Beds of sub-rounded boulders of limestone cover the surface of the outcrop. However, on closer inspection, it is seen that these boulders are formed in situ by the exfoliation of bedded limestone (see also Fig.15). Whether this is caused solely by temperature changes, by differences in hydration, by chemical action, or a combination of all three, is uncertain.

The original bedding is still discernible in the layering of the boulders. No transport has taken place.

Fig.5 shows a limestone boulder which has fallen from a nearby cliff within the Fahud anticline. The way in which the surface layers are flaking off to produce a somewhat smaller, sub-rounded boulder is clearly seen1.

The finer sand surrounding the boulder of Fig.5 is angular (Fig.6). Fragments of both limestone and associated chert are present up to a diameter of roughly 3 mm. Transport by the wind has been practically nil, although some grains may have been blown back and forth over a comparatively small area and become almost spherical in the process. A small percentage may have been derived from outside the area. As seen from Fig.7, the sorting is very poor, and cannot compare with that of an ideal dune sand (cf. Fig.66, 136).

DEFLATION SURFACES

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1 With a different lithology, insolation can result in angular pebbles formed by the splitting, rather than flaking, of boulders. Borreck (1959, pp.36, 37) considers the widespread angular pebbles of the “dasht” in central Iran to have been formed by splitting.
be important in certain coastal areas and in desert regions of inland drainage. It is possible that gypsum acts in this way since it has been seen associated with highly fragmented sandstones east of the Umm as Samim in Oman (Enclosure 1). Areas which are more resistant to these forms of erosion stand slightly proud of the surface and break into rock slabs that are often the nearest thing to an outcrop for many miles around. When these lag deposits of pebbles and boulders cover large areas they are sometimes called "stony desert". They are referred to by the Arabs of North Africa as "serir" (Libya) or "reg" (Algerian Sahara), and as "gibber plains" in Australia.

A serir, then, is a more or less horizontal deflation lag resulting from the erosion of an older rock surface. The term applies equally well to the deflation surface of older alluvial sediments provided that it is not deeply dissected by erosion gullies. Because of the lack of a covering of pebbles, the term "serir" is not normally extended to include the deflated surface of peneplaned aeolian sediments.

Outcrops, and the pebble lags of deflation surfaces in desert areas, develop with time a shiny dark mineral coating known as desert varnish. According to Engel and Sharp (1958), desert varnish forms as the result of a weathering process that involves solution, transportation and deposition of manganese, iron and a variety of trace elements. They believe that the elements are derived from within the rock or, in the case of pebbles of a deflation lag, from the underlying weathered rock material. Moisture is almost certainly required for the varnish to form. Since rain is an infrequent visitor to the desert, it is thought that dew, which is fairly common in the desert after dusk, is important in providing a moist film along which ionic diffusion can take place. Opdyke (1961) assumes that the moisture is brought to the surface by capillary action.

The undersurfaces of pebbles do not receive a desert varnish but possess, rather, an orange, yellow or brown matt surface. Desert varnish does not form on soft friable rock, and can be destroyed by abrasion with wind-blown sand, burial, and presumably also by a climatic change to a wetter climate. Because desert varnish is destroyed as a result of burial, it is unlikely to be a factor that can be used for the recognition of arid conditions in ancient sediments.
The dark surfaces of the outcrops seen in Fig.15 are the result of desert varnish. In the deflation hollow below (see p.27), desert varnish has darkened areas of wadi gravel that have probably not been reworked by water for several decades. In the lighter-coloured areas, sediment has been transported in recent years and the varnish has not had time to form.

**VENTIFACTS**

Sand-blast abrasion of pebbles of a deflation lag is considered to cause facets to form on the windward surface of the pebbles and thereby forms ventifacts. If there are two dominant winds, facets will be cut on two surfaces. A pebble with one sharp edge forming the intersection between the two facets is known as an "Einkanter", one with two such edges, a "Zweikanter", and with three edges, a "Dreikanter". An analysis of the trends of such edges on some ventifact pebbles, and the apparent direction of the sand-laden wind that is presumed to have formed them, has been made by Higgins (1956). Deflation of the surrounding sediment undermines the pebbles and they topple over to expose another surface to the wind. In this way, a Dreikanter may be formed. Walther mentions having seen Dreikanters in his 1888 article, and described them at some length in his book *Das Gesetz der Wüstenbildung* (1900). Ventifacts have also been recorded by Bigarella and Salamuni (1964) from the Early Mesozoic Botucatu Sandstone of Brazil and Uruguay.

Ventifacts that have been formed solely by the action of the wind are probably not so common as is generally believed, however. So far, the writer has not found any preferred orientation of the facets of angular pebbles on a desert surface, and certainly none that could be referred to known prevailing winds either existing at present or inferred for earlier times.

Sand-blast abrasion undoubtedly occurs in the desert, but within the writer's experience, most of the angular pebbles found as a lag on the desert surface were probably formed by splitting as a result of the effects of insolation (see also the references to Sugden and Bobek on p.15 and 16). The polished surfaces that have also been associated in the past with ventifacts (e.g., Holmes, 1965, p.755) are almost certainly formed of desert varnish and, as has been mentioned above, desert varnish is destroyed by abrasion with wind-blown sand; this does not, however, preclude the possibility of earlier shaping by this means.

The writer has examples of pebbles from deflation surfaces whose exteriors show typical desert varnish but whose shapes appear to depend upon an earlier history (such as rounding by transport in a wadi) and the lithology of the rock from which it was derived; siliceous rocks seem to have broken with sharp edges as the result of insolation; argillaceous rocks have flaked surfaces formed by exfoliation, whilst calcareous rocks show a typical intricate pattern of surface grooves and ridges that could well be related to slow chemical solution resulting from dew or rare rain. Similar grooved and ridged limestones are mentioned by Maxson (1940).

The widespread presence of angular pebbles can, in many cases, be considered as indicative of desert conditions, but their universal origin by sand-blasting must be held to be in doubt; splitting by the effects of insolation followed by the formation of desert varnish would appear to be a more likely mechanism for the formation of many "ventifacts".

**SAND REMOVAL BY WIND**

When the wind blows with sufficient strength, it picks up the particles of sand and moves them, first by rolling, and then close to the surface in a low cloud of saltating grains. With a surface consisting of both fine and coarse sand grains, a certain wind velocity is capable of removing exposed fine grains, but is not capable of moving either the coarse grains or the fine grains they protect. With a greater velocity, both coarse and fine grains are set in motion by the wind. Some grains may still be too heavy to be moved by the direct force of the wind, but creep over the desert surface under the force of impact of the bombardment by smaller saltating grains.

At low wind velocities there is a tendency for selective transport of particles of low sphericity (Williams, 1964). This is well seen in the sands of many coastal dunes, where angular shell fragments derived from the adjoining beach have a greater sieve-size than quartz grains from the same lamina (Gibbons, 1967; Yaalon, 1967). Although not visible in the photo, this is also the case in the dune sands from near the coast of the Persian Gulf seen in Fig.122. This difference in grain size is less apparent farther from the coast. At greater wind velocities, on the other hand, transport of more spherical grains predominates and these grains reach a greater height during saltation than the more angular grains (Williams, 1964). This latter character is presumably a reflection of the greater elasticity of a rounded grain.

Fig.9 was taken during a sandstorm in Libya. Most of the sand was seen to move very close to the ground. Saltating grains of 1 mm diameter were collected about 1 m from the ground. Grains of about 5 mm diameter, too heavy to be moved by the direct action of the wind, were slowly creeping along the surface under the force of impact of saltating grains while grains of 2 mm diameter were rolling continuously over the ground. For sands of more uniform size, Bagnold (1941, p.34) tells us that between 20 and 25% of the total sand movement is by surface creep. Although the 5-mm grains will represent nothing like that percentage when the total weight of sand moved is considered, in some cases surface creep can result in sizeable sedimentary structures (see p.77 and Fig.64). Since no dust was available in the area for the wind to transport, the sky was clear when Fig.9 was
SAND ACCUMULATION

On p.72 of his book, Bagnold states that “a given wind can drive sand over a hard immobile surface at a considerably greater rate than is allowed by the loose sand covered surface” (BAGNOLD, 1941). This means that sand tends to accumulate on areas that are already sand covered, in preference to the bare surrounding areas (see Fig.10 and also Fig.69-71).

Pressure gradients also occur between the interdune and dune areas, caused by the resistance to the wind of the dunes themselves. This also means that sand will tend to be transported from the sand-free interdune areas towards the duties. A combination of these two factors explains why sand concentrates into individual dunes and groups of dunes where there is an insufficiency of sand to cover the whole area. With an excess of sand, the dunes grow in height and the interdune areas also become sand covered. The transport of sand from interdune to dune is then solely the result of eddies caused by the pressure gradient between the two areas. This subject will be discussed more fully in chapter 6.

DIFFERENTIAL DEFLATION

In desert areas of horizontal strata, variations in lithology or types and degree of cementation result in differing rates of deflation. In this way, considerable relief can develop without the necessity for fluvial erosion. Fig.11 shows such an area in the Jabal Uaddan, Libya, where hard limestone beds alternate with softer marls. The harder limestones tend to act as a protective “cap”, but, as the underlying marls weather and are blown away, so boulders of limestone, helped by insolation, become unsupported, break off and tumble down the slope. Further wind-scouring of the soft marl beneath the boulder causes the latter to tip over and move further.
down the slope away from the outcrop. In this way, a limited amount of spreading of a lag gravel is obtained.

Layers that are more resistant to erosion can result from caliche cementation of the desert surface. Such calcite-cemented layers of sediment, a metre or so thick, are common in some desert areas as the result of evaporation of carbonate-rich flood or ground water. If this protective layer is broken through by processes of erosion for any reason, then removal of the softer more poorly cemented sediment beneath can proceed at a faster rate and can give rise to an erosional form that is similar to that seen in Fig. 11. In some cases the protective caliche-cemented cap may overhang the softer sediment beneath to give the outcrop a "mushroom" shape (see Di Cesare et al., 1963, photo 3). Another example of differences in the rate of deflation between well and less-well cemented sediment is mentioned on p. 34 and illustrated in Fig. 19 and 20.

Concerning deflation of the clay and fine silt-sized particles produced by weathering of the marl, Bagnold (1941) has made a most important observation. For any given size of grain, there is a certain velocity of wind (named by Bagnold the "threshold velocity") at which the grains are set in motion. Fig. 12 is a reproduction of two of Bagnold's graphs. Fig. 12A shows that particles below about 80 μ have a higher threshold velocity than rather larger particles: below this limit, the finer the grain size, the greater is the threshold velocity, so that a wind which is strong enough to move pebbles 4.6 mm in diameter is unable to move finely scattered portland cement. He explains this by pointing out that there is a wind velocity gradient which reaches zero close to the ground, so that "once fine solid particles smaller than about 30 μ have settled on the ground after carriage in suspension by a wind, they cannot be swept up again individually, because they sink into a viscous surface layer of air and are out of reach of the disturbing influence of the eddies of turbulence". The same principle applies whether the dust has settled out of the air, or has been derived in situ by processes of weathering. The only way in which these fine particles can be moved—apart from scouring with water—is for them to be attacked by a storm of larger saltating sand grains which eject the dust particles above the viscous surface layer of air and thus force them into suspension in the turbulent air above. Once airborne, silt and clay-sized particles can be readily kept in suspension by virtue of their small size and weight, so that if the wind persists for long enough, this fine material may be carried great distances and be deposited as loess far beyond the desert margins.

DEFLECTION HOLLICWS

An extreme case of differential deflation can result in the wind scooping out a hole in what is otherwise a horizontal stony desert surface, such as occurred at Al Fugaha, Libya (Fig. 13). Here, there is an almost circular hole some 2 or 3 km
across and 60 to 70 m deep. The few short wadis all drain into the hollow but, nevertheless, it is believed by the author to have been formed primarily by wind erosion, aided possibly by the presence of strata which weather to a fine dust and can therefore be carried out in suspension by a strong wind. The depth to which deflation can penetrate is limited by the level of the water table. This is the case at Al Fugaha which, since the water there is not too saline, is also an oasis (see also p.74). The largest known depression thought to have been formed solely by deflation is the Qattara Depression (Fig.14) in northwest Egypt, which covers an area of some 18,000 km² and reaches a depth of 134 m below sea level. The sediment removed from the Qattara Depression by the wind is now found as part of a dune field extending away to the south (BALL, 1927, 1933).

Evaporation at the level of the water table in the base of such depressions often causes precipitation of gypsum and rock salt (cf. Fig.59). Salt marshes occur at several levels near the bottom of the Qattara Depression. They are by no means horizontal, and the lowest point in the depression is actually dry clay. Although evaporation of the groundwater has undoubtedly increased the salt content of the marshes, BALL (1933, p.292) points out that as the depression is approached there is an increase in the salinity of the water found in wells that tap the same aquifer. He suggests that solution of salt from these Tertiary rocks is responsible for this after the water has left its original Nubian Sandstone aquifer. An alternative possibility is that the aquifer that supplies the water to the marshes becomes enriched in salt because of the semi-permeable nature of the overlying Tertiary shales. Deflation brings the aquifer relatively nearer to the surface. Upward percolation of water through this semi-permeable barrier to the bottom of the depression brings about an increase in the salt content of the water still within the aquifer. Although part of the salt is retained in the aquifer, a percentage will pass through the shales and be precipitated at the surface as a salt crust. This process is discussed at greater length on p.60 in connection with other areas of salt precipitation.

At Fahud, Oman (Fig.15), a flat elongate plain forming the centre of a dissected anticline is surrounded by an escarpment whose beds dip away from the plain. Although there is now a thin partial covering of braided wadi sediments, the whole area is suspected of having been formed partly by deflation. At one point
there is a small gap in the escarpment on the southern side of the structure (Fig. 4),
but this is at a level which is a few metres higher than the lowest portion of the plain.
Although Tertiary gypsum was plentiful, this lowest part of the plain consisted
only of marl and occasional bands of limestone undergoing deflation. Even though
there had been some fluvial drainage, it was not sufficient in volume to carry
sediment to the lowest part of the basin and much of the evidence of its presence
might yet be removed by deflation before the next rainfall. It is apparent that the
wind is a very effective agent of sediment removal during the process of deflation
in a desert environment.

Most of these deep deflation hollows (Qattara, Fahud, Al Fugaha and others)
are known to be associated with calcareous rocks. ANDERSON (1947) has suggested
that in these cases the process of deflation is assisted by fragmentation of the rock
by solution and collapse associated with ground-water movement. He may well
be right.

Chapter 4

WADIS AND DESERT FLUVIATILE SEDIMENTS

WADIS IN GENERAL

As has already been stated, some rain does fall in the desert. In hilly areas this can
result in wadis becoming flooded and, especially in the lower reaches where aeolian
sediment is common, the flowing water in a wadi can transport considerable
amounts of sediment for a short period such as a day or two. With a low water/
sediment ratio, the wadi may be filled with a flowing slurry of sand or mud which
can support boulders of considerable size which are then deposited as an unsorted
muddy gravel—the mudflow conglomerates of BLACKWELDER (1928) and BLUCK
(1965, 1967). A higher water/sediment ratio gives normal stream sorting with some
grading of the sediment.

Because the flow of water is not maintained in a wadi, permanent channels,
such as occur with rivers are not found. As the water velocity in the wadi drops
and water is absorbed into the dry sediment over which it flows, so the sediment,
whether it was transported in suspension or by traction, is deposited. In this way,
the wadi channel may become partly filled with its own sediment or, later, by wind-
blown sand. Since deposition often takes place over a more or less planar deflated
surface, the next flood finds its old route partially or completely blocked. It must
then seek alternative routes and, in so doing, builds up a braided alluvial fan.

The use of the word "wadi" in the title of this chapter is significant. It im-
mediately indicates a form of fluviatile transport which is sporadic and abrupt.
This particular means of intermittent transport of sediment has many implications
which are discussed in the following pages.

Many of the landforms in modern deserts have been partly inherited from
a wetter past. Thus, many wadis now occupy erosional channels in areas of out-
crop, which were cut at a time of greatly increased water flow (e.g., during Plei-
stocene "pluvial" periods). This is well seen on the southern edge of the Rub al
Khali (Enclosure 1). Here, the wadi channels are partly filled with sediment, but at
present the flow of water is not sufficient to carry sediment beyond the areas of
outcrop. The existence of a more extensive system of rivers or wadis in former
times can be inferred from the fluviatile gravels now occupying the long sand-free
interdune corridors (or feidjs) along the southern edge of the Rub al Khali.

Much of the sediment found in wadi channels is deposited under normal
stream-flow conditions of deposition. The bed forms that result from different
regimes of stream-flow have been described in some detail by such workers as Simons and Richardson (1963, 1966), Jopling (1963, 1965), Allen (1963, 1968a, b) and Middleton (1965). At lower stream velocities small ripples form that are transverse to the flow direction. At a higher velocity of flow, the small ripples are abruptly replaced by larger ripples (dunes, or sand waves) that are also arranged transversely to flow. A further substantial increase in flow strength will commonly lead to the destruction of the large ripples and the formation of an apparently flat bed over which there is intense sand transport (Allen, 1968a). At still larger flow velocities (the upper flow regime of Simons and Richardson, 1966) the plane bed is replaced by standing waves (antidunes) that are roughly in phase with the waves of the water surface. Irregularities in the flow strength may result in scouring of the underlying sediments.

When water in flood breaks the banks of its containing channel, it may spread out over the surrounding countryside in a broad sheet. The sediments that are deposited during such a phenomenon have been described by McKee et al. (1967). They consist of a combination of steeply foresetted sands, horizontally laminated sands and sets of climbing ripples. The foresetted sands are interpreted as being the result of the formation of a sediment front where the water became deeper and was unable to maintain massive sand transport over a flat surface. The formation of foresets with relatively low angles is attributed, however, to current movement that was still strong at the time of deposition; the horizontally laminated sands are thought to have been deposited under conditions of the upper flow regime; the climbing ripples are considered to indicate overloading of the water with sediment as the water velocity falls (Allen, 1963; McKee, 1966b).

When sediment-laden flood water flows across sand that is saturated with water, the laminae of underlying cross-strata can be dragged forward into a recumbent fold (McKee et al., 1962; McKee, 1962). This, apparently, can occur when flood water flows into a standing body of water such as a shallow lake or the sea (see Fig. 94, 95).

**BRAIDED PATTERN OF WADI CHANNELS**

Wadis usually exhibit a braided-stream pattern of sedimentation. Braided streams develop in the continuously flowing rivers of non-desert areas when they are heavily laden with sediment and there is a rapid fall in stream velocity. This often occurs where rivers emerge from mountainous regions and flow out over gently sloping alluvial plains (see also Blissenbach, 1954).

The same principle applies to wadis. They, however, suffer from additional factors peculiar to their desert environment. The water in a wadi does not flow continuously all the year round. Normally, after rain, the water in the wadi will soak into the ground only a few hours, or at the most a few days, after it first starts to flow following rain. All the sediment carried in the water—and this may, in some instances, have almost the consistency of a mud flow—has to be deposited. Much of this sedimentation occurs in the vicinity of the downstream limit of water flow, so that locally, the wadi channels become clogged with gravel and sand. During the succeeding dry period of several months or even years, deflation will result in the removal of the finer sediment from the more exposed parts of the wadi channels and wind-transported sand may be deposited in the more sheltered parts. In some places, the channels may be partly, or even completely, blocked by wind-blown sand. The next time that the wadi again flows with water, it may flood too rapidly for the water to immediately remove all the aeolian sand blocking its path. The water in the wadi overflows its bank and scours new, or modifies old, channels.

Hörner (1936) describes similar processes of deflation and deposition in Central Asia that result in the strange phenomenon of wandering lakes, some of which, such as Lop Nor, migrate over wide areas.

Any tendency for the channel of a wadi to be filled with sediment is counteracted during the next flood by the cutting of new channels that spread out fanlike. This spreading of alluvial sediments is possibly also helped by deflation of the pre-existing sediments on either side of the wadi.

The structure of the sediments of braided streams in deserts is, in general, similar to that described by Doeglas (1962) for two tributaries of the Rhone in southern France. An important exception, however, is the change brought about by the wind; removal of the finer sediment from some areas and deposition of aeolian sand over others.

Fig. 16 is a photo of an alluvial fan occupying the area of a narrow coastal plain between the Oman Mountains and the shores of the Persian Gulf. It measures roughly 2.5 km from its apex to the coast. The braided pattern of the channels is clearly visible. Because the flow of water in a wadi is only sporadic, there is no continuous supply of sediment into the sea. Marine longshore currents rapidly destroy any small delta which may have built up during a flood, and the sediment of this temporary delta is redeposited as marine gravels, sands and clays. Beach sands are often blown inland to form a low dune covering of, in this case, largely carbonate sand.

Many of the large alluvial fans seen around the edge of the Oman Mountains had their origins, like the wadis of the southern edge of the Rub al Khalii, during the Pleistocene (or even earlier?), when at times rainfall must have been considerably greater than now. The present rainfall is rarely heavy enough to account for the great volume of coarse sediment contained in them. The alluvial fan shown in Fig. 16 must have extended much farther from the mountains during the last Pleistocene Glacial period. The post-Glacial sea-level rise has resulted in the sea cutting a low cliff that reaches a height, in some places, of 3 to 4 m above sea level.

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1 See Glossary.
It is reasonable to assume that this fan extends westwards beneath the younger marine sediments of the sea floor in the same way as the conglomerates of the alluvial fan east of Ras al Kaimah extend down beneath the coastal sebkha sediments that are now being deposited (see also Fig. 98, p. 60 and 121).

With a decrease in rainfall, sediment is transported over shorter distances and in smaller amounts, and the wadi tends to cut deeper into its own channels rather than extending the limits of the fan. The dissected remnants of older alluvial fans are quite a common sight around the flanks of the Oman Mountains. Along the northern side of the Oman Mountains, these deeply dissected fans are not necessarily the result only of a decrease in rainfall. Arguments are presented on p. 94 in support of the hypothesis that in this part of the world, Pleistocene "pluvials" coincided with interglacial periods of high sea level. A later lowering of the sea level as ice caps again extended would result in considerable lowering of the base level of the wadis and consequently increased downward erosion. On the other side of the mountains where such changes in base level did not occur, downward erosion is much less apparent.

DEFLECTION OF WADI SEDIMENTS

Over the bulk of the fan seen in Fig. 16, deflation has caused the removal of the exposed wadi sands, silts and clays to leave a boulder-strewn surface (Fig. 17). The boulders on this surface are continually broken down by insolation and other forms of weathering and the fine fragments removed by the wind.

EARLY CEMENTATION OF WADI SEDIMENTS

In areas where the mountains being eroded are rich in limestone, cementation of the wadi gravels can take place almost immediately after deposition. The water evaporates and calcium carbonate precipitates at the contact points of the boulders.
and finer sediment. Rapid cementation is an additional factor aiding the blocking of wadi channels and the formation of a braided distributary pattern and fan development. The cementation of gravels in an alluvial fan of a wadi system is uneven. The upper part of a bed is poorly cemented because the water soaks rapidly away rather than slowly evaporating to dryness. The strongest cementing action appears to be confined to the main wadi channels where the flow of underground water may continue for many months. Cementation takes place at the water–air interface as the level of the water flowing underground slowly falls.

In several instances in the Oman Mountains, the writer has noticed that the calcareous cement holding the limestone boulders together gave the rock such coherence that large pot holes, that were later cut into the conglomerate during a succession of floods, were formed in boulder and cement alike as if the conglomerate were completely homogeneous.

Examples are known of cemented wadi-channel systems resisting erosion by deflation more strongly than the more poorly cemented sediments on either side (differential deflation), with the result that they have later been exhumed and now stand proud of the desert floor (see, for example, Stokes, 1961, fig.15). One such system of probable Pleistocene age occurs to the west of the Wahiba Sands in

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It must not be assumed, however, that rapid cementation of fluvial sediments can only occur in a desert environment. It can take place in any area of hot climate where river waters are rich in dissolved carbonates and the water level falls slowly during the dry season to permit a long period of evaporation.
WADIS AND DESERT FLUVIATEL SEDIMENTS

RELATION BETWEEN WADIS AND DUNE FIELDS

The routes followed by the wadis which intermittently flow with water around Jebel Hafit are strongly controlled by the presence of dune sand, such as the linear dune seen in the left background of Fig.21. The erosion caused by the short period of flowing water is easily compensated for by the intermittent aeolian deposition during the remainder of the year. In the left foreground of Fig.21, there is an ill-defined patch of low sand hills occupying an area where the wadis are not active. Their morphology appears not to be sharply defined, and possibly results from a multi-directional interplay between the locally predominant west winds and those associated with “updraughts” formed around the Jebel in the heat of summer.

The effects of updraughts associated with mountain ranges are perhaps part of the reason why the Wahiba Sands are oriented north-south and not parallel to the main track of the Southwest Monsoon (Enclosure 1).

The major dunes of the Wahiba Sands—so clearly seen in the frontispiece—

Fig.21. Jebel Hafit encircled by interfingering wadi fans, that control the location of the dunes (d) at southern end of Jebel Oman–Abu Dhabi border.

by wadi and dune sediments is well illustrated on a fairly small scale by Jebel Hafit on the Oman–Abu Dhabi border (Fig.21). Jebel Hafit has a visible relief of over 1,000 m, is some 30 km long, about 5 km wide and is being actively eroded. The relief may formerly have been considerably greater, since the lower parts are now buried beneath an unknown thickness of wadi sediments. The encircling, interfingering alluvial fans coalesce with the distal ends of other fans which spread out from the Oman Mountains some 25 km behind the viewer (Enclosure 2). When the wadis are in flood, the combined water-flow divides to go round either end of the Jebel, to become lost in the sand dunes of the eastern edge of the Rub al Khali (Enclosure 2).

Fig.21 (continued).

Oman (Fig.19, 20). The ridges of conglomerate occur at several different levels down to that of the present wadi system. There is a striking similarity between the two patterns, ancient and modern. Even though the Oman Mountains to the north must have been higher at the time these old wadi sediments were deposited, the alluvial fans occupied sites in roughly the same localities as today.

Cementation of wadi sediments soon after deposition may result in an early loss of porosity and permeability. This could have an important bearing on the later availability of the sands to act as a reservoir for hydrocarbon accumulations. On the other hand, aeolian sands that are interbedded with wadi sediments seem to be relatively poorly cemented and often retain good reservoir characteristics (see also p.43). The widespread, poorly cemented aeolian sandstones of the Lower Triassic of northwestern Europe are gas reservoirs in northern Germany. Associated fluvialite sandstones do not appear to be reservoirs, possibly because of early cementation.

RELATION BETWEEN EROSIONAL HIGHLANDS AND WADI SEDIMENTS

Any highland area within a desert is likely to receive a higher rainfall than the adjacent lowland areas and hence to be surrounded by a fairly well-defined belt of wadi sediments. Since its elevation exposes it to the erosive action of the wind, it will not be covered with dune sand. In contrast, the surrounding plains may have a covering of dune sands which will compete with the water-borne wadi sediments for a site of deposition. This pattern of an erosional highland surrounded

1 The pebbles of the exhumed conglomerates have acquired a dark patina of desert varnish in contrast to the light-coloured gravels of the present wadis.
Desert sedimentation occurs in areas of relatively low relief. This may be on a coastal plain as with the wadi sediments seen on Fig.16 and 17, in continental basins formed by crustal downwarping, subsiding fault blocks or areas which have been earlier lowered by a period of intense deflation.

Fig.22 shows the eastern edge of the Djofra Graben in Libya (Enclosure 4). The highland area to the east is one of erosion and wind or wadi transport. At the escarpment edge, each wadi spreads its sediments out into an alluvial fan which coalesces at its margins with the neighbouring fans. The grain size of the sediment within a fan fines away from its apex. In the Djofra Graben this finer sediment is removed by the wind and accumulates in a belt of dune sands occupying a median position in the graben (Fig.23). These dunes effectively prevent the wadi waters from flowing any further, so that after rain, they are flanked by a series of temporary lakes (Fig.49) which remain until the water has either evaporated or soaked into the ground.

THE RESISTANCE OF AEOLIAN SAND TO EROSION BY WATER

Even though flowing water has considerable erosive power, desert wadis often do not flow for a sufficient length of time to allow them to remove all the wind-blown sand which may have filled their channels. This is well illustrated by Fig.24, where on the inner edge of a curve in the wadi, clean, well sorted horizontally laminated sand has resisted much of the scouring action of the flowing water. This sand is thought to have been deposited by the wind. The deeper channels (not seen in the photo) are filled with pebbles but apart from this the only other evidence of the recent presence of water is a thin lamina of clay covering the upper surface of the scoured aeolian sand.

Another example of the resistance shown by aeolian sand to fluvial erosion is given by Fig.25. Here, a clean aeolian sand, showing sharp differences in grain size between the horizontal laminae, had occupied the wadi channel prior to flooding. The flowing water had sculptured the sand into a ripple transverse to the direction of flow. As the water velocity dropped, the only sediment deposited by the flowing water at this locality was a thin covering of pebbles which, only two months later, at the time of the photo, was already being covered by wind-blown sand (cf. lower part of Fig.26).

Stream-flood deposits of the upper flow regime (Simons and Richardson, 1963, 1966) are also horizontally laminated. Such deposits resulting from a major flood in Colorado have been described by McKee et al. (1967). The horizontal strata figured by McKee and his colleagues is often overlain by rippled, foresetted and convolutely laminated sands; scour and fill structures are also present. The

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Footnote:
1 These grains consist of detrital limestone, abraded Foraminifera and other organic calcareous fragments. See also p.127.
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horizontal bedding is considered to be characteristic of rapid flow, whereas climbing
ripple lamination, convolute structures, festoon bedding and scoured surfaces are
thought to result from a decrease in velocity during the waning stages of the flood.
In the areas of the coarser gravels indicating high-energy stream flow,
aeolian sands cannot withstand the erosive power of the flowing water (cf. Fig. 18).
Further out into the plains where stream-flow is lower in energy, the wadi sedi-
ments continually show evidence of the presence of wind-blown sands which have
not been completely removed by the flowing water. Fig. 26 exemplifies this interplay
between wind- and water-transported sands better than the examples given on the
previous figure. It shows the edge of the present channel of Wadi Al Ayn, Oman,
with, from the top downwards:
(a) Deflation lag of wadi pebbles plus the non-deflated remnant of a wadi
conglomerate.
(b) Thin, horizontally-laminated wind-transported sand.
(c) Steeply dipping accretion foresets of pebbly aeolian sand. This sand
overlies with angular contact in almost horizontally laminated, aeolian sand (d).
(d) Near the base of this sand, quite coarse clay flakes and granules have
been incorporated by wind action. It, in turn, overlies a wadi gravel (not seen in
photo).
Unlike stream-flow sediments, the concentration of pebbles in horizon (c) increases upwards towards what is assumed to have been their source. These pebbles are thought to have rolled, or slid down, from a former bank of a wadi in whose protection the aeolian sands were being deposited. All evidence of the former existence of this bank is assumed to have been removed by deflation prior to deposition of the overlying horizontally laminated sand (b).

The aeolian sands of horizon (c) could have been deposited by a wind that blew upstream (towards the left) in the same manner as the aeolian sands that again covered the wadi bed at the time the photo was taken, about two months after the wadi had been in flood (see also Fig.116).

ANCIENT WADI SEDIMENTS

A rather similar sequence from the Lower Permian sediments of Roundham Head, Paignton, Devon, is interpreted as conglomeratic sands deposited in the braided stream of a wadi (Fig.27). The water in the wadi is thought to have scoured a channel through aeolian sands that initially were deposited with horizontal laminae. Flakes and small pebbles up to 1 cm were incorporated in these well-laminated wind-blown sands from nearby water-laid gravel. Upwards the gently dipping foresets of a large wind-formed ripple indicate wind transport from right to left. Imbrication of the pebbles indicates water transport away from the viewer. The wadi conglomerate is much more firmly cemented than the well-laminated aeolian sand. This is possibly because the water in flood, which produced the later scouring, trapped air in the pore spaces of the aeolian sand instead of replacing the air by a potentially cementing fluid. A rather analogous situation has been described by Stieglitz and Inden (1969) in which air-filled vesicles formed in the sediment of a dam spillway following heavy rain.

Fig.28 illustrates another Permian outcrop a few kilometres away in Maidencombe Bay, South Devon, and shows the braided nature of the wadi channels which were filled with angular pebbles set in a sandy matrix. Here, no aeolian sands were recognised. Imbrication of the pebbles suggests that the current flowed from N 320°E at this locality. A simplified geological map of this part of Devon is given in Fig.29. The diversity in transport directions for water-laid sediments shows a marked contrast to the uniform directions deduced for aeolian sediments. Laming (1966) had described these Lower New Red Sandstone wadi sediments of South Devon in some detail. Apart from describing these sedimentary rocks in considerable detail, he also gives a very good “palaeo-view” of the wadi fans as he imagines them to have been during the Permian (Laming, 1966, p.950). Such a scene can be found in many modern deserts.
It is generally accepted that the sandstones and conglomerates of the Upper Old Red Sandstone of Scotland were deposited in the environments of fluvial and lacustrine deposition (Barrell, 1916; Waterston, 1965). Bluck (1967) has recognised a sequence that starts with alluvial fan sedimentation (mud-flow conglomerates presumed to have been deposited by sheet floods), and is followed by braided stream (stream-flow conglomerates and cross-bedded sands; sandy cross-bedded conglomerates with numerous erosion surfaces; mud-cracks and shale pebbles), and river-channel deposits of a floodplain. The development of this sequence is related by Bluck to accumulation in the area of deposition on one side of a fault, combined with downward and backward erosion of the source area on the other side of the fault.

Significant colonisation of the land by plants did not occur before the Devonian Period, and until the end of the Palaeozoic Era was probably confined to near-shore and coastal-plain environments (Schumm, 1968). It seems likely that the presence of sedimentary characteristics in the Old Red Sandstone of Scotland that are also typical of some modern arid and semi-arid regions, is related to lack of vegetation rather than lack of rainfall. Prior to widespread terrestrial vegetation, Schumm (1968, p.1577) estimates that with more rapid erosion and uninhibited transport, the rate of sediment yield in areas of high rainfall could have been four times greater than at present. Rapid run-off after rain would have resulted in sheet floods; just as rapid a fall in the flow of water would have caused the formation of braided streams because of an overload of sediment, sub-aerial exposure of sediment, and the confinement of the reduced volume of flowing water to river channels of the flood plain. Out on the flood plain, evaporation from the surface of wet sediments that are close to the level of the water table is likely to have given rise to calcite cementation of sandstones; finer-grained rocks that have undergone a similar period of evaporation during a temporary dry period might be recognised by the presence of calcareous concretions similar to caliche. Such an interpretation is given for the Upper Old Red Sandstone in the Tweed Basin of southeast Scotland by Waterston (1965) and Smith (1967). A similar interpretation of the formation of caliche by sub-aerial evaporation is given by Nagtegaal (1969) for parts of the Permian continental Peranera Formation of the Central Pyrenees. Nagtegaal (1969, p.163) believes that this formation was deposited in a semi-arid steppe environment that, like the Permo-Triassic rocks of the Vosges described by Millot et al. (1961) had seasonal rainfall. True desert conditions lay further north in Britain and Germany. Nagtegaal (1969) believes that the alluvial fans described by Bruck et al. (1967) from the New Red Sandstone of Raasay and Scalpay on the west coast of Scotland, were deposited in an environment similar to that of the Peranera Formation.

Uncemented wadi sands are an excellent source of dune sand. There is often a high percentage of grains of a size suitable for immediate wind transport; indeed, many wadi sands are the result of fluvial reworking of aeolian sands which had choked the wadi bed before it came into flood. By natural fluvial grading of these sporadically water-transported sediments, the finer sand fractions tend to be the last to be deposited, and many dry out as a result of water-drainage rather than evaporation. This means that they are less likely to be cemented than the underlying coarser sediment, where the water slowly evaporates from stagnant scour channels.

Norris and Norris (1961) and McCoy et al. (1967) believe that the sands
of the Algodones dunes of southern California were derived from the beach sediments of Lake Cahuilla, lying between the dunes and the Salton Sea. Merriam (1969), on the other hand, points to the similarity between the Algodones dunes and the Sonora dunes across the border in Mexico. He relates their content of volcanic fragments, feldspar types, calcite and dolomite grains and abraded Cretaceous Foraminifera to a source in the delta sediments of the Colorado River. Such sediment underlies the dunes and extends westward over hundreds of square miles.

With a continuously flowing river such as the Colorado, additional sediment is deposited over the delta annually, especially during periods of flood; as the flood level subsides, this loose sediment is exposed to the action of the wind.

Fig. 30 (Wadi Dhaid, Trucial Coast, Arabia) shows the wind-formed rippled surface of a wadi bed a few weeks after the wadi had been in flood. The ripple crests are dark because of wind-concentration of heavy minerals. Fig. 31 gives the grain-size distribution of the wadi gravel, which here contained over 7% of heavy minerals. Fig. 32 gives the grain-size distribution of the wind-formed ripples in which the heavy-mineral concentration had increased to over 22%. The high (14) percentage of grains smaller than 62 μ in this sample from an aeolian ripple probably reflects the nearby presence of clay flakes on the surface of the wadi (see also remarks on pp. 56 and 148 on laboratory techniques†). For comparison, Fig. 35 shows the grain-size distribution of sand from a dune bordering the same wadi at a different locality (for sample localities, see Enclosure 2).

† Note that because the standard deviation is calculated at 16 and 84°, and the extreme grain sizes are ignored, the wadi gravel (Fig. 31) would appear to be better sorted than wind-sorted ripples (Fig. 32). This is obviously not the case, as a comparison of the two graphs shows.
Deflation gravel of some higher level of wadi.

Foresetted slightly pebbly wind-transported sand.

Horizontally bedded gravel with top portion deflated away.

Finely laminated wind-blown sand.

Wind from left to right.

Clay flakes, slightly curled.

Fine ripple lamination in wind-transported sand.

Fig. 36. Interbedded wind- and water-transported sediment. Wadi Amayri, Oman.

INTERBEDDED WADI GRAVELS AND AEOLIAN SAND

Also in Wadi Dhaid, but farther downstream, the pebble content of the wadi sediments is reduced to fairly thin beds between wind-and water-transported sands. The accretion foresets in the gravel are up to 30' or more (Fig. 33). Much of the aeolian bedding overlying the gravel has either been destroyed by plant-root burrows or prevented from developing clearly defined laminae as the sand was deposited around plant stems. The grain-size distribution diagrams of the gravel (Fig. 34) and the sands from the dunes flanking the wadi (Fig. 35) are given for comparison (cf. Fig. 44-48).

A somewhat different association is illustrated in Fig. 36. A predominantly wind-transported sequence is overlain by alternations of wind- and water-transported sediment, the highest bed of which is now represented by a deflation lag of pebbles. Since then, flowing water in the wadi has cut a new channel down through the previously deposited sediments. Note that the clay flakes about one third of the way up the section form a more or less continuous band. These flakes have been preserved by the overlying aeolian sand in the position where they formed. The coarse, slightly pebbly, foresetted sand in the upper part of this sequence becomes more pebbly to the left (beyond the photo) with the pebbles concentrated in the upper slopes of the laminae. To the right, the foresets continue at a steep angle and with a sharp angular contact with the underlying water-laid gravel before passing laterally into horizontally laminated sands that can be followed for at least 30 m. As with Fig. 26, the pebbles in this aeolian sand are assumed to have been derived by rolling or sliding from a former wadi bank, all evidence of whose earlier existence having since been removed during a period of intense deflation.

CLAY LAGS IN WADIS AND MUD CRACKS

The presence of clay in water-transported sediments is very common, and the phenomenon of mud cracks has long been recognised as evidence of subaerial desiccation of such clays. In Wadi Dhaid, Trucial Coast (Fig. 37), a deep water-formed scour was cut into the previously deposited wadi and wind-blown sands. On drying out, it was lined with up to 10 cm of clay which was still damp in the deepest part of the channel at the time of photographing. Desiccation caused broad clay polygons to form (Fig. 38). The intervening cracks will probably become infilled with wind-blown sand; in fact whole polygons may be preserved from further destruction by a covering of aeolian sand. Note the sand dunes seen through the trees in the background of Fig. 37. Small rain-drop pits can be seen on the surface of the clay (Fig. 38). The deeper impressions seen in the photo were made by the paws of a dog while the clay was still soft. In spite of their thickness, the polygons are slightly curved concave-upwards.

Rather larger desiccation polygons with fissures up to 5 m or more in depth are described by Neal et al. (1968), but they are from clay playas in the southwest United States. They interpret them as having formed in thick clays as the result of desiccation in the zone between the hard surface crust of the playa and a water table that had been lowered possibly as a result of man's activities.

PERMO-TRIASSIC MUD CRACKS

Mud polygons with an infilling of sand are known from the Permo-Triassic rocks of northwestern Europe. In Devon, England, examples have been seen which were of the same order of size as the recent case described above. Unfortunately, no good photographs are available. The example shown here (Fig. 39) is from the Volpriehausen Sandstone of the Triassic Buntsandstein of north Germany, where water-laid sands and clays are associated with aeolian sands.

In southern Germany, although sandstones of doubtful aeolian origin are occasionally found interbedded with water-laid sedimentary rocks of the Buntsandstein, their occurrence is not thought to be widespread. Instead, cross-bedded sandstones and conglomerates showing no signs of deflation suggest rivers that flowed throughout the year. The presence of numerous clay-pebble conglomerates and mud-crack horizons, however, indicate considerable fluctuation in the water level of the rivers with sub-aerial exposure of the sediments. Presumably there was a marked dry season in the source area of these rivers further to the south.
Fig. 37. Clay polygons on the surface of a water-scoured hollow, Wadi Dhaid, Trucial Coast.

Fig. 38. Close-up of 10 cm thick clay polygons, Wadi Dhaid, Trucial Coast.

Fig. 39. Clay polygons and sand infill. Volpriehausen Sandstone, Buntsandstein, northern Germany.

Fig. 40. Mudcracks and superimposed flute casts in "Plattensandstein" of the Hunsrücker. East of Eberbach on the Neckar, West Germany.
Near Eberbach, on the banks of the river Neckar, there occurs a combination of mud-cracks with superimposed flute casts in the Plattensandstein. The latter were presumably cut during a sudden flood of water in the Triassic river not long after the mud-cracks had developed. The clay polygons were probably still damp and fairly soft when the flood took place (Fig. 40).

**PRESERVATION OF SUN-CURLED CLAY FLAKES BY AEOLIAN SAND**

The mode of preservation of sun-dried clay flakes by wind-blown sand is well illustrated by Fig. 41, taken in Wadi Amayri, Oman, about two months after the wadi had been in flood. The clay may have been deposited as a continuous sheet covering a large scour hollow such as is seen in Fig. 37. In other instances, clay has been seen as a thin drape covering ripples that were formed in water. On drying, the clay cracks along the crests of the ripples and curls up, the degree of curling apparently being related to the thickness of the clay, with the thinnest clays curling most.

Unless the clay is very thin and fragile, it is capable of withstanding winds that are strong enough to transport quite large sand grains. This sand not only fills the hollows formed by the curled clay flakes, but also fills the spaces between the clay and the underlying water-transported sand. Such clay flakes, when incorporated into a dune sand, are likely to give a grain-size distribution curve and sorting coefficient similar to that shown by the dashed line in Fig. 45.¹

Had another wadi flood followed the formation of the curled clay flakes seen in Fig. 41, instead of aeolian sand transport, then these clay flakes would almost certainly have been removed and destroyed by the action of the water. Water-laid pebbles of clay might have been preserved downstream provided that the original clay was thick and hard enough to withstand this form of transport.² If not, then the clay would have been redeposited elsewhere as part of another clay drape.

¹ For further comments on this particular grain-size analysis, see p. 56.
² Although comparatively delicate curled clay flakes are normally only preserved in the place where the flakes formed by a covering of wind-blown sand, clay pebbles are a fairly common feature in the river sediments of areas that have a marked seasonal rainfall.
Fig. 43. Interbedded wind- and water-transported sediment. From lacquer peel. Locality 34, Wadi Dhaid, Trucial Coast.

a = a few rootlet burrows; b = clay flakes and pebbles; c = clay-filled troughs of ripples, moulds of plant roots, climbing ripples; d = sand-filled troughs of clay-covered ripples; e = clay-covered climbing ripples, clay flakes and burrows, moulds of plant roots; f = laminae disturbed by moulds at plant roots; g = finely laminated sand with curled clay flakes; h = laminae disturbed by moulds of plant roots (sand possibly deposited around living plants—drifting sand)—where laminae will not be well developed; i = laminated sands covered by clay film, moulds of plant roots; j = laminae disturbed by moulds of plant roots.

Similar features can be seen in the Permian sedimentary rocks of Devon, England (Fig. 42). At the time of deposition, the upper part of the sequence shown was pre-

Fig. 44. Grain-size distribution curve and sorting coefficient of water-transported sediment of sample GL 479 seen in Fig. 43.

Fig. 45. Grain-size distribution curve and sorting coefficient of water-transported sediment of sample GL 480 seen in Fig. 43.

Fig. 46. Grain-size distribution curve and sorting coefficient of wind-transported sediment of sample GL 478 seen in Fig. 43.

Fig. 47. Grain-size distribution curve and sorting coefficient of wind-transported sediment of sample GL 481 seen in Fig. 43.

Fig. 48. Grain-size distribution curve and sorting coefficient of dune sand of sample GL 483 taken from nearby the same locality as Fig. 43.

dominantly sand with clay pebbles (water-borne?) and curled and cracked clay drapes (sun-dried and preserved by wind-blown sand?). The lower part of the sequence contains a higher proportion of clay with many thicker clay drapes. The clays have obviously for the most part been exposed to sub-aerial drying and are curled with the concave surface upwards. Wind-blown sand, incorporated from above, now occupies the space beneath the curled-up edges. In other parts of the exposure, especially with thicker clays, the waterlogged sand beneath may have flowed-up between the clay polygons as a wet "slurry" to form a "sand dyke". This possibility will be explained in greater detail on p. 64.

INTERPLAY BETWEEN WIND AND WATER IN WADI SEDIMENTS

Fig. 43 is a photo of a lacquer-peel section made on the banks of Wadi Dhaid, Trucial Coast. The close relationship between what are thought to be alternations of
wind- and water-transported sediment is apparent. Features such as climbing ripples—which here suggest overloading of the water with sediment as the water velocity in the wadi falls (see Allen, 1963; McKee, 1966a)—clay lags in ripple hollows, clay flakes and pebbles, all point to sediment transport by water, with the clay flakes and pebbles indicating sub-aerial exposure. The wind-transported zones are indicated by the clearly defined horizontal or low-angle laminae which generally show better sorting of the grains in the individual laminae than do those that are water-laid. This criterion does not apply to the water-formed climbing ripples, however, where the particular conditions of deposition have resulted in well-graded sets of laminae in which the sorting (Fig.44) is better than that found in the associated wind-blown sand (Fig.46, 47 and 48). All the graphs shown are for bulk samples comprising material from several laminae.

Fig.45 gives a more typical grain-size distribution and sorting coefficient for a wadi sediment. The continuous line on this graph shows some 16% of clay with a diameter less than 2 μ. As sampled, this clay was in the form of clay flakes and pebbles. In other words, although the sorting coefficient is perhaps indicative of its origin as a water-transported sediment, if the laboratory techniques used had not involved the use of water which resulted in breaking down the larger clay fragments, the grain-size distribution curve would have looked more like the dashed line showing a higher percentage of coarse grains.

Fig.46 shows aeolian sorting with a coefficient of around 0.45, with slightly poorer sorting in Fig.47. That these results are caused by the close proximity and reworking of wadi sediments is suggested by Fig.48. This is of a bulk sample taken from a dune sand to windward of the wadi. It is, therefore, largely unaffected by admixture with water-transported sediment. As will be shown in the section on aeolian sands, bulk samples can give very misleading results on the sorting ability of wind. Fine-grained laminae will probably give a sorting coefficient in the range of 0.25-0.3.

Chapter 5

DESSERT LAKES AND INLAND SEBKHAS

SALINITY OF LAKES IN BASINS OF INLAND DRAINAGE

As was pointed out on p.8, many basins of inland drainage are also the sites of deserts.

In regions of active mountain building, new ranges of hills may be raised across the path of previous drainage. If these hills are elevated more rapidly than the rate at which the rivers can cut down through them, then a basin of inland drainage is formed. When there is sufficient rainfall, a temporary lake will extend within the basin until such time as the water level reaches the lowest point in the barrier; a new drainage channel is then cut (see, for instance, Glennie and Ziegler, 1964), the lake reduces in size and any temporary increase in the salt concentration is reversed as the water again flows to the sea.

When the rate of evaporation equals or exceeds the supply of water to the basin, its centre of drainage will never acquire sufficient water to flow over the top of the newly formed barrier. The water level in the lake will fluctuate in response to rainfall on the surrounding hills. During dry periods, the lake shrinks in area leaving a thin salt crust over the exposed lake bed. In Iran, the salty and often seasonally marshy nature of such lakes is expressed in some of their names (Daryacheh-ye-Namak—lake of salt; Kavir-e-Namak—salt desert; Batlaq-e-Gav Khunt—marsh of Gay Khunt; Hamoun-e-Jaz Murian—ephemeral lake of Jaz Murian). With increasing accumulation of salts left behind by evaporation, vegetation in the areas surrounding the lakes becomes more sparse and a desert develops, such as around Jaz Murian in southeast Iran or Hamoun-i-Helmond on the Iran–Afghanistan border.

In some regions of Iran, the natural concentration of salts in the river water is increased by the addition of salt from exposed salt plugs. This is the case in the Great Kavir of north central Iran, where Eocene salts are involved. Cambrian salts are added to the natural content of lakes Neyriz and Bakhtegan, east of Shiraz (see also space photo in Wobber, 1967, plate I and fig.19) thus rendering the lake water increasingly unsuitable for irrigation. If it were not for the high concentration of salt in these lakes, the fertile region east of Shiraz would be much more widespread. But it is not only in basins of inland drainage that the Cambrian salt prevents the full use of water for irrigation. Many rivers that flow through the southeast Zagros Mountains to the sea acquire much of their salt content from the
Tertiary Fars Formation over which they partly flow, but they are also contaminated with dissolved Cambrian salt from salt plugs. As a result, the valleys through which these rivers flow are commonly vegetated only along their upper flanks.¹

WADI-FLOOD-WATER ORIGIN OF SOME DESERT LAKES

In an earlier chapter, it was mentioned that temporary lakes may form when wadi flood waters fill a desert depression and further progress is barred by sand dunes. This has happened in the Djofra Graben, Libya, where sand dunes, occupying a median location in the graben, prevent the wadi waters from flowing beyond them (Fig. 22, 23, and Enclosure 4). In the initial stages of the wadi flood, fluviatile erosion will occur. Later, however, additional influx of water to this area will result only in a rise in water level, so that the dune pattern is preserved as it slowly becomes drowned in the temporary lake (Fig. 49). Since water velocity in these desert lakes becomes virtually zero, a graded bed, fining up to the clay which was held in suspension, will be deposited. Clay drapes cover the lake floor. As with wadi clays, they will break into polygons or crack and curl into flakes on drying out. They will be preserved if protected by a covering of aeolian sand. The writer has only identified one clay drape in ancient aeolian rocks which was thought to have such an origin. It was found at the lowest point of a set of dune laminae—i.e., an interdune area—in Permian dune sands at Gatelawbridge Quarry, Thornhill, southwest Scotland. The lack of mud-cracking suggested that this clay was rapidly covered by dune sand. The low boron content (89 p.p.m.)² of this Permian clay is similar to that found in recent non-saline desert wadis.

The calcareous shales that cover parts of the Late Stampian sandstones of the Paris Basin are compared by ALIMEN (1936, pp. 187-218) to the clays deposited in the interdune “bahrs” found around Lake Chad on the southern edge of the Sahara. A bahr is an interdune area which in that region becomes flooded during seasonal rise in the level of the water in Lake Chad. These Stampian sandstones form ridges of very constant orientation and regular spacing. They have a relief of some 15 m and are considered to be ancient sand dunes. As is the case with the bahrs of Lake Chad, the calcareous shales cover only the interdune areas and lower slopes of the dunes.

GROUND-WATER ORIGIN OF SOME DESERT LAKES

Fig. 50 shows a desert lake occupying an interdune area in the Ubari Sand Sea, Libya. There is no surface flow of water into the lake, and yet, judging from the concentration of date palms around its shores, it has existed for many years. It is thought that the lake exists because the water table is higher than the surface of the interdune area. Further away (but visible in Fig. 50) are some salt pans that presumably indicate that the ground-water level there is just too low to sustain a permanent lake. Clays are not likely to be found in these lakes unless deposited from occasional dust storms, but the salts precipitated from the evaporating ground water could be important in providing local cementation of the sands with resulting low porosity and the formation of permeability barriers.

¹ For further discussion on the salt domes and salt deserts of Iran, see STÖCKLIN (1968).
INLAND SEBHKAS

The term "sebkha" (see Glossary) is used by the Arabs of North Africa and Arabia to denote flat areas of clay, silt and sand which are often encrusted with salt. Sebkhas fitting the above definition can be found in both inland continental and coastal environments, but since the reasons for their existence and the criteria for their recognition are different in the two environments, it is necessary to add the words "inland" or "coastal" to the word sebkha (see also SHEARMAN, 1966).

HOLM (1960, p.1378) wishes to restrict the use of the word sebkha (sebkah) to coastal regions, for he says that in parts of Arabia the salt-covered flats of the interior are known by another Arabic word "mamlahah". POWERS et al. (1966), however, use "sebkah" in the sense described above and say that "mamlahah" refers to sebkhas that have been excavated for salt. Colleagues who have worked in the interior of Algeria, however, were well versed in the meaning of sebkha (sebkra) in its "inland" sense, but were surprised to learn that in Arabia it is also used in coastal regions. It is used to designate both inland and coastal salt-covered flats on published maps of Arabia (e.g., U.S.G.S. Miscellaneous Geological Investigations Map 1-270A), and is used in this sense by the writer with the addition of the appropriate prefix "inland" or "coastal", because of the environmental situation implied by its use.

In North America, the words "playa" and "salina" have both been applied to sebkha-like areas in the interior of a desert. HOLM (1960, p.1379) states that "playa" is synonymous with "mamlahah" (inland sebkha). VON ENGELN (1942, p.412), on the other hand, says that in parts of Arabia the salt-covered flats of the interior are known by another Arabic word "mamlahah". POWERS et al. (1966), however, use "sebkah" in the sense described above and say that "mamlahah" refers to sebkhas that have been excavated for salt. Colleagues who have worked in the interior of Algeria, however, were well versed in the meaning of sebkha (sebkra) in its "inland" sense, but were surprised to learn that in Arabia it is also used in coastal regions. It is used to designate both inland and coastal salt-covered flats on published maps of Arabia (e.g., U.S.G.S. Miscellaneous Geological Investigations Map 1-270A), and is used in this sense by the writer with the addition of the appropriate prefix "inland" or "coastal", because of the environmental situation implied by its use.

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Inland sebkhas develop where water flowing in wadis intermittently floods low-lying depressions to leave behind damp, salt-encrusted sediments. They are also found in depressions where, for one reason or another, the water table reaches, or almost reaches, the surface. The water table may reach the surface in the unconsolidated sands of a dune-filled depression as seen in Fig.50. Another way was explained to the writer by A.A.E.A. Coffinier (personal communication, 1966).

In a stable, clastic, sedimentary basin, water flows from the periphery of the basin towards its centre and returns to the surface mainly by percolating through the overlying shales and clays. Though these shales and clays are only slightly permeable to brines, their horizontal section is so large that considerable amounts of water may move upwards. In the aquifer itself, because the overlying shales and clays are semi-permeable, the water increases in salinity along its flow path. This relatively simple process was suggested by DE SITTER (1947), and has been shown to be responsible for subsurface water-salinity patterns in various areas by


Coffinier also mentioned the coincidence of inland sebkhas overlying an anticline in the Great Eastern Erg of Algeria. These interdune sebkhas occupy sites that are topographically higher than other interdune areas in the neighbourhood, thus implying that the water and salt both came from an underground source. It is assumed that the aquifer that supplies the salt and water comes sufficiently near to the surface in the anticline for its water to percolate upwards to the surface and so create sebkha conditions. In some cases, this upward movement of artesian water may be aided by small crestal faults.

A similar relationship between inland sebkhas and anticlines has been noticed on aerial photographs by E. Th. van der Bent (personal communication, 1967) on the north flank of the Ighma anticline, Algeria. In Tunisia, Sebkha el Kourzia is described by COQUE and JAUZEIN (1967, p.241) as occupying a hollow that is being deepened by wind erosion and which also overlies an anticline.

In inland sebkhas, the salt crust forms as the result of concentration of salts caused by evaporation of the water. Gypsum crystals are common in the sediments of inland sebkhas. Algae are known, but the algal mats so commonly associated with coastal sebkhas have not been recognised. At present, inland sebkhas cover a vastly greater area than do coastal sebkhas (see Enclosure 1).

A coastal sebkha, on the other hand, is characterised by marine flooding and evaporitic conditions. It is a diagenetic environment whose sediments are of continental and adjacent marine origin.

In the following sections, various features associated with inland sebkhas are described.

THE INLAND SEBKH A UMM AS SAMIM

Fig.51 is an aerial photo of a part of the surface of the giant Oman inland sebkha called Umm As Samim (mother of poisons) (see also Enclosures 1 and 3). The foreground is covered with salt polygons which appear large, even at a flying height of 500 ft. above the surface. The force of crystallisation of the salt probably causes the roughly polygonal shapes and result in their being surrounded by walls of salt up to 4 ft. high (as may be the case with the large polygons in the aerial photo).1

Fig.52 shows the surface of a road across the salt crust of the Umm As Samim about one year after it had been bulldozed smooth. Already the salt (mostly halite) has grown up about 5 cm at the polygon edges. Beyond the bulldozed area, the salt-polygon walls are between 10 and 20 cm high.

The salt has incorporated in it considerable amounts of wind-blown sand.

1 See also BORÉEK (1959, plates IV and VI) for similar salt polygons from Iranian inland sebkhas.
Under normal circumstances, the salt is either deflated away or dissolved by rain or wadi-flood water, thus permitting the sand grains to sink into the sediment beneath. In desert areas, sodium chloride in solution in ground water is normally brought to the surface by capillary flow that is balanced by evaporation and formation of a halite crust. This means that when the halite is covered by a permeable sediment such as sand, the upward flow of less saline water from below causes solution of the salt and its recrystallisation at the surface. Water flows intermittently from wadis into the eastern end of the Umm as Samim, however, and carries considerable quantities of clay into the area both above and below the existing salt. This can result in salt lenses being permanently trapped in the impervious clay (Fig. 53).

The salt crust is up to 30 cm or more thick. After a wet season, it may have between 6 and 8 m of salt water between it and the underlying sediment (H. van Deventer, personal communication, 1964). When visited by the writer, no free
water existed; the salt was in contact with the underlying gypsum-rich, rather structureless sediment.

When the salt is dry, sand dunes migrate across its surface. It very seldom rains over the Umm as Samim, but when it does, the surface may be pitted with solution hollows which become the starting point for a new series of salt polygons. The Umm as Samim exists essentially because its surface is at, or below, a fluctuating ground-water level. Whether the original basin shape was caused by deflation or structural warping is not known. The salts are probably derived from outcropping Tertiary strata by solution in rare rain water or water-filled wadis, although there is the possibility that the area also represents a relict arm of the sea left behind as the result of relative sea-level changes; its surface is less than 70 m above the present sea level.

SOME SEDIMENTARY STRUCTURES IN INLAND SEBHAS

The temporary desert lake shown in Fig.49 is caused by water flowing down a wadi and into a continental basin in which the water is prevented from flowing farther by a barrier of dune sand. Similar conditions can give rise to inland sebkhas. Salt-rich water slowly percolates into the ground and evaporates leaving a surface crust of salt. Salt polygons may or may not grow, depending upon the relative concentration of salt. Fig.54 shows the surface of such an inland sebkha on the northern edge of the Ubari Sand Sea, Libya, where a heavy truck has broken through the thin hard crust and sunk into the soft wet sediment below. Here, salt polygons are not well formed, because each year the local people remove the salt crust in order to plant crops in the relatively salt-free sediment beneath. However, the clayey sands beneath the salt crust are rich in gypsum crystals (the shiny spots in Fig.55). The wavy lamination is typical of sedimentation in such sebkhas. It is not certain whether the lamination results from deposition from flowing water, or from deformation of horizontal laminae by growing gypsum crystals. Another, and perhaps more likely possibility, is that they grow by the adhesion of wind-blown sand and silt to the damp sebkha surface. Capillary flow of water between the grains permits adhesion of additional laminae to the surface of the structure (see Fig. 61, 62).

SAND DYKES IN INLAND SEBHAS

In the environment of the inland sebkha one can also find the “sand dykes” encountered by Oomkens (1966) at the northern edge of the Ubari Sand Sea. Water-transported clay on the sebkha surface broke into polygons as it dried out; rifts between the polygons either became filled with wind-blown sand, as would be the
case in Fig.37, 38, or were injected from below with water-saturated sand which flowed out onto the edges of the polygons (Fig.56). This latter phenomenon seemed to occur where the clay rested on beds consisting mainly of aeolian sand.

In Fig.57, a slight up-arching of the sand laminae can be seen at the base of the dyke. The sand at the top of the dyke flowed over the edge of the dark clay; its pattern is outlined by the up-arched crust of salt (Fig.56, 57).

Oomkens gives alternative explanations for the mechanics of formation of these dykes as follows:

1. When a hard, impermeable, clayey surface bed is fractured as a result of shrinkage, underlying sandy slurry might rise through the cracks and flow over the clay because of a difference in density between the dry and watery sediments.

2. The weight of isolated dune bodies that migrate and locally load an impermeable hard crust will increase the pore pressure in the underlying mobile sediments, and thus facilitate upward injection through cracks.

3. A slow downslope flow of soft sediment below an impermeable rigid crust, for instance towards the centre of a depression, will increase the pressure in the flowing sediments.

4. Examples of mud and sand welling up from fissures during earthquakes have been observed (Horns, 1907). In such cases, the earthquakes, in addition to rupturing the thin hard surface bed, also increased the pressure of the formation water. Most ancient sand dykes so far described have been explained in this manner.

5. Upwelling of sand from (man-made) fissures under the influence of large variations in atmospheric pressure during storms have been observed (Buhs, 1892).

Oomkens (1966) suggests that the first explanation is the most likely, but it is probable that explanation (3) also plays an important part.

Siltstone polygons that are bounded by walls of sandstone are described by Harshbarger et al. (1957) and Tomkins (1965) from the Jurassic Carmel Formation of Utah. Tomkins believes that the mud polygons formed as the result of desiccation and were then infilled from above by aeolian sand. On the other hand, Harshbarger and his colleagues noted that the sandstone infill appeared to have flowed, and they postulated that the mudcracks may have formed under subaqueous conditions and were then filled by the upward flow of underlying quicksand. Although fossils were lacking to support their ideas, they thought that sedimentary structures within the Carmel Formation suggested an estuarine or lagoonal environment. However, associated regional features suggestive of an arid climate (dune sands, etc.) leave open the possibility that although some mud cracks may have been infilled from above by wind-blown sand, others may have been filled by injection of a quicksand from below in the manner inferred for the inland sebkhas of Libya.

In the Permian sediments of Devon, England, similar sand dykes were seen at several localities. One has already been referred to in Fig.42. Fig.58 shows...
another from Roundham Head, Paignton, in an environment where clay-pebble conglomerates with a sand matrix and possible aeolian sand are interbedded with polygonally cracked clay. The top of the sand dyke has probably been removed by deflation prior to the deposition of the succeeding aeolian sand. Oomsrens (1966) figures an excellent hand specimen of a sand dyke of similar age from Germany. Lamming (1966, p.925) also refers to Permian sand dykes from south Devon and figures examples associated with conglomerates. These, however, are likely to have required some more violent mechanism, such as the shock-wave of an earthquake, to cause their formation. Peterson (1966) describes sandstone dykes that are up to 8 ft. wide and several miles long. He relates their formation to strike-slip faulting in the underlying basement rocks. Even bigger sandstone dykes up to 300 ft. wide and 1,000 ft. in vertical extent penetrate Precambrian crystalline rocks in Colorado (Harms, 1965). Their mode of origin has, in common with those described by Oomsrens (1966) from an inland sebkha, the necessity for tension in the intruded medium, and the presence of a sandy slurry that is capable of flowing into the tension gap.

SEDIMENTATION VERSUS DEFLATION IN INLAND SEBKHAS

It has already been mentioned in connection with the deflation hollows of Qattara, Fahud and Al Fugaha, that the depth to which deflation can penetrate is limited by the level of the water table. The presence of moisture inhibits deflation and encourages sedimentation. Deposition occurs in two main ways, by water-filled wadis bringing the sediment into the inland sebkha, and by the adhesion of wind-blown sediment to the damp surface of the sebkha. As soon as sedimentation brings the sebkha surface above the limit of permanent capillary flow from the water table, then deflation can become active and the surface is again lowered until the inhibiting effect of the ground water is met. Permanent sedimentation will occur when the rate of sediment supply exceeds the rate of deflation or when a rising water table (such as in a subsiding continental basin) permits the addition of further sediment by adhesion (see also p.71).

HORIZONTAL BEDDING IN INLAND SEBKHAS

Fig.59 shows a pit dug into the horizontally bedded sediments covering the floor of a wind-deflated basin of inland drainage in Libya similar to that at Al Fugaha (Fig.13). Erosion by deflation is assumed to have ceased once the ground-water level had been reached (if the impervious cap of an aquifer was breached, this may have produced a temporary lake). Thin horizontal beds of brown and green clays and streaks of red sand alternate with green gypsiferous clays and white bedded gypsum. They are overlain by about 25 cm of reddish-brown, non-laminated gypsiferous sandy clay. The former are thought to have been deposited during an early period as a temporary lake; the latter, in which no laminae were visible, are thought to have been deposited by adhesion of wind-blown sand and dust to the damp salt-encrusted surface. No algal mats were recognised. The salt crust had a very uneven surface and possibly played a part in preventing the formation of laminae in the sandy clay.

NEAR-COAST INLAND SEBKHAS

It will be shown later that wadis which once drained into the sea can have their exits blocked by longshore bars and coastal dune barriers. Under these circumstances, if there is an insufficient flow of wadi water to break through this barrier, tempo-
ary lakes and inland sebkhas, having no marine influence, may form near the coast (see Fig.62, 101).

Fig.60 shows laminated Quaternary gypsiferous clays and sands known to be pre-Roman as the foundation of a Roman dam cut through the clays. Nearer to the sea they intertongue with gypsum-cemented dunes. It is assumed that inland sebkha conditions developed in the lower reaches of Wadi Tareqat behind a coastal dune barrier in Late Quaternary time. By the time of the Roman occupation of this part of Libya, the wadi channel was again in connection with the sea. (In the photograph the holes in the clays are artefacts caused by the removal of clusters of gypsum crystals when an attempt was made to smooth the cut surface with a knife.)

On the Gulf of Oman coast of Oman (Batinah Coast), a series of near-coast sebkhas occurs just above sea level behind the protection of the coastal dune barrier. Although wadis occasionally drain into these sebkhas, the salt water is usually, in this case, supplied by percolation of sea water through the coastal dunes. The surface of the sebkha is dampened twice daily at high tide when water rises by capillary action and makes the salt surface of the road that crosses them exceedingly treacherous. At low tide, the salt dries out to make a good firm road surface. Although supplied with sea water, these sebkhas are cut off from direct access to the sea. No marine fauna lives in the sediment, but the tests of dead marine Foraminifera become incorporated from the nearby beach and coastal dunes by the action of onshore winds.

ADHESION RIPPLES

Sedimentation by adhesion of wind-blown sand to a damp surface has already been referred to (Fig.55). Adhesion ripples were first described by REINECK (1955)

1 See Glossary.
from the northern German coastal flats. Where the ground surface is so damp that deflation is inhibited, as in many sebkha environments, both coastal and inland, sedimentation by adhesion can become important.

Sebkha Matti, Trucial Coast (see Enclosure 1), is a large plain some 120 km long and 60 km wide. It slopes gently northwards to sea level from an altitude of 50 m or more at the southern end. Most of the area is at ground-water level. Dunes and outcrops above this level are subject to deflation, and sediment is carried south by the strong shamar winds. Towards the south, sand dunes begin to accumulate. At the windward end of the dune fields large ripples begin to form on the damp surface by adhesion (Fig. 61), and so form the base for further dune accumulation. Still farther north, other adhesion ripples form the sites of temporary sand sedimentation. A balance between sedimentation and deflation probably coincides with the upper limit of capillary action, which is itself affected by seasonal changes in the level of the water table. Adhesion ripples will be preserved when overlain by a permanent accumulation of dune sand or fluvial sediment deposited from non-scouring water (e.g., temporary lake as in Fig. 49). The base of the shallow pit seen in Fig. 61 coincides with the start of gypsum-cemented sand—itself with a ripple-like surface. Surface relief, attributed to adhesion ripples, has been seen in the order of 30-40 cm.

West of Tripoli, Libya, inland sebkha conditions occur similar to those deduced for Fig. 60. A wadi had its exit to the sea blocked-off by the formation of a longshore bar over which a coastal dune developed. Fig. 62 shows the sediments seen in a pit dug into the sebkha that formed behind the coastal bar and dune. The dark bands consist of brown clay brought into the sebkha (annually?) by the wadi. The white laminae consist of both water-borne and wind-blown carbonate sand derived from the nearby coastal dunes (themselves formed from carbonate beach sand). The upper white surface consists of 2 cm of wind-blown carbonate sand in the form of fine adhesion ripples (the adhesion warts of Reineck, 1955). 30-40 m nearer to the coastal dunes, adhesion ripples were 5 cm or more in height. Many of the diapir-like sedimentary structures seen in the pit are thought to have been formed as large adhesion ripples.

Ripples, similar to those seen in Fig. 55 and 62, and ascribed to a wind-blown origin by adhesion, have been recognised in Permian sandstones from North Sea cores.

OASES

By way of a slight digression from the geological aspects of deserts, this is, perhaps, the place to discuss one of its modern features, the oasis. It is not necessarily connected with either desert erosion or sedimentation, but it is intimately connected with the presence of water and vegetation.
An oasis can be defined as an area in the midst of a desert which is made fertile by the presence of water (Moore, 1949). Its existence is dependent upon a continuous supply of water flowing at or near the surface. Apart from the rare occurrence of more or less permanent surface water in the form of a desert lake, spring, or restricted portion of a wadi, the water usually has to be brought to the surface from a well. Most oases are, therefore, artificial. The former existence of a natural oasis might be recognised in Cenozoic desert sediments by the widespread presence of dikaka (see p.113) provided that the oasis was later covered by aeolian sand.

In Iran, and parts of Arabia and North Africa, water is commonly carried to an oasis from its mountain source in tunnels dug into the wadi sediments. In Iran, where the system is thought to have originated, each tunnel is known as a qanat; in Oman and Libya each is referred to as a falaj. They form a very important means of providing a local permanent supply of water to an area that would otherwise have none within reach of the surface.

Qanats became widely developed in Iran during the 6th century A.D. and may have been introduced into southern Arabia at about the same time. Their importance today as a means of providing fresh water to an otherwise arid area can be imagined from the numbers seen from the air around the ancient Iranian city of Isfahan which is only about 50 miles from the edge of the Great Kavir. The Buraimi oasis at the northern end of Jebel Haft (Enclosure 2) is irrigated, and gets its supply of drinking water from such a source.

There is also evidence of the former use of qanats (falajes) on the southern edge of the Ubari Sand Sea in northern Fezzan, and in Morocco, where they are known as foggara. The introduction of qanats to North Africa probably followed the advent of Islam and the spread of the Arab Empire, with Persian slave labour and skill utilised for their construction. The Libyan qanats appear now to have fallen into disrepair and many of the oases that they used to supply with water have ceased to exist. In Oman, some of the tribes still maintain their qanats (falajes) and even occasionally build new ones.

Some examples of natural oases, together with the probable geological reasons for their presence, are given below:

(a) Deflation hollows: As already explained on p.26, the depth to which deflation can cause removal of sediment from a hollow is limited by the water table. Provided that the ground water so exposed is not too saline, an oasis such as that at Al Fugaha (Fig.13) can exist. Famous Egyptian oases of this type such as that at Kharga, derive their water from the underlying Nubian Sandstone.

(b) Wadis: Permanently flowing water often exists in gravels below the surface of large wadis. In areas of considerable annual fluctuation in the level of the water table, wells have only seasonal value and permanent irrigation is impossible. Within and flanking the Oman Mountains, the presence of cemented wadi gravels coupled with the cutting of new channels and the redistribution of uncemented gravel sometimes leaves natural pools of flowing water. These are invariably the sites of oases, provided that there is sufficient flat ground nearby on which to grow palms or other plants. If not, then a falaj, or a surface irrigation canal, is constructed to carry the water downstream to the nearest place that can be cultivated—often a gravel and sand terrace. In other cases, unevenly cemented wadi gravels may act as a natural barrier that dams back the water to form a natural shallow reservoir.

(c) Ground water associated with wadis is less saline, and therefore less dense, than the salt water found in coastal sediments. When fresh water within the gravels of an alluvial fan flows towards a coastal sebkha, it may be forced to the surface by the denser ground water of the sebkha. This appears to have happened where a large oasis fringes the lower perimeter of a wadi fan between the Oman Mountains and the coastal sebkha at Ras al Khaimah (Fig.98, and Enclosure 2). On the other hand, such natural oases do not normally occur where ground water associated with a wadi flows towards an inland sebkha such as the Umm as Samim. This, presumably, is because the ground water is already too saline by the time it reaches the Umm as Samim to support cultivated plants.

A high proportion of the water that flows in a wadi eventually seeps into the ground. With modern drilling and pumping methods, much of this water can be utilised for agriculture, and so the creation of further oases.
Chapter 6

AEOLIAN SANDS

Some important aspects of the mechanics of transporting sand by the medium of the wind were mentioned briefly in chapter 1. Unsupported criteria for the identification of wind-blown sand were given in chapter 2, and in the succeeding chapters the importance of wind as a deflating and transporting agent and the almost universal presence of aeolian sand in a wide variety of sedimentary environments were made obvious. We must now consider some of the factors that control the accumulation of aeolian sand into drifts and ripples, individual sand dunes and sand seas with a continuous sand cover. We must also provide more evidence for the criteria by which aeolian sands may be recognised.

SAND RIDGES

When a source of sediment has a very wide range in grain size, as in a deflation plain, the finer grains will be blown away at a certain wind velocity leaving a lag of coarser grains. As the wind velocity increases, so coarser and coarser grains are removed until finally even the coarsest may move either by saltation or surface creep. It was in this way, by a combination of deflation and surface creep, that the dark ridges, seen below the dune in Fig. 63, grew up. Bagnold (1941, p. 149) states that the wavelength of such ridges may increase in time, as the surface grading grows coarser with removal of the finer sand. This ridge has formed on an inter-dune deflation plain which, as can be seen from the foreground, has a cover of pebbles left behind as a lag during earlier deflation. The lag is underlain by weathered outcrop containing a wide variety of grain sizes. Beneath the pebbles covering the ridge is similar weathered outcrop. The ridges—there are several of them—are transverse to the prevailing wind direction. The ridges are not a simple product of deflation, but rather the result of a balance between deflation and sedimentation. Surface creep carries the largest grains up the windward slope to occupy the exposed crestal position. These large grains need to be in the order of 3 to 7 times the diameter of those in saltation according to Bagnold (1956, p. 141). As long as the wind is not so strong that it dislodges the crestal grains, and as long as deflation is predominant over sedimentation, the ridge can continue to “grow” in size, partly by surface accretion of coarse grains and partly by deflation of the inter-ridge hollows. The average size of the grains covering this ridge is about 1 cm.
Fig. 64 illustrates the grain-size distribution in a section cut through a large sand ridge about 50 cm high. It has an accretion covering of coarse grains of up to 5 mm diameter, which move by surface creep under the impact of smaller grains in saltation. Note how the grains in the laminae become progressively finer to the right, representing deposition on the then lee slope of the ridge. The sharp crest line is better seen in the continuation of the ridge into the distance with fine sand deposited in its protection on the lee slope. That the ridge has been in process of migration can be seen from the way the windward accretion layer overlaps the eroded, gently sloping foresets. The grain-size distribution for a bulk sample from the middle of this ridge is given in Fig. 65. The same phenomenon can occasionally be seen in sand-covered inter-dune areas in a sand sea. The maximum grain size, which controls the relative height of the ridge, will be that of the coarsest available sand.

The sand ridges described above occurred as individual features, or in groups of two or three ridges on the desert surface. Sharp (1963) figures a sequence of smaller ridges that are regularly spaced which he names “granule ripples”. He considers them to be a coarse equivalent of sand ripples that form where lag concentrates of larger particles are subjected to deflation and surface creep by strong winds.

Because of the size of the grains that cover and protect these large ridges from wind erosion they probably move very slowly, and may be preserved by burial under finer-grained wind-blown sand. Note the scattered vegetation.

AEO LiAN SAND RIPPLES

Ripple formation appears to result from a natural tendency for the surface of a bed of sand grains to pucker into raised features. Bagnold (1956) has shown that during the formation of ripples, a tangential stress is applied to the surface of a bed of grains, and some grains are eroded. Since these grains cannot be supported in suspension without an increase in the applied stress, they are redeposited again. The conditions at the surface of the bed are unstable. “An ultimate steady state can, however, be achieved if the eroded grains are redeposited in such a way that some new and additional tangential resistance is created” (Bagnold, 1956, p.256). This additional tangential resistance is provided by the drag caused by vorticity created in the hollows between newly deposited ripple crests. The principle applies

1 On the other hand, Sharp (1963, p. 624) points out that an eddy forms on the lee side of a ripple only at low wind velocities. At higher velocities no such eddy occurs and if grains on the upper part of the steep lee slope are dislodged, they merely roll down the slope.
equally to sand ripples formed in water or in air.

Aeolian sand ripples develop with the axes of their crests transverse to the wind that formed them. Bagnold (1941, p.149) has defined wind-formed sand ripples as “those surface forms whose wavelength depends on the wind strength and remains constant as time goes on”. He explains that in ripples, as in sand ridges, the coarsest grains collect at the crest and allow it to rise into the region of stronger wind. The troughs tend to fill up, but “if deposition is not too rapid, a balance is achieved between height of crest and trough” (see foreground of Fig.26). With an increase in the rate of deposition, balance is achieved at lower ripple heights until it finally disappears. This can be seen on the ripple-free accretion slopes of a growing dune. Ripples also flatten out and disappear when the wind rises above a certain strength.

The ratio of wavelength to height is known as the ripple index. For aeolian sand, the ripple index is commonly between 15 and 20, but when ripples flatten out at high winds, the index may be as high as 50 or 60. The index for the granule ripples measured by Sharp (1963) in the Kelso Dunes of the Mojave Desert ranged from 2 to 12 with a mean index of 15. He believes that the ripple index varies inversely with the grain size and directly with the wind velocity.

For sand ridges, the wind strength is below that required to remove the largest grains from the crest. Transport of the coarse grains is by surface creep. For ripples, an intercrest wavelength is developed which is proportional to the characteristic path of the grains in saltation; the variation in the distribution of the descending saltation causes a corresponding variation in the rate of surface creep. For a given grain size, wavelength increases with wind strength until a point is reached when the ripple has virtually flattened out. The height of the ripple also depends on the range in size of the grains forming the ripple, and the ability of the coarsest grains to remain in their crestal position for a given wind strength. As with sand ridges, a wide variation in grain size is necessary to produce large ripples, which are always asymmetrical. According to Sharp (1963) the degree of asymmetry varies directly with grain size.

Asymmetric ripples are well seen in Fig.26 of Wadi Al Ayn, where the asymmetry of the recently formed ripples shows that the wind that formed them blew upstream towards the Oman Mountains. Ripples are, in fact, an approach to the sand ridge. With uniform grain size, only low ripples are possible (Bagnold, 1941, p.151).

Although the accretion slopes of some dunes may be free from ripples because of strong winds or a sand supply of uniform grain size, other dunes may have their windward slopes and their flanks covered with ripples, and these ripples may be preserved (see Fig.120, 121). Ripples also develop on the steep avalanche slope of a dune when the wind temporarily blows across that slope. These ripples are preserved when an avalanche of sand covers them from above after the wind has reverted to its prevailing direction.

Wind-formed sand ripples of Permian age have been seen in dune sands in both the quarries shown in Fig.79 and 80, at Houghton-le-Spring, Durham, and at Locharbriggs, Dumfriesshire, and have also been described by McKee (1934, 1945) from the Permian Coconino Sandstone of northern Arizona. Almeida (1953) refers to the preservation of ripples in Triassic aeolian sands from South America, and aeolian ripples of Jurassic age have been described from the Navajo Sandstone of Utah by Kiersch (1950).

SAND DUNES

In contrast to the upward coarsening of the grain size seen in the sand ridges and ripples of Fig.63, the dune in the background shows a progressively finer grain size from the level of the deflation plain to its crest. At the same time, there is an upward improvement in the degree of sorting. This can be seen from the grain-size distribution graphs given in Fig.66.

The size of a ripple is controlled by the strength of the wind and the coarseness of the grains occupying the crestal position. The size of a dune is dependent upon the supply of sand and the ability of the wind to carry grains of any size up to its crest—and keep them there.

BARCHAN DUNES

It was stated on p.23, that sand tends to accumulate on areas that are already sand covered. Such an accumulation of sand is commonly either in the form of long strips or low, oval, ripple-covered mounds, depending on the velocity of the wind that formed them.

Under conditions of constant wind direction and accretion of sand, the highest point in an oval-shaped mound is found towards the down-wind end. As the patch of sand increases in height, so the angle of the lee slope increases until it reaches the angle of repose which, for dry sand, is about 34°, and a slip face (or avalanche slope) forms. This situation appears to be reached at a minimum height, for what may now be called a dune, of about 30 cm. The lee slope of the dune is fed with new material blown over the crest and the whole dune advances down wind. Any tendency to exceed the angle of repose causes sand to avalanche down the slip-face until the slope of 34° is regained. The largest and roughest grains are found at the bottom of the slope.

Bagnold (1954) has shown that this sorting on the avalanche slope is the result of an internal dispersive pressure within the flowing sand that varies with the diameter of the grains for any given shear stress. The largest grains tend to drift towards the zone of least shear strain which is found at the surface of the
migrates, so older slip-face slopes become covered at the leeward end of the dune and are exposed to erosion to windward.

With a growing barchan, rather more sand is deposited on it than is removed. Since there is a greater bulk of sand to be moved in the centre of the barchan than at its "horns", the latter migrate more rapidly, and it is from here that the greatest sand loss occurs. On the horns, there are no slip-faces. The laminae will all have dips of less than 34° in directions up to 90° to that of the wind, although a spread of less than 15° is normal. McKee (1966a) gives excellent cross-sections and photos of dune-bedding relationships not only through barchans, but also through other dune types. The cross-sections were prepared by cutting trenches with a bulldozer right through dunes composed of lightly cemented gypsum dune sands. Barchans are known from most deserts, but for some reason do not appear to occur in the Kalahari or Australia (Hills et al., 1966). They do, however, occur in the Namib Desert of Southwest Africa (Kaiser, 1925). Detailed studies of the barchans of southern Peru have been made by both Finkel (1959) and Hastenrath (1967).

When the attitude of the preserved bedding of barchan dunes is plotted on a polar net (Fig.68), the resulting scatter of points can be used to indicate approximate palaeo-wind directions. This method has given the writer reasonably consistent palaeo-wind directions for Permo-Triassic dune sands in Britain and the Pliocene to Recent cemented dunes from eastern Arabia. The subject is discussed more extensively on p.100.
Aeolian sand will be deposited in any location where there is a reduction in the velocity of a sand-laden wind. A wind of reduced velocity is no longer capable of transporting the larger grains that were previously in saltation. These grains therefore fall to the ground where, by increasing its roughness, apply a further drag resistance to the wind and so further reduce the effective velocity at ground level. We have already seen how wind-blown sand accumulates in the lee of the bank of a wadi. Given a sufficient supply of sand, dunes may fill the wadi bed. Dunes may also form as the result of a simple reduction in wind velocity: "Aeolian sand tends to accumulate on areas already sand covered..." (p.23 and Fig.10). Once started, dune build-up is often self-propagating.

Most dune fields appear to form in areas where there is an adequate windward supply of sand and where the wind velocity is not so great that only deflation can take place. Dunes do not, as a rule, form on elevated landscapes where wind velocities are generally higher than on the planes below. The supply of sand is sufficient for dune formation along many low coastlines of prevailing onshore winds such as parts of the Pacific coast of North America (COOPER, 1958, 1967), The Netherlands coast (VAN STRAATEN, 1961) or the Trucial Coast of the Persian Gulf. As we have already seen, areas of sporadic fluvial or lacustrine sedimentation in a region of arid or semi-arid climate provide an abundant supply of sediment for incorporation into sand dunes further down wind. Two such examples are the Simpson Desert in Australia (MADIGAN, 1946) and the dune fields of the Great Kavir in Iran (BOBEK, 1959).

Sand transport is more effective over a hard immobile surface than over a sand-covered surface (p.23). Sand is therefore transported over and away from regions where active erosion of outcrop is taking place; it accumulates in areas where the surface roughness is greater and where wind velocities are lower. Erosion over the Hamada al Hamra and Harug el Hasued in Libya is compensated by deposition in the depressions or the Ubari and Murzuk Sand Seas (Enclosure 4). Erosion and transport of sediment over the highlands of northeast Arabia are followed by deposition to the southeast in linear depressions such as the Dahna of Saudi Arabia (HOLM, 1960).

Fig.69 shows the leeward edge of a barchan in a small dune field in Qatar, Persian Gulf. The windward end of this dune field starts as isolated barchans near the centre of the peninsula. As the dunes travel to the southeast under the action of the Shamal wind, they combine into more complex barchanoid forms which, on the east coast of Qatar, slowly migrate into the sea. At this coast, however, over much of the year there is an opposing daily onshore wind caused by the fact that...
the air over Qatar heats up more rapidly than over the water of the Persian Gulf. This daytime wind is strongest during the late morning and early afternoon and is far more effective in moving sand grains than the nightly offshore winds. Inman et al. (1966) have noticed a similar variation in wind strengths and directions along the coast of Baja California, Mexico. The result is that south of Umm Said (Enclosure I) the dunes tend to build up along the coastline with only relatively small quantities of sand being carried out to sea.

The source of the sand that is found in a dune field such as that just described on Qatar, is not always obvious. The sand grains consist predominantly of quartz and yet the surface of the Qatar peninsula is composed essentially of limestone and dolomite. It is believed that the quartz sand was supplied to the peninsula from Saudi Arabia during the last Pleistocene glaciation when the shallow Gulf of Salwa (Enclosure I) would have been above sea level. Since the post-Glacial rise in sea level, the supply of quartz sand has been cut off and the dunes are now concentrated near the southeast coast of Qatar.

DUNE COMPLEXES

Not all dunes are simple barchans. In Fig. 70, from the eastern edge of the Rub al Khali, Arabia, small barchan-like complexes oriented transverse to the wind, are being slowly added to the larger stellate dune forms. These latter, by virtue of their size, provide considerable resistance to the wind with an accompanying increase in the wind gradient. As a result, the wind tends to veer towards the larger dunes and so deposits additional sand on and around them. In Fig. 71, from the southern edge of the Umm as Samim, it can be readily imagined that the high stellate complex originally formed as the result of smaller barchans overtaking a larger one during migration from left to right. This description is, however, oversimplified; it applies only to the relationship between the large stellate and small
bar chan dunes as seen at the time the photograph was taken. According to Holm (1960, p.1371), pyramidal (stellate) dunes form as the result of winds that “beat around the compass”. McKee (1966, p.68) also assumes that alternating winds from several directions are required to build a stellate dune. Dune complexes, such as those shown in Fig.70 and 71, can migrate over the salt surface of the Umm as Samim.

Fig.72. Giant bar chan complex. The Liwa, Trucial States.

OPPOSING WIND DIRECTIONS

Barchans can continue to multiply until they merge with one another and the substrate is completely covered, as seen in Fig.72 of the Liwa Sand Sea, Abu Dhabi, Trucial Coast. The major slip-faces of these barchan-like dunes are formed by one dominant wind and are reported to be in the order of 500 ft. high. In Fig.72, a small crestal “lip” occurs at the top of each of these giant slip-faces, with its avalanche slope (see Glossary) facing the opposite direction. This is assumed to result from a temporary wind directly opposed to the prevailing direction. A similar occurrence was observed in the Rajasthan Desert, India (Glennie and Evamy, 1968, plate 3D). There, the major foreset slopes result from the Southwest Monsoon but two months after the monsoon, a milder wind blows from the northeast and forms small crestal lips whose foresets are directed towards the southwest.

SEIF DUNES

Not all dunes are of the barchan type. Many are roughly linear, with the line of elongation apparently parallel to the prevailing wind direction. They are referred to by Bagnold (1941) as “seif” dunes, from the Arab word for a sword. Bagnold (on p.223) has suggested that one type of seif dune may be formed by the elongation of one horn of a barchan which then fuses with the next dune en echelon with it. This, he claims, is caused by winds which are oblique to the prevailing transport direction. He is supported in this interpretation by McKee (1966a) and McKee and Tibbits (1964) who consider that the sel-dune structure seen by them near Sebha, in the Ubari Sand Sea, Libya (see Enclosure 4), is largely controlled by winds from two directions about 90° apart. The present writer is not convinced by their arguments. Although the general pattern of wind directions over desert areas is now known, there is little information available on the strengths and directions of wind at ground level and virtually no continuous (24 h/day) records are kept outside North America. It seems, therefore, that we can only speculate on the effect that a wind of varying strength and direction may have on dune formation.

McKee and Tibbits (1964, pp.6, 7) quote wind records for the Sebha oasis in support of their interpretation. The figures given are of readings of wind directions and velocities that were recorded daily for several years at 6 a.m. and 6 p.m. The morning wind is dominantly from the southeast and shifts during the day to give an evening wind from the northeast. These times, and especially the earlier hour, are, however, often the calmest moments of daylight. In the desert, wind velocities normally increase during the day and die away again in the evening. From personal observation, winds often appear to be stronger in the afternoon than at any other time during the day or night. This may be the result of convection effects associated with the heat produced by the sun in the middle of the day. It is during the heat of the afternoon, when winds are assumed to be at their strongest and capable of transporting the greatest volume of sand, that they will probably blow from about north-northeast in the region of Sebha, or roughly parallel to the dune axes. This direction would appear to agree well with the sand distribution given by McKee and Tibbits (1964, fig.2).

Linear dunes can also undoubtedly form parallel to one wind direction, as can be seen from the results of a single storm in the dunes of the coastal areas of The Netherlands. Bagnold (1941, p.171) says “a strong wind causes an accretion of sand on an existing sand patch together with an extension up-wind of the border… this action lasts only as long as there is a plentiful supply of sand…” He goes on to say (p.178), “in a strong sand-laden wind a uniform drift of sand over a uniform rough surface has a transverse instability, so that sand tends to deposit in longitudinal strips”. It is here suggested that the combination of these two factors, related essentially to a strong wind of uniform direction, causes the formation of many seif dunes.
In a region of seif dunes, pressure gradients exist between the axes of the interdune areas and the crests of the dunes; these pressure gradients are caused by the resistance to the wind of the dune itself. This results in the formation of "wind cells" in the inter-dune areas, which give the wind an overall spiral motion directed outwards at ground level towards the dunes (Fig. 73).

Support for this hypothesis comes from a study of linear sand ridges in the North Sea that are parallel to the marine currents that formed them (Houbolt, 1968). From the distribution of lines of flotsam over the axes of the inter-ridge areas and the absence of flotsam over the crests of the ridges, Houbolt deduced that these ridges were built as the result of a spiral motion superimposed upon a tidal current that had an overall linear direction of movement. Unlike the winds in areas of seif dunes, the direction of these tidal currents is reversed four times each day.

Additional support for this hypothesis can be found in Fig. 74, a vertical aerial photograph of part of the northern end of the Wahiba Sands in Oman. The major linear dunes are a part of the same pattern as that seen in the frontispiece, only a little further to the west. As was mentioned on p. 38, they are thought to have been formed by Pleistocene winds that blew from roughly south to north (from about N180°E to N200°E). Since the Pleistocene, there has been a shift of about 20° to the east in the predominant sand-transporting wind direction (now from about N160°E to N180°E) so that these large dunes are no longer in equilibrium with the wind. The crests of the dunes are now being eroded and the sand is being redeposited in the old broad interdune areas as similar, but much smaller and more closely spaced linear dunes. The active dominant sand-transporting wind—the Southwest Monsoon—is here blowing parallel to the small linear dunes seen spreading over the sands and gravels of Wadi Batha. It is also sub-parallel to many of the small linear dunes that cross obliquely from the interdune areas up the eastern flanks of the large Pleistocene seifs. Many of them, however, show a distinct curvature westwards towards the crest of the large dunes. This curvature is even more pronounced, only this time towards the east, in those small dunes that stretch from the interdune areas towards the steeper western side of the Pleistocene seifs. This distribution of large and small dunes is thought to be in conformity with the idea that seif dunes are built by spiral eddies associated with a strong wind from one dominant direction, as outlined in Fig. 73.1

Had the Pleistocene seifs been parallel to the present dominant sand-transporting wind, the dune pattern of smaller seifs could not have existed. It is because the dune orientation is out of equilibrium with the sand-transporting wind, that erosion of the crests provides sand in the interdune areas that can be used to

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1 The formation of seif dunes in a similar way by means of one strong dominant wind is proposed for the Sahara by Dubief (1952) and for the Simpson Desert (Australia) by Madigan (1946, pp. 57–62). Madigan (on p. 61) also has noticed that eddies can carry sand grains obliquely up the side of a self.
trace the path of the major dune-building eddies. Because the Pleistocene dune orientation is only slightly out of equilibrium with the present sand-transporting wind, these large seifs still affect the winds in much the same manner as a self orientation that is in equilibrium with the wind. The large, moderately dipping surfaces of seif dunes will consist of tightly packed accretion bedding—in contrast to the loosely packed avalanche bedding of the foresets of barchan dunes. The accretion bedding is deposited by saltation and surface creep of sand grains moving over the surface of the dune at an angle only slightly oblique to that of the predominant wind, and sub-parallel to the long axis of the dune. Some characteristics of the bedding associated with seif dunes are given in Fig. 68 and McKee and Tibbits (1964) figure the style of cross-bedding found in seif dunes as seen in test pits.

Bagnold (1941, p.69) states that sand movement is proportional to the cube of the excess wind velocity “above that at which sand begins to move. For a given sand, a wind of 16 m/sec will move as much sand in 24 hours as would be moved by a wind of 8 m/sec in 3 weeks.” This fact lays great stress on the transporting power of strong winds which will control the shape and distribution of sand dunes. Gentle winds, on the other hand, are only capable of slight modification of the surface laminae of dunes.

Fig. 75 is an aerial photograph, taken in the month of February, of a number of recently formed seif dunes in the central Wahiba Sands. They all have well developed slip-faces on their western flanks in response to a light easterly wind that blows throughout much of the winter. Further to the south, dunes occur as north-south rows of small barchans migrating slowly towards the west. A few hours of a strong wind blowing occasionally from the north in winter, or more regularly from the south in the summer, is sufficient to obliterte these slip-faces and convert the barchans into long seif dunes parallel to the path of the wind. The steep western flanks of both the large dunes seen in Fig. 74 and the smaller ones that are superimposed on their surfaces have also been formed by the same light easterly winter wind.

It is thought that although the present southwest monsoon is strong enough to build small seif dunes with a crestal spacing of 100-500 m, it is not strong enough to have built large dunes of the size seen in Fig. 74. To form dunes of this size it is probably necessary to have much stronger winds than exist at present. The occurrence of very strong winds would have been possible during glacial periods when the large areas of polar high pressure would bring about a concentration of the other air-pressure belts towards the equator (Lamb, 1961). Hare (1961) believes that the easterly winds of the desert belts would probably also be more strongly developed during glaciations.

The coincidence of active dune formation in desert areas with polar glaciation is proposed by Fairbridge (1964, p.128). He states that vegetated sand-dune ridges of the seif type can be observed disappearing below sea level in North Australia, Arabia and West Africa. Friedman (1964) has sampled cemented dune sand from below the high-tide zone off the Yucatan Peninsula in Mexico, Kendall and Skipwith (1969) report the presence of Quaternary dune sand below sea level off the coast of Abu Dhabi, Trucial Coast, and the writer has traced cemented dune sand to below low-tide level near Muscat, Oman. The correlation of the formation of West African desert dunes with a time of lower sea level is also proposed by Tricart et al. (1957).

Traces of linear dunes of probable Pleistocene age occur within what is now the zone of tropical low pressure (roughly 10°N-10°S; see Fig. 1). They have been
mentioned by Flint (1959) and Fairbridge (1964) as occurring in the Congo (4°S) and southern Sudan (5°N). Other old systems of linear dunes in Africa are found further from the equator, but their orientations are not compatible with the present prevailing winds (Flint, 1959; Fairbridge, 1964; Flint and Bond, 1968). These occurrences provide evidence of the shifting of air-pressure belts since these old dunes were formed.

The shorter distances involved between areas of high and low pressure during a Pleistocene glaciation would give rise to a much stronger global air circulation. Evidence of this can be seen in other areas of large seif dunes such as in the central Rub al Khali or the Rajasthan Desert, and in dunes that have already undergone or are now undergoing some modification of their original orientation because of shifting air-pressure belts. These latter can be seen in the dunes of the northern Trucial States (Enclosure 2) and possibly also in the periglacial dunes of Nebraska (Smith, 1965).

Conversely, during the short "warm" interglacials, the areas of polar high pressure would be reduced to a minimum. All global air-pressure zones would spread out towards the poles and the earth's air-circulatory systems would be weaker than at present. In mountainous areas in a desert, winds caused by convection would probably be dominant over any trade-wind pattern. When these "convection winds" pass over a sea such as the Gulf of Oman, thunderstorms and heavy rain over the mountains are likely to follow: this, indeed, can occasionally be seen on the Batinah coast of Oman today. In the past, it could have been a common occurrence and might account for the widespread distribution of gravels that now flank the Oman Mountains.

This over-simplified explanation of "pluvials" is given a slightly different interpretation by Butzer (1961, 1963). He points out (1963, p.212) that in the Mediterranean area, the "Würm pluvial par excellence" (italics in original version by Butzer) coincided with the period of glacial advance; the later period of maximum glaciation was dry. From this, he concludes that subtropical pluvials cannot be interpreted as secondary effects of the presence of polar ice sheets, but rather that they are associated with a primary change in the atmospheric circulation that is coincident with polar extension.

In its simplest form, therefore, it is suggested that, other conditions being equal, barchans (or transverse dunes, depending upon the supply of sand) will tend to develop at lower wind velocities, and seif dunes will form when the wind velocities are higher. The higher the wind velocity, the larger the seif dune and the greater the interdune spacing, is a statement that may well explain the distribution of the large and small seif dunes of the Wahiba Sands in time and space.¹ The wind velocities experienced today are probably incapable of causing the formation of the large seif dunes found in many deserts.

In the southern Rub al Khali, seif dunes may be 150 m high (Beydoun, 1966). In the Ubari Sand Sea, Libya, they attain a height estimated at around 100 m and a length of 100 km or more and are often covered by small migrating barchan-like dunes. In the conditions under which these large seifs were formed, barchans are unlikely to have been able to survive. These barchans are, therefore, indicative only of the present wind regime that bears little relation—apart, perhaps, from direction—to the older and stronger winds that almost certainly formed the large seifs.

**TRANSVERSE DUNES**

In the eastern part of the Rub al Khali, there exists a large area of dunes known as Usuq al Mu'taridah (Enclosure 1). The long axes of these dunes are roughly perpendicular to those of the north–south oriented seif dunes found further to the west. These transverse dunes are separated by inland sebkhas. Although they have not been visited in the field, their morphology, as seen in aerial photographs (see Fig.76), suggests that they are formed by a wind blowing approximately from the

¹ This statement may be taken a stage further. For a given particle size, the strength of the wind is a factor that controls the character and spacing of aeolian sediments from transverse sand ripples and barchan dunes at low wind velocities to linear sand strips and seif dunes at high wind velocities. Similar relationships occur in an aquatic medium: low current velocities result in small transverse ripples; sand ridges form parallel to strong tidal currents on shallow continental shelves (Allen, 1968a).
north; they are probably built of more or less unidirectional foresets of barchan type dipping to the south. McKee (1966a) describes similar transverse dunes from New Mexico in which the axes of the dunes are at right angles to the dominant wind direction and the dune laminae dip down-wind. He does not, however, mention interdune sebkhas or playas.

The reason why these dunes have not developed as seifs, with an almost north-south axis parallel to the dominant sand-transporting wind, is probably concerned with the position of the area relative to the main path of the winds of Shamal origin. The Uruq al Mu'taridah lies to the east of the main Shamal track, and, as a result, the wind velocities may well be too low for seifs to develop.

If there had been a more limited supply of sand, it is thought that simple barchan dunes would have formed. With a more plentiful supply of sand, the barchans should grow in size and coalesce. When this happens, the resulting morphology is commonly that of a complex sand body whose long axis is oblique to the wind but which still bears some resemblance to the barchan forms from which it was derived. Why then, should these transverse dunes possess such regular spacing at right angles to the predominant wind?

From observations aided by the use of black smoke, Cooper (1958) came to the conclusion that the path followed by the wind across a series of transverse dunes has a wave-like form. According to him, it climbs parallel to the windward surface of the dune and descends again to the windward edge of the succeeding dune, leaving a zone of dead air in the region of the slip face. Wind vortices occasionally develop in this zone of dead air and presumably account for the low velocity reversed wind directions measured close to the foot of a slip-face by Inman et al. (1966) in the coastal dunes of Baja California, Mexico. Cooper (1958) believes that no deflation takes place in the intervening interdune area. The writer, on the other hand, suggests that deflation of the interdune area does occur. He believes that over the transverse dunes that cover a flat desert surface, a wave-like path is followed by the wind that conforms closely to that sketched in Fig. 77; it causes successively, transport and then deposition of sand, followed by deflation of the interdune area.

Since such a wind path is unlikely to develop over a hard desert surface with only occasional barchan dunes, it seems possible that transverse dunes are derived from areas that are plentifully covered with sand. Once started, the wave path of the wind perpetuates the size and spacing of the dunes for any appropriate wind strength provided there is no gain or loss of sand. In the case of coastal dunes, the initial supply of sand is provided by the beach; with the dunes of the Uruq al Mu'taridah, the source of the sand is probably the southern edge of the Liwa Sand Sea (see Enclosure I). As with the large seif dunes of the Wahiba Sands and Rub al Khali, the time of original formation of these large transverse dunes is thought to be during Pleistocene glaciations. The annual rate of sand transport is now probably much lower than formerly.

In Fig. 77, the situation is depicted in which the almost planar surface of an inland sebkha is overlain by transverse dunes. Initially, the ground-water level is assumed to coincide with the sebkha surface. The transverse dunes that migrate across the surface of the sebkha will be regularly spaced because the near-surface path of the wind will follow a distorted wave pattern. The sand will be transported over the dune, deposited behind the dune and deflated from the interdune sebkha surface. If the wind is from the north, the dunes will slowly migrate to the south. The damp sebkha surface will inhibit growth of barchan-like horns and deflation will tend to lower any "dry-season" accumulations down to the level of the water.
If sand supply exceeds sand removal, the dunes will grow in height and therefore in surface area. Given sufficient time, the interdune sebkhas will be reduced in size until eventually the area becomes a sand sea of transverse dunes. The final remoulding is that the transverse dunes of the sand sea may finish up with a complex barchan-like morphology resembling that seen at present in the Liwa (Fig. 72).

With later burial and a rising water table, most of the salt and gypsum cement of the dune sands, at present seen in sebkhas, is likely to be dissolved by the comparatively fresh ground water brought in from below. An original association of transverse dunes and inland sebkhas may not be recognised in a fossil desert sequence.

The orientation of the almost linear dunes seen in the back half of Fig. 78 does not appear to conform with the present wind direction as deduced from the small barchans migrating over both them, and the large barchan-like structure in the foreground. This, perhaps, can be explained from a study of Enclosure 1.

The dunes are some 160 km south-southwest of Jebel Hafit. Their axes are aligned roughly west-northwest—east-southeast. They were probably formed by strong Pleistocene winds that blew parallel to the axes. They are assumed to have been originally formed as seif dunes but are now slowly undergoing modification under the influence of the present milder winds that blow from the north. This milder wind is not capable of building seif dunes, so that the large barchan-like structure is being modified into a crescent shape. Similar, large, distorted, barchan-like dunes are also being formed at present from the remnants of the Wahiba seifs now in the middle of the Wadi Batha alluvial plain a few kilometres due east of the locality seen in Fig. 74. The foresets that have formed in recent time are all of barchan type and dip downwind. The dunes seen in Fig. 78 are seifs that are being modified into transverse forms because the wind velocities are now too low to build seifs. Without a complete analysis of the bedding of the sands in the core of the dunes, they may easily be confused with normal transverse dunes if the new wind blows at right angles to the old.

**ANCIENT DUNE SANDS AND PALAEO-WIND DIRECTIONS**

Dune bedding is preserved in the sediments of ancient deserts in many parts of the world, and excellent photographs have been given in many publications on the subject.

The bedding of these ancient dunes was laid down in a fairly regular pattern as the result of deposition of sand grains following aeolian transport. From a study of the orientation of the dune bedding, the approximate direction in which
the wind was blowing at the time of deposition of these beds can be deduced. This has been done by many workers employing different statistical methods.

Shotton (1937) noted that in the Permio-Triassic dune sands of the west Midlands of England, the “false-bedding” with steeper slopes showed no more variation in attitude than that found on the down-wind side of a barchan. He extended this (Shotton, 1956) to show that barchan dunes should have a maximum in the direction of the wind that formed the dune when the measured azimuths were plotted on a rose diagram. By contrast, longitudinal (seif) dunes showed maxima that were almost opposed to each other, but with a small component in the direction of the prevalent wind (Shotton, 1956, fig.3E). Wright (1956) measured the maximum angle of dip in each available sedimentary structure in an area of outcropping Tertiary Chuska Sandstone in Arizona and New Mexico. He found that the maximum percentage of recorded dips lay in the range 23°–27°. The azimuths of these dips also gave an indication of the palaeo-wind direction when plotted onto a rose diagram. Lamming (1966) also used a rose diagram of the azimuths of dips for deducing the palaeo-wind direction of the Permian dune sands of Devon, England, and Bigarella and Salamuni (1961) used rose diagrams to show the overall palaeo-wind directions for extensive areas of the Early Mesozoic Botocacu Sandstone of Brazil and Uruguay.

When measuring the attitude of cross-stratification in aeolian sandstones, Opdyke (1961, p.52) states that “in general the magnitude of the angle of dip is a reasonable guide to nearness of approach to the true (palaeowind) direction”. In other words, low dip angles have little significance for palaeo-wind measurements. Opdyke then deduces the palaeo-wind direction from the statistical mean of a sufficient number of readings. Runcorn (1961, 1964) carries this a stage farther. He ignores all values of dip less than say 15° and then plots the frequency of dip directions on a histogram. After smoothing out a curve through his histogram plot, he concludes that the mean direction of the wind that formed the dune bedding coincides with the maximum frequency percentage. Shotton, in a discussion on Runcorn (1964), points out that if the bedding attitudes of longitudinal dunes were measured, two dominant dip directions would be recorded, each at right angles to the wind direction. If only one side of a longitudinal dune were exposed, then the deduced palaeo-wind direction would be almost 90° from the direction of the wind that actually formed the bedding. Poole (1964) also plots his dip azimuths on histograms using 50 readings per histogram sheet.

Following the method of vector analysis described by Curray (1956), Mackenzie (1964a) was able to reproduce remarkably consistent palaeo-wind directions for the Bermuda Pleistocene eolianites. He used between 2 and 11 dip measurements for each cross-bedded unit; the majority of dip angles measured were 30° or greater.

Reiche (1938) plots both the direction and the angle of dip of cross-bedded strata on a stereographic polar net. This has the advantage that the attitude of a bed can be represented by a point (the pole to the bedding plane). All dip attitudes may be shown but visual emphasis is still placed on the importance of the larger dip angles as indicators of the palaeo-wind direction. Polar nets have also been used by McKee (1940, 1962) and Kiersch (1950). The writer prefers to use the polar net for deducing palaeo-wind directions. In addition to direction, he believes that the distribution of the poles of dip planes can also indicate the type of dune that was formed (barchan-type or seif; see Fig.68) and may also suggest the local presence of more than one palaeo-wind direction.

Fig.79 illustrates large-scale dune bedding in the Permian “Yellow Sands” of Durham, England, possibly deposited as part of a linear or seif-like dune. The distribution of bedding attitudes and interpreted palaeo-wind direction is shown on the polar diagram, Fig.80. Dip attitudes from the other flank of the seif dune are poorly represented in this quarry.

Crests, probably associated with the “horns” of a barchan, are exposed in a quarry in southwest Scotland (Fig.81). They trend roughly N230°E and N250°E. The distribution of bedding attitudes for the crestal area and for the non-crestal remainder of the quarry are given in Fig.82 and 83, respectively. Note that the dip attitudes tend to be concentrated on either side of the wind trend in Fig.82, and that the additional points from the rest of the quarry, representing the centre portion of the barchan, fill in this gap in Fig.83. The distribution of dip attitudes on the horns of a barchan is rather similar to that found on the linear seif dunes (cf. Fig.80), but some of the dip attitudes found in the middle of the distribution seen in Fig.82, perhaps more correctly belong to the central part of the barchan.

Fig.84 shows the planed-off festoons of a cemented barchan-dune from locality 54, east of Abu Dhabi town, Arabia (see Enclosure 2). The distribution of dip attitudes (Fig.85), measured on a number of such dunes spread over a small area, does not show the gap seen in Fig.80. Some of the dip angles have been increased after deposition and cementation by the growth of gypsum crystals. Fairly steep dips almost normal to the prevailing wind have also been developed, possibly by wind-formed scours. This results in over-emphasis of the seif-like dip attitudes on the flanks of the barchan dune.

The distribution of the major seif-like sand dunes of the northern Trucial States is indicated on Enclosure 2 by crest lines. This pattern is emphasised by the presence of large feidjes. Together, they show trends that are roughly east–west in the south and progressively become more northerly to the north. This large-scale dune pattern is thought to have been formed by Pleistocene winds that blew from the west in the southern part of the area and that had a more northerly component further north. In the area covered by a broad strip bordering the coast and coastal sebkhas, the lower parts of these dunes have been cemented by calcite or gypsum cement. These cemented beds are occasionally exposed as the result of wind-removal of the underlying uncemented sands. At more than twenty localities,
palaeo-wind directions have been deduced from the bedding attitudes of outcropping cemented dune sands. They indicate that the wind that formed these dunes blew roughly parallel to the axes of the dunes in the directions suggested above.

The fine, closely spaced crest lines seen on Enclosure 2, east of Dubai and Sharjah, represent the axes of small self dunes that are being formed today. They are roughly at right angles to the axes of the large Pleistocene seifs and are formed largely from sand derived from the Pleistocene seifs by wind action. They are now extending across the surface of many of the feidjes, thus indicating that here, the orientation of the Pleistocene seifs is no longer in equilibrium with the present dominant wind direction.

Confirmation that the winds that formed the major self dunes of the Wahiba Sands were parallel to the axes of these dunes (p.90), also comes from outcrops of cemented dune sand. The deduced palaeo-wind directions are consistently towards the north, ranging from about N360° E to N20° E (Fig.86).

Some of the polar diagrams shown in Fig.86 have concentrations of points that are indicative of barchan as well as self dunes. Barchan-like characteristics are well seen at locality A, where high-angle dips (up to 30°) have developed directly down wind, and to a lesser extent at localities B and C.
Fig. 84. Placed-off festoons of a Cenozoic barchan. East of Abu Dhabi, Trucial Coast.

Fig. 85. Distribution of dip attitudes and deduced palaeo-wind direction of Cenozoic barchan dune. East of Abu Dhabi, Trucial Coast.

Any temporary lowering of the wind velocity over an area where seif dunes are forming will result in modification of the surface of the seifs and the building and migration of superimposed barchan-like forms. This is happening over many areas of seif dunes today. Even though strong winds may later destroy the superimposed barchans, some of their foresets that are directed downwind will be locally preserved. In a case like this, the relative importance of the two types of dip concentration seen in the polar diagram can only be assessed correctly if the outcrop areas are sufficiently well exposed. Apart from quarries and recently eroded dune sands in present desert areas, this is rarely possible.

OPDYKE and RUNCORN (1960), RUNCORN (1961), OPDYKE (1961) and POOLE (1964) point out the palaeoclimatological significance of desert sands, and show how the wind directions deduced from their bedding may be used to reconstruct the climatic conditions of the past. There is apparently, still sufficient evidence preserved in present day dune orientations to enable deductions to be made about deserts of probable Pleistocene age. By analogy, they point also to the possibility of local complexities in Permian desert dune orientation that reflect the influence of a Late Palaeozoic polar ice cap.
As shown in Fig. 68, the bulk of the bedding on a barchan dune dips at angles of up to 34° (the maximum angle of repose for dry sand) in the direction of the wind. In contrast, the bedding of linear self-dunes is oriented more nearly at right angles to the wind direction but with a variable component which is downwind.

From the arguments presented on the formation of self-dunes, it follows that bedding that is at the angle of repose should be confined to only a small percentage of the total volume of a self-dune. It is likely, however, that if an orientation study were made of the bedding attitudes found across the unconsolidated surface sands in the Wahiba (Fig. 74, 75), the resulting plot of points on a polar diagram would be asymmetric. This asymmetry would result from formation of dunes by two winds of different character; the strong south winds of fairly short duration that give the overall linear shape to the system, and the more persistent but gentle east winds that add the transverse or barchan-like character of widespread avalanche slopes (see p. 92).

The foreset bedding seen in Fig. 87 from the northern edge of the Ubari Sand Sea, Libya, was "fixed" by soaking with water and then exposed by digging. The 34 foresets continue down, parallel to the visible slip-face, almost to the base of the dune some 8 m below. The thin low-angle accretion laminae on the surface were destroyed by the flow of water.

HORIZONTAL LAMINAE IN DUNE SANDS

Horizontal and low-angle accretion bedding is found on the tops, windward slopes and rounded flanks of dunes. The horizontal bedding seen in Fig. 88 is in a pit dug on the flank of the same dune as that seen in Fig. 87.

HORIZONTAL LAMINAE OF SHEET SANDS AND INTERDUNE AREAS

Similar-looking but much more widespread horizontal bedding may be found in interdune areas and also in some sheet sands, as is the case shown in Fig. 89, also from the Ubari Sand Sea.

Baggsold (1941, pp. 158, 244) suggests that horizontally laminated sheet sands develop as the result of the distribution of protective layers of pebbles by periodic flooding. The sheet sand seen in Fig. 90, however, contains no pebbles; the grain sizes are similar to those found in nearby dunes; the well-sorted laminae are typical of wind-blown sands; there is little in either the sediment or its structures to suggest any fluvial transport; all features point to an aeolian origin. An alternative explanation for the origin of some sheet sands can be derived from
Bagnold's remarks (pp.151-153). He shows that ripples flatten out and disappear above a certain wind strength, that the ripple-forming tendency is reduced with sand grains of nearly uniform grain size, and that an increase in the rate of deposition lowers the ripple height so that with very rapid deposition the ripples disappear altogether.

This interpretation of the origin of horizontally-bedded aeolian sands is analogous to horizontally bedded fluvial sands. Simons and Richardson (1966, p.311) state that "in the Upper Flow Regime, resistance to flow is small and sediment transport is large. The usual bed forms are plane bed or antidunes". Allen (1968a, p.174), after discussing large ripple forms at moderate water velocities, continues: "a further substantial increase in flow strength will commonly lead to the destruction of the large ripples and to the formation of an apparently flat bed over which there is intense sand transport." A similar conclusion for the origin of horizontally-bedded flood deposits was reached by McKee et al. (1967).

Perhaps a combination of rapid deposition, high wind velocities and fairly uniform grain size of the transported sand may also result in sheet sand formation in deserts (see Bagnold, 1941, pp.149-153).

On the other hand, in the broad interdune areas, deflation tends to be dominant over deposition. The grain size at the surface will tend to be the largest which can withstand the interdune wind strength. Grains will be transported by both saltation and surface creep. Horizontal accretion laminae will form during periods of lesser wind strengths. Sorting may be expected to be poorer than in the adjoining dunes (see McKee and Tibbitts, 1964, fig.4).

McKee and Tibbitts (1964, p.8) suggest that the horizontally laminated sands of interdune areas, and the low-angle laminae of the lower slopes of self dunes are both formed by saltating sand grains. Whatever the explanation used for their formation, horizontally bedded sheet-like sands occur over wide areas of the desert without the presence of interbedded pebble beds of fluvialite origin.

CONROTATED BEDDING IN DUNE SAND VERSUS TEMPORARY WADI DELTAS

The contorted bedding seen in Fig.91 is in the slightly damp sand of a dune photographed about 3 km from the Mediterranean coast in Libya. The low-angle accretion bedding which truncates the slumped beds is clearly seen. The light-coloured laminae consist of carbonate grains, including Foraminifera, and shell fragments, etc., blown in from the coast, while the darker laminae consist of quartz grains.

How the slumping occurred and became preserved is partly explained by Fig.92. The surface of this coastal dune in Libya had become damp and heavy after rain. Being relatively unstable on the steep flank of the dune because of its water content, it slipped down over the underlying less-cohesive dry sand and retained a record of this movement in the form of folds and fractures because the sand was...
still damp at the time of slumping. A pit dug into the dune on the line of fracture revealed broken and contorted bedding similar to that seen in Fig. 91. On the other hand, Bigarella et al. (1969) believe that such strata became deformed as the result of oversteepening of the profile. Oversteepening (slopes greater than about 34') can only occur on sand surfaces that are already damp. Under moist conditions, sand transport should be minimal, therefore oversteepening by sedimentation is unlikely to occur. Apart from the illustrations given by Bigarella et al. (1969), contorted bedding from the gypsum dunes of Whitesands National Monument, New Mexico, is figured by McKee (1966, plates IVB and VE).

Slump structures have also been recognised in ancient dune sediments in Britain. Fig. 93 shows slumped aeolian sandstone at the top of the Permian "Yellow Sands" of northeast England (looking down over the edge to the quarry floor some 25 m below). At this locality, there are two slumps, one above the other. Later, these slumped sands were covered by marls of the transgressing Permian "Zechstein Sea", suggesting that the formation of these Permian slumps was probably also due to dampness, either rain, or the effect of the transgression itself.

Similar slumps have been seen at Ledstone Quarry, near Leeds, Yorkshire, where again, the yellow dune sands are overlain by fossiliferous marine marls. About 20 cm of sand between the undoubted aeolian and marine horizons appears to have been reworked by water and contains some moulds of brachiopods.

At Hilton Beck, in the Vale of Eden, slumps also occur at the top of the Permian Lower Penrith Sandstone. The overlying sediments, in this case, are wadi conglomerates known locally as the Upper Brockram. These slumps were presumably caused by the flooding wadi or associated rain. Whether or not water scouring of dune sand was also involved is not certain. Bedding attitudes in the sands of the Lower Penrith Sandstone at this locality suggest that they were part of a self dune formed by a wind that blew from the east. A little farther to the east, imbrication of the pebbles of the Upper Brockram indicates that here the wadi flowed roughly from south to north.

Wind, especially when associated with rain, can cause scouring followed by slumping of the sand. The result can be mistaken for contorted fluviatile or marine sands.

In North America, contorted bedding has been recorded from sandstones of aeolian origin from the Permian Coconino Sandstone (McKee, 1944), the Triassic Moenkopi Formation (McKee, 1954) and the Jurassic Navajo Sandstone of Utah (Kiersch, 1950).

Contorted bedding occurs in the Lower Cretaceous Nubian Sandstone of the northern Fezzan, Libya, and has been discussed in some detail by McKee (1962). He, probably rightly, ascribes their formation to "forward drag and overturning of the upper part of the beds... by a sudden rush of water and sand as demonstrated in laboratory experiments" (McKee et al., 1962, pp. 156-159). The same type of origin is assumed for the recumbent folds seen in Fig. 94 and 95. These cores
Earlier, the concept of cementation of dune sand was considered to be related to aeolian origin. Similarly, silicified sandstones may also reflect diagenetic changes possibly involving an aeolian sand and its alteration in a water-associated environment. The presence of silicified wood suggests that the sediment-laden water of a wadi in flood enters a temporary desert lake. They are included here for comparison with slump folds of undoubted aeolian origin.

In the Nubian Sandstone of the Fezzan, the recumbent folds sometimes form part of a graded bed. Associated quartz-pebble and curled clay-flake horizons indicate sedimentation under water. The curled clay flakes indicate that these beds have also been subjected to sub-aerial exposure. In such a sequence, especially where there are other features indicative of a desert environment, one may expect to find aeolian sand. McKee (1962, p.551) states that no aeolian sands occur in the Cretaceous Nubian Sandstone. On the other hand, when working in the same area, the present writer inferred the presence of aeolian sand on the basis of the criteria set out in chapter 2, p.11. Water-laid sediments undoubtedly exist in the Nubian Sandstone of this area, but then they are also a common feature in some parts of modern deserts and can still be associated with aeolian sands. Although the writer has not visited the other localities in North Africa and Arabia that are described by McKee in his paper, he could infer, from McKee's own descriptions, that many of these other areas possibly also possess aeolian Nubian Sandstone. McKee himself (1962, p.576) appeared to have difficulty in finding a suitable water-associated environment for cross-stratification that "in most respects... matches the structure of barchan dunes". Some of McKee's "recumbent folds" might well conform to slumps in dune sands. Perhaps further study of the Cretaceous Nubian Sandstone will permit its mode of deposition to be defined more precisely.

Another type of structure that is fairly common on modern stabilised dunes and is occasionally seen in Cenozoic dune sands is that of plant-root moulds. The writer,

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1 The Cretaceous Nubian Sandstone of southern Egypt was named by Russegger (1837). Use of the term "Nubian Sandstone" has since spread greatly both geographically and stratigraphically. The name has been applied to clean and conspicuously cross-stratified sandstone bodies in most countries of northeast Africa and on the Arabian Peninsula, and has been used for rock of varied lithology ranging in age from Cambrian to Cretaceous. Although ideas concerning the genesis of the Nubian-type sandstone have been numerous, the theory of dune or aeolian origin, first presented by WALTHER (1888), was one of the most popular. Since then, the concept of a fluvial, estuarine or deltaic origin has been promoted by palaeontologists impressed by the mixture of marine and fresh-water fossils in certain beds (NEWTON, 1909; BURLOLET and MANDERSCHD, 1963) and by geologists who noted the association of plant remains in many places (LYONS, 1894; BOWMAN, 1926). The theory of marine deposition has likewise had numerous supporters (SHUKRI and SAID, 1944; ATTIA, 1955). Others (KLITZSCH, 1966) describe the Nubian Sandstone (Messak Sandstone) as consisting of continental strata without further subdivision.

2 Mud-cracks, shale pebbles and boulders, mud-flakes, ventifacts, cellular dolomite (BUSSON, 1967, p.143). Apart from these features, the widespread occurrence of silicified sands and the presence of silicified wood (Di CESARE et al., 1963; KLITZSCH, 1966) is believed by the writer to be related to diagenesis in a desert environment. Although the time and process of silicification is not known, petrified wood is sometimes found buried beneath the dune sand and its alteration may be connected with the depth of the water table below the desert surface. Similarly, silicified sandstones may also reflect diagenetic changes possibly involving an earlier history of cementation by calcite (calcrete) and the formation of silicretes.
in conjunction with Evamy, has discussed the significance of the occurrence of these structures (GLENNIE and EVAMY, 1968). They use the word "dikaka" for them after an Arab word used on modern maps of Arabia to designate scrub-covered accumulations of dune sand (see also Bramkamp's dune classification in POWERS et al., 1966, p.1100).

All plants require water in order to live, and none more so than in a desert. Desert plants have leaf adaptations to reduce evaporation to a minimum. Many also have root systems that penetrate deeply in order to reach a permanent supply of the moisture that is so necessary to life (see KASSAS, 1966, for a more general description of desert plants).

On the surface of active sand dunes, plants have little chance to establish themselves because they are either smothered by drifting sand, or the sand is blown away from around their small young roots and they wither and die. Seeds that have been incorporated more deeply within a dune, however, can sprout after the dune has been dampened by rain. Provided its roots can reach a more permanent supply of water than is afforded by a surface wetting, the plant may survive and grow to maturity. Since these plants grow upwards, and extend their roots downwards through aeolian sand, they tend to disturb the previously formed aeolian bedding. When the plant dies, its tissue appears to be oxidised fairly rapidly to carbon dioxide in the arid environment of a desert, and sand from above fills the cavity so formed. Some of the dikaka found growing at the surface of stabilised dune systems possibly originally established itself from seed during the last pluvial phase of the Pleistocene. Once the dune is stabilised, it is relatively easy for the plants to spread. The large seif dunes of the central Rajasthan Desert, the northern Trucial States and the Wahiba Sands are all more or less stabilised and have a covering of dikaka. Over much of these areas there are no wadis to provide a shallow source of water. Apart from sporadic rainfall and dew, the life-giving moisture must be reached by long root systems.

Another way in which plants can become established on dunes is for pre-existing vegetation to become enveloped by drifting sand. Already in possession of a supply of ground water from a nearby wadi or coastline, the plant is often capable of maintaining upward growth as the sand drifts around it. Plants may also spread colonially by developing from near-surface lateral root systems. Since sand is trapped among the roots and branches of the plants, well-defined laminae are unlikely to form in the sand except in the lee of scrub bushes. Dikaka of this type is common on coastal dunes and sand drifts and where dunes migrate across old wadis.

Modern dikaka is most prevalent near to coastlines and wadis where a constant supply of water is available, and on stabilised dunes where their deep root systems possibly first became established during a late Pleistocene or early Holocene pluvial period.

Some plants had already adapted themselves to a desert environment by the Permian (e.g., *Walchia piniformis* and *Walchia jiliciformis*; MAGDEFRAU, 1956, pp.177–179), but appear to have grown in water-transported sediment. Their root systems, if any, seem not to have penetrated deeply into the underlying sediment. The earliest dune plants with sediment-penetrating roots (*Nathorstiana, Weichella, Hausmannia*) are known from Lower Cretaceous dune sands of possible coastal origin from Germany (MAGDEFRAU, 1956, pp.279–286). By Tertiary time, however, several root-forming plants had adapted themselves to living in the dune sands of arid deserts. Root penetration of Late Tertiary dikaka can be measured in lengths of at least tens of centimetres.

At Jebel Baraka, at the northeast corner of Sebkha Matti on the Trucial Coast (see Enclosure 1), the Late Tertiary 1 aeolian sands show a considerable amount of fossilistic evidence. Teeth, part of a jawbone and other bones assigned to *Mastodon* (*Tetrhalophodon*) sp. of Pontian age were found in a wadi conglomerate interbedded with the dikaka-rich aeolian sands. They were identified by Professor R. Dehm of the Institut für Paläontologie und Historische Geologie, Munich.
development of dikaka, but the original dune bedding of the sediment is still clearly visible (Fig.96). In other dune sands from the same locality (Fig.97) the dikaka root moulds are so numerous that the original dune bedding is only discerned with difficulty. Many of the moulds are oriented parallel to the sand laminations (almost horizontal in Fig.97), since this presumably was the direction of greatest permeability, and therefore the direction of flow of the intergranular water, so necessary for root-forming plants. Note the pebble horizon cutting across the top of the dikaka-riddled dune sand. These pebbles, deposited in a Pliocene wadi, are another indication of the relationship between dikaka and a nearby source of moisture.

The roots themselves are not preserved, apart from examples that were still living or had only recently died. The former distribution of roots in sediment deposited in an arid climate is, however, commonly exhibited by preferential cementation of the sand grains which had encased the roots. If, after such preferen-

tial cementation, wind deflation removes the more poorly cemented sand between the former roots, the root pattern becomes exhumed and stands out in relief.

Modern plants living along the Trucial Coast are generally halophytes which can live in soils having ground waters of high salinity. The fossil dikaka found in the same area is typically cemented by gypsum euhedra arranged in concentric patterns around former roots. The cement of the host rock, however, need not be gypsum. Along the Trucial Coast, for example, Quaternary dunes, away from localised gypsum-cemented dikaka, are cemented by calcite. Preferential cementation of fossilised mangrove roots has been reported from Florida by Hoffmeister and Multer (1965). It is interesting to note that in Florida, where the climate is relatively humid, the cementing mineral is calcite, whereas along the arid Trucial Coast of the Persian Gulf, the cement is gypsum. Plants that grew on dune sands away from mineral-rich ground water are unlikely to be recorded in fossil sediments by the preferential cementation of their root moulds.
Reference has already been made (p.24) to the fact that as particles decrease in size below about 80$\mu$m, the wind velocity that is required to set them in motion becomes greater. Once airborne, however, these silt and clay-sized particles can readily be kept in suspension by strong turbulent winds by virtue of their small size and weight. They can be carried great distances over the earth's surface by persistent winds before being deposited as a layer of dust or loess.

Loess is an accumulation of such wind-blown dust ranging in size from fine sand to clay-sized particles. Most of the particles fall within the range 20$\mu$m to 100$\mu$m with the bulk varying between about 30$\mu$m and 80$\mu$m.

According to Bryan (1945) and Obruchev (1945), loess has two main areas of origin; one is associated with the outwash of glacial rivers and the other is in deserts: around desert highlands extensive deposits of water-laid sands and silts are exposed to deflation. Both areas of origin can have very dry climatic conditions. The dust picked up in periglacial regions is deposited in marginal areas by a combination of polar anticyclonic winds and the prevailing westerlies. Dust is normally carried out of desert areas as dust storms and it settles in adjacent steppe lands. Dust, blown from the Sahara is occasionally found covering snow on the southern slopes of the Alps.

Although not common in desert areas, Beydoun (1966) mentions the presence of loess in the tributaries to Wadi Hadramaut in southern Arabia. Dust particles blown across the southern Rub al Khali or derived from the highlands of the western Hadramaut are presumably trapped in these deep gorges. Pleistocene loess from the northern Negev (Israel) is occasionally reworked during storms associated with dry easterly winds (Yaalon and Ginsbourg, 1966).

Barbour (1936) believes that the loess of China, although deposited predominantly during the Pleistocene, ranges in age from the Late Pliocene to the present. He considers that the loess was derived from the continental basins of the Gobi Desert. Smalley and Vita-Finzi (1968) suggest that the loess of China may have been derived, at least in part, from the glacial moraines of the Tien Shan Range. They are not in favor of deserts as major source areas for the production of loess. They believe that very little silt will be formed by mechanical reduction of sand-sized particles during aeolian transport. With this, the writer concurs, but as has been described by Hörner (1936), there remain extensive areas of alluvial sediments in the deserts of Central Asia that are subjected to deflation during the long intervals in which the rivers do not flow with water; similar areas of alluvial sediments occur in most other deserts of the world.

Loess is typically soft and porous, and yellow or buff in colour. It usually has little or no bedding, but is often riddled with small tubules that are thought to have been occupied by rootlets. The tubules are often lined with calcite. The loess appears to be trapped between blades of grass that, like dikaka, prevent the formation of well-defined laminae.

Large accumulations of loess in the U.S.A. and Europe appear to have been deposited during the Pleistocene beyond the margins of ice sheets. Dust storms also occur over the Great Plains of the United States when dry soils, lacking sufficient protection of grass or other vegetation, are picked up by strong winds. Swineford and Frye (1945) collected dust from the roof of a Kansas hotel and found from its analysis that it compared well with the grain-size distribution of Pleistocene loess from Kansas State. They concluded that the wind is competent to sort material to the degree represented by loess deposits.
DESSERT COASTAL SEDIMENTS

Apart from isolated “desert islands”, deserts are not usually surrounded by a coast line. Many deserts border coastlines along at least one edge, however (Fig. 1). From the point of view of palaeogeographic reconstruction it is important to realise that continental (desert) shoreline and marine facies may all occur in close proximity.

DEVELOPMENT OF COASTAL SEBKHAS BEHIND LONGSHORE BARS

Fig. 98 shows part of the Arabian coast near Ras al Khaimah, where the Oman Mountains approach close to the Persian Gulf (see Enclosure 2). Spreading out over the narrow coastal plain is a large alluvial fan built up by intermittently flowing braided wadi channels whose sediment is derived from the mountains (cf. Fig. 16, 17). The dark line along the periphery of the fan is caused by the presence of palm groves situated in a zone of maximum supply of water of tolerable salinity (see also p. 75). A coastal sebkha had developed between the conglomerate fan—which it partly also overlies—and the protection afforded by a long-shore sand bar. Much of the sand which went into the formation of the bar was probably obtained by erosion of coastal dune sands and longshore transport from areas of active carbonate-sediment production to the south. Wave action has built up this bar to above sea level and the wind has reworked the sediment into low dunes.

A coastal sebkha is an almost flat land area that occurs just above the level of normal high tides on the coasts of some low-lying hot, arid deserts. Its sediments consist of sand, silt and clay. Its surface is often covered by a salt crust that results from evaporation of water drawn to the surface by capillary action and from occasional marine inundations. The sediments of a coastal sebkha are the same as those of adjacent lagoonal and intertidal areas (and therefore commonly carbonate in composition) with an admixture of wind-blown sand and silt from offshore islands or more landward regions, and rare detritus from nearby hills carried in desert water courses (Evans et al., 1964a, b). Because of its environment, the carbonate-rich sediments of a coastal sebkha are liable to early diagenetic alteration as the result of reaction with brines of high chlorinity; they are characterised by the presence of algal mats (Kendall and Skipwith, 1968, 1969) and by the formation of evaporites (nodular anhydrite, gypsum and halite; Shearman, 1963,
The surface of a coastal sebkha is subject to deflation down to the level of the water table; the material so derived is often blown inland to be incorporated in nearby dunes. Large lenticular crystals of gypsum up to 20 cm across are exposed by deflation; they show a preferred orientation in that they grow with the plane of flattening in a vertical or near-vertical position. Many of these crystals enclose sand grains of the host sediment poikilitically (Shearman, 1963, 1966).

PRESERVATION OF DUNES BEHIND PROTECTION OF SPIT/COSTAL-SEBKHA COMPLEXES

South of the Ras al Khaimah sebkha an extensive tract of dunes occupies the broader area of coastal plain between the peri-montane alluvial fans and the coast (Fig. 99). The long, narrow coastal dune1, which is here higher than any dune directly inland, is now being eroded by the sea except where protected by the recent northward extension of another spit/costal-sebkha complex. This coastal dune probably owes its greater height to the combination of the strong onshore shaml winds and an almost unlimited supply of both quartz and carbonate grain from the beach and coastal sebkha.

The distribution and morphology of the coastal dunes found along the Pacific coast of U.S.A. has been described by Cooper (1958, 1967). He shows that there, coastal dunes form only where there is a good supply of beach sand that is replenished by longshore drift, and where the dominant wind is essentially onshore. Many coastal dunes form where coastal indentations are cut off by the formation of a marine sand bar that thus shortens the coastline. The predominant dune types appear to be transverse and parabolic, the latter being always associated with vegetation and partial stabilisation.

A feature that appears to be characteristic of some coastal dunes is the occurrence of bedding at the downwind side of the dunes that is convex upward. It has been reported by Mackenzie (1964a, b) and Ball (1967) from Bermuda, and by Bigarella et al. (1969) from the coast of Brazil. According to these latter workers, it appears to form as the result of deposition on the windward side of the slip face because the wind is neither so intense nor dry enough to cause complete removal of previously deposited beds. Ball (1967) notes the occurrence of root

1 It has been noticed in the field that the dunes which are nearest to a coastal beach tend to be oriented parallel to the coast line in contrast to those further inland whose orientation is commonly controlled by the prevailing wind. Similarly, Mackenzie (1964a) and Ball (1967) have both noted that the cross-bedding of Pleistocene coastal carbonate dune sands in Bermuda and Florida is always directed inland. Swad and Freshman (1968) have made a similar observation for Pleistocene and Recent dune sands on the coast of West Pakistan.
casts together with these structures and suggests that the dunes may have been partially stabilized. The presence of vegetation on the top of a dune and the resulting lowering of wind velocity near to the surface would certainly assist in the deposition of sand to windward of the slip face and the development of bedding that is convex upward. The coastal dunes described from Brazil are also at present stabilized by vegetation.

ABSENCE OF FLUVIAL DELTAS ALONG DESERT COAST LINES

Except in the case of large rivers, such as the Colorado, Nile, Indus and Tigris-Euphrates, that derive their water from beyond the desert, desert coasts do not have fluvial deltas. Wadi sediment which reaches the sea during flood may build a temporary delta, but redistribution of the delta sediments by wave action and longshore currents soon destroys it. An example of this observed by the writer was the delta of a Libyan wadi which was built some 500 m out into the Mediterranean Sea within a few days. Two months later, with no further flow of water or sediment from the wadi to maintain it, the delta had been so reduced that it protruded a mere 50 m beyond the original coastline. Fig.94 and 95 are cores from this temporary delta.

TIDAL DELTAS

Although fluvial deltas are generally absent along desert coastlines, tidal deltas may exist off the entrances to lagoons (Evans et al., 1964b). In the tropical shallow-water environment of the coastal lagoons of Abu Dhabi, the sediment of the tidal deltas is largely carbonate (Foraminifera, ooliths and fragments of Bryozoa, coral and shells). Storm and tidal oscillations of sea level result in strong currents at the entrance to large lagoons. Sufficiently concentrated backward and forward flow of water results in the formation of both inward and outward facing deltas (although not necessarily in the same channel), which in some cases are built entirely of ooliths (Butler et al., 1965).

PROGRADING DESERT COASTS

Fig.100 shows the development of an outward-facing submarine delta south of Ra's Ghanadah, Abu Dhabi (Enclosure 2). On the north side of the main channel there is an extensive area of coastal sebkha where large quantities of calcium-carbonate grains are trapped behind the protection of a sand spit. The spit is surmounted by a low linear dune consisting mainly of carbonate sand. South of the
main channel, a dune field covers sediments which probably represent an earlier coastal sebkha.

The manner in which this desert coastline has prograded may be imagined to a certain extent from a study of Enclosure 2. Off-shore bars have permitted the development of coastal lagoons and then coastal sebkhas behind their protection. Aeolian transport of carbonate sand has helped to raise the sebkha surface above storm-tide level and has resulted in the formation of carbonate dunes. The coast line will remain temporarily static until a further development of an offshore bar—perhaps a long-shore extension from a submarine delta—permits another phase of lagoonal sedimentation and coastal-sebkha development with its inevitable cover of aeolian sand.

When making a detailed study of the Quaternary, one must differentiate between Pleistocene sediments—aeolian sands and wadi gravels and sands that were deposited during a glacial lower stand in sea level—and the sediments that have been deposited since the post-Glacial rise in sea level. According to FAIRBRIDGE (1962, fig.9, 10), the Würm Glaciation resulted in a world-wide sea level that was something like 100 m lower than now. In the post-Glacial period, the sea level rose by a series of oscillations until about 6,000 years B.P. when it reached about the present level. In many parts of the world, the result of this post-Glacial rise in sea level is seen in widespread transgressions across what is now the continental shelf. In the warm, shallow, clear seas found along some river-free desert coasts, however, carbonate sediment can be rapidly "manufactured" by marine organisms living under optimum conditions.

The sediment sequence seen near Ras al Khaimah is complicated not only by the post-Glacial rise in sea level, but also by sub-Recent tectonic subsidence of the Musandam Peninsula. Further southwest, however, along the Abu Dhabi coast, Sebkha Matti, the Qatar coastal sebkhas, Bahrain Island and the eastern coast of Saudi Arabia, the coastline is prograding in areas where it can be assumed with reasonable certainty that there has been no marked change in sea level for the past two thousand years or more (Enclosure 1).

Although the Mediterranean is, in general, a region of coastal erosion, the coastline appears to be prograding south of Misurata, at the western end of the Gulf of Sirte, and on both sides of the Libyan/Tunisian border (Enclosure 4). This progradation is, no doubt, assisted by the rapid production of carbonate sediment associated with a river-free desert coast.

COASTAL CARBONATE DUNES

Coastal dunes which form in close proximity to a marine province of carbonate sediment production are themselves, in desert areas of prevailing onshore winds, often made up of the same carbonate grains. Early calcite (vadose?) cementation can result in the formation of a limestone which, when examined microscopically, is extremely difficult to differentiate from marine limestones since these are commonly composed of similar grains (Fig.127, 129). Foraminifera are well preserved, and a few of the more resistant of them (e.g., miliolids) may still be recognisable in dune sands after tens of kilometres of aeolian transport from the coast.

According to KUENEN (1960), aeolian abrasion of limestone proceeds 2 to 4 times as rapidly as that of quartz. With an increase in transport distance from a common quartz/limestone source area, the ratio of quartz to limestone will increase. If the carbonate fraction consists of diagenetically unaltered discrete skeletal fragments from the coast, then the rate of abrasion is probably even faster. Over the horizontal distance of 100 km, the quartz/carbonate ratio in the dune sands changed from 19:81 at locality 139 (see Enclosure 2) near the Abu Dhabi coast, to 76:24 at locality 137 near Buraimi. The situation here, however, is not that of a simple reduction in the carbonate content of the sand proportional to transport distance, because there are secondary sources of both quartz and carbonate sediment. Near to Abu Dhabi, the interdune areas undergoing deflation consist of the cemented carbonate sands of former coastal sebkhas, whilst near Buraimi, the interdune areas are rich in both quartz and limestone of older wadi deposits derived from the Oman Mountains.

Away from the coast, the bulk of the quartz sand in the dunes was possibly transported into the area during the Pleistocene, or even the Pliocene, to judge from dated aeolian sands from Sebkha Matti (GLENNIE and EVAMY, 1968). During these periods, the shores of the then Persian Gulf probably lay many tens of kilometres farther north than now, and there is a good chance that the westerly winds that moulded the major dunes during Pleistocene glaciations, blew parallel to the then coastline and transported only quartz sand derived from farther west in Arabia (Enclosure 1). With no knowledge of the earlier history of the area as a desert, a false impression might be gained of the resistance of carbonate grains to abrasion.

The Wahiba Sands, however, are rich in carbonate grains throughout their length. Much of this carbonate fraction is thought to have been blown from coastal sebkhas and marine terraces that existed between the mainland and Masirah Island. These areas are assumed to have been exposed to the action of strong winds during a Glacial lower stand in sea level. The present strait is very shallow. Although further evidence is still lacking, it seems likely that the influence of a rich coastal supply of carbonate sand can still be detected at least 170 km from its source, even though Foraminifera can no longer be recognised.

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1 According to M. Hughes Clarke (personal communication, 1963), Foraminifera may be better preserved in a coastal carbonate dune than in a marine environment, since in the former they are not attacked by marine boring algae.
A cemented coastal carbonate dune can form a very effective barrier between the sea on one side and the desert on the other. It can prevent wadis from discharging their load of water and sediment into the sea and thus result in the creation of an inland sebkha on its landward side (Fig.60, 62); it can protect these desert sediments—whether of aeolian or fluviatile origin—from erosion by a (potentially) transgressing sea if it can maintain a rate of vertical growth equivalent to that of the relative rise in sea level (Fig.3). Under these circumstances, the coastal carbonate dune may tend to migrate landwards over penecontemporaneous non-coastal sediments, which are thus preserved from marine erosion. Early—perhaps almost contemporaneous—cementation of such carbonate dunes strengthens their resistance to erosion.

Fig.101 shows the present coastal carbonate-dune barrier on the coast of Libya west of Tripoli. The recent uncemented dune is overlying an older Late Quaternary cemented carbonate dune, part of which is now submerged beneath the sea. Behind the dune barrier are an inland sebkha and wadi (extreme right) which have no direct access to the sea.

Dunes that are rich in carbonate grains, border the low northern (Batinah) coast of Oman. They also form a barrier between the sea and the wadis that flow from the Oman Mountains. In many places the coastal dunes are cut by wadis, but occasionally wadis have failed to maintain a channel to the sea, and inland sebkhas have developed on the landward side of the dunes (see p.69). Carbonate sand is often carried inland from the beach by the daily onshore winds, and so the dunes are built up and maintained.

**CEMENTED CARBONATE DUNES**

Fig.102 shows a typical eroded outcrop of the cemented carbonate dunes west of Tripoli. The bedding is both dipping and horizontal; alternation of finer- and coarser-grained carbonate laminae is typical of aeolian sands; the foresets exhibit crumpling of some laminae, which is very suggestive of the slumping seen on the flanks of recent dunes from nearby (Fig.91, 92). A thin section of the limestone is given in Fig.127.

Occasional cemented plant-root moulds and borings by recent pholad molluscs (at present living in the splash zone of the sea), combined with Bryozoa, Foraminifera and shell fragments, make it likely that a marine environment of deposition will be interpreted for outcrops such as this unless the aeolian nature of the bedding is recognised. The chances of making an incorrect deduction of the depositional environment are high in some outcrops where distinctive dune bedding is poorly developed. These chances are greatly increased if the deduction has to be made from cores. With only drill cuttings available, these carbonate dunes are almost certain to be considered as representing a marine environment of deposition. Thus their true palaeogeographic significance will not be recognised.

A misinterpretation of the environment of deposition of carbonate dune sands may result in the geological necessity of invoking fictitious changes in sea level. The Pleistocene quartz-carbonate dune sand seen in Fig.128 was deposited on the hills of Kutch over 100 m above the present sea level (for location, see Fig.109). Both Feeden (1884, p.56) and Auden (1952, p.59) believed that these and other similar sands represent a rise in sea level because of the marine microfaunas that they contain. On the other hand, Evans (1900, p.566) and the writer...
believe that these sands are of aeolian origin. Other similar Pleistocene carbonate dune sands have been found in western India at over 300 m above the present level of the sea (Evans, 1900), but the only evidence to connect the sands with the marine environment of their origin is their rich faunal content of miliolids and other skeletal fragments. There is no necessity for invoking a 300 m Pleistocene rise in sea level to account for their presence; a slight lowering of sea level, to expose a broad area of the present continental shelf to the wind during a polar glacial period, seems a much more reasonable hypothesis to account for their presence.

Segerstrom (1962) describes how a prevailing onshore wind in Chile carries sand inland from a deflated marine terrace for a distance of some 35 miles to a height of 2,800 ft. above sea level.

Coastal carbonate dunes are rich in particles consisting of complete or fragmented marine organisms that had first been washed onto the beaches and coastal sebkha and then transported to the dunes by the wind. These organisms had once lived in shallow sea water. Analyses by Friedman (1964) show that in the marine environment, their skeletons were generally composed of aragonite or high-magnesian calcite.

Many workers (Ginsburg, 1953; Emery and Cox, 1956; Friedman, 1964; Stoddart, 1965) have shown that carbonate sands can lithify as "beach rock" in a tropical intertidal marine environment. The cementing material commonly consists of acicular crystals of aragonite.

Friedman (1964, has shown that if carbonate fragments of marine origin are brought into a sub-aerial environment, they may lithify and the aragonite and high-magnesian calcite alter to low-magnesian calcite. This is possibly brought about by solution of the aragonite and the precipitation of low-magnesian calcite as cement; high-magnesian calcite possibly alters to low-magnesian calcite by solution of magnesium or solution-deposition on a micro-scale that leaves the original texture intact. Friedman (1964) has also shown that under sub-aerial conditions, aragonite grains have a metastability sequence in which the order of increasing stability is skeletal grains-ooids-pellets-cryptocrystalline grains. Aragonite grains are commonly leached to form moldic porosity and the molds are infilled by a drusy mosaic of low-magnesian calcite. Matthews (1968) points out that ancient carbonates are composed of the stable minerals calcite and dolomite. He has demonstrated that subaerially exposed Pleistocene carbonates exhibit mineralogical stability, thus suggesting that the process of stabilization proceeds rather rapidly.

From studies of eolianites on the coastal plain of Israel, Yaalon (1967) was able to infer that cementation must take place above the water table as the result of wetting by rainwater. From analyses of sands, he found that a minimum of about 8% CaCO3 was probably necessary to initiate cementation under the environmental conditions prevailing locally. He, like Friedman (1964), believes that precipitation of the interparticle cement takes place close to where carbonate was dissolved. In general agreement with Yaalon’s findings, the writer noticed that in the northern Trucial States (Enclosure 2), calcite-cemented dune sands were not exposed more than 30 or 40 km from the present coast. It is possible, of course, that cemented sands are present farther to the east but buried beneath a cover of uncemnted sands.

As was mentioned on p.70, dune sands can also be cemented by gypsum. In that particular case, it seems likely that coastal dunes prevented the flow of water from a wadi and inland sebkha to the sea. The dune sands probably became cemented with gypsum as the result of slow evaporation of the ground water associated with the inland sebkha.

Gypsum cemented quartz dune sands of probable Pliocene age are also found in Sebkha Matti (Enclosure 1) close to the locality seen in Fig.61. The quartz grains are poikilitically enclosed within large gypsum crystals in a fashion that is reminiscent of the Fontainebleau Sandstone of the Paris Basin. In this latter case, however, the quartz grains are locally enclosed in calcite rhombohedra. Alimen (1936, p.236) believes that these Oligocene dune sands were cemented recently as the result of percolation of lime-rich water from the overlying Beauce Limestone. In support of this idea, she mentions the growth of small concretionary nodules on the surface of freshly exposed sandstone in quarries. Similar nodules

1 Carbonate minerals can be identified by X-ray diffraction methods (Friedman, 1964) or by staining techniques (Friedman, 1959; Evamy, 1963, 1969).
have been seen by the writer on freshly exposed surfaces of the Permian "Yellow Sands" at Houghton-le-Spring in Durham, but he does not believe that the cementation of the Fontainebleau Sandstone is of such recent origin. The Oligocene dunes of the Paris Basin were apparently partially flooded by lime-rich lake water prior to the deposition of the Beauche Limestone (see also p.58). The writer suggests that evaporation of this water within the dune sands could result in cementation by the growth of calcite rhombohedra. Had the area been more evaporitic, the dunes might instead have been cemented by gypsum.

TIME OF CEMENTATION OF CARBONATE DUNES

The conditions required to cement the Libyan carbonate dune sands are thought to be similar to those described by YAALON (1967) for Israel. They presumably involve rainfall and subaerial exposure. The cement is low-magnesian calcite. The dunes were already cemented in Roman times, for not only is the ancient Roman city of Sabratha built on the cemented dunes, but the already-formed limestone was quarried as a rough building stone which was then faced with ornamental limestone and marble imported from Italy. The time of deposition of similar cemented carbonate dunes on the coast of Cyrenaica can be correlated with the Early Würm Glacial of the Alps (McBuRNEY, 1960, pp.163-165)—125,000 to 100,000 years B.P. In Morocco, archaeological dating of two generations of carbonate dunes shows that the older ones were deposited after the onset of the Mindel Glacial (780,000 years B.P.) and were probably already cemented by the start of the Riss Glacial (360,000 years B.P.) (McBuRNEY, 1960, pp.115-121). The younger dunes probably have an Early Würm age for the time of their deposition, like those of Cyrenaica.

There has been adequate time for slow cementation. A mechanism for the cementation of these carbonate dune sands could be solution of aragonite and high-magnesian calcite from skeletal particles and ooids by rainwater, followed by precipitation of low-magnesian calcite cement as the water evaporates. This explanation is supported by the archaeological dating of the artefacts from a cave which was carved out of the previously cemented dune sand (McBuRNEY, 1960). The dune sand was cemented under subaerial conditions before an interglacial rise in sea level resulted in the cave formation.

An oyster bed on the coast of West Pakistan has yielded a radiocarbon age of about 23,000 years. The oyster bed is overlain by a dune sand that is lightly cemented by calcium carbonate and which, in turn, is covered by another dune system that is still not cemented (SNEAD and FRISHMAN, 1968).

1 Dates taken from HOMES (1965, pp.698-707). According to FAIRBRIDGE (1962, fig.10), however, the Mindel Glacial took place about 220,000 years B.P.; the Riss about 110,000 years B.P. and the Early and Late Würm glacials roughly 50,000 and 20,000 years B.P., respectively.

Fig.103. Suspected dune bedding in Permian sucrosic dolomite. Magnesian Limestone, Derbyshire, England.

DIFFICULTIES OF RECOGNISING ANCIENT CARBONATE DUNES

Unfortunately, undoubted limestone dunes have not been recognised in ancient sediments which have also been subjected to deep burial and diagenesis. In Britain, foreset bedding in a sucrosic dolomite of the Permian Magnesian Limestone from Scarcliffe, Derbyshire, suggests the possibility that this sediment was originally deposited as a carbonate dune on the desert shores of the Zechstein Sea, and was later dolomitised (Fig.103). Microscopic examination of a thin section cut from this dolomite, however, gives no indication as to either its original lithology or its mode of deposition (Fig.130).

THE SALTS OF COASTAL EVAPORITIC LAGOONS

Deposits of bedded evaporites (halite, gypsum, anhydrite) are generally associated with seas and lagoons having a very restricted connection with the open ocean and are largely bordered by deserts. Water is evaporated from the surface of these seas and is largely replenished by an intermittent or even steady influx of salt water through a narrow strait; such conditions exist at present in the Gulf of Kara Bogaz. Water of low salinity (13°/00) flows continuously from the Caspian Sea into the gulf, which acts as a gigantic evaporating pan. Mostly gypsum was deposited in the past, but the salt concentration is now sufficiently high for a little halite to be precipitated in the hot season (GREEN, 1961; HOMES, 1965, p.138; DICKEY, 1968).

1 These evaporite minerals can be identified by their crystal form (SHEARMAN, 1966) and by X-ray diffraction methods.
According to Dickey (1968), the rate of evaporation in the Gulf of Kara Bogaz is such that almost four times as much total salt is precipitated annually in the gulf as is carried into the Caspian Sea by rivers. The Caspian Sea is annually becoming less salty. On a smaller scale, gypsum and halite are also being deposited in the narrow estuary of Bocana de Virrila on the northern coast of Peru (Morris and Dickey, 1957).

Salt deposition on the same scale as that of past geologic periods (Cambrian to Tertiary) is not seen today. The writer suggests that this is, perhaps, due to Pleistocene climatic changes that brought about geologically rapid changes in sea level associated with repeated growth and reduction in size of the polar ice caps. The world is probably only slowly returning to a climatic and sea level situation conducive to evaporite formation in peri-desert seas.

The salinity of normal sea water is taken to be about 36/oo. Although the Red Sea has a narrow marine connection with the open ocean at its southern end, salinities of about 42/oo have been recorded from the northern end (Lotze, 1957). A slightly increased restriction of the already narrow strait at the southern end of the sea would, because of its hot, almost rain-free climate and lack of rivers, convert it once more into the basin of evaporite sedimentation that it was during parts of the Tertiary. A similar restriction of the Strait of Hormuz would soon turn the Persian Gulf into an evaporite basin similar to that which existed in northwest Europe during the Upper Permian.

On a much smaller scale, deposition of thin, bedded evaporites in continental basins of inland drainage, have already been referred to (Fig.49, 50, 59, 60). Any single bed of evaporite in these cases is usually the result of complete evaporation to dryness of lake water. Its thickness is limited by the volume of water evaporated and the original salinity of the water prior to evaporation.

On shores and in lagoons along desert coasts, intertidal and supratidal evaporation results in the precipitation of gypsum between grains of the existing sediment, and the formation of a thin crust of salt. The salt crust is usually either dissolved by the next high tide or is deflated by the wind and redeposited elsewhere.

### RECENT ANHYDRITE AND CONTORTED BEDDING

Anhydrite has recently been described from the modern sediments of desert coasts (Shearman, 1963; Butler et al., 1965; Butler, 1969). On the sebkhas of the Trucial Coast, gypsum is the stable calcium sulphate mineral in contact with brines of chlorinities less than 145/oo, and anhydrite at chlorinities greater than 145/oo. Anhydrite will be hydrated into gypsum where brine chlorinity falls to less than 145/oo (Butler, 1969).

1See, for instance, Kozary et al. (1968)

Fig.104 shows a pit dug into the sediments of a coastal lagoon at Pisida, near the Libya/Tunisia border (Enclosure 4), that is cut off from the open sea during all but the highest storm tides. The sediment consists largely of quartz sand, halite and gypsum crystals and there is a thin crust of halite on the surface. The gypsum crystals are likely to be recognised in the subsurface as nodules of anhydrite in a sandy matrix (cf. Fig.113).

The contorted bedding seen near to the surface of a coastal sebkha in Fig.104 is not fully understood. There are several possible explanations for its presence. Two are given here; one, however, is particularly related to evaporites. Although the sediment was still wet from recent marine flooding at the time the pit was dug, there were traces of anhydrite in the sediment. This suggests the possibility that anhydrite had formed after gypsum above the water table (or from a brine with a chlorinity greater than 145/oo) during the hot dry summer season and that, because of local permeability barriers within the sediment (illite is also present), not all of it had been converted back into gypsum. There is a 38% volume reduction when gypsum alters to anhydrite. Volume changes connected with the alteration sequence gypsum–anhydrite–gypsum, may have caused lateral flow of wet sediment beneath a fairly hard and rigid surface crust. Another possible explanation is that sediment flow occurred because of changes in hydrostatic pressure due to various causes.

The dark colour of much of the sediment suggests that the environment was reducing. A core was taken from the sediment. Since exposure to the atmosphere,
all the sediment has become white. Potash feldspars (microcline) and traces of dolomite are also present in the sediment.

GYPSUM DUNES

The flooding of another lagoon south of Misurata, Libya, was witnessed by the writer. Storm waves were breaking over the barrier beach and flooded the dried-up lagoon behind. The influx of water allowed stromatolite-forming algae to start growing along the edges of the polygonally cracked lagoon floor (Fig. 105). Dunes along the landward side of the lagoon were composed almost entirely of gypsum grains derived, presumably, from the sediments of the lagoon during periods of deflation. The lagoon is dry for the greater part of the year. A small gypsum dune is seen in the background of Fig. 105. Farther along the lagoon they attain a height of at least 10 m. Gypsum dunes occur along the edges of the evaporite-producing Ranns of Kutch in India, and dunes built almost entirely of gypsum grains are also described by McKee (1966a) from White Sands National Park, New Mexico. The source of the gypsum in this case, however, was an inland "playa" known as "Lake Lucero". Cross-laminated gypsum of possible aeolian origin occurs in the Triassic Moenkopi Formation of Arizona and is illustrated by McKee (1954). Here, the gypsum appears to have been deposited in a near-coastal environment and may well be analogous to the dunes described from Libya and India.

Cross-bedded gypsum also occurs in the Jurassic Todilto Formation in New Mexico. It is interpreted by Tanner (1965) as having been formed by the wind, the gypsum in this case possibly being derived from a desert lake.

THE RANNS OF KUTCH

An area of more widespread deposition of evaporitic minerals is formed by the Ranns of Kutch on the west coast of India (see Fig. 109). Like the coastal sebkhas of the Persian Gulf, the surface of the Ranns of Kutch is at, or slightly above, sea level. During the three months of the Southwest Monsoon, storm tides, aided by the wind, force water from the Arabian Sea over the flat surface of the Ranns. Rainfall is fairly low, so that as the waters recede and evaporate, they leave behind a crust of halite, and gypsum crystals grow in the clays and sands. The few rivers (wadis) which flow into the eastern ends of the Ranns carry only limited fresh water and sediment for about one week during the monsoon. For the rest of the year they are dry. The result is that an area of about 30,000 km² (about the same size as The Netherlands) is subjected to annual flooding with the succeeding formation of evaporites.

Fig. 106, from Great Rann of Kutch, shows a thin (2-3 cm) salt crust some
Fig. 107. Wavy lamination in gypsiliferous argillaceous sands on the beach at Pachham Island, Great Rann of Kutch, India. (See Fig. 109.)

Fig. 108. Thin salt crust overlying horizontally bedded gypsiliferous sands and clays. Tidal estuary of Kori Creek, Lakhpat, Kutch, India. (See Fig. 109.)

Fig. 109. Sketch map of the Ranns of Kutch, India.

150 km from the open sea. The crust is buckled and thrust, probably by the forces of crystal growth of the salt. In the left background are low, scrub-covered dunes of the north end of Pachham Island. Wind deflation during the dry season has resulted in the incorporation of halite, gypsum flakes and one species of Foraminifera into the dune sand.

Fig. 107, from the shore of Pachham Island, shows wavy lamination which

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1 BOECK (1959, p. 22) ascribes the shape of salt polygons to horizontal expansion of the salt crust. "...the polygons are not due to some process of shrinking in the course of drying out..." He also has noticed thrusting of the edges of some polygons over those of others, when studying the salt flats of the Great Kavir of Central Iran.

2 WEDMAN (1964) has estimated that of the 30,000 tons of salt produced in Sebkra Sedjoomi, Tunisia, each year, 90% is removed by wind deflation.
is very reminiscent of the lamination seen in sediments of inland sebkhas (cf. Fig. 55, 62). It is not certain whether this lamination is the result of current action, gypsum crystal growth, or the formation of adhesion ripples, although the last mentioned is thought the most likely.

In a tidal estuary much nearer to the open sea, the horizontally bedded clays and sands are rich in gypsum crystals (Fig. 108). The surface of the sediment is covered by a thin halite crust as the water recedes and evaporates after high diurnal or spring tides.

In contrast to the coastal sebkhas of the Persian Gulf, the sediments of the Ranns of Kutch are poor in carbonate minerals. This is thought to reflect an influx of sediment from the Arabian Sea which, in the vicinity of the Indus Delta, is rich in fluvial sand and clays.

The evaporitic clays and sands found in the Ranns of Kutch have many similarities with the Triassic Keuper sediments of northwest Europe. Perhaps some of the Keuper evaporitic sediments were deposited under similar conditions of storm tide over broad flat areas at, or slightly above, normal high tide in an area of hot, arid climate.

Chapter 8

SUBSURFACE RECOGNITION OF DESERT SEDIMENTS

GENERAL

In the foregoing chapters, an attempt has been made to relate different desert sediments to their specific environments of deposition and to show some of the relationships that exist between these environments. The examples of ancient desert sediments that have been given so far all occur in outcrop, where recognition of the environment of deposition is sometimes relatively easy. The broader picture as seen in Enclosures 1–4 may help in envisaging possible palaeogeographic relationships of rocks in which their desert origins have already been recognised. Now, however, we must consider in more detail those features of desert sediments that may be recognised from cores, drill cuttings and the individual grains of sand of which they are composed. The criteria for the recognition of subsurface desert sediments are still essentially the same as in chapter 2. The scale on which these criteria must be recognised, however, is very small. Diagenetic changes may make the recognition of some criteria more difficult, and if they lead to the obliteration of primary depositional features, even impossible. Some criteria by which desert sediments may be recognised from cores, and more doubtfully from drill cuttings, are given in Table I.

Before the recognition of a desert environment of deposition can be inferred from a particular sequence of rocks, all the available petrographic, structural and stratigraphic data must, of course, be considered. For a well sequence, some of this evidence may come from lateral differences or correlations seen in other wells, or from inferences made from a regional study of the area.

SUN-DRIED SEDIMENT

Four pieces of core of roughly the same age and from the same area have been chosen by E. Oomkens to illustrate one aspect of sedimentation in a desert environment (Fig. 110–113). The slightly curled and cracked clay flakes in Fig. 110 indicate sub-aerial exposure and drying of water-transported sediment. This, with perhaps a conglomerate below, might suggest a wadi environment, but would not
prove it. It could also have been deposited by a river with a high annual variation in water level (cf. Fig.40).

EVAPORITE ENVIRONMENT

The wavy silt and sand laminae in Fig.111 suggest adhesion ripples formed by wind-blown sediment adhering to a damp surface. They could represent a temperate climate, and were, in fact, first described from the recent sediments of the north German coast (Reineck, 1955). The presence of small white specks of anhydrite, however, makes one think of the possibilities of an associated evaporitic environment—either coastal or inland. The interpretation of a desert environment of deposition is possible.

A similar analysis was made by Oomkens with Fig.112 and 113. The cracks in the clay flakes (Fig.112) have been filled in from above by (windblown?) sands and also injected from below by (wet?) sand in the form of a sand dyke that locally forced a layer of clay to arch upwards. Sand dykes have been recognised in inland sebkha environments (cf. Fig.56–58). The well-developed anhydrite nodules in the sand-laminated silt of Fig.113 confirm a slightly evaporitic environment. Evidence leading to a fuller environmental interpretation might be provided by cores from higher or lower in the sequence. These evaporitic sediments are likely to be overlain by dune sands, wadi conglomerates or a marine sequence.

The above example shows that a desert environment can be inferred from an association of sediments that does not include typically foresetted dune sands. It also emphasises that not all desert sediments are dune sands.

RECOGNITION OF DUNE SANDS IN CORES

Dune sands are sometimes easily recognised in a core (see Table 1). The charac-
teristic sharp differences in grain size between different laminae, and the clean, often uniformly dipping sand beds with dips of up to 34°, are highly suggestive. Some dune sands, however, have an almost uniform grain size, and may be very difficult to distinguish from a well-sorted beach sand. SHEPARD and YOUNG (1961)

### Table I

**SUBSURFACE RECOGNITION OF DESERT SEDIMENTARY ROCKS IN CORES**

<table>
<thead>
<tr>
<th>A. Wind-deposited sands</th>
</tr>
</thead>
<tbody>
<tr>
<td>1. Sequences of sands that may vary in thickness from a few centimetres to several hundred metres and whose laminae dip at angles from horizontal to 34° (after allowing for hole deviation or structural tilt). Dips may be of constant or multiple orientation (Fig. 114-117, 120-122).</td>
</tr>
<tr>
<td>2. Laminae commonly planar, but ripples occasionally seen on steeply-dipping foresets (Fig. 120).</td>
</tr>
<tr>
<td>3. Individual laminae well sorted, especially in finer grain sizes; sharp differences in maximum grain size between laminae common (Fig. 124-126).</td>
</tr>
<tr>
<td>4. Larger sand grains tend to be well rounded (Fig. 126).</td>
</tr>
<tr>
<td>5. Percentage of silt and clay generally well below 5%, or even absent (authigenic clay may modify this criterion, but if present, should be recognisable by X-ray diffraction analysis or in thin section).</td>
</tr>
<tr>
<td>6. Clay drapes very rare, and when present should be accompanied by evidence that it was water-laid (Fig. 114, 115, 117).</td>
</tr>
<tr>
<td>7. Quartz sands at shallow depths commonly friable or lightly cemented with haematite. Local discoloration of red haematite-coated grains to green or white is not uncommon (see also chapter 9).</td>
</tr>
<tr>
<td>8. Presence of adhesion ripples with associated increase in clay content and common presence of gypsum or anhydrite cement.</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>B. Water-laid sands</th>
</tr>
</thead>
<tbody>
<tr>
<td>1. Most sedimentary features similar to those of water-laid sediments from non-desert continental environments but modified by the presence of one or more of the following:</td>
</tr>
<tr>
<td>2. Common cement—calcite-cemented or locally cemented by gypsum or authigenic clay.</td>
</tr>
<tr>
<td>3. Many grains coated with haematite.</td>
</tr>
<tr>
<td>4. Conglomerates may be common, and sometimes with several cycles of deposition that lack a sand-sized fraction at the top of the cycle (deflation of the sand and silt).</td>
</tr>
<tr>
<td>5. Presence of mud-flow conglomerates (Fig. 118).</td>
</tr>
<tr>
<td>6. Sharp upward decrease in grain size (especially from sand to clay) indicating a rapid fall in water velocity.</td>
</tr>
<tr>
<td>7. Common presence of clay pebbles and curled clay flakes (Fig. 110).</td>
</tr>
<tr>
<td>8. Presence of mud cracks with sandy infill (Fig. 112).</td>
</tr>
<tr>
<td>9. Presence of sandstone dykes (perhaps implying interbedded aeolian sands) (Fig. 112).</td>
</tr>
<tr>
<td>10. Calcite cement stained with a solution of Alizarin red S and potassium ferricyanide in hydrochloric acid commonly stains red (had the cement formed in a reducing environment, the rock might have contained ferrous iron and would stain violet: EVANS, 1963).</td>
</tr>
</tbody>
</table>

Criteria marked + in right-hand column can also sometimes be recognised in drill cuttings (see remarks in text on p. 156).

1. This general principle is well known to helicopter pilots who fly in mountainous areas. The writer once made the mistake of camping opposite the mouth of a gorge cut through the Siwalik Hills on the northern edge of the Ganges Plain. In the late afternoon, very strong "upstream" winds prevented the cooking of the evening meal. At dawn, almost equally strong "downstream" winds upset the timing of a hot breakfast. In this area, the periods of calm air were around midday and midnight. In deserts, where the lack of vegetation permits much more rapid temperature changes, the times of calm air are shortly after sunrise and sunset.

and SCHOCK (1965) have presented evidence to show that there are only slight statistical differences between beach sands and sands of nearby dunes. In this case it may be possible to differentiate between them on the assumption that the grain size of beach sands tends to increase upwards, whereas the opposite is true of dune sands (see Fig. 66, 134-136). Ancient aeolian sands are also often more poorly cemented than other sands from the same sequence which were deposited in water.

Continuous coring through a thick sequence of sediments naturally gives the geologist the best chance of determining the environment of deposition of that sequence correctly. If the cores consist mainly of dune sand, the investigator will not only recognise the characteristics of individual laminae, but will be able to build up a picture of the changes in bedding attitude with changes in depth. He may even measure the dips and deduce palaeo-transport directions.

If cores have been taken at only rare intervals, these must be used for determining the characteristics of the sand—uniform bedding generally with no ripple structures or evidence of scouring or slumping; the bedding is either horizontal or dipping at angles of up to 34°; the orientation of the bedding is either uniform in direction or shows changes in direction; each set of laminae tends to show a slight upwards reduction in grain size.

The cores seen in Fig. 114 and 115 show both sub-horizontal and foresetted Permian aeolian sandstone. The aeolian sandstone in Fig. 114 overlies a fine sandy conglomerate (not shown) some 2 m below, that is interpreted as having been deposited by water. The dip of the bedding in the sandstone slowly increases from sub-horizontal, at its contact with the conglomerate, to about 17° when it is truncated by a second set of aeolian sandstone laminae dipping gently in an opposing direction (wind at right angles to the plane of the photo, or winds from two different directions?). This latter sandstone has a few larger granules incorporated in the lowermost laminae. Another sequence of fluvial sandstones and clays starts 1 m above the laminae seen at the top of the photo. By analogy with Recent desert areas, one assumes that here we have a wadi in which at least one of the local channels has been filled with wind-blown sand. When thin sequences of aeolian sands are associated with wadi sediments, care must be taken in assuming any regional palaeo-wind direction from the aeolian foresets. In the vicinity of hills or mountain ranges, irrespective of the regional prevailing wind direction, there is often a strong wind that blows "upstream" towards the hills during the day, and a milder wind that blows "downstream" at night. Fluvial transport directions are often opposite to apparent wind-transport directions, and accretion slopes along the...
protected banks of the wadi may dip at almost 90° to the local wind direction (i.e., towards the centre of the wadi). See the arguments applied to the formation of a seif dune. p.90 and Fig.73.

Fig.115, on the other hand, shows the central portion of a sequence of aeolian sandstone 6 m thick that started with sub-horizontal laminae at the base and ends with laminae dipping at over 20° at the top (not shown). It represents a very small portion of a more or less continuous aeolian sequence well over 100 m thick and is interpreted as representing dune sands of a sand sea. Each dune sequence usually starts at the base with horizontal or low-angle dips that increase in angle upwards. Often, the dip may be constant in both angle and orientation through several metres of vertical sequence; it may be at about 34°, thus implying slip-face conditions of deposition, or at any lesser slope. It is in major dune-sand sequences such as this that, from an analysis of the bedding styles and orientations, an interpretation of dune type can be made. Once this is decided, then the palaeo-wind direction can be worked out with reasonable accuracy.

A reconstruction of the history of sedimentation, as seen from the cores taken from a single well, may follow the following lines. The lowest cores show opposed fluviatile and aeolian transport directions. The fluviatile sediments can be assumed to have been transported away from a range of hills and the aeolian sands towards them. If, higher in the sequence, the attitudes of the more continuous dune bedding seen in the cores suggests that the wind came from another direction—perhaps at right angles to the earlier wind and water-transport directions—then this later direction probably coincides with the prevailing wind. If the cores have all been oriented with respect to the north, then directions become available for reconstructing the palaeogeography as suggested in Fig.116.

This situation of differing wind and water-transport directions is, perhaps, easier to follow if the southwest corner of Enclosure 1 is studied.

When the wadis are in flood, sediment is transported from the highlands of the Hadramaut towards the north. The Pleistocene to Early Recent gravels, now exposed in the feidjes (interdune corridors) of the southern edge of the Rub al Khali, probably show the same transport direction. The linear dunes of this part of the Rub al Khali are oriented east-northeast–west-southwest in response
to the dominant east-northeast sand-transporting winter wind of this area. According to BEYDOUN (1966), however, many of these dunes have slip-faces directed to the south. These slip-faces are probably formed by local winds that blow “upstream” towards the hills in response to convection currents that build up over the Hadramaut (see also Fig. 75).

An example of contrasting wadi and aeolian transport directions has been described from the Permian desert sediments of Devon, England, by LAMING (1966). See also Fig. 27 and 29.

Fig. 117 and 118 are of Triassic cores from Algeria. In Fig. 117, uniformly oriented steeply dipping foresets suggest an origin on a dune. A grain-size analysis, however, might be rather confusing, for 19% of the sediment would appear to have a grain size of less than 32 μ (Fig. 119). Dune sands should be almost free from silt and clay. The explanation is given by a microscope examination of the sandstone. The silt and clay is in the form of sand-sized particles composed of silty clay which were deposited on the dune as such. They are probably the deflation products of a nearby interdune (feidj or inland sebkha?) lightly cemented soil. Laboratory techniques involving water in a sedimentation balance\(^1\) resulted in the disintegration of these grains into their constituent particles (cf. Fig. 45, 132).

The mud-flow conglomerate seen in Fig. 118 is overlain by sharply contrasting foresets of a slightly pebbly sandstone. This sandstone is gradually replaced upwards by sandstone that possesses more typical aeolian characters. The presence of pebbles in the sandstone and an obviously water-laid conglomerate beneath, could lead to an interpretation in which a stream-flow origin is assumed for the sandstone; and yet, we have seen with Fig. 26 and 36, that the presence of pebbles in an aeolian sand can be explained. The sharp line of contact between the conglomerate and the overlying sandstone is possibly more easily explained if the sandstone is considered to be of aeolian origin and deposited at the foot of the avalanche slope of a dune. When undoubted aeolian characters develop upwards with no obvious lithological break, then an aeolian origin for the pebbly portion of the sandstone is also likely. Other evidence, including the nature of the laminae and the degree of sorting, could also affect the final interpretation.

\(^1\) See also PLANKEE (1962) and SENGUPTA and VENSTRA (1968).
DIFFICULTIES OF DEDUCING PALAEO-WIND DIRECTIONS FROM CORES

In order to stress the limitations imposed by the dimensions of a core, oriented "cores" have been cut from the lacquer peels of dune sands shown in Fig.120-122. The bedding seen on Fig.120 has almost unidirectional dips. The corresponding core has been given a hypothetical tectonic tilt which has almost removed the appearance of foresetted sands. The assumed original transport direction of the sediment is now much less certain. Without great care, the set of foresets, now apparently dipping to the right nearly half way up the core, could be mistaken for "climbing ripples" in a clean river sand (cf. Fig.43). Since the dips are uniformly oriented, it is impossible to deduce an accurate palaeo-wind direction (Fig.123A and B), because it cannot be known whether the dips were formed on a barchan or a self-like dune. The error in the deduced palaeo-wind direction could exceed 90°. The lacquer peel was actually made in the centre of a barchan dune. The horizontal surface accretion dips are not shown. The same uncertainty in interpretation will be found with a non-oriented core from a hole that had deviated from the vertical. With a hole deviation of 15°, apparent horizontal bedding could, in reality, dip in any direction. If the visible dip in the core from this deviated hole is 15°, the true dip of the bedding could vary from 0° to 30° with vastly different interpretations of environment of deposition within the desert and of possible palaeo-wind directions.

A THREE-DIMENSIONAL MODEL OF BARCHAN BEDDING

Fig.121 is of lacquer peels made at right angles to each other in the same pit dug into a barchan dune. A three-dimensional model of the bedding can be imagined by folding the two photos at right angles to each other. Note that for directional orientation, these photos should face outwards since they represent sections peeled off the inside walls of the pit.

Changes in the dip of the bedding are self-explanatory. The thin laminae that appear to stand out, do so because they are made up of the finer sand. When lacquer peels are made, the greater capillarity of the finer sand permits greater penetration of the lacquer. This should not be confused with the better permeability which exists in the coarser sand laminae. The few small burrows must not be confused with similar phenomena occurring in, say, estuarine sands and formed by burrowing animals. The burrow-like structures seen here were made by plant roots. The peel represents an area of limited dikaka.
The palaeo-wind direction can be deduced from the dip attitudes measured on the oriented cores. The distribution of dip attitudes (Fig. 123C) roughly fits the pattern to be expected for a barchan dune. High-angle dips oriented in the direction of the palaeo-wind are absent, and there is a tendency for lobate asymmetry. This is because the "core" was taken from the broad right-hand horn of the barchan where a tendency to develop a self-like pattern is present (Fig. 82). The higher dips are directed towards the slip face on the inner edge of the crescentic dune.

The lacquer peel and "core" seen in Fig. 122 are taken from a seif dune. Normally, a core taken on the flank of a seif will show almost unidirectional dips. As may be gathered from the opposed directions of dip (gently to the right, above; steeply to the left, below) a location near the crest of the seif may be assumed. This is partly confirmed by the rare 34° dip near the base of the core. It suggests the presence of an occasional slip-face which is sub-parallel to the predominant wind direction. The asymmetry in the distribution of the dip attitudes (Fig. 123D) also reflects a location to one side of the crest line.
It should be realised, however, that for instance, 10 small seif dunes with an inter-crest spacing of 100 m could be replaced upwards, in time, by 5 larger seifs with an inter-crest spacing of 200 m, the larger seifs probably representing a long period of stronger winds. The presence of these two stages in the development of dune systems may not be readily recognised in a vertical sequence. Dip measurements taken on an oriented core or series of cores, from well down the flank of the larger dune may include dips belonging to the opposite flank of an older and smaller dune.

Possible range in deduced palaeo-wind directions taken from the almost unidirectional dip attitudes measured on the oriented core Fig. 120 (after correction for tectonic tilt).

A. Assuming the dips to be representative of a seif dune. Wind from N 265° E ± 37°.
B. Assuming the dips to be representative of a barchan dune. Wind from N 210° E ± 45°.

Note: The lacquer peel was actually made on a barchan dune formed by a wind which blew from about SSW (N 205° E).

Palaeo-wind direction deduced from dip attitudes measured on the oriented cores figure 121 (reconstructed 3-dimensionally). The distribution of dip attitudes fits a pattern comparable to that found in barchan dunes. Deduced palaeo-wind from ca. N 190° E.

Note: These lacquer peels were made on the broad horn of a barchan where dip attitudes are not unidirectional.

Palaeo-wind direction deduced from dip attitudes measured on the oriented core figure 122 (reconstructed 3-dimensionally). The distribution of dip attitudes fits a pattern comparable to that found in seif dunes. Deduced palaeo-wind from ca. N 310° E.

Note: The lacquer peel was made on the flank of a seif dune not far from the crest line. The dune was formed by the Shamal which blows from the NW.
Fossil dune bedding is rarely as simple as the examples given here. With the present state of our knowledge, it should be sufficient to recognize that the sediment is aeolian sand, that the dune is, perhaps, of barchan or seif type, and if oriented cores have been taken, that the palaeo-wind probably blew in a certain approximate direction.

**PERMEABILITY IN DUNE SANDS**

In order to make lacquer peels of dune sands in the arid climate of a desert, it was found necessary to first wet the sand with water so that the sub-vertical walls of a pit dug into the dune would remain standing during the process of peel making. It was noticed that owing to increased capillarity, the water travelled further along the fine grained laminae than along those of coarser grain. It was also noticed that vertical penetration of water through horizontal sand laminae was very poor. In fossil dune sands, the finer grained laminae tend to be better cemented than coarser-grained laminae. It is suggested, therefore, that in an oil-bearing fossil dune sand, these finer sands will, because of their greater cementation and capillary action, act as semi-permeable barriers to oil flow; the better permeability will be found parallel to these lamellae in the coarser grained sands. It follows, therefore, that an oil field producing from what is essentially a barchan-type dune complex, may have different production characteristics from one producing from a semi-like reservoir because of differences in permeability distribution related to the bedding styles found in the two dune types.

**DRILL CUTTINGS**

If only drill cuttings are available for examination, the problem of recognizing desert lithologies is naturally increased. Comparison of different cuttings may give some indication of the degree of sorting within sand lamellae. The roundness of larger sand grains and the presence or otherwise of frosting on the surfaces of the grains (provided that the cuttings have not been cemented with calcite—see p.166) may give some indication as to whether or not the sediments could have been deposited in a desert, and if so, whether they could include dune sands. The presence of Foraminifera or other calcareous skeletal fragments need not preclude the possibility that the environment of final deposition was that of a desert. As pointed out on p.127, Foraminifera have been recognized in dune sands many tens of kilometres from the coast, and have also been found in the fluvial sands of wadis. They are typically abraded and normally free of signs of boring by algae.

A clean sandstone that is cemented with anhydrite may lead to an interpretation that the sands were deposited in an area where the ground water was rich in Ca$^{2+}$ and SO$_4^{2-}$ ions. Here also, a desert environment of deposition might be suspected. If the sandstone chips contain no clay particles, then the presence of dune sands is also a possibility.

With such a series of "suspicions" concerning the possible environment of deposition, one would naturally wish to core the interval if another well is to be drilled nearby. Support may also be found in the interpretation of well logs. This book is not the place for discussing their possible uses for interpretation of lithology, but it is perhaps pertinent to refer to the dip meter and the possibility it offers for distinguishing between sands deposited on dunes and sands deposited in other environments (see Gilreath and Maricelli, 1964, and Campbell, 1968).

**BEDDING AND SORTING IN DUNE SANDS**

It has been previously stated that individual laminae are well sorted in a dune sand. This is illustrated in Fig.124 by a sandstone of Permian age. It is typical of what can be seen through a microscope of the surface of an unpolished piece of sandstone. Each lamina is really the thickness of one grain. A group of laminae may contain grains of roughly the same size (well sorted). Alternating laminae (or groups of laminae) can differ in grain size from say 100 µ to 1,000 µ thus giving a high (poor) sorting coefficient for a bulk sample. Grain-size differences are not usually so great, however. The coarse white patches (in Fig.124) consist of kaolinite, altered post-depositionally from what are presumed to have been sand-sized grains of alkali feldspar.

Alteration of groups of laminae of markedly different grain sizes is characteristic of some dune sands. This is illustrated in Fig.124–126 where groups of laminae rich in coarse grains alternate with those that are much finer. By contrast, some dune sands are of remarkably uniform grain size. This presumably occurs where either the source area provides grains of uniform size or, more likely, where the source area is far distant from the site of deposition and original variations in grain size have been selectively reduced during transport.

Fine sand and silt and even clay particles can become trapped in the interstices between larger grains. By contrast, large individual grains are not normally found in the middle of fine grains unless they are relics of temporary deflation. It follows, therefore, that sorting is usually better in bulk samples of fine-grained dune sands than in coarse-grained dune sands. This can be seen more clearly in the photomicrographs made from thin sections of Permian aeolian sandstone (Fig. 125, 126) and in the grain-size distribution graphs of modern dune sand (Fig.66). The sandstones illustrated in Fig.124 and 125 came from the locality shown in Fig.81. The sandstone seen in Fig.126 is from the locality shown in Fig.79.
THIN SECTIONS OF CARBONATE DUNE SANDS

The difficulties of recognising carbonate dune sands in outcrop and core has already been mentioned. It is even more difficult to recognise the aeolian origin of a carbonate dune sand from cuttings. The individual grains can consist of well-preserved Foraminifera, Bryozoa and other skeletal fragments (Fig. 127), which give little indication of their aeolian mode of deposition. If cementation, leaching and other diagenetic changes have not gone too far, it may be possible for a palaeontologist to recognise a surface polish on the Foraminifera. This could be diagnostic if a marine limestone of similar age and depth of burial were available for comparison, but the chances are that the sediment would be interpreted as wholly marine. The thin section (Fig. 127) was made from a sample taken from the outcrop shown in Fig. 102. It contains about 5% scattered quartz grains.

The grains seen in Fig. 128 consist of a mixture of quartz and Foraminifera. The range in grain size is much greater than that seen in Fig. 127. This Pleistocene example of a quartz-carbonate dune sand from Kutch, India (for location, see Fig. 109) was found about 50 km from the present coast. The carbonate grains are thought to have been transported inland by the wind when a broad expanse of the present continental shelf in the Gulf of Kutch became exposed during a glacial period of lower sea level; the quartz could have been derived more locally. It is assumed to be a correlative of the purely calcareous Porbander Limestone (Miliolite) found on the southwest coast of Kathiawar, India (Wadia 1957, p. 409).

Fig. 129 shows typical carbonate grains taken from a dune 3 km inland from the coast at Dubai (see Enclosure 2). The grains, which are well rounded and polished, include occasional recognisable Foraminifera such as the specimen near the right-hand edge of the figure.

DOLOMITISED CARBONATE DUNE SANDS

Recognition of carbonate dune sands which have been diagenetically altered may well be impossible. Ordinary cementation with associated leaching of some grains can make it difficult to recognise an aeolian origin in a purely carbonate rock (Fig. 127). After dolomitisation, however, there may be no characters left which indicate aeolian deposition. Fig. 130 is a photomicrograph of a thin section made

1 The term "miliolite" was first used by Carter (1849) when describing granular calcareous rocks from the coast of southeast Arabia, Sind, Kutch and Kathiawar that contain ooliths and miliolids. Evans (1900) and Chapman (1900) produced arguments to support an aeolian origin for many of these miliolite occurrences, including some on the islands of the Persian Gulf. Pilgrim (1908) also assumed an aeolian origin for the miliolite of the Persian Gulf and that of Kathiawar. The term is still used in India and the Persian Gulf, but its aeolian character is not always recognised (see, for instance, Holm, 1960, p. 1377). The Persian Gulf "miliolite" is often rich in ooliths, Foraminifera and other skeletal debris.
Fig. 125. GL.143. Photomicrograph of Permian dune sand illustrating bedding and sorting. Southwest Scotland (location seen in Fig. 81). Plane polarised light.

Fig. 126. GL.241. Photomicrograph of Permian Yellow Sands illustrating bedding and sorting in dune sand. Durham, England (location seen in Fig. 79). Plane polarised light.
Fig. 127. GL. 69. Photomicrograph of Pleistocene (pre-Roman) cemented carbonate dune sand. Zanzur, West of Tripoli, Libya (location seen in Fig. 102). Plane polarised light.

Fig. 128. GL. 360. Photomicrograph illustrating alternations of quartz and carbonate dune sand. Pleistocene, Kutch, India (for location see Fig. 109). Plane polarised light.
from the dolomite seen in Fig.103. Its aeolian origin is suspected only on the basis of its dune-like bedding. It is unlikely that this would be recognised in a core, and would be impossible, with present techniques, to determine from drill cuttings. In the subsurface, the horizon would be described as a dolomite and interpreted as of marine origin.

FROSTING OF QUARTZ GRAINS IN A DESERT

The well-rounded sand grains that are considered by many authors as typical of dune sands acquire their shape after long-distance aeolian transport and abrasion (see Fig.131). Well-rounded grains are, however, usually those of larger diameter. With decreasing grain diameter, a point is reached - TANNER (1956) places this point at about 100 μ - where there is a change in degree of angularity. Either the saltating grain is so light that it is incapable of self-abrasion on impact with other grains, or it is so small that it falls between larger grains and is thus protected from the abrasive action of other saltating grains. For any particular grain size above a certain minimum limit, the degree of roundness will depend essentially upon the distance of travel; that is, upon the number of impacts with other grains. Because of its greater mass (and therefore, potentially, greater kinetic energy) a larger grain should achieve a certain degree of roundness in a shorter travel distance (fewer saltations) than is required by a smaller grain. KUENEN (1960) has carried out some experiments on aeolian abrasion, but it is still not possible to estimate the distance which any particular grain has travelled, since its history prior to its aeolian existence will not be known in sufficient detail. The roundness of larger quartz grains may be indicative of an aeolian environment, but large quartz grains also become rounded in other environments (see, for example, INMAN et al., 1966, p.800). Dune sands, however, also possess many angular grains that are derived from larger particles as the result of insolation (see Fig.132-136).

FROSTING OF QUARTZ GRAINS IN A DESERT

The large grains seen in Fig.131 are frosted as well as rounded. They have possibly been reworked from older Tertiary dune sands. The frosted appearance, typical of many aeolian sands, appears under the normal optical microscope to result from light diffraction associated with small pits on the surface of the grain. KUENEN and PERDOK (1962) claim that the frosting is not the result of pits made by the impact

1 It is assumed here that the grains under discussion became rounded solely as the result of aeolian transport and abrasion. Many grains that occur in dune sands, however, have probably also had an earlier history involving transport and rounding in entirely different environments (e.g., fluvial or marine).
I have acquired characteristic markings in different geologic environments: glacial and fluvo-glacial, rivers and beaches; aeolian sands from both hot deserts and non-desert areas; sands whose surfaces are chemically etched or altered by authigenic overgrowths (Bramer, 1965; Krinsley and Funnel, 1965; Waugh, 1965; Soutendam, 1967; Wolfe, 1967; Krinsley and Donahue, 1968a, b).

This type of study is still in its infancy; the reasons for the development of certain markings on the surface of grains is still under active discussion (see, for instance, Soutendam, 1967, and discussion by Krinsley and Donahue, 1968b). Further detailed studies of this type may eventually reveal patterns of surface markings on quartz grains that conform to particular climatic and depositional environments. This, in turn, may lead to a better understanding of the diagenetic changes (quartz overgrowths as well as corrosion) that can alter the surfaces of grains after burial.

**Frosting and Calcite Corrosion of Quartz Grains**

Frosted quartz grains are typical of many modern dune sands. If, however, the quartz grains have been cemented by calcite in a fossil sediment, the surface of the grains are corroded by the calcite to give the grain a frosted appearance (Alimen, 1944; Walker, 1957, 1960). Under the optical microscope these corrosion pits appear as light diffraction rings similar to, but often bigger than, those associated with the frosting of Recent dune sands. In the early stages of corrosion, however, the pits will be about the same size as those associated with desert frosting. It may be possible to recognise the differences between desert frosting and calcite corrosion by use of an electron microscope. Without such means of differentiation, it is at present very dangerous to use the frosting of a carbonate-cemented quartz sand as evidence for a desert environment of sedimentation. Electron-microscope investigations may help to resolve this problem. Excellent electron-microscope photographs have been published of the surfaces of quartz grains that are thought to have been deflated from nearby interdune desert-soil crusts (cf. Figs. I.117, I.119).

A finer fraction, from the same locality, is formed from almost equal amounts of carbonate grains and quartz together with grains of heavy minerals. The grains tend to be more angular than those in the coarser fraction. The “shiny” surfaces on many of the quartz grains are unfrosted conchoidal-fracture surfaces. This lack of frosting on the surfaces of grains of fine angular aeolian sand has also been noted by Snead and Friishman (1968, p.1673). The angularity of aeolian sand grains increases with a decrease in grain size. The sand grains illustrated in Fig. I.32 and I.33 are from the same locality as the lacquer peel and core, Fig. I.20.

**Frost-Free Conchoidal Fractures on Desert Quartz Sand and Relict Desert Soils**

One grain in Fig. I.31 (top, right of centre) is about to calve a flat flake of quartz along a fracture formed by the effects of insolation. The conchoidally fractured surface which results will not be frosted. The elongate grain below it has earlier lost a similar flake. Its fracture surface has already undergone further abrasion and frosting.

Most of the sand grains seen in Fig. I.32 have probably not suffered extensive aeolian transport. A few quartz and carbonate grains are very well rounded and almost spherical, but some are angular. Brown, irregularly shaped grains form almost 50% of this size fraction. They consist of silt, or fine sand-sized particles cemented together by an earthy looking brown oxide of iron. They are thought to have been deflated from nearby interdune desert-soil crusts (cf. Figs. I.117, I.119).

**Dune Sand from Deflation Plain to Top of Dune**

It has already been remarked that the grain size of dune sand tends to diminish...
Fig. 132. GL.426. Rounded to angular grains of quartz and carbonate with irregular iron-cemented aggregates representing former interdune desert soil. 480-500 μ. Bandah, Rajasthan Desert, India. Reflected light.

Fig. 133. GL.426. Sub-rounded to angular quartz, carbonate and heavy-mineral grains. Many grains are partly frosted, with unfrosted conchoidal-fracture surfaces. 250-297 μ. Bandah, Rajasthan Desert, India. Reflected light.

Fig. 134. Sample taken from a dune at 27.5 m above the deflation plain (cf. Fig. 66). Locality 92, near Barik, Oman. GL.559. 53-220 μ. σ = 0.22. Reflected light.

Fig. 135. Sample taken from a dune at 1.5 m above the deflation plain (cf. Fig. 66). Locality 92, near Barik, Oman. GL.555, 53-260 μ. σ = 0.36. Reflected light.
from bottom to top of the dune. The three examples of sand shown on Fig.134-136 are comparable to three of the grain-size distribution curves given in Fig.66. It will be noted that there are several angular grains in the example taken from the top of the dune (Fig.134). As with the examples from India (Fig.132, 133), many of the sand grains have only partial frosting of their surfaces. At the foot of the dune (Fig.136), locally derived deflation products of the limestone interdune area make up the bulk of the sand. They resemble in form the similarly derived sand seen in Fig.6. The presence of large angular grains at the base of an undoubted aeolian sequence could therefore suggest that they represent the products of a nearby deflation surface.

POSSIBLE PRESERVATION OF ENVIRONMENTAL HISTORY ON A QUARTZ GRAIN

Fluvial transport is slow to polish the surface of an already frosted grain. Grains of wadi sand are usually just as frosted as grains of aeolian sand. This is not surprising since much of the wadi sand has probably been airborne at some stage in its history. As KUENEN and PERDOCK (1962) have already pointed out, however, it may be that sub-aerial corrosion in a desert climate is a more important
cause of frosting than aeolian transport. Fig. 137 is of a river sand from the Niger Delta. Although the surfaces of the grains are fairly well polished, microscope examination shows that they have relict markings reminiscent of desert frosting. This might imply an earlier history of aeolian transport in the Sahara Desert, all traces of which have not been removed by its subsequent fluviatile transport, or the markings may reflect an earlier period of calcite cementation. The environment of final deposition was fluvial, in a hot, humid climate. For the present, one can only speculate as to the reasons for the markings. The possibility of recognition of relict environmental features has recently been presented by KRINSLEY and FUNNEL (1965) and KRINSLEY and DONAHUE (1968a) in papers describing electron-microscope photomicrographs of the surfaces of some quartz grains. CAILLIEUX (1937) believes that relict peri-glacial aeolian features can be recognised in the present North Sea sands.

There are few criteria that can be used as direct evidence of a fossil desert environment when only drill cuttings are used. In the absence of bedding features, well-rounded, frosted sand grains are perhaps the best indicator of the environment (Fig. 138, 139), but the possibility that these grains have been reworked into a non-desert (marine?) environment must not be overlooked. At present, frosting on calcite-cemented quartz grains cannot be accepted as evidence of a desert environment.

THE SIGNIFICANCE OF RED-STAINED QUARTZ GRAINS

The red iron oxide coating found on quartz grains is often accepted as evidence of deposition in a continental environment. Recognition of this iron oxide on the grains can be taken as additional evidence in support of a hot arid or seasonally wet climate. Here again, however, later reworking into a marine environment need not result in the removal of all the red coating. Grains that are covered with patches of iron pyrites altered from haematite indicate a colour change caused by an alteration from an oxidising to a more reducing environment. This could be caused by reworking into a marine environment, a change in the ground-water chemistry, or even, perhaps in some instances, the migration through the sands of undersaturated hydrocarbons. The whole question of the significance of red beds in desert sediments is dealt with in some detail in the following chapter.

Chapter 9

THE RELATION BETWEEN RED BEDS AND DESERTS

INTRODUCTION

No account of the various aspects of desert sedimentation would be complete without considering the significance of red beds.

Since the beginning of the century there has been much controversy concerning the interpretation of red beds. The only subjects on which all geologists agree are that the cause of the colour in red beds is the presence of red-stained clays and a coating of the red oxide of iron (haematite) on grains of sand and silt, and that most red bed sequences are terrestrial in origin.

The conditions under which the red pigment formed and then coloured grains of sand or particles of clay red are assumed by various authors to be related to different environments ranging from hot, humid tropics to hot, arid deserts. Some geologists believe that the red pigment is detrital; others believe that it forms penecontemporaneously in situ. Still others point out that some beds appear to have been reddened long after deposition. The history of this controversy over the origin of red beds, which appears to have been gathering in momentum since the late 1940's, is outlined in chronological order in Table II.

In the following pages, it is hoped to establish the probable conditions under which sediments—especially desert sediments—can acquire a red colour, and also to consider other factors related to the formation or loss of colour in sediments that, apparently, were deposited in similar environments.

HOT, WET TROPICAL ORIGIN FOR RED SEDIMENTS

WALTHER (1900) thought that desert sediments inherited the colour of the rocks from which they were derived. If desert sediments were red, then they were derived from a source that became reddened in a hot, humid climate associated with the formation of lateritic soils. He, like DORSEY (1926) and RAYMOND (1927), was impressed by the common yellow and dun colours of the desert. These latter geologists concluded that red colour is not produced in Recent deserts.

GÈZE (1947) and WAHLSTROM (1948) state correctly that red laterite soils form in hot, humid climates. Pigment derived from such areas should stain red a sediment that is otherwise non-stained. This is the argument followed by KRYNINE
### Table II

**History of the Controversy Over the Origin of Red Beds**

<table>
<thead>
<tr>
<th>Year</th>
<th>Author</th>
<th>Nature of hypothesis</th>
</tr>
</thead>
<tbody>
<tr>
<td>1900</td>
<td>WALTHER</td>
<td>Red beds formed in a hot, humid climate and preserved in an arid desert climate.</td>
</tr>
<tr>
<td>1908</td>
<td>BARRELL</td>
<td>Red beds formed in a hot climate with alternate wet and dry seasons.</td>
</tr>
<tr>
<td>1916</td>
<td>DORSEY</td>
<td>Secondary reddening of previously deposited beds by penetration of oxidising conditions resulting from overlying arid desert.</td>
</tr>
<tr>
<td>1926</td>
<td>BAYLEY</td>
<td>Red colour is not produced in (dune sands of) Recent deserts.</td>
</tr>
<tr>
<td>1926</td>
<td>DORSEY</td>
<td>Red colour is not produced in (dune sands of) Recent deserts.</td>
</tr>
<tr>
<td>1927</td>
<td>RAYMOND</td>
<td>Red beds formed in a hot, humid climate and preserved in an arid desert climate.</td>
</tr>
<tr>
<td>1937</td>
<td>DE LAPPARENT</td>
<td>Red colour of ancient sedimentary rocks is the result of intermittent humidity in a desert environment.</td>
</tr>
<tr>
<td>1937</td>
<td>BOURCART</td>
<td>Red colour of ancient sedimentary rocks is the result of intermittent humidity in a desert environment.</td>
</tr>
<tr>
<td>1938</td>
<td></td>
<td>Intra-stratal alteration of iron-bearing minerals as a source of iron for haematite pigment in red beds.</td>
</tr>
<tr>
<td>1947</td>
<td>GEZE</td>
<td>Intra-stratal alteration of iron-bearing minerals as a source of iron for haematite pigment in red beds.</td>
</tr>
<tr>
<td>1948</td>
<td>WAWLSMANN</td>
<td>Red colour is produced during laterite formation in hot, humid areas.</td>
</tr>
<tr>
<td>1949</td>
<td>ROBB</td>
<td>Red colour is produced during laterite formation in hot, humid areas.</td>
</tr>
<tr>
<td>1949</td>
<td>KRYNINE</td>
<td>Red colour is produced during laterite formation in hot, humid areas.</td>
</tr>
<tr>
<td>1949</td>
<td></td>
<td>Intra-stratal alteration of iron-bearing minerals as a source of iron for haematite pigment in red beds.</td>
</tr>
<tr>
<td>1953</td>
<td>CHUBERT</td>
<td>Intra-stratal alteration of iron-bearing minerals as a source of iron for haematite pigment in red beds.</td>
</tr>
<tr>
<td>1953</td>
<td>DUNHAM</td>
<td>Intra-stratal alteration of iron-bearing minerals as a source of iron for haematite pigment in red beds.</td>
</tr>
<tr>
<td>1953</td>
<td>TROTTER</td>
<td>Intra-stratal alteration of iron-bearing minerals as a source of iron for haematite pigment in red beds.</td>
</tr>
<tr>
<td>1954</td>
<td>MILLER and FOLK</td>
<td>Intra-stratal alteration of iron-bearing minerals as source of red colour. Pigment lacking at contact point between grains, therefore in situ reddening.</td>
</tr>
<tr>
<td>1957</td>
<td>DUNBAR and ROGERS</td>
<td>Source area will determine whether or not a sediment will be initially red. Environmental conditions in basin of deposition determine whether or not it remains red.</td>
</tr>
</tbody>
</table>

**Table II (continued)**

<table>
<thead>
<tr>
<th>Year</th>
<th>Author</th>
<th>Nature of hypothesis</th>
</tr>
</thead>
<tbody>
<tr>
<td>1960</td>
<td>BUNDELL and MOORE</td>
<td>Red Carboniferous strata of South Wales secondarily reddened by oxidising conditions of overlying Permian desert.</td>
</tr>
<tr>
<td>1960</td>
<td>MYKURA</td>
<td>Similar secondary reddening of Carboniferous strata in southwest Scotland.</td>
</tr>
<tr>
<td>1961</td>
<td>VAN HOUTEN</td>
<td>(a) Ferric oxides originated in red lateritic upland soils in a tropical or subtropical climate.</td>
</tr>
<tr>
<td></td>
<td></td>
<td>(b) Oxidising conditions necessary in place of deposition -- a drier climate, possibly local desert conditions, but not in great desert like those of today.</td>
</tr>
<tr>
<td></td>
<td></td>
<td>(c) Possibility of age of hydrohaematite to haematite after deposition.</td>
</tr>
<tr>
<td>1963</td>
<td>WALKER</td>
<td>In situ formation of red beds in an arid to semi-arid climate.</td>
</tr>
<tr>
<td>1964</td>
<td>VAN HOUTEN</td>
<td>(a) Most of ferric oxide pigment found in red beds brought from source area in colloidal suspension or in solution and precipitated as cement.</td>
</tr>
<tr>
<td></td>
<td></td>
<td>(b) Some pigment supplied by alteration of iron-bearing minerals.</td>
</tr>
<tr>
<td></td>
<td></td>
<td>(c) Diagenesis has played a significant role in modifying components inherited from source area.</td>
</tr>
<tr>
<td>1965</td>
<td>MILLOT</td>
<td>Red beds deposited in hot, seasonally wet climate. Inferred from Permian-Triassic rocks of the Vosges, France.</td>
</tr>
<tr>
<td>1965</td>
<td>DOWNING and SQUIRREL</td>
<td>Red colour in Upper Carboniferous strata of South Wales contemporaneous or penecontemporaneous and either: (a) derived from source areas where red soils or red rocks existed; or (b) resulted from subaerial processes that produced red soils from grey sediment in situ by weathering of iron-bearing minerals.</td>
</tr>
<tr>
<td>1965</td>
<td>ARCHER</td>
<td>Formation of red beds and other associated phenomena in South Wales resulting from penecontemporaneous lowering of water table.</td>
</tr>
<tr>
<td>1966</td>
<td>FRIEND</td>
<td>(a) Redness develops in situ and does not result directly from presence of red soils in source areas.</td>
</tr>
<tr>
<td></td>
<td></td>
<td>(b) Alteration of red to non-red occurs post-depositionally as result of differences in oxidation-reduction potential, which is determined by position of water table—applied to water-laid Devonian sediments.</td>
</tr>
<tr>
<td>1967</td>
<td>WALKER</td>
<td>Red colouration in desert sediments is due to presence of haematite that forms in sediments after deposition. Critical factors that control formation of haematite are: (a) Presence of iron-bearing grains in original sediment.</td>
</tr>
<tr>
<td></td>
<td></td>
<td>(b) Post-depositional conditions favouring intra-stratal alteration of iron-bearing grains.</td>
</tr>
<tr>
<td></td>
<td></td>
<td>(c) Eh-pH interstitial environment that favours formation of ferric oxide (probably as limonite in initial stages of formation).</td>
</tr>
<tr>
<td></td>
<td></td>
<td>(d) Absence of subsequent reduction of ferric iron.</td>
</tr>
<tr>
<td></td>
<td></td>
<td>(e) Enough time for alteration of iron-bearing minerals and formation of haematite from limonite.</td>
</tr>
<tr>
<td></td>
<td></td>
<td>(f) Possibly an elevated temperature.</td>
</tr>
</tbody>
</table>
(1949; (a) in Table II). DUNHAM (1953), DUNBAR and ROGERS (1957) and VAN HOUTEN (1961, 1964). Deposition of organic-rich sediment that has been stained red by lateritic muds can, however, result in reduction of the sediment to a greyish or even a grey colour (see, for example, core description by NORTA, 1956, p.23 etc., of the sediments off the mouth of the Orinoco River).

That red continental sediments can be reduced when transported into a marine environment is supported by BORCHERT (1965). In a discussion on marine sedimentary iron ores¹, BORCHERT (1965, p.180) assumed from earlier arguments "that most of the source material for marine iron ores originated from the continents, and that the separation of iron from this detrital sial material took place principally not on the continents, but rather in sea basins by the leaching of ferrous iron from clay minerals". Later, he says: "In both the continental and the marine environments, iron can be dissolved from the sialic detritus only under conditions in which ferrous solutions are stable (i.e., the low pH and Eh values of the CO₂-zone)." "A CO₂-zone can develop in certain confined sea basins in which...stratified water masses can develop." This condition of stratified water masses is found off the mouth of the Orinoco River where the river water flows over the more dense sea water. "In the main CO₂-zone there exists a moderately strong reducing environment (Eh = +0.05-0.2 V and pH 6-7.5) and iron may be dissolved from the bottom sediments" (Borchert, 1965, p.183 and fig.9). Ferric minerals carried in suspension in the river water will become reduced to the ferrous state as the clay particles flocculate on contact with the sea water and settle down in the underlying CO₂-zone.

Further evidence is provided by HINZE and MEISCHNER (1968) who show that although red detritus from the Istrian Peninsula is carried into the Adriatic Sea, red marine sediments do not result. The water at a depth of 30 m has a seasonal fluctuation in temperature of between 10° and 19° C, in salinity of between 37°o and 38°o, and in oxygen saturation of between 70% and 100%. Cores show a transition with depth of penetration into the sediment from red-brown (at the surface near the source area), through brown-grey to grey (at a depth of a few centimetres below the surface of the open-sea floor). Hinz and Meischner believe that the haematite is partly destroyed during fluviatile transportation. Because of burrowing, these ferric hydroxides that reach the sea bottom are gradually carried down into the zone of negative Eh and low pH, where they are reduced and partly fixed as siderite and pyrite.

¹ Throughout this discussion, the writer assumes that the colour of red beds is very close to 5R or 10R on the Geological Society of America Rock Color Chart (Goddard, 1951), although sediments that have other colours (yellow, brown, green etc.) may be related to red beds in some aspects of the process of redening. The iron ores that form in the marine environment discussed by Borchert (1965) are mainly yellow-brown limonite and brown siderite; they cannot be considered as representing a marine equivalent of the continental red beds discussed in the remainder of this chapter.
wet and dry seasons to account for the red colouration of these sediments of the New Red Sandstone. He points out that humidity is necessary for the hydrolysis of the ferromagnesian silicates that must precede the liberation of iron, which, during the dry season, can be converted into the sesquioxides of iron.

The origin of red and non-red Devonian fluviatile sediments is discussed by FRIEND (1966). He considers that the reason for the differences between red and non-red rock sequences is the presence of fine-grained haematite, often of hexagonal crystal form, in the red sediments. He says (on p. 285): “The redness developed in situ, and did not result directly from the presence of red soils in the source areas.” Although the haematite is often intimately connected with clay minerals, the clays are not themselves a cause of the red colour, for the same type of clay minerals occur in non-red sediments.

Friend considers that the localised presence of organic material causes reducing conditions below the water table, which result in ferrous solutions. The ferrous–organic complexes so formed can be carried away in solution, thus causing a drop in the total iron content of the rock. Under strongly reducing conditions associated with the presence of organic material (and sulphur), pyrite may form. Localised green patches within the red beds are often associated with dark plant fragments. The formation of the green colour is also post-depositional, and follows the formation of red colour.

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### TABLE III

<table>
<thead>
<tr>
<th>Criteria for deducing palaeoclimates (From Millet, 1964)</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Climatic characters recognised in Permo-Triassic sediments of the Vosges</strong></td>
</tr>
<tr>
<td>Presence of polycrystalline grains</td>
</tr>
<tr>
<td>Alkali feldspars preserved</td>
</tr>
<tr>
<td>Quartz grains not corroded</td>
</tr>
<tr>
<td>Quartz grains not micro-fractured</td>
</tr>
<tr>
<td>Quartz grains ferruginous</td>
</tr>
<tr>
<td>Clay fraction essentially illite</td>
</tr>
<tr>
<td>Accidental presence of arid-type soils</td>
</tr>
</tbody>
</table>

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The association of some red beds with evaporites, and locally with dunes, underwent many earlier workers to consider red beds as the product of an arid or semi-arid environment. DUNBAR and ROGERS (1957), like WALKER (1900), were more impressed by the presence of dun and buff colours and the apparent lack of red sediments in Recent deserts. There is, however, mounting evidence to support a hypothesis that supposes post-depositional in situ reddening of sediments under suitable conditions. As we have just seen, FRIEND (1966) proposes in situ reddening for the Devonian fluviatile sediments of the Catskill Mountains. Both DE LAPPARENT (1937) and BOURCART (1937, 1938) considered the colour of red fossil sediments to be the result of intermittent humidity in a desert environment. They were supported by CHOUBERT (1953) who pointed out that, even in deserts, some humidity is necessary to produce the red stain. Moreover, the lack of pigment at the contact points between grains suggests reddening of the sediment after deposition (MILLER and FOLK, 1955).

The clearest reasons for an in-situ, post-depositional origin for the red colour of ancient desert sediments so far published are given by WALKER (1967a). They are based on a study of Pliocene to Recent sediments in the Sonoran Desert of Baja California, Mexico, and their Late Palaeozoic analogues from Colorado. He provides convincing evidence that the haematite pigment formed after deposition, in a hot arid or semi-arid environment, by tracing stages in the in-situ colour change from Recent non-red desert sediments to red Pliocene desert sediments.

Walker's interpretation is based essentially on the widespread presence of unstable iron-bearing minerals such as hornblende and biotite in the non-red Recent alluvium, and the presence of corroded iron-bearing minerals (e.g., hornblende extensively altered to montmorillonite clay; see Walker et al., 1967) with halos of haematite in red Pliocene desert sediments. The varying degrees of alteration of these iron-bearing minerals appear to be related to an increase in the presence of haematite and the increasing degree of redness of the sediments from the Recent, through the Pliocene to the Pleistocene. Time is needed for desert sediments to become red. Time is apparently also required for the haematite to become sufficiently ordered structurally to allow X-ray identification. The red pigment found in the Pleistocene alluvium seems to be some form of ferric oxide that is either amorphous to X-rays or is too poorly crystallised to give a diagnostic X-ray diffraction pattern.

In the Palaeozoic red beds of Colorado, Walker finds that biotite, with a red halo of haematite, is particularly abundant in the red mudstones and shaly sandstones. Hornblende, however, is almost completely lacking, in spite of the fact that the rocks from which the biotite and haematite are derived—again exposed to erosion—are rich in the mineral. Alteration of iron-bearing minerals as a source of iron for the haematite pigment found in red beds has also been advocated...
to various extents by Robb (1949), Miller and Folk (1955), Shotton (1956), Walker (1963) and Van Houten (1964, 1968). In addition, Walker and Honea (1969) point out that the clay fraction of desert soils contains an average of about 4.5% total iron, most of which is held in the clay-mineral lattices. Under favourable interstitial chemical conditions, the iron-bearing clay should undergo post-depositional alteration and yield additional iron which will ultimately also form haematite pigment.

Walker (1967a) lists six critical factors that he believes control the formation of red pigment in the Baja California beds. They are given in Table II. He includes his last factor, “possibly an elevated temperature”, because the formation of the red pigments studied by him took place in a hot climate where the ground-water temperature is approximately 72 °F (22 °C). He is not certain, however, whether elevated temperatures are necessary for haematite formation in situations where long periods of ageing are involved. Other authors (by implication) also appear to favour a hot climate as an additional factor in the formation of red beds. Van Houten (1961) has suggested the possibility of “aging” brown haematite to red haematite after deposition—the time factor required by Walker’s (1967a) hypothesis for the conversion of limonite into haematite. This is supported by Berner (1969), who shows that goethite is unstable relative to haematite, and given sufficient time, the yellow-brown goethite could dehydrate to red haematite. Obviously following similar lines of thought, Norris and Norris (1961), Price (1962) and Norris (1969) suggest that the colour of dune sands is, in part, a measure of their age.

From studies undertaken both in desert and non-desert areas, Walker and Honea (1969, p.542) conclude that “special types of climate in the source areas are not essential for the formation of red beds”. “Essentially all sediments, regardless of type of parent material or type of source-area weathering conditions (moist or arid), contain enough iron to produce bright red sedimentary rocks if the interstitial environment, either during or subsequent to the time of deposition, favours the formation and preservation of iron oxide.” “The vital factor for the formation and preservation of red beds is the occurrence within the depositional basin of special interstitial conditions (for example, favourable Eh and pH) that favour the formation and preservation of haematite.”

ARE RED BEDS RED THROUGHOUT?

The above remarks by Walker and Honea (1969) and Walker’s (1967a, p.363) statement that the red colour will probably form after deposition in any climate where his six conditions are satisfied, are worthy of further study. Both the Old Red Sandstone of Scotland and the Tertiary Siwaliks of the Himalayan foothills are fluvialite and lacustrine sequences that contain red beds. Neither of these sequences, however, is so consistently red as many fossil dune sands. In the writer’s experience, probably less than 20% of the Old Red Sandstone and Siwaliks can be described as red.1 The remainder of these sequences—for which a hot climate and seasonal rainfall have been invoked to account for the conglomeratic and graded nature of the fluvialite sediments—consist of brown, yellow, green, grey or even white sediments. By contrast, the aeolian desert sediments of the Permo-Triassic New Red Sandstone of southwest Scotland, the Vale of Eden and the Midlands of England are virtually red throughout. It would appear that in the case of the dune sediments a more efficient process of reddening—or of preservation of the red colour—has been active than that affecting the fluvialite Old Red Sandstone and Siwaliks. Walker’s (1967a) criteria would appear to have been met fully in the one case, but not in the other. It is, perhaps, worthwhile considering the environments of deposition and preservation of other sediments that are, or might once have been red, in order to see whether there are other factors that help in the formation of red beds or cause their later alteration to other colours.

THE “BARREN RED MEASURES”

There are areas of Britain (southwest Scotland, north and central England, South Wales) where Carboniferous strata are red. Some of these sequences can be traced from normal productive Coal Measures into areas that are barren of coal seams. Bailey (1926) suggested that the Carboniferous sediments of Arran were reddened during the Permo-Triassic. This hypothesis is extended to the mainland of southwest Scotland by Mykura (1960) who considers that there, overlying arid desert during the Permo-Triassic. This hypothesis is extended to the mainland of southwest Scotland by Mykura (1960) who considers that there, overlying arid desert conditions caused the reddening of Upper Carboniferous strata to a maximum depth of 1,950 ft. He has also suggested (Fig.140) that, at the same time, the deeper Westphalian coal seams in the Mauchline Basin of Scotland were replaced by dolomitic limestone2 and the shallower ones by ironstone. The oxidation (and replacement)

1 Jones (1965) describes the three sequences of red beds from the Cretaceous Bima Sandstone of northern Nigeria. This sandstone is assumed to be of fluvialite and lacustrine origin and deposited in a climate which was seasonally wet. In one section, 9% of the beds were red, in another, 4% and in a third, only 3%. On the other hand, Arndt et al. (1962, Fig.72.2) show two sequences of the Catskill Formation of Pennsylvania, one of which has 68% “red beds” and the other 40% “red beds”. Included in the red beds, however, are rocks that are “grayish-red, pale brown, brownish-gray and yellowish-brown” (Arndt et al., 1962, p.353). In the Catskill sediments studied by Friend (1966, p.281), 73% of the thickness of coarse beds in a local section of 3 cyclothems are non-red and 98% of the fine beds are red. His non-red group includes grey, greenish-grey, olive-grey and brownish-grey. The reds fall in the range 5R, 10R, on the Geological Society of America Rock-Color Chart (Goddard, 1951). See also Gill (1951, pp.382-384) for description and colours of Siwalik sedimentary rocks.

2 Desert erosion of early New Red Sandstone volcanic lavas is thought to have given rise to a concentration of Ca2+ and Mg2+ ions in the ground water of the area. This is assumed to have been the reason why the coals were replaced by dolomitic limestone.
of the coal seams—which were originally formed in a tropical humid climate—is also believed to be associated with overlying arid desert conditions which resulted in a great lowering of the water table.\(^1\)

Sediments associated with the Carboniferous coals were probably not red when deposited, and even though red lateritic clays may originally have been deposited, the presence of so much organic matter in the environment of deposition must soon have caused the reduction of the oxides to non-red colours. For the reddening of these organically rich sediments, there must have been very strong post-depositional oxidising conditions. These are unlikely to have been met by the presence of oxygenated ground water at great depths below the water table (see Walker, 1967a, p.363, reference to Germanov et al., 1959). Something stronger was needed, but once oxidised, there was no more organic matter with which to bring about a return to reducing conditions.

Reddening of Carboniferous strata in Permo-Triassic time is also proposed for the north of England by Trotter (1953, 1954). The same hypothesis was applied to South Wales by Blundell and Moore (1960) but is opposed by Downing and Moore (1958) or moisture brought to the surface by capillary action (Opdyke, 1961). The most strongly coloured sediments are normally those of wadis, where the older wadi sands and gravels exhibit darkish-brown to pink shades, whilst the youngest wadi sediments are usually light brown or yellow. Pebbles on the deflated surfaces of wadi gravels that have not been covered by flood waters for many years—perhaps many decades—also show the dark brown or almost black colouration of desert varnish (Fig.19). Water may only be present in these wadis seasonally or at longer intervals, but the intermittent soaking of the wadi sediments with long

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\(^1\) In some deserts, the water table is at a considerable depth below the surface. This depth may be as much as 700 m in mountainous areas such as the Hoggar Mountains in the Sahara (A.A.E.A. Coffinier, personal communication, 1966). In depositional basins, however, the water table is usually somewhere between 10 and 100 m below the surface of wadis or interdune areas. The water is at, or very close to, the surface in inland sebkhas. The level of the water table may fluctuate several metres annually, or even, in some cases, daily (A.A.E.A. Coffinier, personal communication, 1966). Tidally controlled fluctuation of the water level occurs in water wells in Dubai, Trucial Coast.

\(^2\) The Permian aeolian sands of the Midlands of England have long been shown on Geological Survey maps as Triassic. According to these maps, Permian strata are virtually absent, an age classification that is not accepted by the writer, Shortton (1956) and many others.

Squirrel (1965) and Archer (1965) who present convincing evidence to support contemporaneous or pene-contemporaneous reddening of these sediments. They point out that the red beds are both overlain and underlain by typical grey sediments of the Coal Measures. Downing and Squirrel assume that the red colour found in the Upper Coal Measures of the eastern part of the South Wales Coalfield were either derived from source areas where red soils or red rocks existed, or it resulted from subaerial processes that produced red soils from grey sediments in situ by the weathering of iron-bearing minerals. Archer, who worked in the western part of the South Wales Coalfield, concluded that the red beds and other associated phenomena were related to pene-contemporaneous lowering of the water table.

In the Midlands of England, there is similar conflicting evidence. Analysed, it might well be suggested that some of the rock sequences that are classed as "Barren Red Measures" may have been reddened in post-Carboniferous time, but the majority of the red beds are probably part of a sequence of terrestrial rocks that reflect increasing aridity laterally, away from low-lying coal-forming areas, and upwards into passage beds that herald the approaching Late Carboniferous (Stephanian) and Permian desert conditions (Edmunds and Oakley, 1958, pp.40-42, 48-50).\(^2\)

Recent Desert Sands

Some modern desert sands are red, but most are buff coloured, yellow or even white. The darkest colours in a desert are usually associated with areas of outcrop where the yellow, dark brown or even almost black "desert varnish" (p.19) is thought to be associated with the frequent presence of dew (Engel and Sharp, 1958) or moisture brought to the surface by capillary action (Opdyke, 1961). The most strongly coloured sediments are normally those of wadis, where the older wadi sands and gravels exhibit darkish-brown to pink shades, whilst the youngest wadi sediments are usually light brown or yellow. Pebbles on the deflated surfaces of wadi gravels that have not been covered by flood waters for many years—perhaps many decades—also show the dark brown or almost black colouration of desert varnish (Fig.19). Water may only be present in these wadis seasonally or at longer intervals, but the intermittent soaking of the wadi sediments with long
by the end of the Pleistocene. Why then, is there a difference in degree of reddening between the Pliocene dune sands. They have been dated by means of mastodon teeth and bones found in an interbedded wadi conglomerate (see p.115 and GLENNIE and EVAMY, 1968). These sands are red and had probably already acquired much of their red colour by the end of the Pliocene. sands of similar age were probably remoulded by strong winds during the Pleistocene to give rise to the present dune distribution in the Buraimi area. Loss of pigment due to aeolian abrasion would be kept to a minimum because during the process of reshaping, the bulk of the sands were probably not moved, and for the rest, transport distances were possibly small. On the other hand, Norriss (1969) suggests that in spite of minor losses by abrasion, sand grains can add to their coating of iron oxide even during aeolian transport. By analogy with the present, some lightening in the colour of the surface sands probably occurred during the Pleistocene. It is thought that this local loss of colour is either masked by the stronger colour of the mass of sand that was not reworked, or, it is possible that a slight addition of red oxide to partly coloured sand grains is sufficient to bring back the full colour in a shorter time interval.

The winds that formed the large dunes of the Wahiba crossed no known source of red sediment; all the iron-oxide now found in these sediments is assumed to have been forming and ageing since sometime in the Pleistocene and they are still not red. The wind that formed similar-looking large dunes in the Buraimi area, however, did cross a known source of probably already-red dune sands that, away from the coast, were unlikely to have been cemented. Pleistocene reworking of older red sands seems a very plausible explanation for the occurrence of modern red dunes.

PERMIAN RED BEDS OF BRITAIN

In the Vale of Eden, the centre of the elongate Permo-Triassic basin is occupied by nearly 400 m of Permian aeolian sandstones that are red throughout. Wadi conglomerates are found on the flanks of the basin (Fig.141). All the quartz grains of aeolian sand have a coating of red haematite and yellowish-brown goethite. At the contact points between grains, no haematite exists (Fig.142) although haematite can be traced around the points of contact. The surfaces of individual grains have circles free from pigment at the contact points. This suggests that the formation of the haematite coating is post-depositional.

Grain-like patches of white kaolinite are also coated with haematite and goethite. The kaolinite is thought to be a post-depositional product of alteration from feldspar. The feldspars were probably derived by erosion from Late Carboniferous (Stephanian) to Early Permian volcanics which, in southwest Scotland, are interbedded with dune sands (MYKURA, 1965), or from older igneous outcrops. They may be found in other beds as feldspar, in which case kaolinite need not be present, although feldspar, partly altered to kaolinite, is common (Fig.142, 143, 145). A similar situation of no pigment at the contact points between grains, and partial alteration of feldspar to kaolinite is found in the Permian dune sands of Devon (Fig.143) and the Midlands of England.

One of MILLOR's (1964) criteria for deducing palaeoclimates (Table III), implies that feldspars are destroyed in climates that are hot and humid. Kaolinite

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1 From regional evidence, it seems possible that desert conditions existed over much of southeast Arabia from the Miocene onwards.

2 THIESSEN (1959, p.307) noted that locally in the Wahiba Sands, the sand of the interdune areas was "rusty red whereas the dunes on either side were honey-coloured, both colours becoming paler as we travelled farther north". The same conditions seem also to apply to the large seif dunes of the Rajasthan Desert in India. They are also thought to have formed during the Pleistocene, and are slightly reddened, and yet have no known source of red sediment. This was found to be the case in all thin sections examined. The evidence, as seen in thin section, does not, of course, permit one to say that none of the grains had any haematite coating prior to deposition, but it certainly implies that much, if not most of the haematite is post-depositional.
Fig. 142. Haematite-coated grains of Permian dune sand with haematite-free contact area between the grains (a). Occasional partly altered feldspar (b). Pore space impregnated with plastic. GL. 313. Hilton Beck, Vale of Eden, England. Magnification × 45; reflected light.

Fig. 143. Haematite-coated grains of Permian dune sand with haematite-free contact area between the grains (a). Occasional partly altered feldspar (b). Pore space impregnated with plastic. GL. 97. Mamhead House, south Devon, England. Magnification × 70; reflected light.

Fig. 144. Carbonate-cemented Permian dune sand. Except for the contact area between grains (a) the grains are coated with haematite which in turn is encrusted with dolomite crystals. The dolomite crystals also, appear to have a thin coat of haematite, but it also has a crust of iron-oxide coated dolomite. The remaining pore space (c) has been filled with calcite. GL. 229. Downhill Quarry, Durham, England. Magnification × 45; reflected light.

Fig. 145. Carbonate-cemented Permian dune sand. The grains are coated with haematite and some dolomite. The grains are encrusted with iron-oxide coated dolomite. A grain of feldspar (f) is partly altered to kaolinite. GL. 228. Croe Rigg Sand Pit, Durham, England. Magnification × 50; reflected light.
apparently forms from feldspar under these conditions. It seems probable that below the arid surface of a desert, the humid and relatively hot conditions directly associated with the ground-water/air-filled pore-space interface reproduces a "physico-chemical climate" suitable for the alteration of feldspar to kaolinite, and for the oxidation and precipitation of ferrous solutions to haematite. That some feldspar-bearing horizons remain unaltered is the result, perhaps, of chance circumstances of a locally more rapid rise in the permanent level of the ground water, or of local differences in the permeability of the sands.

On the flanks of a continental basin where wadi sedimentation predominates, post-depositional reddening will occur, but since there is likely to be more organic matter associated with wadi sediments than with dune sands, it is possible that in the former, patches may occur locally which have been permanently reduced to a green or even a white colour. Such discoloured patches are common in the wadi conglomerates of south Devon and the Vale of Eden. Sporadic patches of iron pyrites on the surfaces of some fossil desert sand grains may also indicate post-depositional organic reduction of a previous red coating.

THE APPARENT ANOMALY OF YELLOW ANCIENT DUNE SANDS

An area which does not appear to fit the pattern of reddening of dune sands is that of the Permian "Yellow Sands" of northeast England, which are shown in Fig.79. They directly overlie Carboniferous strata of Namurian and Westphalian age. The sands are themselves overlain by marine shales and dolomites of the Magnesian Limestone (Zechstein equivalent). The "Yellow Sands" are dune sands which, as their name implies, are yellow and not red, and yet they represent part of the same desert that filled other Permian continental basins with sands that are now red from Devon to the Highlands of Scotland (Elgin), and from Britain to Germany.

Many of the quartz grains from the "Yellow Sands" have little or no coloured oxides on their surfaces. X-ray analysis shows that part of the yellow colour comes from a thin coating of goethite and a trace of haematite on the surface of grain-like patches of white kaolinite. The sands are lightly cemented by calcite and dolomite. Occasionally, a yellow iron mineral (siderite or limonite?) is found disseminated through the calcite cement. If this yellow mineral is limonite, this might suggest early calcite cementation of the sands that arrested its alteration to haematite. Small patches and aggregates of haematite also occur in the intergranular spaces.

How is it that grains of kaolinite sometimes have a thick coat of iron oxide whilst many quartz grains have only a thin coat or no coat at all? One possible explanation involves the ground water chemistry that gave rise to a carbonate cement. In Fig.144 the lack of iron oxide at the contact points of the grains implies that the deposition of a film of oxide on the sand grains was post-depositional.

A feldspar grain b has been altered to kaolinite. Hall (1967) has shown that plagioclase feldspars can contain 15 to 30 times as much iron in their molecules as alkali feldspars. Plagioclase feldspars are generally more readily destroyed by weathering than alkali feldspars, but they have been found preserved in desert sediments. K, Ca and Na ions are removed from feldspar during kaolinitisation. It seems reasonable to assume that ferrous iron, resulting from post-burial alteration of plagioclase feldspar, could be trapped in the intercrystallite spaces on the surface of the kaolinite grains and later oxidised. Similar alteration of alkali feldspars would give rise to a much thinner zone of iron oxide. Quartz has no such intercrystallite spaces, so ferrous iron cannot be trapped. If a change to a reducing environment follows the original deposition of iron oxide, much of the oxide, if not all of it, will be lost from the surfaces of the quartz grains. This explanation would appear to fit the observed relationship between grains of much of the Permian "Yellow Sands".

In Fig.145, iron oxide is again absent at the contact point a between two grains. It also appears that the calcite cement has locally removed the iron oxide from around the central grain b and the upper surface of grain c.

When the calcite cement of Fig.145 is treated with acidified alizarin red S and potassium ferrocyanide (Evamy, 1963), calcite free of ferrous iron stains red (Fig. 145, d) and calcite containing ferrous iron stains violet (Fig.145, e). After staining, it can be seen that the iron-oxide films around the quartz grains are absent next to former pore space only where the cement contains ferrous iron.

It requires reducing conditions to incorporate ferrous iron in carbonates (Evamy, 1969). B. D. Evamy (personal communication, 1965) suggests that the reducing environment concerned might first locally have taken the iron-oxide films into solution and that the ferrous iron so produced might then have been incorporated in the calcite cement e of Fig.145.

In Fig.144, a rather similar situation occurs, but is not so clear. The iron-oxide coated quartz grains are encrusted with a thin layer of dolomite crystals. Some solution of the iron oxide may have occurred, but is not so obvious. Some of the dolomite crystals also appear to have a very thin crust of iron oxide. The final stage of cementation was the deposition of pore-filling calcite c.

The pore-destroying calcite cementation seen in Fig.144 and 145 has resulted in a partial loss of colour subsequent to the diagenetic reddening. In Fig.146, the development of quartz overgrowths has had no effect on the iron-oxide films around the grains, suggesting that the diagenetic environment for their precipitation remained constantly oxidising. Fig.144 and 145 are of dune sands deposited on the continental margin of the transgressive Zechstein Sea. The ground water which gave rise to the calcite and dolomite cementation and subsequent loss in colour of the "Yellow Sands" is almost certainly associated with the Zechstein Sea. Fig.146, on the other hand, is, like Fig.142 and 143, of dune sands deposited in a continental basin where oxidising conditions prevailed continuously after deposition.
Further study may show that there is a more widespread palaeogeographical significance attached to some yellow and white fossil dune sands, in that they may imply deposition on a continental margin whose ground waters became saturated with respect to calcium carbonate derived from a transgressing sea. Had a former continental basin of desert deposition been transgressed by the sea, the red colour of the topmost sands might have been reduced as the result of carbonate cementation, but those beneath, whose ground waters remained oxidising, would still retain their red colour.

Non-red dune sands are also known that are overlain and underlain by red desert sediments. The diagenetic processes that brought about these local colour differences are still not fully understood. Much has yet to be learned about the authigenic formation of clay films and quartz and feldspar overgrowths on grains of fossil desert sand, and their relation to colour and diagenetic environment.

THE NECESSITY FOR WATER IN THE FORMATION OF RED DESERT SEDIMENTS

WALKER'S (1967a) criteria for the in-situ reddening of desert sediments appear to be sound. Although he implies the necessity for the presence of water to bring about the diagenetic changes that result in the reddening of desert sediments, he says little about it. Whilst concurring with Walker, the present writer also believes that much of the oxidation of iron in solution takes place at the interface between the ground-water saturated sediment below the water table and the air-filled pore space above. In support of the idea, oxidation of ferrous iron has been noticed in the laboratory by BLOOMFIELD (1964) at the interface between air and water in a water-saturated soil (see also the discussion SCHMALZ, 1968; WALKER, 1968). This principle also applies to the seasonally flooded wadi and river sediments and locally, to dune sands that may acquire a perched water table following rain.

Rain water is known to remain in the pore spaces of mobile dune sands for many years. THESIGER (1948, p.4) refers to vegetation deriving moisture from sand dunes in the Rub al Khali which had received no rainfall for at least four years. The writer has also encountered moist sand when digging deeply into the base of desert sand dunes. Such moisture, in the form of perched water tables or adhering around the contact points between grains even without free ground water, could well account for some minor reddening (such as is seen in the darker colours of the older dunes of the Rajasthan Desert or the Wahiba Sands), but is thought unlikely to cause strong reddening of the total dune sediment as seen in the Permian dune sands of southwest Scotland or the Vale of Eden. A similar argument may be applied in connection with the frequent and often heavy dews encountered in desert regions. The surface grains which are dampened are those which are most likely to suffer further transport and abrasion. There is, moreover, less chance of the grains, coming into contact with a solution containing ferrous iron.

For dune sands to become strongly reddened throughout their total thickness, they must first be buried to the level of the water table in order to come into contact with ferrous solutions. Then, repeated daily and annual fluctuations in the level of the water table will permit the formation of the red oxides of iron over a zone of sediment a metre or more in thickness. As the desert basin slowly fills with sediment, this fluctuating water table will also slowly rise and will soak buried dune sands that have not previously been wet, thereby assisting in the alteration of iron-bearing minerals and permitting ferrous solutions to come into contact with the sand grains. Annual fall in the level of the water table will permit drying of the sand grains and oxidation of ferrous iron to ferric iron. In time, all the sediment through which the interface between the ground water and air-filled pore-space passes will be reddened.

The same principle should apply to annually flooded wadi sediments, or sediments that are affected by the annual rise and fall of the water level in tropical rivers which are controlled by seasonal rainfall.

In contrast to the above ideas, BERNER (1969) supports WALKER (1967a, 1968) in thinking that yellow limonite can be converted into red haematite within watersaturated sediments.

The solution which provides the ferrous iron is reducing. It must have, however, a very low reducing capacity, so that contact of this solution with any previously deposited iron oxide will not lead to re-solution of the oxide. This is thought to be the case for the following reasons:

The most common reducing agents encountered in sediment are dissolved sulphide ($S^{2-}$) and organic matter. If dissolved sulphide is present, this would react with ferric oxide to form pyrite. Since this, in general, is not found, the presence of sulphide in quantities large enough to redissolve the ferric oxide may be excluded. Plants are generally absent from desert sediments except on stabilised dunes and wadi sediments, where they are usually oxidised soon after they die. This lack of organic matter, together with the absence of sulphide, leads one to assume that there are insufficient reducing agents in desert basins to reduce ferric oxide to ferrous oxide. The sediment will, therefore, normally retain its red colour after it has passed permanently beneath the water table.

CRITERIA FOR THE RECOGNITION OF A DESERT PALEOClimATE

The palaeoclimatological significance of desert sandstones has been discussed in some detail by ODVYKE (1961), and some of the problems connected with the recognition of arid and hot climates of the past have been discussed by MCKEE (1964). Following Millo's climatic criteria given in Table III, additional criteria indicative of desert conditions of deposition in ancient sediments are given in Table IV. They are mostly additional to the criteria already given in chapter 2 for the
recognition of ancient desert sediments and are closely allied to the problem of why some ancient desert sediments are red.

TABLE IV

<table>
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<tr>
<th>MILLOT'S (1964) CLIMATIC CRITERIA APPLIED TO DESERTS</th>
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<tbody>
<tr>
<td><strong>A. Modern desert sediments</strong></td>
</tr>
<tr>
<td>Polycrystalline grains occur</td>
</tr>
<tr>
<td>Feldspars are preserved</td>
</tr>
<tr>
<td>Quartz grains are not micro-fracture</td>
</tr>
<tr>
<td>Quartz grains are not corroded; they are usually “frosted” at the time of deposition</td>
</tr>
<tr>
<td>Grains not normally covered with a strong coating of red iron oxide unless they were derived from nearby red outcrop or already red wadi or dune sediments</td>
</tr>
<tr>
<td>Larger detrital aeolian grains are often rounded; no overgrowths present on grains of quartz or feldspar</td>
</tr>
<tr>
<td>The clays are usually detrital, in which case their composition appears to depend entirely on provenance; some desert salt lakes contain authigenic Mg-rich clay minerals (D.H. Porrenga, personal communication, 1966)</td>
</tr>
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</table>
Adhesion ripples: Adhesion ripples and adhesion warts form when dry sand is blown onto humid surfaces and is fixed there by the surface tension of the water that ascends through the sediment. It follows that a dry area of deflation (as the source of wind-blown sand) and a damp area of sedimentation are both necessary for their formation. Adhesion ripples are oriented transverse to the wind direction. The smooth windward slope is steeper than the wart-covered lee slope. Surfaces that are covered exclusively with adhesion warts form preferentially under conditions of strongly varying wind direction. Adhesion ripples and adhesion warts are different from other ripples in that the laminae within the ripples are concave upwards (Reineck, 1955).

Adhesion warts: Small sedimentary adhesion structures that commonly form part of the larger adhesion ripples (see also Adhesion ripples).

Arroyo: American synonym for wadi.

Avalanche slope: That slope which forms on the lee side of a dune at the maximum angle of repose for dry sand, 34°. Any tendency to overload this slope results in an avalanche of sand down the slope. Synonym for "slip-face".

Bahr: Interdune area in the region of Lake Chad, flooded during seasonal rise of the water of the lake.

Bir: Water well (Arabic). Similar to tawi.

Braided streams: A braided river is an interlacing network of distributaries with shoals and islands of gravel and sand between. The river channel as a whole is characteristically wide and shallow (Holmes, 1965, p.538). This definition for braided rivers may be extended to the braided channels of wadis. The sediments of a braided system of wadi channels consist of inter-tonguing lenses of fluvial gravels and sands that have been subjected to deflation, and upon which aeolian sands may also have been deposited.

Calcrete: Horizon of hard, dense calcium carbonate or sediment cemented by calcium carbonate, found in soils of hot, arid and semi-arid regions.

Caliche: Surface sediments in arid or semi-arid areas that become cemented by evaporation of lime-rich ground water.

Deflation: The blowing away of dry incoherent rock material, sand and dust by the wind; a form of transport (denudation) chiefly at work in deserts. (Challinor, 1962, Dictionary of Geology).

Desert: An almost barren tract of land in which precipitation is so scanty and spasmodic that it will not adequately support vegetation, and where the potential rate of evaporation far exceeds precipitation.

GLOSSARY

1 A compilation of terms used in describing deserts and their land forms can be found in "A desert glossary" by Stone (1967, pp.211–268). His terms are listed according to country with a marked emphasis on North America—a reflection of the relatively much greater volume of literature associated with one of the world's less extensive areas of desert.
Glossary

Dikaka: Accumulations of dune sand covered by scrub or grass vegetation, extended to include plant-root burrows in ancient dune sediments.

Eolianite: All consolidated sedimentary rocks deposited by the wind (Stone, 1967). More commonly used in connection with calcite-cemented dune sands of coastal regions.

Erg (pl. uruq): Arabic word for vein. Applied to linear dunes or belts of dunes (Holm, 1960, p.1329).

Falaj (foggarah): A long underground aqueduct used for irrigation in Arab countries. Synonym for the Persian qanat, or the Moroccan foggarah.

Feild: Sand-free interdune corridor in area of linear dunes.

Frosted sand grains: Sand grains, typical of desert and aeolian sediments, whose roughened surface causes a scattering of the light; the opposite of glassy or polished sand grains.

Gibber plain: Australian synonym for "serir". A gibber is an aboriginal word for pebble, sandstone or sandstone pavement.

Hammada: Rocky desert. A tableland or plateau of rock denuded by wind erosion.

Insolation: Exposure to radiation from the sun. More particularly, the term implies the wide diurnal temperature range found in desert areas, and the temperature differences encountered between the hot surface of rock exposed to the sun's rays and the cooler interior of the rock, which are liable to cause splitting.

Jebel (jabal, djebel): Hill or mountain.

Kavir (kawir): Desert (Persian).

Loess: Soft, porous, yellow or buff coloured accumulation of wind-laid particles, predominantly of silt size.

Mamlahah: Inland sebkhas that have been excavated for salt. (Powers et al., 1966, p.D100).

Monsoon: A monsoon is the type of wind system in which there is a complete or almost complete reversal of prevailing direction from season to season. It is especially prominent within the tropics on the eastern sides of great landmasses (Moore, 1949). The Southeast Trade Winds in the southern Indian Ocean are warm and saturated with moisture. Because of intense heating of the land in summer, a low-pressure area develops over northwestern India and West Pakistan which deflects the Southeast Trades northward. Part of these deflected trades travel parallel to the coast of Africa and then swing parallel to the South Arabian coast (as the Southwest Monsoon) where the hill slopes facing the sea receive some rain. These moist winds bring torrential rainfall to much of central and northern India. This summer monsoon lasts between April and September, although its duration at any one place depends upon its geographical location. The winter monsoon blows (north) across the Wahaiba Sands, the elevation of the sands above sea level is too low to induce rainfall over this hot area. The winds are, however, strong. Similarly over the Rajasthan Desert, the Southwest Monsoon brings little rainfall because of the desert's relatively low elevation.

Sebkha, sabkha(1), sabkra (pl. sibakh): Flat area of clay, silt, or sand often with saline encrustations. Subdivided into:

(1) coastal sebkha: a coastal flat that occurs just above the level of the normal high tide in a hot, arid, desert climate. Its sediments consist of sand, silt or clay and its surface is often covered with a salt crust that results from evaporation of water drawn to the surface by capillary action and from occasional marine inundations. Its sediments are mostly derived from adjacent marine environments (commonly carbonate in composition) with admixture of material of continental origin; they are liable to diagenetic alteration as the result of its environment. These sediments of the coastal sebkha are characterised by the presence of algal mats and the formation of evaporites (nodular anhydrite, gypsum and halite) and dolomite. During much of the year, the surface of the coastal sebkha is subject to deflation down to the level of the water table.

(2) inland sebkha: a flat area of clay, silt or sand, often with saline encrustations. Their salts may be derived from intermittently flowing wadis that have no access to the sea, from fluctuating saline ground water, or from relict salts left behind by an advancing coastline or from salt crystals blown from another locality, often the coast. The sediment may consist of clays and sands carried in wadis, wind-blown silt and sand trapped on the moist surface as adhesion ripples, or of small sand dunes and sporadically drifting sand that has migrated over the dry salt crust. (Often referred to in North America as a playa or salina.) Algae are known but algal mats have not been recognised.
Serir: Deflation lag of pebbles, often of alluvial origin, on the desert surface.

Shamal: Arabic word for north; applied to north or northwest wind that blows down the Persian Gulf.

Silcrete: Horizon of siliceous material found in hot, arid and semi-arid regions; possibly formed by silicification of a sediment.

Slip-face: The slope which forms when wind-blown sand from the windward side of a dune passes into relatively calm air of the leeward side. The maximum angle of repose for dry sand is 34°. Also referred to as “avalanche slope”.

Tawi: Water well (Arabic). Similar to bir.

Wadi (pl. widyen): Desert watercourse, dry except after rain. The American equivalent is “arroyo”. In mountainous areas, little sediment is deposited. Beyond the mountains, or in broad intermontane valleys, alluvial fans are formed by rapid deposition of sediment resulting from sporadic flow of water over a lessening slope; here, this flow of water may be in the form of a spreading sheet-flood, or may be confined to an interlacing network of distributaries. Beyond the limits of the fan, the wadis spread out over the desert plain unless confined by the presence of outcropping strata or sand dunes. The water in a wadi may flow into the sea, a desert (temporary?) lake or sebkha (inland or coastal); often there may be insufficient rain for the water to flow the full length of its channel. The unconsolidated fluvial sediments of a system of wadis may be deflated and so form an important source for aeolian sand. Conversely, aeolian sand may be deposited on the bed of a wadi and so form a part of the sedimentary sequence.

REFERENCES


CROSS REFERENCES TO FIGURES, TABLES AND ENCLOSURES

Many of the figures in this book have been used to illustrate more than one aspect of a desert environment of sedimentation, sedimentary structure or other feature from which the existence of a former desert can be inferred. In many cases, they also suggest more than one means of sediment transport. The following list of pages, in which individual figures, tables and enclosures have been referred to, will assist the reader in making a more complete analysis of any particular figure.
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PERMIAN ROTLIEGENDES OF NORTHWEST EUROPE
INTERPRETED IN LIGHT OF MODERN DESERT
SEDIMENTATION STUDIES

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Permian Rotliegendes of Northwest Europe Interpreted in Light of Modern Desert Sedimentation Studies

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Abstract. The Permian redbeds of northwest Europe, termed the "Rotliegendes," are continental clastic sediments laid down under desert and semidesert conditions. Recent drilling in the North Sea, Netherlands, and north Germany has shown that these beds are in a sedimentary basin up to 2,000 km long and 500 km wide. This basin is in the foreland of the Variscan mountains (Hercynian orogeny) and is of part-orogenic origin and partly fault bounded. Volcanic rocks are present locally, especially in the east. The Rotliegendes strata attain a maximum thickness of approximately 1,500 m. A central shaly and halite facies (Hallsengebirge facies) corresponding to a salt lake. Paleowind directions were essentially from east to west. Deflation of the alluvial fans spreading northward from the Variscan mountains resulted in accumulation of up to 200 m of dune sands in the southern North Sea area. The dune sands form the reservoir rock for important accumulations of gas. At present, the proved and probable reserves of Rotliegendes gas in the North Sea, Netherlands, and western Germany are about 2,500 x 10^9 cu m (85 x 10^12 cu ft).

The interpretations of Rotliegendes facies are based on a study of modern deserts and their sediments (wadi sediments, dune sands, and the sediments of inland and coastal sabkhas).

INTRODUCTION

The Permian redbeds of northwest Europe, known generally in Germany as the "Rotliegendes" (the "redbeds" that underlie the Zechstein), are continental clastic sediments deposited under desert and semidesert conditions. The presence of Permian rocks of Rotliegendes facies has been recognized in England and Scotland (Sherlock, 1926, 1928, 1947), southern Norway (Holte, 1934), the Rhine graben in France and western Germany (de Lapparent, 1906; Harras, 1926), Thuringia and Harz in central and eastern Germany (Gagel, 1926; Born, 1926), and in Poland. Red continental sedimentary rocks of the same age also are present in other parts of France, Germany, the Pyrenees, and Russia (Sherlock, 1947), but will not be discussed. Recent drilling in the North Sea, the Netherlands, Denmark, and western Germany has shown that most of the scattered outcrops form part of one large basin, the northwest European Permian-Triassic basin, of which the Rotliegendes forms the initial sedimentary sequence.

Oil companies first became interested in these rocks in 1963 when the magnitude of the Groningen gas field in the north of the Netherlands (discovery well, Slochteren 1, 1959) became apparent. This immediately focused attention on neighboring parts of the Netherlands and on the nearby areas of the North Sea. Since then, additional important gas fields have been discovered beneath the North Sea (Lemans, Indefatigable, West Sole, Viking) and smaller accumulations of gas in the Netherlands and northwest Germany. At present, proved and probable reserves are estimated at 1,650 x 10^9 cu m for the Groningen field (at 95 percent confidence level), 700 x 10^9 cu m for all the gas fields of the North Sea, and 240 x 10^9 cu m for the smaller Rotliegendes gas fields of the Netherlands and western Germany.

It has long been realized that the outcropping Rotliegendes represents the sediments of a former desert or semidesert environment. However, the significance of its internal facies changes could only be properly understood after an extensive study of recent desert depositional environments. The detailed results of this investigation have been published elsewhere (Glennie, 1970). Following are its salient points and their application to the Rotliegendes facies interpretation.
SEDIMENTS OF MODERN DESERTS

A classical desert is an almost barren tract of land over which rainfall is too limited and spasmodic to support vegetation adequately. The upper limit for rainfall in a desert is about 25 cm a year, and because of the high temperature and general lack of humidity, the potential rate of evaporation far exceeds precipitation.

Deserts cover about one fifth of the world’s present land surface and tend to be concentrated into the regions of prevailing trade winds that blow roughly between the latitudes of 10 and 30° north and south of the equator. Deserts exist in these latitudes because the relative humidity of the descending high-pressure air of the horse latitudes (30° north and south of the equator) decreases as the air is compressed adiabatically and gives rise to cloudless skies. Part of the air flows toward the equator and is deflected westward (the trade winds) by the Coriolis force. As the air blows across the desert under clear skies, it heats and absorbs additional moisture. The land thus is subjected to increasing desiccation. When rain does fall over the desert—and this may be annually or at less frequent intervals—it commonly does so because of meteorologic disturbances initiated beyond the limits of the desert. In areas of outcrop, the rock surface continually is subjected to weathering processes. During the cooler hours before dawn, the relative humidity in such places as the desert interior of Oman may be 100 percent and heavy dews frequently cover the rock surfaces, leading to slow but regular chemical corrosion, especially of carbonate rocks. Moisture—even though in small quantities—is drawn to the surface from the water table by capillary action. Evaporation close to the surface results in the growth of gypsum and halite crystals which exert an expansion force on the surrounding host rock, causing it to split. Rapid diurnal changes in temperature, especially if accompanied by a rare rainstorm, cause differential expansion and contraction stresses between the surface and the directly underlying rock, resulting in spalling of the rock surface or even the splitting of boulders. Siliceous rocks and fine-grained homogeneous limestones tend to form angular boulders, but more argillaceous boulders, granites, and dolerites become rounded by exfoliation. Many of these processes of weathering have been discussed in greater detail by Ollier (1969).

The wind forces silt- and sand-size products of weathering into motion and results in transport in suspension, by salivation, or by surface creep (Bagnold, 1941). With removal of the finer weathering products by the wind, the larger, more resistant pebbles and boulders are left on the desert surface as a lag deposit that locally may be abraded by a sand-laden wind into ventifacts.

FIG. 1—Pattern of braided-stream (wadi) channels fanning out from point on western edge of northern Oman Mountains. In foreground, recently flowing water in most active major wadi channel was unable to remove sand dunes in its path and so followed line of least resistance parallel with edge of dune field.

Lag gravels on the desert surface effectively prevent most of the underlying finer grains from being removed by the wind; beneath the gravel, however, weathering processes continue.

Water-Laid Sediments

Desert fluvial sediments—Heavy rainstorms are infrequent. With little vegetation to stay the flow of water, the weathering products are transported by the rush of water from the highlands to the lowlands where they are deposited as slightly graded fluvial sediments. Because there is little or no soil on the barren highlands, much of the rainwater has little chance of soaking into the ground. When the rain ceases, the flow of surface water soon stops and the wadis, along which sediment was being transported, dry out. After heavy rainstorms, flowing water may extend the length of the wadi to reach the sea or to form a temporary lake if its onward passage is stopped by sand dunes. After milder rainstorms, the water soaks into the sediment of the wadi before reaching the lowest point in the channel and the channel then locally becomes overloaded with sediment. Overloading also occurs when the water velocity is lowered at the nick point between hill and plain; braiding follows with the formation of an alluvial fan (Fig. 1). When there is a high sediment-water ratio, viscous muds capable of transporting large boulders may form in the wadi.
As the supply of water diminishes, much of it soaks into the sediment of the wadi channel. The water commonly is saturated with respect to calcium carbonate and evaporation results in rapid cementation of the sediment at the air/water interface. The surface sands and silts dry out rapidly, however, and generally are not cemented; they can be readily carried away by the wind. The clay lags forming in stagnant pools dry on exposure to the air, and curl and crack. Very thin and fragile curled clay flakes will be blown away by the wind, but slightly thicker and heavier ones can be preserved in place by a protective covering of wind-blown sand (Fig. 2).

Temporary desert lakes and inland sabkhas—Temporary lakes may form in the center of a basin with inland drainage or where dune sands block a wadi channel (see Glennie, 1970, Fig. 49). As the water evaporates, salts are concentrated and eventually the sands, silts, and clays are covered with a crust of halite, and gypsum crystals grow in the sediment. Such flat areas of salt-encrusted sand, silt, or clay are known in Arabia and North Africa as sabkhas. As sabkhas occur in both inland and coastal areas and possess different sedimentary characteristics, the term should be qualified by the words "inland" or "coastal." Not all sabkhas are supplied by surface water. In many wadis, the water percolates through the wadi sands to the sabkha, where it is brought to the surface by the pressure of the head of water or by capillary action, and then evaporates. Other inland sabkhas appear to be supplied solely by groundwater.

A layer of silt and clay commonly is deposited on the floor of a temporary lake that is supplied with water from a wadi. As the water evaporates or soaks into the underlying sediment, the lake reduces in size and the clays become desiccated as they are exposed to the air. Sand dikes (Fig. 3) form when cracks develop in the mud and are filled with a slurry of wet sand from below (Oomkens, 1966). Other cracks are filled with wind-blown sand from above.

Adhesion ripples—The wind may blow grains of sand, silt, or clay onto the damp surface of a sabkha, and as moisture continues to rise to the surface by capillary action additional grains will adhere. In this way, adhesion ripples form (Fig. 4). Adhesion ripples on the coastal flats of north Germany were first described by Reineck (1955). In cross section they have a typical irregular wavy appearance. The formation of adhesion ripples can continue until the upper limit of capillary action is reached. This limit may be achieved
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by the addition of more wind-blown particles or by lowering the source of the moisture—the lake level or the water table. Modified adhesion ripple structures may result from deposition upon the hydroscopic surface crust of halite that covers the sabkha.

**Wind Deposits**

*Transport of sediment by wind*—Wind is capable of transporting very fine (<100 μ) particles in suspension. Consequently, much of the clay and silt originally deposited in wadi channels or temporary lakes is removed by the wind and carried beyond the limits of the desert where it is redeposited as loess. Coarser sand grains (100-2,000 μ) move by saltation under the influence of the wind and cause still larger grains, up to six times the diameter of the saltating grains (Bagnold, 1941), to move over the desert by surface creep.

*Interbedded water-laid and eolian sediments*—The uncemented fluvial sands on the surfaces of wadis form a very important source of wind-blown sand. During the long intervals between rainstorms when the wadi channels are dry, the fluvial sands are reworked by wind and thus become eolian sands with sedimentary structures of eolian origin. When the wadi is next in flood, much of the eolian sand in the wadi channel is removed by the flowing water. The water, however, commonly is not capable of removing all the eolian sand, especially from protected parts of the channel, so that wind-blown sand can become interbedded with fluvial sediments (Fig. 5).

Although eolian ripples form on the surface of wadi channels, the most characteristic features preserved are fairly well-sorted, subhorizontal laminae of alternating coarser and finer sand grains that are interpreted as being wind-laid. These laminae commonly dip gently upstream, especially at places fairly close (tens of kilometers?) to hills or mountain ranges. This dip is thought to be in response to the convection of air over nearby hills during the hottest part of the day which gives rise to a strong breeze blowing up the wadi channel toward the hills. This breeze blows toward nearby hills irrespective of the direction of the regional prevailing wind.

*Eolian sands*—Eolian sands are deposited when the velocity of the driving wind is reduced to the point where it can no longer transport the sand. On the grand scale, extensive areas of dune sands are present in the relative protection of broad structural depressions and over low-lying plains. Locally, eolian sands may be found part...
way up the flanks of mountains, and along coastlines and protective banks of wadis or cliffs (Fig. 6). On the open plains, eolian sands tend to accumulate in areas that are already covered with sand rather than in the surrounding areas of bare rock. Such accumulations are in the form of linear strips when the sand-transporting wind velocities are high, or of low oval mounds when the force of the wind is less.  

Barchan and transverse dunes—Sand accretion, influenced by winds of moderate velocity and uniform direction, results in the building of oval mounds which have their highest point in a downwind direction. As the height of an individual sand patch becomes greater, the lee slope increases until it reaches the angle of repose for dry sand, at about 34°, and a slip-face (or avalanche slope) forms. This situation seems to be reached at a minimum slip-face height, for what may now be called a dune, of about 30 cm. Sand grains are transported up the windward slope of the dune. Any tendency to exceed the angle of repose of the lee slope causes sand to avalanche down the slip-face until the slope of 34° is regained. The largest and roundest grains are found at the bottom of the slope.

Because of the resistance presented by the dune itself, the rate of sand transport is greater along the flanks of the dune than over its crest. The flanks thus are drawn out into "horns" that are directed downwind, and the dune assumes a crescent shape commonly referred to as a "barchan." Barchans migrate downwind as the windward slopes are eroded and the sand is redeposited on the horns and leeward avalanche slopes. The sand grains found on the crest of the dune are finer grained than those at the foot of the avalanche slope or along the low flanks of the dune. Barchans increase in size if more sand is deposited on the dune than is removed. With a considerable increase in the supply of sand, barchans coalesce and commonly form long sand ridges transverse to the dominant sand-transporting wind (Fig. 7).  

Seif dunes—Bagnold (1941) made three important observations:

1. In a strong sand-laden wind a uniform drift of sand over a uniform rough surface has a transverse instability, so that sand tends to deposit in longitudinal strips (p. 178).
2. A given wind can drive sand over a hard immobile surface at a considerably greater rate than is allowed by the loose sand-covered surface (p. 72).
3. A strong wind causes an accretion of sand on an existing sand patch with an extension upwind of the border...this action lasts only as long as there is a plentiful supply of sand (p. 171).

These factors are thought to be responsible for the formation of linear dunes (seifs) whose axes are roughly parallel with the dominant sand-transporting wind. When the wind velocity is so great that there is a transverse instability in a patch of sand, there will be a slight pressure gradient between the hard immobile surface of the sand-free area and the loose sand-cover of the growing dune. Parallel with the dominant wind direction, spiral wind-cells develop and transport the sand obliquely up the flanks of the dune (see Glennie, 1970, Fig. 73). So long as there is a continuous supply of sand, there will be accretion of sand upwind as well as an extension of the dune downwind. Because these spiral winds are not directed at right angles to the axis of the seif, avalanche slopes normally will not form; they may form, however, if there is another (seasonal?) wind that blows across the axis of the dune, or if the dominant high-velocity wind veers slightly oblique to the dune's axis so that the wind-cells on either side of the seif are of unequal strength.  

Influence of wind velocity on dune type—We distinguish two basic types of sand dunes, the transverse dunes (or barchans where the supply of sand is limited) formed by winds of moderate velocity, and seif dunes built by strong winds. Each dune type has its own characteristic distribution of bedding attitudes. Transverse dunes are composed mostly of beds deposited as avalanche
slopes; much of the bedding will dip at angles of 30° or more in the general direction of the prevailing wind. Seif dunes, however, have most of the bedding dipping away from the axis of the dune at almost right angles to the direction of the dominant sand-transporting wind. A plot of the characteristic orientation of bedding attitudes for seif and transverse dunes can be made on polar stereographic nets (Fig. 13). In ideal circumstances, it is possible to deduce from the plots not only the direction of the dominant wind that formed the dune, but also to indicate the probable dune type (Glennie, 1970, Fig. 68).

Eolian ripples—Although sand ripples are common on the surfaces of dunes, they appear to be formed only when the wind velocity is moderate. During periods of maximum sand transport, it is thought that the velocity of the sand-laden wind generally is too great for ripples to form on the accretion slopes of seif dunes. The windward slopes of barchans are subjected to erosion and transport of sand, so that ripples are not normally preserved. The avalanche slopes of transverse dunes are not formed by the direct action of the wind. Ripples will form on the avalanche slopes, however, if the wind blows across these slopes. As observed in Arabia these ripples are preserved when avalanche conditions recur with a return to the prevailing wind direction.

Interdune inland sabkhas—Many of the world’s major areas of sand dunes are in basins where they are thought to develop to considerable thicknesses (e.g., 300 m south of Bir Zelten in Cyrenaica; McKee and Tibbits, 1964). In a subsiding basin, the water table rises and may reach the surface if the rate of subsidence exceeds that of sedimentation. When this occurs in a desert that is covered by dune sands, the surface of the low-lying interdune areas is moistened by the groundwater and, following evaporation, a crust of halite forms and gypsum crystals grow in the sediment. An interdune inland sabkha results, and adhesion ripples develop on the damp interdune surface. Annual fluctuations in the water table caused, for example, by seasonal rainfall in areas marginal to the desert, can give rise to an alternation of adhesion ripple formation (rising water table) and their destruction (falling water table). Adhesion ripples will be preserved if permanently overlain by dune sands.

Desert Coasts

Because in deserts there are no rivers that have a continuously flowing supply of water, coastlines that border deserts have no fluvial deltas. Notable exceptions to this rule are those large rivers...
such as the Nile and the Indus that are fed with water from beyond the limits of the desert.

Because areas of major accumulation of desert sediment are those of subsiding basins, the coastlines of these basins also are likely to be low and subject to marine transgression. This appears to have been so at the southern end of the Persian Gulf where it borders the northern end of the Rub al Khali basin. According to Evans et al. (1969), the last marine transgression over older dune sands of the Abu Dhabi coast began about 7,000 years ago, but for almost the past 4,000 years, progradation of intertidal and supratidal sediments has taken place.

The clear, inland tropical seas that border many hot deserts are precisely those areas where organisms produce large quantities of calcium carbonate. Therefore, a continuous supply of carbonate material is available for transport by tidal and longshore currents. Such material is washed onto the beaches where it is blown inland by onshore winds. Any indentation of the coast is likely to be cut off from the open sea by the formation of a submarine bar or a sand spit. If the bar builds up to sea level, the dry capping sediment can be blown into the lagoon behind. Strong tidal flow of water into and out of the lagoon will cause the formation of oololiths and build oolite deltas out to sea (or into the lagoon in the case of predominantly inflowing tidal currents).

Large low-lying coastal areas are subjected to flooding by high spring tides and storm tides. As the floodwaters recede or soak into the ground, a crust of halite crystallizes on the sandy, silty, or clayey surface and gypsum crystals grow in the sediment (Kinsman, 1969). As mentioned, such areas are referred to in Arabia as "sabkhas." The outer edge of the coastal sabkha is marked by a belt of dark rubbery algal mat that binds the underlying sediment and which is soaked by each high tide; when exposed to desiccation, it will harden, curl, and crack as it dries. The sediments of a coastal sabkha are commonly rich in the skeletal carbonate forming just offshore (marine Foraminifera, gastropods, and more rarely, fragments of reef-building organisms). After burial, these sediments are liable to diagenetic alteration to dolomite. Adhesion ripples may form over the damp sabkha surface. Once the surface of the sabkha is dry, deflation may predominate over sedimentation and large gypsum crystals can be exposed (Fig. 8) with their long axes almost vertical; they also may be exposed by marine erosion of the unconsolidated sands in which the crystals are embedded (Kendall and Skipwith, 1969, p. 888).

**Rotliegendes Facies**

The sedimentary rocks of the Rotliegendes of northwest Europe have much in common with the sediments of modern deserts.

The Rotliegendes disconformably overlies pre-Permian rocks and is conformably overlain by the Copper Shale (or Kupferschiefer) that marks the base of the Upper Permian Zechstein evaporites. The sediments of the Rotliegendes are virtually devoid of fossils and have been divided arbitrarily into a lower and an upper unit on the presence or absence of associated volcanic rocks. As Permian volcanic rocks apparently are absent over most of the investigated parts of the North Sea and the Netherlands, descriptions of the rocks of the Lower Rotliegendes will not be given in this part of the paper.

**General Sequence**

Figure 9 presents two generalized sedimentary sections as found in the Upper Rotliegendes of the Netherlands and adjacent North Sea area. The section in column A consists of crossbedded eolian sandstones with minor amounts of fluvial sandstones and conglomerate. In column B,
Anhydrite and halite
Dolomite
Copper Shale - base Zechstein
Homogenized (non-laminated) sandstones
some slump structures
Well-sorted horizontal and large-scale planar cross-bedded sandstones,
adhesion ripple beds increasing in importance upward
Some beds of sandstone with small-scale cross-bedding
Homogenized sandstones; sandstones with small-scale cross-lamination,
locally developed beds of conglomerate, mud-cracked clay or adhesion ripples
Well-sorted horizontal and planar cross-bedded sandstones, some beds of adhesion ripples
Sandstones (small-scale cross-bedding or homogenized), clay-pebble conglomerates and mud-cracked clays
with sandstone dikes.
Homogenized sandstones or sandstones with planar cross-bedding at base
Dark shales, cross-bedded sandstones.
Coal seams - Carboniferous, mainly Westphalian
Eolian sand
Adhesion ripples in eolian sand
Quartz and clay-pebble conglomerates
Fluvial sands, shales, curled clay flakes, S-shaped dikes
Approximate location of figure
crossbedded eolian sandstones are far less important. Quartz-rich conglomerates with clay pebbles, fluvial sandstones and clays with minor eolian sequences dominate the lower part of the section. In the middle, eolian sands increase in importance. In the upper part of the sequence clay beds with anhydrite nodules grade vertically laterally into rock salt with laminated red clays and practically no anhydrite (Haselgebirge facies).

The predominantly sandstone and conglomerate sequence has been referred to as the Slochteren Member of the Upper Rotliegendes, after a well in the Groningen area; the overlying Ten Boer Member, is named after another well in the same area. The type sections for both the Slochteren and the Ten Boer Members are in the Slochteren 4 well, which was cored throughout the Rotliegendes sequence (see te Groen and Steenken, 1968; Bungener, 1969; Stäuble and Milius, 1970). The relative proportions of these different facies vary from place to place. Columnar sections showing correlation of the two members of the upper Rotliegendes have been given for four wells in the Groningen area of the Netherlands by Bungener (1969, Fig. 7) and Stäuble and Milius (1970, Fig. 5).

The total thickness of the Rotliegendes along the southern margin of the basin ranges up to about 250 m (Gill, 1967; Kent and Walmsley, 1970). However, in the central, north German, part of the Rotliegendes basin, where the evaporitic "Haselgebirge" facies is most fully developed, Kent and Walmsley (1970) suggested that the Rotliegendes thickness exceeds 5,000 ft (1,500 m). Other publications concerning the facies and thickness of the Rotliegendes found in both well and outcrop in Germany and Poland include Fabian et al. (1962), Hecht et al. (1962), Helmuth (1968), Reichel (1968), Lützner (1969), Grodzicki et al. (1967), Klapcinski (1967), Krason (1967), and Lacka and Moliety-Rakowska (1968).

In the following section, the facies development of the Rotliegendes is described in terms of modern desert environments.

Wadi Environment

In the northern part of eastern Netherlands and Germany, the Rotliegendes sequence disconformably overlies Carboniferous rocks. The basal Rotliegendes clastics consist of fluvial sandstones and conglomerates commonly cemented by calcium carbonate.

Figure 10 presents a typical sequence from the basal Rotliegendes of the southern North Sea. The cores are made up of alternations of conglomerate, laminated and nonlaminated brown-red sandstones, and dark-red clay. At the base of the sequence is a sandy conglomerate (c) containing subrounded to rounded light-colored quartzite pebbles and dark, angular to rounded shale pebbles. The latter were probably derived from the Westphalian shales that lie about 8 ft below. The conglomerate almost certainly was deposited by the action of flowing water.

The red clays (black in Fig. 10) have been cracked and curled (concave side up, cf. Fig. 2), by subaerial desiccation. However, other clays (D), that are also fractured and curled, seem to have been penetrated from below by sand. Although not textbook examples, these sandstones are interpreted as sandstone dikes (Fig. 3). The generally slightly argillaceous sandstones (F) directly beneath the clay beds display low dip, commonly discontinuous laminae. They are interpreted as having been deposited by flowing water.

The dip of bedding of the well-laminated sandstones (A) is both subhorizontal and fairly steeply inclined. In both cases, the grain size of the sands ranges from very fine to medium, and locally even coarse (> 500 μ). The larger grains are subrounded or rounded, but the finer ones are angular. Frosted grains are common. The grains in each lamina generally are well sorted, although there may be marked grain-size differences between adjoining laminae. At porosity-plug 739, steeply inclined laminae are truncated and overlain by coarser grained, finely laminated horizontal sandstones that become more steeply inclined upward and also develop relatively thick sets of laminae of constant grain size. Above porosity-plug 716, a clay zone is sharply overlain by steeply inclined (25°) sandstone. All these sandstones display many of the characteristics of eolian sand. The dipping sandstones apparently are inclined in about the same direction and are interpreted as having been deposited by a wind of fairly constant direction. The horizontal sandstones likewise are considered to be of eolian origin; their apparent lack of ripple- or dune-bedding is tentatively ascribed to their having been deposited on the floor of a dry stream bed by a strong sand-laden wind.

An important part of this sequence is composed of almost structureless, relatively poorly sorted, very fine- to fine-grained, or locally medium-grained, sandstones (H) with some clay pebbles and flakes. These apparently homogenized sandstones are believed to be composed of sands from nearby that had been reworked by the wind or flowing water. Overloading and short
FIG. 10—Sequence of cores starting about 8 ft above Carboniferous (Westphalian) shales. Sequence comprises alternations of conglomerates (c, whose components include clay and shale pebbles and flakes), curled and broken clay layers (black) that are probably still in their site of original deposition; sandstone dikes (D), fluvial sands (F), poorly sorted sands that are probably former eolian sands homogenized during water transport (H); and undisturbed well-sorted subhorizontal to inclined eolian sand laminae (a). Other sands (w) are of more doubtful origin and could have been deposited after either eolian or water transport. Numbered holes are of plugs cut in cores for porosity/permeability determinations. Rotliegendes, North Sea.
distance of transport did not permit better sorting. The clay pebbles probably were derived from desiccated clay drapes deposited earlier on the floor of the wadi, and ripped up by a sandcharged flow of water.

These structureless sequences may grade up into sandstones that cannot be assigned with any degree of certainty to either a fluvial or eolian origin. In most cases, however, they develop weak subhorizontal or gently inclined laminae (w) whose mode of formation is difficult to ascertain. These sandstones could have been either wind- or water-laid. In either case the poorly developed laminae could be interpreted as the result of overloading of the transporting medium and a transport distance from its source area that was just sufficient to permit the first signs of sorting (Fig. 5).

The sequence seen in Figure 10 thus is comprised of alternations of wind-laid and water-laid sediments that show evidence of subaerial exposure. Therefore it probably represents an ephemeral stream environment in which there has been considerable reworking of the sands by both wind and water. This phenomenon most commonly is seen in deserts where the fluvial sediment deposited in a wadi channel is eroded and partly overlain by eolian sands (Figs. 2, 5, 6).

Sabkhas Associated with Temporary Desert Lakes

In some sequences of the Rotliegendes, conglomerates are virtually absent and clay, with a much reduced proportion of sandstone, is dominant. Minor amounts of nodular anhydrite are present. The horizontally bedded clays are, in places, mudcracked and curled, indicating subaerial exposure and desiccation. In other places, systems of sandstone dikes (Fig. 3) cut through several layers of clay (Fig. 11A, B). The anhydrite tends to be concentrated in the sandy layers (including sandstone dikes). Whereas many of the interbedded sandstones (not shown) have sedimentary structures that indicate fluvial transport, others are more likely to be of eolian origin. These sequences are interpreted as having been deposited on mudflats in an evaporitic environment, i.e., a sabkha.

From the local association of fluvial sands with considerable thicknesses of bedded clays, the relatively low concentration of nodular anhydrite, and the lack of evidence suggestive of a nearby marine environment (stromatolites, marine faunas), these sediments are interpreted as having been deposited in temporary desert lakes that were supplied with sediment and water by ephemeral streams. These lakes undoubtedly

were temporary, because some of their mud-cracked sediments also were interbedded with eolian sands. In fact, the cause for the existence of such lakes could be that the wadi channels became blocked by a belt of sand dunes. The hydrostatic pressure required to inject a sandy slurry up through cracks in an otherwise impervious clay cover could have been provided by a flow of water within the wadi sands; such water is known to remain within the sands and gravels of wadis long after surface signs of flowing water have ceased.

Dune Environment

The well-laminated sandstones seen in Figure 12 are representative of the sedimentary sequence that generally overlies that of the wadi environment. The sequence comprises several units of red sandstone separated by intraformational unconformities (u). Each sedimentary sequence commonly starts at the base with subhorizontal, finely laminated sandstones that truncate the steeply dipping (20-27°) sandstones of the underlying unit. Upward, the dip of the laminae increases and sets of laminae are more uniform in grain size. Just below 24 (Fig. 12), subhorizontal laminae seem to be overlain directly by laminae that are inclined at a relative angle of 17°. Although a part of the core is missing, this is believed to represent the original relation of a foreset sand that was deposited over a subhorizontal surface (compare with the foreset laminated sands overlying the clay zone at a depth of 5 ft in Fig. 10).

The grain size of these sands ranges from very fine to medium and, locally, particularly near the base of an intraformational sequence, may be coarse. The finer grains are subangular and the coarser ones subrounded or, rarely, rounded. Frosted grains are common. No argillaceous material is present apart from authigenic clay. The sandstones generally are cemented with hematite and authigenic clay, but locally dolomite and anhydrite are important as cements, together with minor authigenic quartz. Depending on the amount of cement present, the sandstone may be hard or quite friable. Where primary porosity is preserved, these sandstones form the main reservoir rock for the Rotliegendes gas.

The whole sequence seen in Figure 12 is interpreted as consisting of dune sands deposited in a desert environment. The dip direction is rather uniform. The lack of dips that approach or exceed 30° suggests that the cores could have been taken through the flank of a self dune. However, the regular occurrence of subhorizontal laminae at the base of most intraformational units renders
Fig. 11—Slabs cut from cores of same sequence from two different localities. They comprise alternations of sandstone and shale in which shales are cut by sandstone dikes. Small nodules of anhydrite, especially in sandstone, suggest an evaporitic environment. In right slab, photographs of matching ends of two consecutive pieces were spliced together. In left core, poorly developed adhesion ripples start a few centimeters above 3-cm gap. Rotliegendes, Netherlands.
Permian Rotliegendes of Northwest Europe

Fig. 13—Stereographic polar nets (upper hemisphere) of poles to Rotliegendes dune bedding. A and B are from two areas of outcrop in northern England. C and D are from wells in southern North Sea. Deduced paleowind directions are indicated. Corrections for low-angle tectonic tilt have not been made. Dashed lines indicate areas of typical barchan (a) and seif dune bedding attitudes (b).

This unlikely, because horizontal laminae seem to be associated most commonly with barchan or transverse dunes. Continuous dipmeter measurements taken from wells spread over a large area of the Netherlands and the North Sea indicate a westward regional dip for these dune sandstones (Fig. 13).

The slip faces of modern transverse dunes have a maximum slope of 33-34°. In the Rotliegendes, the dune bedding commonly reaches a dip of only 25-27°. Much of this difference can be accounted for by compaction.

The plotted distributions of bedding attitudes found in both outcrop and well cores (Fig. 13) indicate that probably both barchan (or transverse) and seif dunes formed in the Rotliegendes desert (Glennie, 1970, Fig. 68). There is at present insufficient evidence to be certain which dune type is the more common, but from a study of the distribution of bedding attitudes measured at many different localities, seif dunes are more numerous in parts of Britain (Devon, Vale of Eden, Durham) and west Germany (Bad Kreuznach) where high local relief may have resulted in the strength of the prevailing wind being accentuated by convection over adjacent mountain peaks. Such convection winds, especially in areas close to the Variscan mountains, are also likely to have caused eolian foresets to dip upstream, irrespective of the direction of the prevailing wind.

Away from the mountains, however, barchan or transverse dunes seem to be the dominant type. Figure 12 is typical of uniform sequences (Fig. 9A) that are more than 100 m thick. It is likely, therefore, that these sequences were deposited as transverse, rather than barchan, dunes in a sand sea (cf. Fig. 7). Because most of the preserved bedding in these transverse dunes probably was formed as avalanche slopes, there is little likelihood of many ripples (R, in Fig. 12) being preserved.

An intraformational slump is present just below 39 ft (Fig. 12). Preservation of this feature in dune sands indicates that the surface of the dune probably had been dampened, possibly by rain, become heavy and unstable, and slid down over the underlying dry dune sand (Glennie, 1970, Fig. 92).

Interdune Sabkhas

Some sequences of dune sands, especially in the upper part, are interbedded with horizons of small, irregular, and more or less horizontally-bedded adhesion ripples (Fig. 14). The sand grains within the ripples range from fine to coarse and generally are poorly sorted. Some laminae are slightly silty, but few are argillaceous. In the left-hand core (Fig. 14A), the underlying steeply dipping eolian sandstones are truncated and sharply overlain by the ripple horizon. In core B, gently dipping eolian sandstones merge upward into subhorizontal low-relief ripples that, in turn, grade almost imperceptibly into the overlying finely laminated eolian sandstone.

From the considerable thickness of continuous dune sand underlying the adhesion-ripple sequence, the general lack of argillaceous material within the ripples and the absence of ripples in correlated intervals in nearby outstep wells, it is inferred that these adhesion ripples formed in an interdune area of a sand sea far from any direct fluvial influence. The moisture probably was derived from groundwater whose table remained close to the interdune surface of a subsiding continental basin.

The irregular truncation of dune sands in Figure 14A could have resulted from differential deflation of a salt-cemented sabkha surface. P. J. C. Nagtegaal (personal commun., 1970) has noticed similar irregularly deflated surfaces in the Namib Desert that are covered by adhesion ripples. Such a process is coupled with a relative rise
Fig. 14—In left core, steeply dipping eolian sandstones are truncated and overlain by 9 in. of adhesion ripples followed by gently dipping, finely laminated eolian sandstones. In right core, gently dipping eolian sandstones are overlain by low-angle adhesion ripples that grade up almost imperceptibly into finely-laminated eolian sandstones. Rotliegendes, North Sea.
in the water table, the solution of cement in earlier surface deposits, and reprecipitation of the cementing halite near the new air/groundwater interface. That gypsum crystals also were precipitated in some places is evidenced by the presence of small anhydrite nodules or early anhydrite cement in some Rotliegendes sequences of interdune adhesion ripples. The presence of salt-encrusted sand grains implies that these sequences of interdune ripples were deposited in interdune sabkhas (Fig. 7).

In parts of the northwest European Rotliegendes basin, dune sands, with or without horizons of adhesion ripples, are continuous to the top of the succession, where they are overlain by the Kupferschiefer and succeeding Zechstein evaporites. In most places, however, between the Kupferschiefer and underlying undisturbed dune sands, considerable thicknesses of eolian sands seem to have been homogenized by flowing water in much the same way as described for those shown in Figure 10. The sandstones commonly show slump structures within the homogenized sequence (Fig. 9A) and, in places, two or three episodes of slumping can be seen to have succeeded each other (see also Glennie, 1970, Fig. 93). This implies that there was considerable local dune relief prior to flooding. In several places, temporary development of dune sand between two stages of slumping was noted, suggesting that the flooding associated with the Zechstein transgression did not take place as one single catastrophic event.

Desert Lake and Its Marginal Sabkha Deposits

In the southeastern part of the Rotliegendes basin, the main eolian interval is succeeded by a monotonous succession of red claystone with minor sandstone and siltstone, and local concentrations of nodular anhydrite (Ten Boer Member). Some thin sandstones possess sedimentary structures that are characteristic of aqueous transport; others are probably of eolian origin.

Some of the horizontally bedded claystone sequences are mudcracked and curled, indicating subaerial exposure and desiccation. In other places, systems of sandstone dikes cut through several layers of claystone (cf. Fig. 11A, B). Anhydrite tends to be confined to the sandy layers (including sandstone dikes) and is rarely bedded. Adhesion ripples are common (Fig. 15A, B) and are associated with nodular anhydrite, especially toward the top of this sequence. Their amplitude is usually up to 5 cm or more and their appearance is more contorted than those of the almost anhydrite-free interdune sabkhas.

The combination of fluvial sands, silts, and clays, together with anhydrite nodules and the presence of sandstone dikes, curled clay flakes, and adhesion ripples, leads to the interpretation that these sediments were deposited in a sabkha environment. The lack of indications of a nearby marine environment (stromatolites, Foraminifera), implies that the sabkha was probably of the inland variety.

The greater amplitude and complexity of the adhesion ripples found in this sequence possibly is related to a sabkha surface that was repeatedly subjected to deflation. Exposure of large gypsum crystals with their long axes subvertical is a common sight on some modern sabkha surfaces (Fig. 8). Later adhesion of clay, silt, and fine sand to the sabkha surface would involve deposition against and over the crests of these crystals. The ensuing volume reduction (38 percent) as the gypsum crystals altered to anhydrite nodules after burial would result in further distortion (and even fracturing) of the adhesion laminae.

The sabkha sediments illustrated in Figure 11 contain many sandstone dikes, but adhesion ripples, although present, are not common and never exceed a few centimeters in thickness. There seems to be an important difference in environment of deposition between the two sequences seen in Figures 11 and 15.

To obtain a thick succession of adhesion ripples, the level of the water table must keep pace with sedimentation. Where adhesion ripples are scarce, one can assume that either there was no eolian supply of sand or silt, or the water table was at a greater depth below the desert surface than the height to which water could rise by capillary flow. It seems likely that the sabkha environment of deposition implied for the sequences seen in Figure 11 was temporary and probably was associated only with the flooding by surface waters of a relatively small desert depression. The near-surface groundwater required for the formation of continuous and widespread sequences of adhesion ripples implies the existence of more permanent sabkha conditions.

The inland sabkha sediments of the Ten Boer Member grade laterally (and locally vertically) into a clayey sequence in which halite is a characteristic component. This sequence has been referred to by several authors (Schott, 1944; Deecke, 1949; Richter-Bernburg, 1955; Kent and Walmisley, 1970) as the Haselgebung facies of the Rotliegendes. According to Kent and Walmisley, the Haselgebung is the dominant facies in the central (north Germany) part of the Rotliegendes basin where the thickness of the Rotliegendes may exceed 1,500 m.
Fig. 15—Slabs of very argillaceous adhesion ripples cut from cores of Ten Boer Member from two localities. Small anhydrite nodules suggest evaporitic environment. Rotliegendes, Netherlands.
Permian Rotliegendes of Northwest Europe

In this sequence, halite may account for up to about 30 percent of the total thickness. Anhydrite is almost wholly lacking except as anhydritic clays and as cement. Indeed, Richter-Bernburg (1955) has reported the absence of any calcium or magnesium sulfates and carbonates associated with the halite. In the south, the halite is gradually replaced by red claystone and siltstone that take on as aspect similar to that of the Ten Boer inland sabkha deposits.

With one exception, the sediments of both the Haselgebirge facies and the associated Ten Boer sabkha deposits are completely devoid of marine faunas. The exception has been described by Plumhoff (1966) and consists of one thin marker bed (Bänderschiefer) of yellowish or olive-gray marl about 2.5 m below the base of the lowest Zechstein limestone; it contains a rich marine fauna which is identical with that found in the overlying Zechstein and thus should be related to the Zechstein marine transgression.

From the almost total lack of marine fauna and the nonmarine composition of the Haselgebirge evaporities, this sequence is interpreted as the sediments of a desert lake. The salts of such a lake are unlikely to have been derived from a nearby ocean. Following Brunstrom and Walsley (1969), the writer rejects the interpretation of Richter-Bernburg (1955) that calcium and magnesium salts were precipitated from marine water onto a shallow shelf farther northwest and that the residual unevaporated water, now enriched in Na and Cl ions, was then carried into the Rotliegendes continental basin where only halite was deposited. With the widespread spread though limited evidence of fluvial transport of sediment within the Ten Boer Member, and the alternating wadi and eolian sediments of the Slochteren Member in the eastern Netherlands, north Germany, and Poland, it seems likely that a desert lake occupied the site of a large depression marking the center of the Rotliegendes basin that was supplied—perhaps annually—with relatively fresh water derived from the Variscan mountains in the south.

Support for this interpretation is given by R. J. Murris (personal commun., 1970), who noticed that claystones within the Haselgebirge facies are correlative with clearly fluvial zones of the southern basin-margin facies (Slochteren Member). The halite zones of the Haselgebirge facies would thus correlate with periods of eolian sedimentation in the basin marginal areas. Under conditions of a seasonal supply of water and high rates of evaporation, the level of the desert lake would fluctuate considerably, and exposure of the marginal sabkha flats during the dry season would result in desiccation and mudcracking of the clays and widespread deflation of the sabkha surface. This interpretation appears to be consistent with the lithologies and sedimentary structures observed in well cores.

In this respect, Shearman (1970) in a study of salt deposits of desert coastal areas, observed, "Nodular anhydrite requires the presence of a pre-existing host sediment, whereas halite does not; nodular anhydrite develops in raised positions relative to sea level and the groundwater table. Layered halite, on the other hand, is formed in depressed environments where marine waters are impounded." The respective environments of crystallization of nodular anhydrite and bedded halite are therefore mutually exclusive.

Shearman (1970) has shown that single halite layers can result from several inundations by salt water, and that beds of halite can develop in shallow depressions that are evaporated repeatedly to dryness. It is not necessary, therefore, to invoke a deep-water origin for bedded halite. Each layer of salt need not represent an annual precipitation cycle. Although Shearman's observations concern sediments that were flooded by marine waters, the same principles are thought to apply to the salts associated with desert lakes.

The Ten Boer sabkha facies has already been shown to have resulted from alternate flooding and desiccation. To accumulate a thick succession containing nodular anhydrite like the Ten Boer facies, relative subsidence is necessary. It is likely, therefore, that the adjacent Haselgebirge lake facies was deposited in a broad shallow depression, where subsidence kept pace with sedimentation; because this facies was not subjected to subaerial desiccation, the lake probably was supplied with water seasonally, and it partly dried out annually.

The arrival of the marine Zechstein fauna of the Bänderschiefer marks the first incursion of ocean water into the Rotliegendes continental basin. The homogenized and slumped sandstones at the top of the dune sands farther southwest seem to correlate with the Bänderschiefer in stratigraphic position and thus also mark the earliest marine incursion of the Zechstein Sea.

Rotliegendes History and Paleogeography

Structural Framework

Toward the end of the Carboniferous (Stephanian), the Saalien phase of late Hercynian (Variscan) orogentic movements resulted in the uplift of a range of highlands (Variscan mountain chain) over an area that stretched from the southern
Fig. 16—Generalized Rotliegendes facies distribution, northwest Europe. Boundaries between areas of different facies are somewhat arbitrary, being based on their relative abundance. Major gas fields include (1) Groningen, (2) Indefatigable, (3) Leman, and (4) West Sole.

British Isles (St. George’s Land, Fig. 16), through Belgium (Brabant Highlands) and northern France, to central Germany.

Coincident with denudation of the Variscan mountain chain and its block-faulted northern foreland, sedimentation started in a WNW-ESE-striking basin which covered large parts of northern Europe and which crosses the Variscan chain in Germany. This basin is bounded on the north by the Mid-North Sea high and the Fyn Grinsted (Ringkøbing-Fyn) high extending from England to the Baltic.

North of this positive trend, other Rotliegendes basins may be present, but their limits are virtually unknown. Although Rotliegendes-type sediments have been cut during drilling, no clear picture has emerged of their environment of deposition, paleogeography, or age relations with the Rotliegendes sediments found farther south.

The broad outline of the Early Permian basins of northwest Europe has been indicated by Bar tenstein (1968, Fig. 10); they seem to conform with the Zechstein salt basins as shown by Hey broek et al. (1967, Fig. 3). The Rotliegendes facies distribution shown in Figure 16 is based on the maps of Bartenstein (1968), Krason (1967), Helmuth (1968), Brunstrom and Walmsley (1969), Sorgenfrei (1969), and others.

Lower Rotliegendes

Widespread volcanic activity in the Late Carboniferous and Early Permian coincided with the Variscan orogeny. Volcanic activity was particularly pronounced in the area east of the Netherlands and in the Oslo graben (Fig. 16). Volcanism also occurred in smaller depositional basins in southwest England (Devon) and in the Central Valley of Scotland (Mauchline basin).

In Devon, the Exeter Volcanics have a radiometric age of about 280 m.y.; this places the time of deposition of some of the associated wadi conglomerates and dune sands in the Stephanian (Laming, 1965, 1966). In southwest Scotland, plant impressions from the sediments associated with the Mauchline Volcanics of the Central Valley also suggest a Stephanian time of volcanicity (Mykura, 1965, as modified by Wagner, 1966). There, the lavas are partly interbedded with, and are succeeded by, dune sands.

At the same time there was almost no volcanic activity in the Midlands of England. The Westphalian Coal Measures are succeeded by the Etruria and Keele Beds of the “Barren Coal Measures.” According to Edmunds and Oakley (1958), the Keele Beds are composed of red and purple sandstones and bright-red marlstones and claystones that show evidence of desiccation.
cracks and precipitation of calcium carbonate following evaporation of carbonate-rich water. This sequence is now known to be of Stephanian age. A brief return to coal-forming conditions was followed by deposition of the Lower Permian (Autunian) Enville Beds, consisting of locally thick accumulations of conglomerate in a succession of red marlstones and calcareous sandstones overlain by another conglomeratic sequence with angular clasts. These rocks are the products of erosion of the nearby highlands and were deposited as conglomerate fans. On the basis of desiccation-cracked surfaces, raindrop impressions, footprint beds, etc., the whole sequence was considered by Edmunds and Oakley (1958) as having been deposited under semi-arid conditions. Fossils from the Enville Beds include impressions of the leaves of Waichia and Asterophyllites and silicified wood of Dadoxylon and Cordaites. The Enville Beds (or local conglomeratic sequences having other names) are overlain by the “Lower Mottled Sandstone,” a sequence of predominantly dune-bedded rocks that have been classified by many authors as Lower Triassic “Bunter” (Sherlock, 1947; Edmunds and Oakley, 1958), but which are believed by Shotton (1956) and the writer to include Permian desert sedimentary rocks.

Meanwhile, in the eastern Netherlands and northern Germany, Stephanian to Early Permian volcanicity resulted in the deposition of considerable thicknesses (up to 1,000 m or more) of quartz porphyries, spilitic lavas, and tuffs. Their penecontemporaneous products of erosion gave rise to interbedded conglomerates and conglomeratic sandstones that, in the Thuringerwald area of central Germany, are associated with coal seams (Lützner, 1969); boulders of similar volcanic rock form an important part of the components of the lower Slochteren conglomerates in east Netherlands (de Booy, 1968), central Germany (Lützner, 1969), northern Germany (Fabián et al., 1962), and in Poland (Krason, 1967).

Over the Fyn Grinsted high, Rotliegendes sediments are generally absent except for thin conglomerates; along the north flank of the high about 500 m of Zechstein rocks unconformably overlie Silurian and Ordovician graptolitic shales, whereas over the high the Zechstein may be totally absent. On the southern flank of the high, however, Lower Rotliegendes volcanic rocks are present (Sorgenfrei, 1969; Sorgenfrei and Buch, 1964). A similar situation is found on the Mid-North Sea high, where only relatively thin Zechstein deposition took place (Heybroek et al., 1967), but Rotliegendes volcanic rocks are not known.
North of the Fyn Grinsted high, Lower Rotliegendes volcanic rocks, conglomerates, sandstones, and shales are present in the Oslo graben (Holtedahl, 1934). In southeast Germany, Early Permian (Autunian?) coal seams associated with volcanic rocks (Schwab, 1968; Reichel, 1968) are evidence of locally prolific vegetation in basins within the Variscan orogenic belt that received relatively high rainfall: the sediments of these basins grade laterally into red sandstones, however.

Rotliegendes Paleogeography and Climate

As mentioned, the Lower Rotliegendes volcanic rocks have not been found in the southern part of the North Sea and west Netherlands, even though volcanic rocks are components of some Slochteren conglomerates. The available evidence suggests that for the most part the desert sediments that make up the Rotliegendes in this area were deposited over a continental surface consisting mainly of Carboniferous (Westphalian) sandstones, shales, and coals. It seems likely that this pre-Rotliegendes desert surface had a somewhat block-faulted structural configuration like that described by Kent (1949, 1966, 1967) for eastern England and suggested for the southern North Sea. A similar pattern of faulted blocks in Germany, uplifted in the south and subsided in the north, has been described by Hoyer (1967). Although continental deposition may have been almost continuous from the late Westphalian through the Stephanian to Early Permian time in some areas, other parts of the basin were subjected to erosion far into the Permian. In this respect it is significant that, in the eastern Netherlands, Westphalian D and Stephanian strata are overlain by lower Slochteren sediments containing components derived only from Westphalian C or older Upper Carboniferous strata (de Booy, 1968).

From grain-size distribution of the desert sediments in the Rotliegendes one gains an impression of deposition in a Permian desert with a seasonal fluvial influx from the Variscan mountain chain, into a large desert lake whose water level fluctuated considerably over its surrounding sabkha-covered margins. This lake (Haselgebirge facies) must have been at least 1,000 km long, with its maximum extension (Ten Boer facies) reaching 1,500 km. It fluctuated between 300 and 400 km in width. This involves an area greater than that of the present Caspian Sea.

Much of the sediment carried northward in water-filled wadi channels probably was deposited before it reached the lake because of the lack of a continuous flow of water. The un cemented surface sediments within and on either side of the wadi channels were subjected to deflation by a prevailing wind from the east or northeast, and sand-size particles were redeposited in sand dunes farther west. Some sand grains in the Slochteren dune sands consist of high-grade metamorphic minerals that were not derived from the same source as the Slochteren conglomerates (de Booy, 1968); these grains may have been transported by the wind from outcropping areas of the Baltic shield. As the prevailing wind blew from the east or northeast (Figs. 13, 16), deflation of the northern borderlands of the Rotliegend basin could be the reason for some of the silts found in the Haselgebirge facies.

From the limited amount of fluvial sediment shown in the generalized section (Fig. 9A), some of the wadis probably terminated within the dune sands of the western part of the basin, under local inland sabkha conditions of deposition. The greater percentage of fluvial sediment found in the eastern part of the basin (Figs. 9B, 16), the coarseness of some of the components found in this sequence, and their correlation with the clays of the Haselgebirge facies suggest that the wadi channels terminated at the edge of the Haselgebirge desert lake. The wadis of Germany and Poland (Fig. 16) were possibly the main routes along which the lake's supply of fresh water was maintained. In northern Poland, away from the rain-producing influence of the Variscan highlands, sequences of Rotliegendes sediments, in the form of red gypsiferous sandstone and claystone, give evidence of an arid climate, and carbonate-cemented conglomerates (Lacka and Motyl-Rakowska, 1968) indicate deposition in a fluviolacustrine environment that was probably subjected to some eolian sand transport (Krason, 1967). The spatial interpretation of these two contrasting sequences is shown schematically in Figure 17. For simplicity, only four halite zones are shown in this figure, although many more exist.

The general impression gained from a study of literature concerned with the Rotliegendes, and from a comparison of these sediments with those of modern deserts, is that, coincident with the formation of the Variscan mountain chain during the Late Carboniferous, the climate became more arid; during Early and mid-Permian time the area north of the Variscan highlands could be considered as a desert, but within the highlands there was sufficient rainfall to permit probably seasonal transport of water and sediment northward to a very large desert lake.

The transgression of the Zechstein sea did not
Permian Rotliegendes of Northwest Europe

bring about an end to desert conditions in the marginal areas of northwest Europe. The fact that the Zechstein sea was the site of a major evaporite basin corroborates this statement. The influx of Zechstein water into the southern Rotliegendes basin probably took place over or around a low northern barrier formed by the Mid-North Sea and Fyn Grinsted highs. It flooded an area that, for a large part, was probably well below sea level.

The Zechstein transgression transformed the area of the former Haselgehirge continental desert lake into a basin within the huge, almost enclosed evaporitic sea. In the area of sand-dune development, the transgression seems to have taken place in two or three stages with local reversion to dune formation between each stage. Although the first breakthrough of the Zechstein sea may have been locally catastrophic (basal clastics), the finely laminated and bituminous nature of the Copper Shale and the base of its Lingula-bearing lateral equivalent in northeast England (Marl Slate), suggests quiet-water, even stagnant, conditions of deposition in an almost tideless sea.

The Zechstein did not cover the highland areas of the Variscan mountains, but did penetrate into some of the basinal areas of central Germany. Central and northern England east of the Pennines also was flooded by the Zechstein. However, palynological studies indicate that the Late Permian microflora found in the Hilton plant beds of the Vale of Eden (Eastwood, 1953) lack the open marine environmental microflora typical of the Zechstein, so that the Zechstein sea probably did not extend west of the Pennines (personal commun., R. F. A. Clarke, 1965). Thus the Late Permian evaporitic marls of the Vale of Eden and Lancashire are possibly the deposits of a continental desert environment. Across the western parts of north and central England and southern Scotland, continental desert conditions with local sand-dune formation probably were continuous from the mid-Permian until the Late Triassic (Rhaetic) transgression.

Analogies with Quaternary Deserts

Although the conglomerate and sandstone facies of the Rotliegendes are analogous to the sediments of modern deserts, there is no modern analogue for a desert lake as large as that indicated by the Haselgebirge evaporites. There is, however, also no modern analogue for the even larger inland sea that later gave rise to the Zechstein evaporites. As can be seen from a study of the distribution of evaporites during past ages (Kozary et al., 1968) salt deposition today is not on the same scale as during some periods in the past.

The northwest European Permian desert approaches in scale the area of the present Sahara, especially if one considers its possible former extension into North America. The present major deserts of the world are mostly between the latitudes of 10° and 30° north and south of the equator. The prevailing trade winds that cross these deserts blow towards the equator, but are deflected westward by the Earth's increased velocity at the equator (Glennie, 1970, Figs. 1, 2). The pattern of prevailing winds deduced from the Rotliegendes dune sands indicates that this Permian desert might have existed in the northern hemisphere between the paleolatitudes of roughly 10°N and 30°N, and that, since then, northern Europe has drifted north. Paleomagnetic data from northwest Europe support this hypothesis. Although the exact location of the Permian equator is open to some doubt (compare, for instance, the Permian isocline maps of Rutten and Veldkamp, 1964, and van der Voo and Zijderveld, 1969), it almost certainly lay south of the Rotliegendes desert (Fig. 18).

The Permian isoclines for northwest Europe also imply that this region has been rotated in an anticlockwise direction since that time. If this is correct, it means that the deduced Rotliegendes
paleowind directions must also be considered relative to the Permian equator with a resulting improvement in the analogy with modern north-ern Hemisphere trade wind deserts (Fig. 18).

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A reconnaissance of the Recent sediments of the Ranns of Kutch, India

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ABSTRACT

The Ranns of Kutch are flat salt-covered desert areas (sahkhas) which are just above the normal tidal range and may be regarded as supratidal flats. They are flooded annually by the storm tides of the Southwest Monsoon. As the waters recede and evaporate they leave behind a crust of halite, and gypsum crystals grow within the clays and sands. The increase in salinity of the interstitial waters as they are traced inland is reflected in the higher boron content of the clays. The Mg\(^{2+}/Ca^{2+}\) ratio of these waters increases from 3 on the coast, to 240 in the evaporite environment of the shoreline of Pachham Island.

Much of the sediment of the Ranns was probably once derived from the Indus and Nara rivers which used to flow into the western end of the Great Rann. Clay is now probably carried in by the monsoon storm tides after being transported along shore from the Indus, and also from the rivers of Kutch and Kathiawar. Coarser siliciclastic sediment is carried into the eastern inland portion of the area by the Luni and other intermittently flowing rivers and streams. Some sand and silt is blown into the Ranns from the surrounding hilly areas. Foraminifera are found concentrated in the sandier fractions of the sediments, having been transported there primarily by tidal currents and also by the wind. One species of *Ammonia* is almost the only fauna found in the sediments of the very saline environment around Pachham Island.

INTRODUCTION

Little or no work has been carried out on the Recent sediments of the Ranns of Kutch since Wynn's (1872) memoir, although several papers have referred to its geography and to earthquake phenomena in the area (see Platt, 1962, for references). The writers undertook a short reconnaissance in the autumn of 1964 to study these little known Recent sediments. Lack of time, great distances and few roads meant that only scattered locations could be visited. However, cursory as the results are, due to the lack of knowledge of the area, they were thought to be worth recording.
The regional setting and earlier history

The Ranns of Kutch (Figs 1 and 2) consist of two large plains separated by the hills of Kutch which rise to nearly 400 m, with Pachham Island possessing the only peak to exceed that figure: the greater is situated on the Indo-Pakistan border just east of the mouths of the Indus, the lesser, or Little Rann, lies between the mainland of Kutch and Kathiawar. They are analogous to the delta-flank depressions commonly found on the margins of deltaic areas (Russel & Russel, 1939). A fairly broad shallow continental shelf exists to seaward, stretching from off the Indus delta to the Gulf of Cambay southeast of Kathiawar.

The Great Rann extends more than 300 km from east to west, with a width of 80–100 km. The total area of the two Ranns that is subject to annual flooding by marine and land derived waters during the Southwest Monsoon is roughly 30,000 km².

The surface of the Ranns appears, for the most part, to be just above normal high tide level and are thus best considered as forming supratidal areas. They are flooded annually with sea water, to a depth of up to 2 m, when winds of the Southwest Monsoon, blowing from July to September, force the waters of the Arabian Sea up the narrow tidal estuary of Kori Creek into the Great Rann, and up the Gulf of Kutch into the Little Rann. As the monsoon dies away, the waters recede or evaporate and leave behind a crust of halite which covers large areas of the Ranns.

Flowing into the Ranns from the east are a number of rivers which rise in the Aravalli Range. They bring both sediment and some fresh and brackish water into the eastern part of the Ranns during the Monsoon. The hills of Kutch have small intermittent streams, some of which drain into the Ranns, but these only flow after rain brought by the Southwest Monsoon.

North of the Ranns lies the Rajasthan Desert, an area of more or less stabilized sand dunes but with some areas of mobile sand, which is bounded to the east by the Aravalli Range and to the west by the Indus valley.

Rainfall in this area is largely confined to the months of July–September when the Southwest Monsoon is blowing. The area of lowest rainfall—less than 4 in (10 cm) a year—is centred over the Indus valley southwest of the confluence of the Indus and Sutlej (Fig. 1). Over the Ranns of Kutch, annual rainfall ranges between 8 in (20 cm) and 15 in (38 cm) per year (Pramanik, 1952). Because of the high average temperature and low humidity, the potential rate of evaporation is exceedingly high.

The structure of Kutch appears to be controlled by faults which are aligned roughly west-northwest–east-southeast. They have resulted in a series of horsts and grabens with the hilly areas of Kutch forming one horst and the ‘islands’ of Pachham, Khadir, Bela and Chorar (Fig. 2) forming another. This area is complicated by north-northeast trending faults which separate the horsts into individual fault blocks. The intervening areas of the Bunnee, and the Great Rann to the north, occupy the sites of grabens. The relationship of the Little Rann to this structural pattern is not known.

Parts of Kutch and Kathiawar began to stand out as islands during the early Tertiary and have maintained that situation until the present, apart from modifications caused by changes in sea level during the Pleistocene, and the influx of much sediment into the area. A considerable part of this sediment was probably derived from the combined Indus and Nara rivers which spread southward towards the site of the present Indus delta with some of its waters escaping to the Arabian Sea via what is now Kori Creek. However, following artificial damming of the distributaries of the Indus which began around 1764 and, apart from disastrous floods in 1826 when the
Fig. 1. Map of the Ranns of Kutch, Rajasthan Desert and Indus River system.
upper Indus broke its banks and burst every artificial dam in the river, was fully effective by 1802, neither water nor sediment have flowed directly from the Indus into the Ranns.

There is strong evidence for believing that the River Sutlej, which now flows into the Indus, used once to flow into the Nara, the old water course of which can be traced through the western Rajasthan Desert to the Ranns of Kutch at Kori Creek (Fig. 1). Similarly, the Jumna, before it was captured by the Ganges to flow past New Delhi, once joined the Chautang, the Ghaggar and legendary Sarasvati and also contributed to the flow of the Nara. The Chautang and Ghaggar now lose their waters in the sand dunes of the northern Rajasthan Desert, but their dried-up water courses can be traced far beyond the limits of present water flow (Geddes, 1960, Fig. 4). Such rerouting of the rivers possibly accounts for a change from a plentiful to a less certain supply of water to the Harappa and Mohenjodaro river cultures between 3000 and 2000 BC (Ghosh, 1952). These changes have resulted in a loss of water to the Nara of about 20% of the combined Indus and Jumna flows (see flow volumes in Fig. 1) which was probably sufficient to bring about the drying up of the Nara in the arid regions marginal to the desert.
Sediments of the Ranns of Kutch, India

There has been considerable sedimentation in the two Ranns during the recent past as, in the 4th century BC, when Alexander the Great visited India, the Ranns were still marine gulfs and had several ports (Krishnan, 1952). As a result of sedimentation at the heads of these gulfs, the sites of the ports have since migrated westwards.

The area has suffered considerable earthquake activity in historical times. Wynne (1872) has shown that prior to a severe earthquake in 1819, the Bunnee and a large tract of land north of Lakhpat were above normal high tide level. During the earthquake, however, a large basin—Sindree Basin—and the area north of Lakhpat subsided by about 1–5 m and the Bunnee was flooded by a tidal wave. The northern limit of the Sindree Basin was marked by a 50 mile (80 km) long west-northwest–east-southeast fault scarp some 3–5 m high at the southern edge of ‘Allah Bund’ (‘Built of God’). Since Allah Bund provided no barrier to the Indus flood waters of 1826, it was probably not uplifted, but would have had that appearance when viewed from the salt-encrusted surface of the Sindree Basin.

THE RECENT SEDIMENTS OF THE RANNS OF KUTCH

Various environments of deposition found around and within the Ranns of Kutch were visited during the reconnaissance. Surface samples, short cores and water samples were collected. Some in situ measurements of pH and sediment temperature were made. The localities visited and their salient features are described in the following section. Some analyses of the sediments and water samples are shown in Table 1 and their locations are given in Fig. 2.

Mandvi

The upland mass of Kutch is bounded on its Arabian Sea side by sandy beaches and dunes which extend along the coast line, either directly fringing alluvial outwash plains or partly enclosing and protecting narrow intertidal and supratidal flats from the storms of the Southwest Monsoon.

At Mandvi, the outwash plain is cliffed and fronted by a wide sandy beach: this has a smooth low-tide terrace with ‘Cerithid’ type gastropods that produce trails, and some stranded starfish. This terrace is succeeded landwards by a wide smooth beach-face with burrows and the feeding pellets of a crab very similar to Scopimera sp., so common on the Persian Gulf beaches. A low ridge and runnel (similar to those on most North Sea beaches) are developed at the top of the beachface. Inside the runnel the beach rises to a crest, where there is a concentrate of shell debris with fragments of gastropods, echinoids and corals. The displaced fauna contrasts strongly with the indigenous fauna of crabs and small gastropods. Landwards of the beach crest is a wide berm without dunes.

The sediments of the beaches are clean sands with less than 2% silt and clay. The low-tide-terrace sample (GL 361) is composed of fine to very fine sand with a small admixture of coarser material. Both the beachface sample (GL 362) and the sample from the beach crest (GL 363) are composed of mixtures of coarse and very coarse

* Wentworth size grades have been used throughout.
sand with some medium-fine and very fine sands. These samples are noticeably bimodal and presumably represent sands that still inherit some of the ill-sorted nature of the outwash material from which they are undoubtedly partly derived.

The samples are only moderately rich in calcium carbonate: the low-tide-terrace sample and beach-face sample contain less than 10% \( \text{CaCO}_3 \) and the sample from the beach crest contains 14% \( \text{CaCO}_3 \) (see Table 1).

The sands contain large iron cemented rock fragments, well-rounded quartz grains, chert fragments, a little feldspar and well-rounded and polished shell fragments. There are some Foraminifera, spines and fragments of echinoid plates, polyzoan fragments, a few sponge spicules and some ostracods. Mica is present and probably also some chlorite. Crab burrows and feeding pellets are common. The chlorinity of the sea water is only slightly above normal at this location.

Khandla

Within the shelter of the Gulf of Kutch, the shoreline contrasts markedly with that of the open coast of the Arabian Sea at Mandvi. Wide mud flats, cut by tidal creeks with patchy salt marsh, characterize the coastline (Fig. 2). They are generally very reminiscent of the intertidal flats of the sheltered regions of the North Sea coast. Several large intermittent streams run into this embayment from the upland mass of Kutch and also from the Kathiawar Peninsula to the south. These must carry some sediment when they flow during the monsoon. The finer grades of this sediment are probably recirculated by the strong tidal currents within the embayment and finally settle on broad intertidal flats such as are found at Khandla. The composition of the clay grade of the sediment is not the same as that found in the Great Rann: montmorillonite is present in higher percentages in the Khandla samples (Table 1) and appears to be derived from the volcanic lavas of Kutch and Kathiawar (Stewart & Pilkey, 1966). However, the similarity in composition of GL 418 with samples from the Great Rann suggests that some Indus mud reaches this part of the coast to mingle with locally derived material.

The silty clays and clayey silts of the Khandla mud flats contain less than 10% sand. The sand fraction is made up mainly of colourless, brown and green micas, some of which are probably chlorites. Mixed with the mica are small Foraminifera, small lamellibranchs, ostracods, echinoid plates and spines, prismatic calcareous grains (probably produced by the breakdown of larger mollusc shells) and some sponge spicules. There are also a few quartz grains and the samples have a carbonate content very similar to that of the sands of Mandvi (Table 1, GL 413, 415).

In places, the intertidal flats have been reclaimed. Sections show brown oxidized silty clay with a sandy admixture, overlying a blue-grey silty clay with gypsum crystals and old plant roots. These silty clays are similar to the sediments exposed on the modern intertidal flats. Organic debris is, however, less obvious in them than in the latter, and the surface brown clay is richer in quartz. The lower blue-grey clay is richer (GL 317—16%), and the surface brown clay poorer (GL 318—6%), in calcium carbonate than the sediments of the present intertidal flats. This anomalous carbonate distribution is possibly due to leaching of calcium carbonate from the surface sediments and subsequent precipitation in the lower horizons during the periods of monsoon rains. The boron content of samples GL 413, 415 and 417 in all cases (see Table 1) indicates deposition from approximately normal sea water. This agrees with the
### Table 1: Mineral and water analysis of recent samples from the Burns of Kutch and their relation to environment

<table>
<thead>
<tr>
<th>Location</th>
<th>Lithology</th>
<th>Environment</th>
<th>Sample number</th>
<th>Carbonate</th>
<th>Calcite</th>
<th>Dolomite</th>
<th>Mg calcite</th>
<th>Gypsum</th>
<th>Quartz</th>
<th>Feldspar</th>
<th>Clay</th>
<th>Boron in clays</th>
<th>Carbonate car</th>
<th>Organic X-ray diffraction mineral analysis</th>
<th>X-ray diffraction</th>
<th>Boron analysis of clays</th>
<th>pH</th>
<th>Calcium</th>
<th>Mg</th>
<th>Na</th>
<th>K</th>
<th>P</th>
<th>Si</th>
<th>Cl</th>
</tr>
</thead>
<tbody>
<tr>
<td>Khandla</td>
<td>Slightly calcareous silty clay</td>
<td>Open tidal channel</td>
<td>KGM 398</td>
<td>7.9</td>
<td>42.5</td>
<td>12.2</td>
<td>6.4</td>
<td>10</td>
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<td>Open coast</td>
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<td>KGM 399</td>
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<tr>
<td>Bunere</td>
<td>Nan-calcereous luterpreled as and bioclastic sand</td>
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<td>KGM 400</td>
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<td>42.5</td>
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<tr>
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<tr>
<td>Sand</td>
<td>Slightly calcareous Sandy clays with very restricted toads</td>
<td>Runtle tidal channel</td>
<td>KGM 402</td>
<td>7.9</td>
<td>42.5</td>
<td>12.2</td>
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<tr>
<td>Slightly calcareous Salt covered reins</td>
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<td>KGM 403</td>
<td>7.9</td>
<td>42.5</td>
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<tr>
<td>Channel beta-coo</td>
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<td>KGM 404</td>
<td>7.9</td>
<td>42.5</td>
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</tbody>
</table>

**Notes:**
- Analyses carried out by KSEPL.
- So values for CaCO\(_3\) taken from similar analyses carried out by P. R. Bush, Imperial College, London.
- All water samples are surface samples unless otherwise noted. No clay was found on the beach at Mandvi, but its source could be identified on the beach or at least.
- All values not indicate pH buffer salinity values remain stable when heated under each other.
- Analyses carried out by KSEPL.
- So values for CL taken from similar analyses carried out by P. R. Bush, Imperial College, London.
- Composition of dolomite: Ca\(_{50}\)CO\(_{50}\) determined as well as the calculated are not corrected for about 10% ignition loss. For the spectre evolution by phosphoric acid and absorbed in standard Ba (OH)\(_2\). The percentage organic carbon is calculated from the percentage CO\(_2\) evolved in the assumption that no other carbonates are present.
- All carbonate is removed by treatment with hot phosphoric acid; the organic carbon is then met evolved using chromium trioxide in phosphoric acid. The percentage organic carbon is calculated from the percentage CO\(_2\) evolved.
- Analyses carried out by KSEPL.
- No clay was found on the beach at Mandvi, but its source could be identified on the beach or at least.
- All values not indicate pH buffer salinity values remain stable when heated under each other.
Fig. 3. Restricted environment of an intertidal creek at low water, and mud-flats partly covered with halophytic vegetation and with several 'burrow craters' of the mud-skipper *Periophthalmus* in the foreground. Khandla. Samples GL 413 and 414 are from this locality.

Fig. 4. Small mud flows on the crater rim formed from mud thrown out of the burrow by the mud-skipper *Periophthalmus* (Blegvård, 1944). The smaller holes in the foreground are probably made by small crabs. The pen is 13 cm long. Restricted intertidal mud-flats Khandla.

chemistry of the water from a neighbouring tidal creek which is only slightly concentrated when compared with normal sea water (see Table 1). During the intervening dry season, however, evaporation, particularly in the surface layer, could produce pore waters of very high salinity, and this may explain the higher boron content of the surface sample (GL 418).
Lakhpat

The coastline at Lakhpat, on the eastern side of Kori Creek, is also in marked contrast to the open Arabian Sea coast of Mandvi. Here again, as at Khandla, the environment is extremely sheltered and is the site of extensive mud flats, which abut against the older rocks of Kutch. Only a few small intermittent streams reach the shore near Lakhpat and much of the sediment probably originates from the Indus delta to the northwest.

At Lakhpat, the hills of Eocene limestone are surrounded by a narrow strip of outwash. The latter is worked on its outer edge to produce a poorly developed beach. This passes out into wide mud flats with a smooth surface. Crab burrows and small 'U'-shaped burrows, very like those of *Corophium* sp. found on the North Sea mud flats, are sparingly distributed. The fauna is here much more restricted in species than at Khandla, presumably because of the higher salinity of the waters at this site.

The surface sediments of the mud flats at Lakhpat are clayey silts with less than 10% of sandy material. The sand fraction is composed dominantly of mica, with some angular quartz grains and some gypsum crystals. The samples are noticeably lower in biogenic carbonate, but contain about the same amount of organic carbon as those from Khandla (Table 1). Nearer to the shore line the calcium-carbonate content of the sands is increased by the incorporation of Eocene Foraminifera and other derived carbonate material. The boron content of the clays from this locality (GL 367) indicates deposition under approximately normal sea water (the high salinity of the water sample from this location may well be due to the fact that it was taken from a pool which had been cut off from the sea for several days, so that it had become concentrated).

East of Lakhpat

The Miocene hills of clays, marls and siltstones of Kutch are flanked by a fairly narrow outwash plain cut by gullies of small intermittent streams. This outwash plain passes out into a sandy halite-encrusted intertidal flat with a crinkled surface (Fig. 5) like that of parts of the coastal sabkhas of the Persian Gulf. A few coarse sand grains are present on the surface and the sediment around the presumed high-water mark is extensively burrowed by crabs. On the inner intertidal flats, east of Lakhpat, the surface sediment consists mainly of fine and very fine sand with a variable mixture of silty clay and medium, coarse and very coarse sand; outwards these grade into silty sands and sandy silts and at the outermost station, the surface is composed of a silty clay. The sand fraction of the sediment is ill sorted, presumably because of the low energy of the sorting processes in this sheltered environment. The surface is smooth on the outer parts but there are some *Corophium* sp. type burrows and gazelle tracks. The halite crust on the flats becomes thicker as traced away from the land but it is everywhere sufficiently thin for the dun colour of the underlying sediment to show through and it soon disappears as the high-tide water line of the inner Kori Creek is approached.

The finer sand fraction consists mainly of clear or tinted subangular quartz, mica, brown earthy grains and other rock fragments with an occasional iron stained echinoid spine and gypsum. The paucity of organic remains, other than plant debris and probable derived material is most striking. The carbonate content is low and that which does occur probably originates mainly from derived skeletal debris and fine-grained carbonate brought in with the clay-grade sediment.
The clays are similar in composition to those found in the sediments of Lakhpat, indicating a similar provenance, although some cores show horizons richer in kaolinite. The latter may be an indication of increased sediment contribution from streams of adjacent outwash fans, but the sharp basal reflections seen in the X-ray diagrams suggest an in-situ formation of kaolinite.

The sands are well laminated (Glennie, 1970, Fig. 108) and silty clay laminae can be seen both in cores and pits at the edge of the intertidal flat, although sand layers are still present in the outer intertidal flats. Eocene Nummulites are scattered throughout the cores and are even found in the sediments of the outer flats. Gypsum, as well as being common in the sand laminae and layers, is also found scattered throughout the silty clays. The gypsum found in the sands appears to be coarser than that found in the clays (Fig. 6). The sediment is usually brown to reddish brown and, apart from a few black silty-clay laminae and patches of grey sand laminae in the sediment of the outer part of the intertidal flat, the environment seems nowhere to be reducing.

Paccham Island

As at the site east of Lakhpat, there is a fairly narrow outwash plain at this location, cut by gullies of small intermittent streams, flanking the older Jurassic rocks of Paccham. This outwash plain has been cliffed at its outer edge to form a small feature 1–1.5 m high. The outer parts of the outwash plain are topped by low dunes.

At the inner edge of the supratidal flat there is a thin sandy beach composed of gypsiferous sands and clays, built up into wavy laminations (Glennie, 1970, Fig. 107);
these are believed to be adhesion ripples, modified perhaps both by crystallization of halite and gypsum (see Glennie, 1972, Fig. 8) and by intermittent current action. This beach passes outwards into a zone with a crinkled surface, dun in colour, similar to that at Lakhpat. This zone is succeeded by a strip of old swash marks now preserved in hardened halite (these presumably indicate the position of the last high water) and form the inner edge of a glistening salt flat which increases to a maximum of 5–8 cm in thickness seawards.

The halite layer has crystallized into cubes. Many old swash lines can be seen on its surface, with occasional fragments of bushes and shrubs heavily coated with halite (Fig. 7) which were probably carried in by the last flood waters. Dome-shaped halite protuberances, which are hollow beneath and up to 20 cm high, are seen in places on the flat, and forces of crystallization of the halite have built up pressure ridges. In places, the thin halite crust has fractured and adjacent sheets have been thrust one over the other (Fig. 8 and see Glennie, 1970, Fig. 106).

The sands of the dunes and beach pass seawards into the sands of the inner intertidal flats and ultimately grade out under the halite into clayey silts containing approximately 40% sand, and then into almost pure silt.

The coarser grades of the sand consist of gypsum, well-rounded and frosted quartz grains, iron-cemented rock fragments, earthy rock fragments, jasper and plant debris. Some silty clays contain Cerithid-type shells, but these are more common in the sands of the inner flats (Fig. 9).

The finer part of the sand grade is composed of subangular, clear and tinted quartz, mica (colourless, brown and green), heavy minerals, carbonate grains, a few fragments of mollusc shells, and some Foraminifera and Ostracoda (mica is less abundant than at Lakhpat).

Generally, as at Lakhpat, the sand content increases upwards and shorewards. The sandy sediments of the inner intertidal flats are well laminated with thin lenses of heavy minerals and plant material. As traced outwards from the edge of the intertidal flats, the sand laminae decrease in importance until silty clays become dominant (Fig. 9). However, some sand laminae are usually present beneath the halite. The sediments appear to have been burrowed in places, the burrows being filled with sand. Gypsum is found scattered throughout the cores, both in the sands and in the silty clays. The gypsum in the sand grades is coarser than that found in the finer sediment (Fig. 9). In the sands, the gypsum crystals poikilitically enclose the sand grains whereas in the clays, the crystals tend to push the sediment apart as they grow. The nearby dune and beach sands are composed almost entirely of gypsum with small cerithid-like gastropods. The dunes have a fairly hard crust, which is thought to have been formed by the cementing action of the incorporated gypsum after wetting by rain.

The carbonate content of the sediments is fairly low and is comparable to that of the sediment found east of Lakhpat, and noticeably lower than that of those at Mandvi and Khandla. The higher percentages found in GL 385 and 387 (see Table 1) are due to local concentrations of Cerithid debris. The carbonate here, as with the sand grade at Lakhpat, is dominantly derived, with a little produced by Recent Foraminifera and other organisms. The fine-grained carbonate in the clayey silts was presumably partly carried in with silt and clay, in addition to being produced by local breakdown of shell debris. The X-ray results indicate the presence of possible Recent dolomite, which appears to have formed by reaction between the sparse calcium carbonate and the Mg²⁺-ion-rich saline groundwater (see Fig. 10).
Fig. 6. Four shallow cores taken from the shore into the Great Rann of Kutch east of Lakhpat and one from the channel of the Inner Rann. The sediments are comprised of alternations of clay with silt and sand (Wentworth size fractions given by cross-hatched columns). Superimposed on these sediment sizes are those of gypsum crystals, which grew post-depositionally within the sediment, and Nummulite tests derived from nearby outcrops of Eocene limestone. Water was available for sampling (Table I) only in the holes represented by cores GL 371 and GL 372.
Fig. 7. Salt-encrusted driftwood near Pachham Island. The 'driftwood' comprises twigs and branches of plants that were probably carried into the area by the monsoon flood waters, and have since been completely encased in halite crystals.

Fig. 8. A part of the thin salt crust near Pachham Island heaved up into 'teepee' structures and over-thrusts as the result of localized horizontal compression (forces of crystallization) caused by growth of halite crystals. The uplifted salt terminates against a linear thickening of the halite crust thought to represent a 'swash' mark as the flood waters of the Rann evaporated and contracted areally. The black camera case is 35 cm long.
Fig. 9. Five shallow cores taken from the shore into the Great Rann of Kutch near Pachham Island. The sediments comprise alternations of clay with silt and sand (Wentworth size fractions given by cross-hatched columns). Superimposed on these sediment sizes are those of gypsum crystals, which grew post-depositionally within the sediment, and of *Cerithium* shells. Analyses of the water samples taken from three of the holes are given in Table 1. The $\text{Mg}^{2+}/\text{Ca}^{2+}$ ratio of the water in sample GL 396 was 240. Water was not available for sampling in the holes represented by GL 386 and GL 401.
The composition of the clay grade of the samples from Pachham are very similar to those of Lakhpat. In some horizons, however, there appears to be more kaolinite, which may be explained by the growth of authigenic kaolinite within the sediment. The boron content of the clays (see Table 1) appears to be a faithful record of the high salinity of the interstitial waters in the one sample examined. Only in the upper few centimetres are the clayey silts black reduced although the sand in the outer parts of the flat may show black patches. Below this, the clayey silts and sands are everywhere brown to reddish brown.

The Bunnee

The upland masses of older rock which forms Kutch are surrounded by alluvial outwash fans which are cut by small intermittently-flowing streams. Between Kutch and Pachham Island, a flat plain supports a sparse vegetation of scrub and grass. It is cut by numerous small channels and pans in which salts crystallize. The surface sediment is extensively cracked and is impregnated with halite and gypsum.

Most of the sediment examined from this locality consists of gypsiferous-sandy clay similar in outward appearance to some of the sediment seen at Lakhpat or Pachham. In general, the sand fraction consists of abundant gypsum crystals with some quartz, a little feldspar, mica and some heavy minerals. The content of calcium carbonate is very low and is entirely due to the presence of tests of Foraminifera.

Although the Bunnee is now above the range of monsoon storm tides and is only flooded by water from streams and monsoon rains, a comparison of the boron content of the clays here with that of the clays from Pachham (Table 1) suggests that they were probably deposited under conditions similar to those now existing over the halite-covered parts of the Rann. This interpretation appears to be corroborated by the similarity in the unusual foraminiferal distributions of the two areas (see Fig. 11).

The Inner Rann

Only a narrow strip of lowland separates upland Kutch from Chorar Island and the mainland of India to the east. Here, the surface of the sediment is cracked and impregnated with halite. The sediments are too hard to allow deep penetration of the core barrel (Fig. 6). They consist of sands and a mixture of sands and silty clays grading into silty clays with an admixture of sand. The sandier sediment is found near Kutch, while the fine-grained material occurs further out on the flat.

The sediment is ill-sorted and grades from very coarse sand to very fine sand, with the major part falling in the fine-sand grade. The coarser grades of sand consist of rounded frosted quartz, white chert fragments, fragments of iron-cemented sandstone, black rock fragments (possibly volcanic) and worn iron-stained derived shell debris. The finer grades are composed mainly of clear and tinted quartz, heavy minerals, rock fragments and some gypsum.

The sediments are notably poor in calcium carbonate. The sand fraction is almost entirely devoid of skeletal debris, except that of molluscs. The clay grade of the samples from nearer Kutch show a higher montmorillonite content and a lower illite content than those from half way across the flat (see Table 1). The higher montmorillonite content may again be due to derivation from the weathered Deccan Traps.
Their boron content is, rather surprisingly, approximately that of a normal marine clay.

These sediments of the Inner Rann are probably fluviatile in origin, since during the monsoon this part of the Rann is covered with brackish fluviatile water, but never with sea water.

DISCUSSION OF THE SEDIMENTS OF THE RANNS

The depressions occupied by the Ranns of Kutch were once gulfs of the sea, immediately following the late Pleistocene rise in sea level. Originally, sediment was carried into these depressions by the Indus, Nara and other rivers of the western part of the Indo-Gangetic plain and during the monsoon by local streams draining upland Kutch and Kathiawar. With the drying up of the Nara and other streams and the displacement of the Indus tributaries, much of this direct supply of sediment was terminated. During this period, the rivers Luni, Banas etc., together with the local streams of Kutch and Kathiawar, supplied some sediment to these depressions during monsoon seasons, but now the Luni almost loses itself in its lower reaches in sands of the eastern edge of the Rajasthan Desert. During the monsoon season, the coastal waters were forced up into the Ranns by high winds and must have brought considerable fine-grained sediment into the area from the mouth of the Indus. It is probable that fine-grained sediment from the latter river was being added to the outer parts of the Ranns by longshore transport and transport by tidal currents throughout the year. The findings of Stewart & Pilkey (1966) would appear to support this view.

It is interesting to note that Oldham (1893) reported the presence of considerable quantities of mica in the Indus sediments, and this seems to be reflected by the common occurrence of mica in the sediments of the Ranns. It appears that these processes, occasionally interrupted by local subsidence and uplift, have ultimately filled these depressions to form the present Ranns.

Today, the processes continue essentially unchanged. The annual flooding by the sea, driven by the Southwest Monsoon, still brings fine sediment into the Ranns. The streams draining the borders of the Ranns transport coarser material from the hills during this season and deposit it where they debouch into the Ranns. The finer material is carried further out onto the Ranns to mingle with the marine derived silt and clay. Small waves rework the outwash plains around the edges of the Rann at these high-water stages and sometimes cut cliffs in the outwash plains. As the water recedes, evaporation proceeds and halite crystallizes out to form the extensive surface crust. The interstitial waters become concentrated and gypsum crystals grow within the sediment. Waters enriched in magnesium ions (Fig. 10), as a result of precipitation of other salts, react with the sparse carbonate present in the sediment to form dolomite. The outwash material is reworked by the wind and some sand, very rich in gypsum, has been built into low dunes along the edges of the Rann. Also, some of the crystallized halite and gypsum is blown even further landwards into the Rajasthan Desert. Aeolian transport of salt northeastwards into Rajasthan, presumably as fine-grained material, has been earlier suspected (see Auden, 1952; Sarin, 1952). Similarly during other seasons (e.g. northeast winds of the ‘Little Monsoon’), considerable dust is probably blown onto the Ranns from the surrounding arid areas.
Fig. 10. A comparison of the ionic composition of interstitial-water samples from the Ranns of Kutch and the Persian Gulf. The Persian Gulf samples were collected during an Imperial College/Socoby-Mobil Expedition in 1965 (see Evans et al., 1969) and analysed under the supervision of P. R. Bush.
Sediments of the Ranns of Kutch, India
Much of the halite crust is probably dissolved each year by the flooding marine waters and it does not therefore appear to attain any great thickness. Halite layers with thicknesses of 90-120 cm have, however, been described by Oldham (1893) from near Sindree (Fig. 2). The gypsum that has grown within the sediment will probably remain and may gradually increase in percentage with time. Although no information is available for the deeper sediments underlying the Ranns, they probably consist of silty clays and clayey silts rich in gypsum, passing into sandier sediments around the edges of the Ranns and also, at depth in the northwest where the Indus and Nara must have flowed into the area. These sediments are probably unfossiliferous or only sparingly fossiliferous, but they could contain an increasing marine fauna with depth. They are probably underlain by the early Holocene marine sediments that must have accumulated when the Ranns were gulfs of the sea.

The carbonate-rich areas shown by Stewart & Pilkey (1966, Fig. 8) as occupying the edge of the continental shelf are apparently too far removed from the effects of the shoreward transport for any of this material to be carried into the Ranns. Presumably, the vast quantities of siliciclastic sediment carried to the sea by the Indus inhibit the production of carbonate-forming organisms closer to the coast. With reduced river flow at the delta mouth caused by recent irrigation work (United Nations Flood Control Journal, 1962, Figs. 9 & 10; Holmes, 1968, p. 375), the areas occupied by carbonate-rich sediment may spread, and in so doing will be brought within range of shoreward transport and the storm tides that flood the Ranns of Kutch. If this happens, the chemistry of the interstitial waters of the Rann sediments is likely to change. In particular, the magnesium content will be reduced as increased amounts of calcium carbonate are altered to dolomite.

The distribution of Foraminifera in the Kutch area (Fig. 11) is very interesting. The fauna found on the open coast at Mandvi can be distinguished from those found in the tidal channels at Khandla; those from the tidal estuary at Lakhpat differ from that near Pachham Island which, in the smaller size ranges, is almost monospecific. Faunal differences between the open and restricted tidal channels at Khandla are not apparent. The similarity between the fauna of the Bunnee and that of Pachham Island strengthens the probability that the former represents an uplifted portion of the Rann.

Whether the animals actually lived in the place where their remains were entombed cannot be decided on the available evidence, except for obviously allochthonous assemblages in beach and aeolian deposits. The assemblages are distinct, which is difficult to explain in terms of selective transport alone. Conversely, since the Ranns are flooded for only about 3 months each year and gradually evaporate to dryness during the remainder of the year, with associated extreme salinities in the pore waters, it is equally difficult to imagine that a permanent population exists near Pachham Island. A logical compromise would be to assume that repopulation takes place annually, following transportation during the time of flooding. The most likely sources for repopulation would be the sea and those environments within the Ranns with a permanent body of water. The reason why only a few species, and especially Ammonia beccarii, assume such dominance in the Rann environment cannot be explained due to lack of knowledge about the exact ecological requirements of this form and the paucity of localities which have been sampled. The scarcity of fauna in the Inner Rann may reflect the inability of the monsoon floods to populate this area with Foraminifera.
### Foraminiferal Faunal Components

<table>
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<tr>
<th>Environment</th>
<th>Sample</th>
<th>Component In Percentages</th>
<th>Foraminiferal Numbers</th>
<th>% &lt; 150μ</th>
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<th>Faunal Components</th>
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<td></td>
<td>0 20 40 60 80 100</td>
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<tr>
<td></td>
<td>384</td>
<td></td>
<td>1,066</td>
<td>1,065</td>
<td>100</td>
<td>III</td>
</tr>
</tbody>
</table>

** = number of specimens per gramme of sediment (see text)

** = from total fractions >70μ

- **a** Pararotalla ozawai
- **b** Milolidae
- **c** Cellanthus craticulatus
- **d** Rotaliidae
- **e** Poroteponides cribrorepandus
- **f** Ammonia beccarii
- **g** Elphidium s.l.
- **h** Nonionella
- **i** Nonionella
- **j** Ammonia annectens

Fig. 11. The distribution of Foraminifera in the sediments of the Ranns of Kutch (determinations made and figure compiled by J. Brouwer).
A COMPARISON WITH OTHER AREAS

The Ranns are similar in many ways to low accretionary coastal plains in other parts of the world. They are similar to those built by intertidal-flat deposition around the borders of the North Sea (Evans, 1970), the coastal Sabkhas around the borders of the Persian Gulf (Evans et al., 1964; Evans et al., 1969; Kinsman, 1969), and the western side of the Gulf of California (Shearman, 1970).

They are similar to the former, particularly around the Wash embayment, in that they are sheltered bays where accretion occurs because of the shelter provided by the embayment, rather than by a coastal barrier as in the Dutch Wadden Sea or on the Trucial Coast (Evans, 1970) although Kutch could have acted as an offshore barrier.

They differ from the North Sea coastal plain and are similar to both those of the Trucial Coast and the Gulf of California in that they are subjected to arid conditions. Consequently, as in the latter two areas, they are the site of deposition of evaporites. As an area of salt-encrusted sand, silt and clay, the Ranns of Kutch may be termed a sabkha (Glennie, 1970, p. 197). Because the greater part of the Ranns is influenced directly by sea water, this area should be considered as a ‘coastal sabkha’. A relatively limited area in the eastern part of the Ranns is not covered by sea water, but is flooded annually by brackish water from ephemeral streams; this area is an ‘inland sabkha’. A transition zone from coastal to inland sabkha will occur approximately at the limits of marine and freshwater flooding. In the Trucial Coast, a similar transition occurs in the Sabkha Matti where the influence of marine flooding gives way to that of groundwater of fluvial origin (see Glennie, 1970, map encl. 1).

The striking difference between the Ranns and the Trucial Coast, however, is that sediments of the former are almost entirely detrital siliciclastic sediments with a low carbonate content, whereas those of the Trucial Coast sabkhas are characteristically carbonate rich with lower percentages of noncarbonate material.

The groundwaters of the Ranns, like those of the Trucial Coast, are probably derived from the nearby sea and diluted by intermittently flowing fluvial water, particularly in the easternmost areas and around their edges. The interstitial waters of the coastal sabkha sediments of the Persian Gulf are similarly diluted by subsurface brackish groundwater (and not by surface flow) which is probably mainly of wadi-origin.

The interstitial waters in both areas become exceedingly concentrated as a result of evaporation and are consequently similar. As might be expected, however, the waters do show some difference because of differences in their host sediment. As is shown in Fig. 10, precipitation of gypsum increases the $\text{Mg}^{2+}/\text{Ca}^{2+}$ ratios in the waters, thus creating favourable conditions for dolomite formation (see Bathurst, 1971, pp. 525–530 for general review). The $\text{Mg}^{2+}$ of the Trucial-Coast waters is lost in converting calcium carbonate into dolomite, whereas in the Ranns, it is largely retained in solution because of the paucity of calcium carbonate to convert to dolomite.

CONCLUSIONS

From the foregoing evidence, the following conclusions may be drawn.

(1) The Ranns of Kutch probably occupy the sites of tectonic depressions. They were once shallow marine gulfs following the post-glacial rise in sea level.
Changes in the course of the Sutlej and Jumna rivers and the drying up of the Nara permitted an extension to the area covered by sand dunes in the Rajasthan Desert, and this former source of fresh water and sediment was denied to the Great Rann of Kutch.

The Ranns of Kutch became infilled, probably about 400 BC or later. They then developed into areas of broad salt flats (sabkhas) as the result of annual flooding by the sea water forced across them by the storm winds and tides of the Southwest Monsoon.

The Ranns of Kutch are in a desert environment where the potential rate of evaporation is considerably higher than the annual precipitation. Most of the rainfall (8–15 in, 20–38 cm) occurs during the period of annual marine flooding and thus has no effect on the evaporites deposited.

The clay found in the Ranns of Kutch has both a detrital and an authigenic origin. The normal association of illite, kaolinite, chlorite and montmorillonite is probably largely derived from the catchment of the Indus drainage system and is found typically in the Greater Rann. The addition of further montmorillonite to the normal clay association, as seen at Khandla and the Inner Rann, is the result of influx of weathered products of the basalts of the Deccan Trap. Some of the kaolinite in these sediments appears to be due to the authigenic growth of this mineral.

Because of the low calcium-carbonate content of the source areas of the sediment that is carried into the Ranns of Kutch and the lack of production of skeletal carbonate within the Ranns, the sediment of the Ranns contains very little of this constituent. Because of the extremely high salinity and periodic desiccation there is very little indigenous life, so that the sediments are poor in all organic matter. However, one species of Foraminifera, *Ammonia beccarii*, occurs in the very saline sediments found near Pachham Island, and the area is possibly repopulated annually during the monsoonal flooding.

The halite crust of the Ranns is precipitated as the result of evaporation of sea water, and gypsum crystals grow in the sediments thereby giving a high $\frac{\text{Mg}^{2+}}{\text{Ca}^{2+}}$ ratio to the interstitial water. In the Persian Gulf coastal sediments a similarly derived high $\text{Mg}^{2+}$-ion concentration in the interstitial waters results in dolomitization of their calcium-carbonate-rich host sediments, and a lowering of the $\frac{\text{Mg}^{2+}}{\text{Ca}^{2+}}$ ratio of these waters. In the Ranns, however, where there is little calcium carbonate to alter to dolomite, the $\text{Mg}^{2+}$ is retained in the interstitial waters.

It seems likely that the salt flats develop only where the Rann surface is at sea level or within reach of the storm tides. Earth movements accompanying earthquakes have been followed by erosion of the uplifted areas and the silting-up of the depressed areas. If temporarily cut off from the open sea to form a lake, such areas may be recognized in section by thin-bedded gypsum overlain by bedded salts resulting from evaporation to dryness of the ponded waters.

It would appear that, in this area, gypsiferous clays are deposited under evaporitic conditions in very shallow water, intertidal or supratidal areas, where the rate of sedimentation exceeds the rate of subsidence, so the sequence of events will produce a regressive cycle. As such, the unit should only be in the order of a few metres in thickness, but possibly of great areal extent, unless further subsidence permits a continuation of the process. The thickness of gypsiferous clays occurring in the Ranns of Kutch, and the number of regressive cycles possibly involved, will not become known without fairly deep coring.

The Ranns are an area of very extensive evaporite deposition and are another
example of the growing evidence for the importance of the supratidal sabkha environment as the site of formation of these sediments.

(11) Because these supratidal sediments are being deposited in a region characterized by considerable fault activity, the graben areas may already be the sites of thick sequences of sediment deposited at or close to sea level.

(12) The extensive evaporitic mud-flats of the Ranns of Kutch could be taken as a possible model for the understanding of the Tertiary gypserous shales of Sind, Kutch and Kathiawar, and some of the gypserous Triassic shales of the Keuper of Northwest Europe.

ACKNOWLEDGMENTS

The calcium-carbonate and organic-carbon contents of the sediments were determined in the sedimentological laboratory of Imperial College under the supervision of P. R. Bush. At the Koninklijke/Shell Exploratie en Productie Laboratorium (KSEPL), X-ray diffraction analyses of some other minerals were undertaken by K. de Groot, analyses of clays and estimation of their boron content by D. H. Porrenza and J. Brouwer determined the Foraminifera in the samples. The water samples were analysed at KSEPL by K. de Groot and duplicate analyses of some samples were undertaken at Imperial College under the supervision of P. R. Bush.

The authors are indebted to the Oil and Natural Gas Commission of India for making the reconnaissance possible, and for providing the assistance of Mr S. Biswas, without whose geological knowledge of Kutch little would have been accomplished. Permission to publish this paper was given by Shell Research BV, The Hague, Netherlands. The figures were prepared at KSEPL under the supervision of the Exploration Department's Chief of Geological Draughting, Mr R. R. J. Davilar, whose assistance is gratefully acknowledged.

REFERENCES


Sediments of the Ranns of Kutch, India


(Manuscript received 25 April 1975; revision received 27 October 1975)
Depositional environment and diagenesis of Permian Rotliegendes sandstones in Leman Bank and Sole Pit areas of the UK southern North Sea

K. W. GLENNIE, G. C. MUDD & P. J. C. NAGTEGAAL

SUMMARY: Early diagenetic features of the Rotliegendes are related to deposition in a desert environment. Wadi sands suffered early calcite cementation as the sediment dried. Aeolian sands were cemented with halite and gypsum where the water table was close to the surface and all grains became coated with ferric oxide and clay after passing below the water table.

Later diagenesis is recognised by the presence of pore-destroying cements (recrystallized clays, authigenic feldspar, quartz, dolomite and anhydrite). Their occurrence is related to deep burial followed by faulting and late Cretaceous uplift. In Leman Bank and Sole Pit, the top Rotliegendes reached depths of about 11000 and 14000 ft (3351 and 4250 m respectively) prior to maximum uplift of 6000 ft (1800 m).

Depositional environment and geological history

The Rotliegendes consists of continental sediments that were deposited in a Permian desert located to the N of the Variscan Mountains (Glennie 1972). The Lower Rotliegendes, which in E Netherlands and Germany is characterised by associated extrusive rocks, is missing in the U.K. sector of the southern North Sea; instead, the Saalian unconformity separates Upper Rotliegendes sediments from the deflated surface of a block-faulted Carboniferous sequence. In U.K. waters, we are thus only concerned with the Upper Rotliegendes.

Conforming to an axis of subsidence that was probably already established in Westphalian time, fluvial sediments, derived from the Variscan Mountains, were transported northwards by ephemeral streams (wadis) to the margin of a permanent desert lake whose fluctuating shoreline marked the limits of a broad inland sabkha (Fig. 2). The lake sediments comprise silty mudstones with horizons of halite (e.g. Rhys 1974, fig. 2) and anhydrite in the form of minor interstitial clay-sized material, whereas similar mudstones of the sabkha environment are characterised by an absence of halite and the ubiquitous presence of anhydrite nodules (formerly gypsum crystals) and adhesion ripples. Together, these mudstone sequences are referred to as the Silverpit Formation (Rhys 1974, 1975). The sabkha sediments also contain stringers of sandstone which merge southwards into a broad belt of aeolian and wadi sands (Marie 1975), the Leman Sandstone Formation (Rhys 1974, 1975).

Van Veen (1975) showed that in Leman Bank, the Rotliegendes is represented entirely by the Leman Sandstone which can readily be subdivided into three parts (Fig. 1):

(a) an upper unit of partly homogenised water-laid sandstones,
(b) a middle unit entirely of aeolian sandstones,
(c) a lower unit of wadi deposits with minor aeolian sandstones.

At the West Sole gas-field 100 km to the NW (Fig. 2), a siltstone unit intervenes between two sandstone units (Butler 1975). The upper sandstone correlates with the water-laid sandstones at Leman Bank and...
has been interpreted as resulting from reworking of Rotliegendes sandstones during flooding of this continental basin by marine waters of the Zechstein Sea (Glennie 1972, van Veen 1975). The lower sandstone appears to correlate with a combination of the other two Leman Bank units and is thus also of mixed wadi and aeolian origin. The aeolian facies is characterised by the common occurrence of adhesion ripples, which implies that the water table was close to the surface (Glennie 1970, 1972). The wadi sandstones are partly conglomeratic with clay clasts, and in part grade up to thin claystones. Butler (op. cit.) inferred deposition of the upper part of this lower sandstone in a shallow-water environment, which may have been in direct communication with the desert lake of the Silverpit siltstone facies, into which it grades.

In the Sole Pit area between the Leman Bank and West Sole fields (Fig. 2) the Rotliegendes facies can be...
Depositional environment and diagenesis, Permian Rotliegendes sandstones

considered as intermediate between the two extremes, but with a strong development of aeolian sands, much of which contains interbedded adhesion-ripple sequences. It is inferred, therefore, that the water table was closer to the depositional surface in the Sole Pit and West Sole areas than in Leman Bank. This conforms with the interpretation made by Marie (1975), which showed wadi sediments spreading N from the present Norfolk coast and becoming interbedded with aeolian and sabkha sediments in the Sole Pit area (Fig. 2).

Rotliegendes sedimentation probably took place in a continental basin whose surface was below sea level. It was terminated abruptly by the inflowing of the Zechstein Sea, which, as pointed out above, resulted in the reworking of the uppermost sands. Continental deposition was again re-established at the beginning of the Trias and continued in the area under consideration until the Rhaetian transgression.

Marie (1975) showed that throughout the time interval early Permian to mid- or late Jurassic, more

![Schematic Rotliegendes facies distribution.](image)

Fig. 2. Schematic Rotliegendes facies distribution. S of the lake-margin sabkha, dune sands (stippled) form the dominant sediment; they are interbedded with a variable proportion of wadi sands along an axis trending N from the present Norfolk coast, and with sabkha sediments, including adhesion ripples, along an E-W axis through the northern Sole Pit area. Superimposed on this map are contours of the estimated difference between present and maximum depths of burial of the overlying Bunter Shale (solid lines) and the amount of porosity destruction attributed to compaction (dashed lines). The Dowsing fault zone was probably active from Rotliegendes time on.
sediment accumulated in the Sole Pit area than to the N or S. ‘Inversion’ (uplift after deep burial) of this area during sub-Hercynian and Laramide movements—late Cretaceous to early Tertiary (Heybroek 1974, Lutz et al. 1975) resulted in erosion of much or even all of the previously deposited Mesozoic strata (Fig. 1). The result of these movements is that the Rotliegendes of this area was formerly buried up to 4000 or 6000 ft (1200 to 1800 m) more deeply than at present. This had a direct effect on the formation of later diagenetic cements.

**Environmental diagenesis**

The earliest diagenetic processes affecting the Rotliegendes are clearly related to its mode and environment of deposition. The two extreme end members are seen in wadi sands and dune sands, transported by water and air respectively.

**Wadi sediments**

Marie’s (1975) maps show that the highest percentage of bedded clay in the Rotliegendes coincides with the distribution of wadi and sabkha/lacustrine sediments. Within the areas of wadi concentration, bedded clay is about 2 per cent of the total thickness of Leman Sandstone in Leman Bank and 15 per cent in the Sole Pit area; the usual thickness of each bed is 20 cm or less. The associated wadi sandstones are poorly sorted but have a relatively low percentage of incorporated clay matrix. This lack of primary argillaceous matter possibly derives in part from the lithological composition of the southerly source area undergoing denudation (Carboniferous and Devonian rocks?) but more as a result of the arid climate of this source area. By analogy with modern deserts, much of the clay-size products of weathering were probably removed by deflation and blown out of the desert. The remaining clay and the coarser detritus were then transported fluvially into the Rotliegendes depositional basin where the exposed clays and silts were subjected to another stage of deflation in areas where the sediment dried out.

Desert streams are ephemeral. Water flows through wadi sediments for much longer periods than it flows over them and, in so doing, may provide a considerable amount of calcium carbonate cement to the sands. In modern deserts, repeated floods over a number of years have resulted in a high degree of cementation of many wadi sands in the vicinity of active channels, and a similar process can be invoked to account for the tightly cemented wadi sandstones of the Rotliegendes, not only in the Leman Bank–Sole Pit areas, but also elsewhere in the Rotliegendes basin. As these sandstones have limited value as hydrocarbon reservoirs, their further diagenesis has not been studied in detail, but it seems in most cases to have involved early dolomitization of the calcite cement and thus some secondary increase in porosity, followed by the probable addition of authigenic clays and haematite. Locally, small gypsum crystals (now patchy anhydrite cement) indicate past evaporitic depositional conditions.

In the broad areas between active wadi channels, sedimentation results mainly from channel overflow and barely keeps pace with deflation. Since the flooded area is initially dry, there is a rapid downward percolation of muddy water from which the clay content is filtered at horizons of lower permeability, or settles out geopetally on the upper surfaces of pre-existing grains on reaching the water table (Walker 1976). Although not specifically recognised in the Rotliegendes wadi sediments, clay infiltration is likely to be another important process resulting in permanent porosity reduction of wadi sands.

**Aeolian sediments**

As mentioned earlier, the best reservoir rock in the Rotliegendes of the southern North Sea was deposited in the form of desert dunes. Its quality is due to its primary reservoir properties of good porosity and permeability, both of which are products of its mode and environment of deposition.

Because clay- and silt-size particles can be carried by the wind in suspension, the bulk of these size fractions are transported beyond the limits of the desert to be deposited elsewhere on the continent as loess or to form a small addition to a sequence of marine beds. Particles of this grain size form only a very small fraction in dune sands.

Coarser sand grains (100–2000μm) travel by saltation, are selectively sorted, and are deposited when the velocity of the driving wind becomes insufficient to transport them further. In the southern North Sea, the bulk of the Rotliegendes dunes seem to be of the transverse type, which are built up largely of sequences of well sorted, westerly dipping, foreset beds, which grade down to sub-horizontal bimodal bottomset beds (Glennie 1972).

Sandstone porosity is, in general, a function of grain sorting; the better the sorting, the better the porosity. Modern dune sands are commonly very well sorted, with Trask coefficients in the range 1.84–1.22 and associated porosities of 37–48 per cent. The Rotliegendes foreset beds probably had porosities in the upper half of this range together with excellent permeabilities at their time of deposition. The porosities of the bimodal bottom-set beds, however, were more likely at the lower end of this range, and because of the greater capillarity associated with the finer-grained laminae, had a much lower permeability than the foreset beds especially when measured perpendicular to the bedding.
Because the dune sands are composed largely of quartz grains, and are virtually free of detrital clay particles at the time of deposition, the effects of diagenesis can be readily recognised. Much of the succeeding discussion is therefore confined to this aeolian facies.

Within the aeolian sequence of the Sole Pit area adhesion-ripple horizons are common. They were probably deposited over the damp surface of a sabkha (Glennie 1970, 1972), either between migrating dunes or adjacent to the basin-centre desert lake. Unlike more normal aeolian sands, some silt and clay, deflated from the lake-margin sabkha surface further to the east, was certainly incorporated into the ripples by adhesion, the amount possibly reflecting proximity to the lake margin. By analogy with modern deserts, early cementation with halite occurred as moisture, drawn to the surface by capillary action, was evaporated, and gypsum crystals enclosed sand grains poikilitically as they grew. Thus the porosity of adhesion-ripple sequences was probably largely destroyed soon after deposition. However, after burial the rising water table caused the halite to go back into solution, thereby improving the porosity. Because of the high Mg"/Ca" ratio of the interstitial waters caused by gypsum precipitation, any early calcite cement in this facies was rapidly dolomitized and thus created some secondary porosity (Glennie & Evans 1976). With increasing burial resulting in formation temperatures in excess of 50°C, the gypsum crystals were converted to nodular anhydrite, the state now recognised in cores, with an associated volume reduction of 38 per cent and the creation of additional secondary porosity. Thus, despite the poor outlook as a potential reservoir rock at the time of deposition, burial resulted in an improvement in reservoir potential to the extent that adhesion ripple horizons do act as reservoirs in some fields, the quality depending on other diagenetic factors, although both porosity and permeability remain poor (Fig. 3).

Red oxide and early clay cements

Another process that starts once desert sediments are buried below the water table is the reddening of sand grains with oxides of iron (Pl. 1A). As can be seen in modern deserts, the sediments are not red when deposited but redden with time. The processes of weathering, which had been largely inhibited in the arid environment of the desert surface, took place in the humid environment below the water table, thereby releasing a variety of ions for incorporation into new minerals; for example, plagioclase feldspars alter to kaolinite, thereby releasing potassium, calcium, silica and alumina.

From a study of Cenozoic desert sediments, Walker (1967, 1976), and Walker & Waugh (1973) demonstrated that a red pigment forms in response to the intrastratal alteration of iron-bearing minerals, especially the more unstable ferro-magnesian silicates such as olivine, augite, hornblende. The absence of such minerals in ancient red-bed sequences resulted from their progressive alteration, with intermediate stages involving haematite aureoles around partly altered ferro-magnesian minerals, and crystals or larger blebs of haematite forming one end-product.

The iron that eventually coats the sand grains (except at their points of contact) is probably transported initially in solution as ferrous ions, but in the oxidising environment of the Rotliegendes, the interstitial water must have been in the Eh-pH stability field of haematite and the sediment became reddened. The initial mineral formed seems to be a ferric hydrate precursor of haematite, which converts to haematite upon ageing (Walker 1967).

Provided these sediments stay in the haematite stability field, they will remain red. In a reducing environment the ferric iron will alter to the soluble green ferrous state and, if flushing occurs, the sediment may change colour to grey or even white.
Although a reddish-brown colour is common in the Rotliegendes of both the Leman Bank and Sole Pit areas, many wells contain green or greyish green horizons and streaks, some of the latter being associated with fractures along which reducing solutions are thought to have penetrated.

By contrast, those sands in the upper part of the Rotliegendes that were reworked by water during the Zeichstein flooding are grey or white in colour, leading to the name 'Weissliegendes' by which they are sometimes known. This suggests that groundwaters associated with this earliest Zeichstein marine environment, which led to the reducing conditions in which the Kupferschiefer was deposited, resulted in the removal of any early reddening that may have been present before reworking. Instead of reddening, these sands became cemented with early calcite (later dolomitized) and gypsum (now seen as anhydrite) so that this unit rarely forms a good reservoir.

Quite commonly, sandstones in the lower part of the Rotliegendes are also grey in colour whereas associated fluvial clays are red. The colour of the clays suggests that the total sequence including the sandstones was probably once red, but that a change in the Eh-pH environment of the formation water brought about reduction of the red oxide in the more porous sandstones. The clays, however, probably because of their original lower permeability, which had already been further reduced by the addition of iron-oxide cements during the reddening stage, largely resisted this process of reduction. In some cases, however, clay pebbles in fluvial sandstones have red cores which grade to pale green rims.

Although the reason for the loss of colour in the lower part of the Rotliegendes is not precisely known, it may be related to the underlying coal-bearing Carboniferous strata. These strata provided a source of acidic formation water formed by the release of CO₂ during the processes of compaction and coalification. Overburden and depth of burial increased continuously throughout the Triassic and Jurassic periods in the Sole Pit and Leman Bank areas, so it is during this time span that decolouration of the Rotliegendes probably occurred.

Almost every detrital grain is coated either with red haematite or, where this has been removed, by a green flaky clay mineral (Pl. 2A). Walker's (1976) work on Cenozoic desert sediments strongly suggests that this clay was originally a mixed-layer illite/montmorillonite, which formed during the reddening stage of the sandstones. Because the sand grains are not coated at their points of contact (Pl. 2A) the coating, both haematite and clay, must be post-depositional and therefore authigenic.

The combined haematite and clay coating on Rotliegendes sand grains averages about 50 per cent of the authigenic minerals in the rocks. The source of all this material is not known.

The mixed-layer illite/montmorillonite described by Walker (1976) and presumed once to have been present in the Rotliegendes, was probably derived from solutions produced by the post-depositional breakdown of other unstable minerals in the oxidising environment below the desert. However, no matter what the original composition of the clays, what we now see is the final product of a later history of recrystallisation and mineral authigenesis. These later events are discussed below.

**Burial-related diagenesis**

**Porosity losses due to compaction**

As mentioned earlier, the Rotliegendes in the Sole Pit–Leman Bank area underwent a maximum depth of burial, prior to the late Cretaceous–early Tertiary uplift, of up to 4000 or 6000 ft (1200–1800 m) below present depths. By comparison of the acoustic velocities of the Triassic Bunter Shale of this area with velocities in areas having a normal burial history (Marie 1975), maximum depths of 11 000 ft (3350 m) in Leman Bank and 14 000 ft (4250 m) in Sole Pit are postulated (Fig. 4).

Burial diagenesis resulted in mechanical compaction and other diagenetic processes involving pressure solution, alteration and dissolution of feldspars, and the interstitial deposition of authigenic minerals (Pl. 1b,c). Porosity losses due to compaction of foreset dune sandstones can be estimated by taking a mean figure of 42 per cent for the original porosity of modern dune sands of similar grain size, and subtracting from it the sum of the present porosity and the percentage of cementing minerals and matrix.

In the Sole Pit area compaction has resulted in the porosity of the dune sandstones being reduced by up to half its original value along a NW–SE trending axis which roughly coincides with the axis of maximum former burial (Fig. 2).

**Porosity losses due to faulting**

The porosity reduction values calculated for compaction effects are valid regionally, but there are local anomalous zones where an extra degree of compaction is found. There is evidence from orientated core studies that this extra compaction can be attributed directly to the compressional fracturing associated with strike-slip faulting. Adjacent to such fractures there is a high degree of physical crushing of quartz grains and a redistribution of the pre-fracture cement with an apparent net movement of authigenic clay minerals and quartz towards the fracture. Anhydrite is a common fracture-filling mineral (Pl. 2b). The movements inferred from core material are small and it is suggested that displacement was taken up by small increments across a wide zone. The resulting porosity...
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Reduction led to the development of zones with poor reservoir qualities in otherwise favourable areas. Tensional fractures also occur. They contain a higher proportion of anhydrite cement, and lack a crushed grain fabric. Tensional fractures with no secondary cements have also been observed, greatly enhancing the overall permeability of the sediment. Such fractures could well account for the productivity of, for example, the West Sole area (see Butler 1975).

Authigenic cements

The authigenic mineral assemblages developed in the Rotliegendes sediments are influenced by the origi-
nal facies and the environment of primary cementation. In the Rotliegendes dune sandstones, to which the following discussion is mainly confined, the cements recognised in their approximate order of crystallization are: illite and chlorite, quartz, dolomite and anhydrite. Some of these diagenetic events were inhibited by the presence of gas within the reservoir, so that both porosity and permeability are commonly better preserved above the gas/water contact than below it.

**Illite and chlorite**

We have shown above the the Rotliegendes dune sandstones had already acquired a considerable haematite-impregnated clay coating every grain soon after deposition. The clay in these coatings is now composed of illite and chlorite, the former being considerably more abundant than the latter (Pl. 1a, 2a).

The degree of illite crystallinity is a function of the former maximum temperature to which it has been subjected (Kübler 1968), which itself is dependent upon the depth of burial if the geothermal gradient is constant. The illite achieved its maximum degree of crystallinity during the Cretaceous, prior to late Cretaceous uplift. These clays therefore represent the end-product of a history of clay diagenesis that began during the Permian (Fig. 4).

The illite and chlorite were formed at the same time (Pl. 2a). The illite is a ferruginous variety with a crystal habit at the contacts with detrital grains of short leaf-like crystals which separate into whiskers (Pl. 2c). The crystal habit of chlorite is that of thin hexagonal plates (Pl. 2n). The development of illite whiskers may explain why some rocks are more affected by permeability losses than others. It has been suggested by Stalder (1973) that the illite whiskers effectively reduce permeability by dividing the intergranular pore spaces.

The main components of illite are silicon, aluminium, potassium and iron. Authigenic quartz precipitated later than the illite so that pressure-solution silicon was possibly first used to form the authigenic illite. The most likely source of potassium is the K-feldspars, which are strongly leached (Pl. 2u). The time of formation of the authigenic quartz is more abundant in the porous top-set beds than in the less permeable bottom-set beds. Authigenic feldspar, on the other hand, is almost entirely absent from the bottom-set beds even though they contain detrital feldspar grains upon which to nucleate.

Authigenic quartz amounts to an average of 20–30 per cent of the total cement (in contrast to 1–5 per cent of feldspar) increasing in amount from SE to NW. The amount of authigenic quartz present in these rocks seems to be too great to be accounted for entirely as the product of pressure solution, so that additional silica must have been introduced into the system. Possible sources include the silica liberated by feldspar leaching (Pl. 2u) and the alteration of detrital montmorillonite and illite/montmorillonite mixed layers in the underlying Carboniferous shales.

The time of formation of the authigenic quartz is likely to be during the Cretaceous, both before and after uplift of the Sole Pit–Leman Bank area (Fig. 4).

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**Plate 1 (facing)**

Photomicrographs of Rotliegendes sandstones

**FIG. A.** Bimodally-sorted aeolian sandstone. Haematite and clay coating to grains and larger patches of haematite (black) are concentrated in the finer-grained laminae. Present depth 6660 ft (2032 m). φ = 11 per cent, K = 0.17 mD. Leman Bank.

**FIG. B.** Coarse-grained aeolian sandstone showing the effects of pressure solution at grain contacts. The rock is cemented by poikilitic anhydrite. Present depth 6851 ft (2088 m). φ = 12 per cent, K = 0.9 mD. Leman Bank.

**FIG. C.** Grains (gr) of aeolian sandstone showing the effects of pressure solution at clay-free contacts (c). Coatings of illite and chlorite occur on the other grain surfaces. Black areas represent porosity. Sole Pit area.

**FIG. D.** Mylonitised and anhydrite-filled fracture. φ = 7 per cent. K = 0. Leman Bank.

**FIG. E.** Sand grains (gr) are coated with authigenic illite/chlorite (i). The remaining intergranular space is occupied by authigenic quartz (o’rz). Sole Pit area.

**FIG. F.** Poikilitic dolomite in dune sandstones. φ = 5 per cent, K = 0.17 mD. Leman Bank.
Scanning-electron micrographs of Rotliegendes authigenic clays

FIG. A. Only the contact areas between detrital grains are free of the authigenic clay (mainly illite) coating. ×100. Sole Pit area.

FIG. B. Intermixed crystals of illite (whiskers) and chlorite (plates) indicate that their crystallization was synchronous. ×1860. Sole Pit area.

FIG. C. At the contact with the detrital grain, illite crystallizes as leaves, which separate outwards into whiskers. ×2255. Sole Pit area.

FIG. D. Strongly leached detrital grain of K-feldspar. This type of alteration of K-feldspar is common in the Rotliegendes of the Sole Pit area. ×1420.
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Dolomite

Dolomite is found throughout the main Rotliegendes facies and accounts for 2 to 15 per cent of the authigenic cement in dune sandstones. In this facies it occurs as poikilitic crystals embracing several grains (Pl. 1a) and shows a preference for the bottom-set beds.

In the lower and middle parts of the Rotliegendes sandstones dolomite occurs as a blocky anhedral cement completely filling pores.

A third type of dolomite cement is confined to the upper dune sandstones and reworked dune sand facies of the Leman Bank area. It occurs as fine euhedral rhombs either as aggregate masses or surrounding grains.

There is petrographic evidence for the presence of primary dolomite grains preserved in well cemented sandstones. It is suggested that most of the original dolomite present in the formation was dissolved and re-precipitated by fluids flowing into the Rotliegendes as it was uplifted during the Cretaceous (Fig. 4). Additional carbonate cement may have been dissolved from dolomitic horizons in the underlying and overlying formations.

Anhydrite

Anhydrite, in both its early and late forms, is the least important in total volume of the authigenic cements in the Rotliegendes, but adjacent to faults and fracture zones it is the most common mineral (Pt. 1b). It occurs generally with a poikilitic texture or as a void-filling cement. Like the dolomite, the anhydrite is envisaged as originating by precipitation from mineral fluids flowing along faults into the Rotliegendes. Many of these faults placed the Rotliegendes formation in juxtaposition to beds of Zechstein anhydrite.

Conclusions

1. The Rotliegendes of the Sole Pit and Leman Bank areas of the southern North Sea comprises wadi, aeolian and sabkha sediments deposited in a desert environment. They are underlain by coal-bearing Carboniferous strata.

2. The dune-bedded Rotliegendes sandstones are assumed to have had excellent permeabilities and porosities of the order of 40 per cent at their time of deposition. The Rotliegendes sandstones became reddened with a coating of ferric oxide stained clay after passing below the water table. They remained red except where the interstitial environment became reducing, in which case they changed colour to green, grey or white.

3. Adhesion-ripple sandstones generally have a higher content of detrital silt and clay than dune sandstones. They were deposited along the sabkha margin of a desert lake and in interdune areas where the water table was at or close to the surface. Early halite cement passed into solution after permanent burial below the water table, and with deeper burial, the volume occupied by environmental gypsum cement was reduced by 38 per cent as it converted to anhydrite. Both processes resulted in some diagenetic improvement in porosity.

4. During the Mesozoic the Leman Bank and Sole Pit areas were subjected to burial some 4000-6000 ft (1200-1800 m) deeper than at present. Porosity in dune sandstones was locally halved by compaction. Local anomalies between wells can be accounted for by invoking wrench faults, which caused juxtaposition of areas of deeper and shallower burial, or by assigning the differences from the regional pattern to the result of stresses induced by faulting.

5. Early calcite cements in all Rotliegendes facies were dolomitized shortly after burial below the water table. Authigenic minerals resulting from burial diagenesis in the Rotliegendes dune sandstones are: illite and chlorite, quartz and feldspar, dolomite and anhydrite.

6. The final phase of recrystallization of authigenic illite and the less common chlorite was during the early Cretaceous, which roughly coincides with the period of maximum burial in the Sole Pit–Leman Bank area. These minerals, like the later authigenic dolomite, preferentially cemented the finely laminated bottom-set dune beds (thus forming strong permeability barriers parallel to the bedding) and are more abundant in proximity to faults.

7. Authigenic quartz and the less common authigenic feldspar occur both as overgrowths and as pore-filling cement. They show a preference for the better sorted dune foresets.

8. Authigenic dolomite and anhydrite, with poikilitic or void-filling habit, were the latest minerals to crystallize. They both show a preference for proximity to faults.

9. The diagenetic history of the Rotliegendes of the Sole Pit–Leman Bank area is the result of changes in the interstitial environment from the time of deposition until the present. This diagenesis was preconditioned by the desert environment in which the Rotliegendes was deposited and was strongly influenced by deep burial followed by sub-Hercynian to Laramide uplift and by faulting.

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DESCRIPT SEDIMENTARY ENVIRONMENTS

A classical desert is an almost barren tract of land, within or bordering the tropics, over which rainfall is too limited and spasmodic to support vegetation adequately. The upper limit for rainfall is about 25 cm/yr, and because of the high temperature and general lack of humidity, the potential rate of evaporation far exceeds precipitation. Tropical deserts cover about 20% of the world's present land surface and tend to be concentrated in the regions of prevailing trade winds which flow roughly between the latitudes 10° and 30° N and S of the equator (Fig. 1). The reasons for their occurrence seem to be largely meteorological. Desert conditions also occur in other latitudes, such as in the rain-shadow of mountain ranges or in areas far from the sea. Only about 20% of the desert surface is covered by sand dunes. The remainder comprises the deposits of ephemeral streams and large areas of outcrop, which are subjected to processes of desert weathering and erosion (Ollier, 1969; Glennie, 1970).

Desert Erosion. Areas of outcrop are continually subjected to weathering processes. The relative humidity in some desert areas may reach 100% before dawn, and heavy dews cover the rock surfaces, leading to chemical corrosion, especially of carbonate rocks (Clark et al., 1974). Moisture, in very small quantities, reaches the surface from the water table, and its evaporation close to the surface results in the growth of...
gypsum and halite crystals which exert an expansion force on the surrounding host rock, causing it to split. Rapid diurnal changes in temperature, especially if accompanied by a rainstorm, cause differential expansion and contraction stresses between the surface and underlying rock, resulting in spalling of the rock surface and even splitting of boulders. Tucker (1974) cites a possible Triassic example of this phenomenon. Siliceous rocks and fine-grained homogeneous limestones tend to form angular boulders, whereas more argillaceous boulders, granites, and dolerites become rounded by exfoliation. Softer rocks can be abraded by wind-driven sand and silt.

The wind can force dry silt- and sand-sized products of weathering into motion. With removal of the finer weathering products by wind (deflation), the larger, more resistant pebbles and boulders form a lag deposit that may be abraded into ventifacts.

Water in Deserts

Fluvial Sediments. Soil is thin or absent on the barren highlands, and little rainwater soaks into the ground. When rainfall ceases, the flow of surface water soon stops, and ephemeral streams, along which sediment was being transported, dry out. These dry watercourses are known as wadis in Arabia and North Africa. After heavy rainstorms, flowing water may extend the length of the wadi to reach the sea or the center of a basin of inland drainage. Overloading occurs at the nickpoint between hill and plain, and braiding follows with the formation of alluvial fans (see Alluvial-Fan Sediments). Over the plains, however, waters of a flash flood may overflow from the channel and cover the surrounding plains. During the Central Australian floods of 1967, for example, all except the crestal parts of a system of dunes were covered by water (Williams, 1970). The bed forms of the resulting deposits of sand and clay were ascribed to both the upper and lower flow regimes (see Flood Deposits).

When there is a high sediment:water ratio, viscous muds capable of transporting large boulders may fill the channel with mudflow conglomerates (see Mudflow Deposits). A stream may also become overloaded as water soaks into the stream bed before reaching the lowest point in the channel, which becomes clogged with sediment.

Desiccation. After surface flow has ceased, the water percolating through the stream bed is commonly saturated with respect to calcium carbonate, and evaporation results in initial cementation of the sediment. The surface sands and silts dry out rapidly, however, and generally are not cemented; they can be readily carried away by the wind. Clay lags in stagnant pools may dry, curl, and crack. Thin and fragile clay flakes are removed by the wind, but thicker and heavier flakes may be preserved in place, at least temporarily, by a covering of windblown sand (Fig. 2).

Desert Lakes and Inland Sabkhas. Temporary lakes may form in the center of a basin of inland drainage or where eolian sands block a stream channel. As the water evaporates, salts are concentrated; and the sands, silts, and clays become encrusted with halite and gypsum crystals (Glennie, 1970). Such saline areas are known in Arabia and North Africa as sabkhas (see Sabkha Sedimentology), in the Western Hemisphere as playas or salars (q.v.). Not all inland sabkhas are supplied by surface water. In many stream channels, water continues to percolate through the sediment long after surface flow has ceased and is brought to the surface of the sabkha by the pressure of the head of water or by capillary action, and then evaporated. Other inland sabkhas appear to be supplied solely by ground water. Two sedimentary structures characteristic of inland sabkhas are sand dikes and adhesion ripples (see Salar, Solar Structures). The former seem to be confined to inland sabkhas and desert fluvial environments, but the latter may also occur in nondesert areas.

Eolian Sediments

Uncemented fluvial sediments form an important source for windblown sand. Wind is capable of transporting very fine (100μ) particles in suspension. Consequently, much of the clay and silt originally deposited by stream action is removed by the wind and redeposited beyond the desert as loess (q.v.). Coarser sand
grains (100-200μm) travel by saltation and cause still larger grains to move over the desert by surface creep (see *Eolian Sedimentation*).

Eolian sands (q.v.) are deposited as wellSorted, laminated sediments (A in Fig. 3) when the velocity of the driving wind becomes insufficient to transport them farther. A given wind can drive sand over a hard, immobile surface faster than over a surface of loose sand, so that sand tends to accumulate in areas that are already sand covered.

The axes of eolian ripples are transverse to the wind, and the coarsest grains are concentrated at the crest. Ripples tend to flatten out at higher wind strengths and greater rates of deposition (Bagnold, 1941); these factors possibly result in extensive areas of horizontally laminated sheet sands. Transverse dunes, or barchans, are produced where the supply of sand is limited, are formed by winds of moderate velocity, and are composed mostly of beds deposited as avalanche slopes at the 34° angle of repose for dry sand; they are unstable in strong winds. The axes of longitudinal (self) dunes are parallel to the dominant sand-transporting wind; their bedding dips away from their axes with a down-wind component, but avalanche slopes are rare (see *Eolian Sedimentation*).

Coastal Desert Sediments

Many deserts border coastlines, but because in most deserts there are no rivers with a continuously flowing supply of water, coastlines that border deserts generally have no fluvial deltas. Exceptions include the Colorado, Indus, and Nile rivers. Subsiding coastal areas are subject to marine transgressions; and, in the clear tropical seas that border them, organisms produce large quantities of calcium carbonate which is available for transport by tidal and longshore currents. Also, strong tidal flow into and out of lagoons formed behind submarine bars causes the formation of oolite deltas (Fig. 4).

Extensive low-lying coastal areas are subjected to flooding by abnormally high tides and seepage from adjacent saline waters, producing a coastal sabkha environment (see *Sabkha Sedimentology*). The coastal sabkhas are characterized by crusts of evaporites produced by

![FIGURE 3. Sequence of Permian Rotliegend cores from the North Sea (from Glennie, 1972). They comprise alternations of conglomerates (C; including claystone pebbles and flakes); curled and broken clay layers (black) that are probably still in their site of deposition; sandstone dikes (D); fluvial sand (F); poorly sorted sands that are possibly former eolian sands homogenized during water transport (H); and undisturbed well-sorted subhorizontal to inclined eolian sand laminae (A). Other sands (W) could have been deposited after either eolian or water transport. Numbered holes are of plugs cut in cores for porosity/permeability determinations.](249)
evaporation of saline ground waters; by a belt of algal mat (Fig. 4) that binds the underlying sediment and which, when exposed to desiccation, hardens, curls, and cracks; and by adhesion ripples, which may form over the damp sabkha surface. Accumulation of these sediments can result in a rapidly prograding coastline (Evans et al., 1964). The sediments of a coastal sabkha are commonly rich in skeletal (high Mg) carbonate, which, after burial, is subject to dolomitization (q.v.). The products of deflation of a coastal sabkha are rich in foraminifera, which become constituents of nearby dunes (Glennie, 1970).

Ancient Desert Sediments

Ancient desert sediments are characterized, as are many continental deposits, by a reddish coloration, caused for the most part by a post-depositional coating of ferric oxides (goethite, hematite) on sand grains (Walker, 1967). Other colors may occur locally.

Ancient desert sediments may be identified from an association of any of the following criteria (see also Glennie, 1970, ch. 8), which are grouped into two categories.

Water-laid sediments are characterized by sedimentary features similar to those of water-laid sediments of nondesert continental environments having a seasonal rainfall, i.e., sedimentary structures of both the upper and lower flow regimes, but modified by one or more of the following:
1. Commonly calcite cemented; locally cemented by gypsum or anhydrite.
2. Conglomerates may be common and, locally, several cycles of deposition may lack a sand-sized fraction at the top of the cycle (deflation of sand and silt).
4. Sharp upward decrease in grain size (especially from sand to clay), indicating rapid fall in water velocity.
5. Common presence of clay pebbles and curled clay flakes.

Wind-laid sediments may exhibit any of the following characteristics.
1. Sequences of sandstones that may vary in thickness from a few centimeters to several hundred meters and whose laminae dip at angles from horizontal to 34° (less if compacted)—dips either of constant or multiple orientation.
2. Laminae commonly planar with only sparse poorly developed ripples (Glennie, 1972).
3. Individual laminae well sorted, especially in finer grain sizes; sharp size differences between larger grains of adjacent laminae are common.
4. Large sand grains tend to be well rounded.
5. Grain size commonly ranges from silt (60 μ) to coarse sand (200 μ) with bulk about 125–300 μ; and, apart from authigenic clay, silt and clay generally well below 5% of sediment.
6. Clay drapes very rare and usually accompanied by evidence that they were water-laid.
7. Adhesion ripples with associated increase in clay or silt content and common presence of gypsum or anhydrite cement (Glennie, 1972).
8. Mica generally absent.
9. Under scanning electron microscope, quartz grains exhibit pattern of meandering ridges and upturned fracture plates; and under optical microscope, quartz grains appear frosted (see Eolian Sands).

From criteria such as those listed above, ancient desert sediments of various ages have been identified in many parts of the world (Bigarella, 1972). The sediments must have accumulated in areas of subsidence and in many cases were preserved beneath the wave base of transgressing...
Desert sediments such as the Nubian type sandstones of North Africa and Arabia (McKee, 1962; Busson, 1967; Klitzsch, 1968) recur interbedded with marine sequences from the Cambrian to the Cretaceous. In NW Europe, deposition of Permian desert sediments (Fig. 3) was succeeded by salts of the Zechstein Sea, followed again by continental desert sediments during the Triassic. Analysis of the bedding attitudes of Permian dune sands indicates that the Permian winds blew over NW Europe from northeast to southwest (Glennie, 1972) and over North America in the same direction (Poole, 1964). These paleowind directions show that the Permian deserts of these two areas probably existed within the trade-wind belt of the Northern Hemisphere (Fig. 1). Desert conditions persisted in NW Europe until the end of the Triassic, as seen, for example, from local development of dune sands (Thompson, 1969) and more widespread continental evaporites. Dune sands were still being deposited during the Jurassic in parts of the western US (Poole, 1964); dune sands in Triassic/Jurassic age also occur in South America (Bigarella and Salamuni, 1961).

The deposits of coastal sabbkhas of widely differing age have also been recognized in different parts of the world (see Sabbkha Sedimentology). They are characterized by sedimentary cycles comprising from bottom (shallow marine) to top (supratidal): algal boundstones and grainstones; dolomites with birdseye structures overlain by dolomitized algal mat with an upward increase in anhydrite cement; nodular or chicken-mesh anhydrite in a dolomite matrix. Unconformities (deflation surfaces) are common at the top of each cycle.

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References


Cross-references: Alluvial Fan Sediments; Alluvium; Braided-Stream Deposits; Eolianite; Eolian Sands; Eolian Sedimentation; Evaporite Facies; Faceted Pebbles and Boulders; Flood Deposits; Flow Regimes; Fluvial Sediments; Sabbkha Sedimentology; Salar, Salar Structures; Syneresis; Weathering of Sediments. Vol. XI: Alluvial Fan, Cone; Desert and Desert Landforms: Playa; Sand Dunes; Wind Action. Vol. XI: Desert Soils.
EOLIAN SANDS

Eolian sands (Fig. 1) are commonly well laminated (horizontal, or inclined at angles of up to 34°), with each lamina comprising moderately to very-well-sorted grains ranging in size from silt (60µm) to medium sand (1000µm; Glennie, 1970) and locally, particularly near the base of an intraformational sequence, to very coarse sand (2000µm or over).

At the base of a dune sequence there are sharp differences in grain size between individual laminae, with the finer-grained laminae being very well sorted and the laminae containing coarser grains being only moderately so because of an admixture of finer sand and silt. Upward, the overall grain size diminishes, and individual laminae of differing grain size tend to become sets of laminae of more uniform grain size (Glennie, 1972). Vossmeyer (1974) found that Quaternary eolian sands from Germany were coarser-grained, less well-sorted, and more skewed than desert sands from Algeria and Iran. Folk (1968) cited a bimodality of sand grains as evidence for eolian action.

The smaller eolian grains are subangular, the coarser ones subrounded to well rounded. When seen under the optical microscope, the majority of eolian sand grains have a typical frosted appearance, apparently caused by light diffraction associated with small irregularities on the surface of the grain. Small angular grains and parts of some larger ones have conchoidal fracture surfaces which are not frosted. These conchoidal surfaces are apparently the result of spalling of smaller flakes as the result of im-
EOLIAN SEDIMENTATION

Examination of eolian sand grains with a scanning electron microscope (see Margolis and Krinsley, 1971; Fig. 2) reveals a pattern of meandering ridges and upturned fracture plates of quartz which possibly dip in a direction consistent with crystallographic planes; it has been suggested that these ridges represent quartz cleavage (Krinsley and Smalley, 1973) resulting from mechanical abrasion. The close placing of these planes is possibly responsible for the frosted appearance of many desert sands. The plates seem to be subdued by chemical solution and precipitation during periods of reduced wind in the desert. In the coastal dune sands of more temperate climates, the upturned plates are larger than those of desert sands and are less affected by solution and precipitation. In periglacial dune sands, these features are associated with others of typical glacial origin (Krinsley and Cavallero, 1970).

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Voissnerbaumer, H., 1974. Grain-size data of some aeolian sands: inland dunes in Franconia (Southern Germany), Algeria and Iran, a comparison, Geol. Pären. Förh., 96, 261-274.

Cross-references: Beach Sands; Desert Sedimentary Environments; Eolian Sedimentation; Glacial Sands; River Sands.
That these tilted blocks are the upper ends of a large conically plunging anticline, with Mountian on the southeast and Tadde on the northwest flank, believe that these will appear in the future as more detailed information becomes available. It is hoped that this method of structural analysis will lead to a full understanding of the geometry, structure and hydrocarbon systems in the North Viking Graben.

**Conclusions**

K.W. Glennie and P.L.E. Boegner


**Discussion**

D. H. Matthews (University of Cambridge): Problems in the plastic deformation of layered sediments applied stress usually lead themselves to computer modelling. Are your ideas based on such an approach?

A. Challenor: The authors appreciate that it may be highly desirable to model by computer the ideas presented in respect of the southern North Sea. Because of the rapid horizontal and vertical variations in rock parameters, however, the resulting model will be extremely complex in three dimensions, and this has not yet been attempted. The ideas themselves evolved progressively and (hopefully) logically from a long appreciation of the problems of mapping complex three-dimensional structures. They are now being used upon computer modelling, once the logic is needed to order to create the model.
CHAPTER 9

Sole Pit Inversion Tectonics

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Abstract. The relatively narrow Sole Pit Basin trended northwest–southeast, straddling Quadrant 48 of the UK Southern North Sea northeast of the Dowsing Fault. Within the basin, sedimentation was probably almost continuous from the Permian to the Late Cretaceous, when final uplift resulted in erosion deep into the Jurassic sequence. Comparison of the sonic velocities of the Bunter Shale, measured in wells both within and beyond the basin limits, indicates that the uplift locally may have exceeded 5000 ft. Positive inversion affected the northeast side of the basin only during the Early Cretaceous, whereas the southwest side was not uplifted until the latest Cretaceous; the basin axis had some effects of both. The Sole Pit area is traversed by many northwest–southeast faults. Both the earlier basin subsidence and the later inversion are believed to have been reactions to stresses generated by events beyond the limits of the North Sea. These have resulted in strike-slip offsets of mostly minor displacement both in the lateral and in the vertical sense.

INTRODUCTION

Sole Pit is an area of locally deep water (91 m) between northwest–southeast-trending sand banks some 100 km east of the Humber Estuary (Figs 1 and 2 a; Houbolt, 1968). It gives its name both to the Sole Pit High, which came into being during the Cretaceous—in fact, the earliest Tertiary, and to the area of basinal deposition (Sole Pit Basin, Figs 2 and 3) that preceded the structural inversion. This process involves conversion of a basinal area into a structural high. The converse is also possible, so that inversion can be considered in both positive (uplift) and negative (subsidence) senses relative to the immediately preceding history. In its positive sense, the term ‘eversion’ (turning inside out), which has already been used by Kingma (1958) to describe such phenomena in New Zealand, is an apt alternative.

Because of its size, the Sole Pit High was the target of some of the earliest exploration activity in the North Sea, which resulted in the discovery of the West Sole and Leman Bank gas fields (Fig. 2 a). Much of the intervening high area seems to be gas-bearing. Because of the small size of some of the closures, however, and the lack of good porosity and permeability in much of the objective Rotliegendes reservoir (Glennie, Mudd and Nagtegaal, 1978), none have yet been proven capable of producing gas commercially.

STRUCTURAL SETTING

The Sole Pit Basin and the succeeding Sole Pit High occupied a northwest–southeast-trending area in the UK portion of the Southern North Sea Basin (Fig. 2 a). The Sole Pit High is limited to the southwest by the slightly arcuate Dowsing Fault Zone (Figs 1 and 2 a). To the east, it is limited by a combination of a steep synclinal flexure and a series of north–northwest–south–southeast-trending en echelon faults, which meet the Dowsing Fault Zone in the vicinity of block 53/15. We suggest that this hinge zone be named after the Swarte Bank submarine sand dune, which is almost parallel to it (Fig. 1). The northern extent of the Sole Pit Basin is defined by northwest–southeast-trending faults that intersect with the Dowsing Fault in block 42/17, and with the Swarte Bank hinge zone in block 43/29. These faults are visible only in the sequences beneath the Zechstein salt, and seem to be the reason for diapirc uplift of the salt along that trend. As delineated above, the Sole Pit High, and the basin that preceded it, have a length of some 250 km and a maximum width of about 60 km (Fig. 2 a).

Just south of the 54°N parallel is an east–west oriented Zechstein salt-wall diapir whose uplift was almost certainly triggered by underlying fault movement. This fault, which we refer to as the Outer Silverpit Fault, separates the greater part of the Sole Pit area from the much smaller northern extremity.

In a broader context, the UK part of the Southern North Sea Basin occupies an irregular four-sided area bounded by the Pennine uplift to the west, by the mid North Sea High to the north, and by the London–Brabant Platform to the south (Fig. 2 a). The UK–Netherlands median line roughly coincides with a north–south zone of Late Cimmerian uplift and erosion deep into rocks of Triassic and, locally, even of Permian age (Figs 3 a and 4).

Fig. 1. Isometric block diagram of part of the Sole Pit area. Each cross-section represents an approximate depth conversion of several en echelon seismic lines of vastly differing quality; thus they are schematic rather than factually precise. Carboniferous reflectors are rarely recognized on older seismic lines. The boundaries of the Carboniferous sequences are dashed, thereby stressing our sparse knowledge of both their distribution and thickness. Orientation and location are by means of the oil concession block boundaries.
BURIAL HISTORY AND PALAEOGEOGRAPHY

We have only limited data on the Carboniferous and older strata of the southern North Sea. Nevertheless, in this region the Variscan orogeny seems to have been responsible for a change in sediment transport direction. Southward transport during Dinantian to early Westphalian time was reversed in the late Westphalian, older strata of the southern North Sea. Nevertheless, in the London-Brabant Platform, thereby reflecting uplift of the Variscan Highlands and the London-Brabant Platform.

Early Permian erosion cut deeply into the Carboniferous sequence (Fig. 4). To judge from what is still preserved elsewhere in the southern North Sea, uplift and erosion over the Mid North Sea High prior to the Westphalian transgression was probably locally in excess of 6000 ft (1800 m), and may have been as much or even more over the London-Brabant Platform.

The Carboniferous sequence within what later became the Southern North Sea Basin is transected by west-northwest-east-southeast-trending faults. Inversion, possibly largely in the form of differential uplift across these fault planes (depicted schematically in Fig. 1), resulted in erosion deep into Coal Measures of Westphalian A, and locally even late Namurian, age along a broad zone trending from the coast (see also Kent, 1980) southeastwards across the Sole Pit area (Fig. 2 b). Erosion of the Westphalian sequence may have exceeded 3000 ft (900 m) in the vicinity of Sole Pit.

The initiation of the Sole Pit Basin is clearly seen from isopachs of the Rotliegendes sequence (Fig. 2 c). The general axis of subsidence trends north-northwest-south-southeast, which is some 30° oblique to the trend of the Carboniferous uplift in the same area. The Sole Pit Rotliegendes sub-basin is roughly parallel to, and en echelon with, a similar basin in Netherlands waters, which suffered Late Cretaceous inversion to become the Broad Fourteens High (Heybroek, 1974; Lutz, Kaaschieter and Van Wijhe, 1975).

The approximately east-west orientation of some Rotliegendes contours in the vicinity of 54°N roughly coincides with a change in Rotliegendes lithology from aeolian and wadi sandstones in the south, to sabkha and desert-lake sediments in the north (Glennie, 1972; Marie, 1975). This change in facies is sub-parallel to the Outer Silverpit Fault, which is on trend with the Market Weighton axis in Yorkshire (Fig. 2 a).

The Zechstein transgression flooded a basin in which the rate of subsidence had probably outpaced that of deposition (Smith, 1979). Virtually the whole of the Rotliegendes surface of the Southern North Sea Basin was below the level of the Late Permian oceans. Perhaps because of the absence of important contemporary fault activity, the regional Zechstein isopachs give no indication that the Sole Pit area had any special significance in terms of subsidence during the Late Permian. The early Zechstein carbonate development was confined to water that was shallow enough to permit photosynthesis. The Haupit dolomit shell edge, however, marked a rapid change to deeper water and is roughly parallel to the isopachs shown on Fig. 2 d. By the end of the Permian, the Southern North Sea Basin had probably been almost filled, much of it with the products of evaporation (Taylor and Colter, 1975).

Because of deep Late Cimmerian erosion in the vicinity of the UK—Netherlands median line, the isopachs for deposits of Triassic age overemphasize the basinal nature of the UK part of the Southern North Sea Basin (Fig. 3 a). Even so, it is clear that the Sole Pit Basin was the site of more rapid subsidence and deposition than its flanking areas. This is confirmed by changes in thickness across the area of individual rock units and especially those of mid to late Triassic age. The Keuper Salt member of the Haisborough Group (Fig. 4) for example, was deposited in a small sub-basin that approximates to the 4000 ft isopach in Fig. 3 a. That this sub-basin had no surface connection with its marine time equivalent in Germany and the Netherlands seems to be confirmed by the existence of an intervening area of anhydritic claystones; the gypsum crystals, which were later converted to anhydrite, probably grew within the zone of capillary movement of water above the water table and were not deposited as the result of evaporation from a body of standing water (cf. Shearman, 1970). Thus, the Keuper Salt of the Sole Pit Basin was separated from its German equivalent by supratidal sabkha or low-lying land.

Jurassic strata of Liassic age are now exposed virtually at the sea bed along the crest of much of the Sole Pit High (Fig. 1). In this axial region, therefore, we have no direct proof of the depositional history from the Liassic onwards, and thus also of the timing of structural uplift (Figs 3 b, c and d). From the seismic profiles, however, it is possible to deduce some of the history of the flank areas. In the vicinity of block 48/9, 48/10 (Fig. 1, Section 1) Lower, Middle and Upper Jurassic sequences all thicken towards the axis of present uplift, and the succeeding Early Cretaceous Cromer Knoll Group does likewise; this may in part, however, be a reflection of adjacent diapiric uplift of Zechstein salt. Farther to the southeast, the history is clearly different. On the flanks of the Leman Bank structure Late Cimmerian erosion has resulted in a thin sequence of the Early Cretaceous Cromer Knoll Group being deposited above the Late Triassic Haisborough Group (Fig. 1, Section 3). Thus, perhaps 3000 ft or more of Jurassic strata were here removed during the Early Cretaceous (Fig. 3 b).

The history again differs within the Dowsing Fault Zone. An apparently complete but condensed Jurassic sequence, some 1500 ft thick, is overlain by over 2000 ft (600 m) of the Cromer Knoll Group along a narrow zone in block 48/23. This indicates that deposition here was probably continuous across the Jurassic—Cretaceous time boundary (Figs 1 and 3 c). It seems then that whereas Jurassic—Cretaceous sedimentation was locally continuous, elsewhere this time boundary forms part of a major hiatus. The stratiographically most complete sequence of Jurassic rocks in the UK southern North Sea occurs between the Sole Pit High and the English coast (Fig. 4), but the thickest sequence is locally preserved along the plunging northern axis of the Sole Pit High (Fig. 3 b).
Fig. 2. Structural and isopach maps of the Sole Pit and adjacent areas. (a) Structural outline of the UK southern North Sea showing faults bounding the Sole Pit area. These latter are repeated on all other maps of Figs 2 and 3. Cross-hatching represents location of diapiric salt walls. (b) Ages of Carboniferous sequences subcropping the Rotliegendes and younger rock units. Modified from Eames (1975). (c) and (d) Isopach maps of the Permian Rotliegendes (contours × 100ft) and Zechstein (× 1000ft) Groups. Isopachs derived largely from well data.
Fig. 3. Isopach maps of Mesozoic rock units of the Sole Pit and adjacent areas. Isopachs derived largely from well data. (a) The Triassic Bacton and Haisborough Groups and Winterton Formation (× 1000 ft). Vertical lines indicate areas where Early Cretaceous erosion has reduced the original depositional thickness. (b) The Jurassic Lias, West Sole and Humber Groups (× 1000 ft). Probably most areas have suffered from some intraformational as well as post-Jurassic erosion. The latter is most marked east of the Dowsing Fault Zone. (c) Lower Cretaceous Cromer Knoll Group (× 100 ft) and depositional area of the basal Cretaceous Spilsby Sandstone Formation. (d) Upper Cretaceous Chalk Group (× 1000 ft). Vertical lines indicate areas where post-Cretaceous erosion has reduced the original depositional thickness.
Early Cretaceous uplift resulted in varying degrees of erosion. Resumption of sedimentation over much of the Sole Pit area seems to have been delayed until the late Early Cretaceous. Deposition of the Late Cretaceous Chalk is known to have taken place over parts of the High, but lack of sufficiently precise palaeontological dating, and the difficulty of accurately correlating the wireline logs of thin Chalk sequences beneath the Tertiary and Quaternary erosion surfaces of the flank areas, render any estimate of its former thickness very speculative. Also, but at the southeastern limit of the area in block 53/7, the Chalk sequence is limited in age to the later Campanian and Maastrichtian, and overlies a Cromer Knoll sequence in which the uppermost part of the Group is absent. Furthermore, just to the south in block 53/12, both the Cromer Knoll and Chalk Groups contain a sandstone facies, which must indicate a nearby source of sand on an uplifted fault block within either the Dowsing Fault Zone or the Sole Pit area.

Some indication of the unknown episodes of the depositional history can be acquired, however, by comparing the acoustic velocities of the Bunter Shale in wells from this area with those of regions of normal continuous subsidence. By this means, estimates can be made of the depth of maximum burial (Marie, 1975; Glennie et al., 1978). These estimates indicate that, in contrast to the flanking areas, the Bunter Shale of parts of the Sole Pit High was once buried more than 5000 ft (1500 m) deeper than at present.

The burial histories of the Rotliegendes sandstones (Fig. 5) indicate the inherent differences between the two flanks of the Sole Pit area. There is clearly a regional change from more or less continuous subsidence until a Late Cretaceous uplift, southwest of the area (Fig. 5 a), to the other flank in which Mesozoic subsidence is interrupted by a phase of Early Cretaceous uplift and erosion (Fig. 5 c). The burial curve for the well 48/7b-4 and, by analogy, also 48/13-2, indicates that the 4000–5000 ft by which the Rotliegendes

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**Fig. 4.** Simplified stratigraphy of the Sole Pit and adjacent areas.

- **Early Cretaceous uplift resulted in varying degrees of erosion.**
- **Resumption of sedimentation over much of the Sole Pit area seems to have been delayed until the late Early Cretaceous.**
- **Deposition of the Late Cretaceous Chalk is known to have taken place over parts of the High,** but lack of sufficiently precise palaeontological dating, and the difficulty of accurately correlating the wireline logs of thin Chalk sequences beneath the Tertiary and Quaternary erosion surfaces of the flank areas, render any estimate of its former thickness very speculative.
- **Also, but at the southeastern limit of the area in block 53/7,** the Chalk sequence is limited in age to the later Campanian and Maastrichtian, and overlies a Cromer Knoll sequence in which the uppermost part of the Group is absent. Furthermore, just to the south in block 53/12, both the Cromer Knoll and Chalk Groups contain a sandstone facies, which must indicate a nearby source of sand on an uplifted fault block within either the Dowsing Fault Zone or the Sole Pit area.

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Fig. 5. Burial histories of the base Rotliegendes at selected well locations. Deduced from the relevant drilled sequence and from estimates of the maximum depth of burial derived by comparison of the acoustic velocities of the Bunter Shale with wells from Southern North Sea areas having a history of normal continuous subsidence.
was formerly more deeply buried, can logically be accom-
modated only during the Late Cretaceous. It is
during this deepest phase of burial that the diagenetic
damage to reservoir properties of the Rotliegendes
sandstones in the Sole Pit area is believed to have taken
place (Glennie et al., 1978). The axial areas of the Sole
Pit Basin therefore seem to have been affected in vary-
ing degrees by both the Early and the Late Cretaceous
phases of uplift (Fig. 5 b). The burial curves also indi-
cate that the probable source area of the basal Creta-
ceous Spilsby Sands was within and to the east of Sole
Pit, and that its depositional area was limited to the
southwest of Sole Pit (Fig. 3 c).

The evidence given in Fig. 3 does not accord in de-
tail with the obviously over-simplified interpretation of
Sole Pit palaeo-burial given by Glennie et al. (1978). It
also indicates a burial history that differs from the
otherwise rather similar and structurally
sub-Hercynian phase of uplift is involved
Lutz et al., 1975).

FAULT PATTERNS

As is clearly shown in Fig. 1, the majority of faults in the
Sole Pit area can be divided into three groups.

(a) The Dowsing Fault and Swarte Bank hinge zones
affect all rock units up to the Cromer Knoll Group
along the total length of the former fault, but only
along a part of the latter. They must reflect the
movement of master faults at depth. In the north-
ern part of the area, where the Zechstein Stassfurt
Halite is very thick, strong activity of these faults
has probably induced the halite to act diapirically
and has resulted in the development of salt walls
(Fig. 1, Section 1). The approximate time of these
fault movements, which seemed to be most active
during the early Late Cretaceous and early Tertiary,
can be deduced from the uplift and subsidence
patterns of the affected seismic reflectors adjacent
to the diapirs. In the southeastern part of the area,
post-Zechstein sequences are not cut by the
Swarte Bank fault system (Fig. 1, Section 3).

(b) Numerically the most common faults are those
that cut nothing younger than the basal Zechstein
rock sequences. It could be argued that following
deposition of the Hauptdolomit, virtually all fault
movement was confined to the Dowsing and, to a
lesser extent, the Swarte Bank fault zones. We
have seen, however, that fault activity can be rec-
ognized by the effect it has on the diapiric move-
ment of salt, which can be dated by the contempor-
ary sedimentation patterns. This suggests that
relatively small movements below the salt are not
propagated as faults above the salt, the induced
stresses being absorbed by minor adjustments
within the salt.

(c) Tensional faults occur over the crest of the Sole Pit
High (Fig. 1, Sections 1 and 2). They also occur
within the boundary fault zones over the narrow
uplifted blocks between upward-diverging fault
planes of a type that Kingma (1958) calls 'pierce-
ment structures', and which we refer to informally
as 'flower structures' (e.g. Fig. 1, Section 2, block
49/16; Section 3, block 53/1).

Most of these tensional faults do not penetrate
below the level of one of the Triassic halite hori-
zons, usually the Keuper or the Muschelkalk (see
also Fig. 4), but occasionally reaches the Rot
Halite, the top of which is the mid-Triassic seismic
reflector shown on all three sections in Fig. 1.

These Triassic salts act as planes of décollement
permitting adjustment to the uplift-induced tension,
and seem also to prevent downward propagation
of the tension faults. Below these halites there
probably is a change from a tensional to a com-
pressional regime; the halites, therefore, must also
approximate to the neutral fold surface between
the overlying youngest folded sediments and the
underlying zone of relatively free deformation at
the top of the Zechstein halite.

Thus the few faults that cut the complete seismically-
visible sequence provide a partial outline to the Sole
Pit province. The many more faults that cut only the
deepest visible seismic reflectors give some indication
of the complex history that this region has passed through.

When analysed statistically on the basis of cumula-
tive length of fault versus azimuth, these deeper faults
seem to fall into certain groupings (Fig. 6). All groups
occur throughout the southern part of the UK North
Sea Basin, but their dominance varies from area to
area. Although the relative density of faults, and the
recognition of their orientation, varies with the quality
and density of the available seismic data, it is never-
theless believed that the percentage groupings pre-
presented in Fig. 6 are meaningful.

We suggest that the groups of faults with different
orientations have a primary distribution and an origin
as follows:

(1) Faults of Group I (trending 110°-130°) are domi-
ant along the southern margin of the North Sea
Basin adjacent to the London-Brabant Platform
(63% of total faults), and decrease in importance
to the north. They also form a system of en echelon
faults in the middle portion of the Dowsing Fault
Zone and along most of the Swarte Bank hinge
zone (Fig. 6). Their orientation is the same as that
of the Westphalian 'A' uplift across the Sole Pit
area shown on Fig. 2 b. It seems likely, therefore,
that they were formed during the Variscan or-
ogeny, and have been reactivated from time to
time since then.

(2) Faults of Group II (trending 130°-160°), with 47% of
the total are twice as numerous in the southern
half of the Southern North Sea Basin as the next
most important group (1), which has only 23%.
They form the dominant group in the Sole Pit
area, but are more important in the Indefatigable
area to the east. Faults of the Swarte Bank zone
belong to this group, as also do some of those in the
middle portion of the Dowsing Fault Zone. This
fault orientation is important throughout the UK
Southern North Sea, and is dominant also in the
Fig. 6. Percentage distribution of four fault-trend groups in the Sole Pit and adjacent areas, and their spatial distribution in the Dowsing and Swarte Bank fault zones. Based on the cumulative lengths of seismically mapped faults below the Stassfurt Halite.

West Netherlands Basin (see maps in Heybroek, 1974; Ziegler, 1978). The same structural trend can be seen on aeromagnetic maps of the East Midlands of England (Anon., 1965), where it presumably reflects the influence of the Late Precambrian Charnian orogeny; it also coincides with the axis of uplift of the Sole Pit High (Fig. 3d). To judge from the orientation of the depocentres shown on Figs 2 and 3, these faults have been active intermittently from at least the Early Permian until the Late Cretaceous, and probably represent reactivation of faults that were perhaps first formed in the Late Precambrian.

(3) Group III faults (trending 160-200°) represent only 14% of the total, and are dominant within the study area only on a very local scale. They are best developed in the Indefatigable area, where their occurrence probably reflects the Late Cretaceous and Early Tertiary subsidence of the North Sea Central Graben region.

(4) The fourth group of faults are oriented east-west (70-110°), and offshore are best developed east of the Yorkshire coast. Here, they are parallel to, and probably form an eastward continuation of faults flanking the Market Weighton uplift, which was a positive area during the Jurassic and Early Cretaceous (Kent, 1980). The latter is also parallel to the axis of the Cleveland Hills Permian structural high, over which Rotliegendes sediments were not deposited (Fig. 2c; Marie, 1975). Faults of this orientation undoubtedly underlie the diapiric salt wall associated with the Outer Silverpit Fault. The regional importance of this fault trend is also emphasized by the east-west orientation of the mid North Sea and Ringköbing-Fyn uplifts, which effectively divided the Permian and early Mesozoic North Sea area into two basins. Faults of this orientation therefore seem to have been active during the Permian (Cleveland Hills), the Jurassic and Early Cretaceous (Market Weighton), and the Late Cretaceous (diapiric salt wall).

A colleague (Peter Kamerling, personal communication) has suggested that the east-west Market Weighton axis, and its eastern projection across the Sole Pit area, may coincide with the boundary between the deeply buried, essentially Precambrian basement to the south, and folded Caledonian basement of geosynclinal origin to the north. This interpretation, however, may involve a minor spatial conflict with that concerning the Charnian structural trends. Even so, some of the east-west faults parallel to the Market Weighton axis may be a near-surface expression of a long history of relative movement between the margins of these two basement types. (See also D. J. Blundell, this volume.)

**DISCUSSION**

The foregoing analysis indicates that the Sole Pit and adjacent areas are underlain by a mosaic of fault blocks. There are also many features that suggest the presence...
of a wrench component in many of these faults (e.g.,
change in sense of throw and change from normal to
reverse fault along strike; abrupt change in formation
thickness across the fault; and occurrence of 'flower
structures', 'Riedel'-type shear faults). Yet when con-
sidered in their regional context, there is no marked
linear offset of geological features across the line of their
prolongation. The subsidence and later inversion of the
Sole Pit Basin therefore cannot have been caused by
large scale transverse movement such as occurred
along the San Andreas fault in California, or the Alpine
fault in New Zealand.

Although described in the previous section in
terms of the possible history of their origins, the fault
groups also comprise two conjugate systems, I/II and
II/IV. However, we do not propose to elaborate on
this aspect of the fault patterns.

Faults are far more numerous below the Zechstein
salt than above, because most fault stresses were not
transmitted through the salt with sufficient intensity to
fracture the overlying sequence. The virtual absence of
wrench faults above the Zechstein over most of the Sole
Pit area indicates, however, that the larger strike-slip
movements must have been confined to the marginal
Dowsing and Swarte Bank fault zones.

The Dowsing Fault forms a zone in which relatively
free small-scale movement has taken place, although a
group of en echelon faults, rather than a clear-cut break,
seems to occupy its middle portion. On the other side
of the basin, extension also largely in the form of small
en echelon horizontal offsets, probably played as im-
portant a role as any other fracture type in the Swarte
Bank hinge zone.

It seems that much of the horizontal stress present in
the Sole Pit area was probably absorbed by very small
offsets between the many fault slivers recognized be-
neath the Zechstein salt. The effect of such movements,
which probably occurred over much of the Southern
North Sea Basin, can be likened to the deformation of a
rectangle into a parallelogram by movement along
closely spaced parallel lines.

The available evidence thus suggests that the origin
of the Sole Pit Basin may be attributed not so much to
brittle fracture, but rather to 'soft-rock attenuation' of
a pull-apart basin in the sense implied by Crowell
(1974) for some of the Cenozoic basins of Southern
California. The lack of associated volcanism, however,
confirms that crustal rupture was not achieved beneath
the Sole Pit Basin.

The pull-apart process and associated subsidence
must have begun during the Early Permian, and con-
tinued intermittently until the Cretaceous. The pat-
terns of faults and isopachs displayed in Figs 2 and 3
indicate that this was most likely achieved by a right-
lateral sense of movement in reaction to east–west ex-
tension. Tensional stresses must have been greater
within the Sole Pit Basin than in its flanking areas.
Using Crowell's criteria, if the crust had a thickness of
around 30 km in this area, then an average subsidence
of 2.5 km (8500 ft) indicates crustal extension of 8°/o.
Crustal rupture, rather than just extension, might have
occurred if local subsidence in the order of 5 km (Fig.
5 b) had been greatly exceeded.

Basinal development is believed to have been ar-
rested, and the Sole Pit High created, by crustal thick-
ening in the area between the boundary fault zones.
These events are thought to be related to the stresses in-
duced by a reversal in the sense of strike-slip movement;
the former system of tension-creating faults became one
of compression.

Uplift of the area to the northeast of Sole Pit began
during the Early Cretaceous and temporarily, though
strongly, affected part of the study area (Figs 5 b and
c); crustal thickening, caused by left-lateral fault move-
ment, is implied. On a smaller scale, lateral compression
is undoubtedly the reason for the origin of the 'flower
structures' in the Dowsing Fault Zone, and also for the
local inversions of both Early and Late Cretaceous age
associated with the Swarte Bank hinge as seen in Fig. 1,
Section 3. In the cases of both the Sole Pit High and the
'flower structures', isostatic adjustment and the main-
tenance of volume were most easily achieved by
positive inversion.

Abrupt termination of the Cretaceous subsidence
and final uplift to form the Sole Pit High took place at
the end of the Cretaceous. Rapid crustal thickening,
caused by the squeezing associated with a reversal in
the sense of fault movement, is presumed to be the
main reason for the inversion. The high compressive
stresses that must have been developed along the axis
of the Sole Pit Basin towards the end of the Cretaceous,
are believed to have been a factor in the diagenetic pro-
cesses that caused porosity and permeability destruc-
tion in the Rotliegendes sandstone. These effects are
thought to have been greatest close to north-northwest–
south-southeast-trending faults. The existence of high
compressional stresses also implies that the depth of
maximum burial of the Rotliegendes in the Sole Pit
area may not be as great as indicated in Fig. 5 b. At
present, however, we have no means of apportioning
the different effects of deep burial and lateral com-
pression on the acoustic velocities of shale.

A clockwise rotation of 30° was involved in the
change from the Variscan axis of positive inversion,
which is assumed to have resulted from north–south
compression, to the axis of Permian and younger sub-
division, for which east–west extension is inferred.
Right-lateral strike-slip movements would have
occurred during both events, the former being compres-
sional and the latter extensional.

There is conflicting evidence for the sense of strike-
lift movement involved in the positive inversion of
the Sole Pit area. Both right-slip and left-slip movement
can be inferred from drag folds and Riedel shears as-
associated with the major faults. On balance, howev-
er, both Cretaceous phases of positive inversion probably
resulted from east–west compression involving left-
lateral fault movement. This seems to have been fol-
lowed, possibly after uplift, by minor north–south
right-lateral movement which, as the later event, left
clear evidence of its presence.

Within the relatively limited confines of the UK
Southern North Sea, there are no obvious reasons for
the events described in the Sole Pit area. To find their
cause, one must look beyond.

The greatest geological forces recognized today are
those involved in the wholesale creation and consump-
tion of oceanic crust at the earth's surface, and the
associated relative movement between crusts of oceanic and continental origin. The two-phase timing of Sole Pit inversion suggests a possible causal connection with North Atlantic events; the first, with compressive stresses generated during the earliest stages of the Early Cretaceous opening of the Rockall Trough, and the second, with the northwest European stress field associated with the transference of a spreading axis from the Rockall Trough to the present Mid Atlantic Ridge. The final right-lateral movements may reflect stresses generated during the Alpine orogeny.

CONCLUSIONS

The present Sole Pit High is the outcome of positive inversion of a Late Palaeozoic-Mesozoic sub-basin. Its axis crosses obliquely the site of a Late Carboniferous uplift.

The faults that bound the Sole Pit area, and also cut all pre-Zechstein strata, have trends similar to those that cut the Precambrian of the East Midlands of England as well as those that controlled the Variscan deformation of the London-Brabant Platform.

Both basin subsidence and later inversion were probably controlled respectively by attenuation and thickening of the underlying crust in response to tensional and compressive stresses induced by events beyond the limits of the southern North Sea.

Positive inversion of the west-northwest–cast-southwest trending Carboniferous sequence occurred during the latest Carboniferous–earliest Permian. The succeeding negative inversion of the Sole Pit Basin began with deposition of the Early Permian Rotliegendes sequences. The history of continuous Mesozoic subsidence was interrupted by positive inversion along the northeastern flank of the basin during the Early Cretaceous, and along the southwestern flank during the latest Cretaceous. The axial parts of the high were affected by both positive phases, slightly by the first, and strongly by the second phase, which immediately succeeded the time of maximum burial of the Bunter Shale. The reasons for these two phases of uplift are not known.

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EARLY PERMIAN (ROTLIEGENDES) PALAEOWINDS OF THE NORTH SEA

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ABSTRACT


The classical desert sediments of the Early Permian Upper Rotliegendes were deposited in a post-Variscan basin that extends from eastern England to the Russo-Polish border, which has been referred to as the Southern Permian Basin. These sediments comprise four main depositional facies; fluvial (wadi), aeolian, sabkha and lacustrine. The lacustrine facies includes important bedded halites.

The Southern Permian Basin is separated from the smaller Northern Permian Basin by the fragmented Mid North Sea–Ringkøbing–Fyn system of highs; and beyond the Grampian Spur lies the even smaller Moray Firth Basin. These latter basins contain rock sequences similar to those of the southern basin, with the exception that bedded halite has not been recognised in the lacustrine facies.

The Rotliegendes of the southern basin apparently conforms to deposition in a Northern Hemisphere tradewind desert similar to the modern southern Sahara. Apart from regional palaeomagnetic considerations and the occurrence of sediments of a non-arid nature south of the Variscan Highlands, the strongest evidence supporting this interpretation is the pattern of palaeowind directions deduced from the orientations of dune bedding. This is seen in both outcrop and core and derived from the continuous dip-meter logs of many North Sea wells by plotting the poles to the bedding attitudes on polar nets.

Palaeowind directions have also been deduced from the dip-meter logs of wells drilled in the Northern and Moray Firth Basins. In general, they indicate winds that blew in a direction opposed to those of the Southern Basin. Thus an area of high barometric pressure seems to have been located over the Mid North Sea High.

When the bedding attitudes in a well are roughly unidirectional, an origin on a transverse dune is normally indicated. With some wells, the indicated palaeowind direction is regionally anomalous. In many cases this can be rectified if the well is presumed to have drilled down through one flank of a seif dune. The indicated palaeowind is thus changed by up to 90° and then commonly fits the regional pattern.

These early Permian desert basins were smaller and closer to the equator than the modern Sahara. Their location in the centre of a continent between the Caledonian and Variscan mountain ranges suggests the analogy of a more tropical equivalent of the Central Asian Takla Makan Desert.


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INTRODUCTION

Since the mid 1960's, the search for liquid and gaseous hydrocarbons has led to very extensive exploration activities over much of the North Sea. Seismic surveys resulted in the delineation of various structures, of which many have been tested by the hundreds of exploration wells drilled so far, and oil and gas is now being produced from some of them. Because of these activities, there has been an explosive increase in our geological knowledge of the area.

One early geological result from this exploration activity was a better understanding of the desert sediments of the Early Permian Upper Rotliegendes. It has long been known that these rocks were deposited in a post-Variscan basin that extended from Germany to the Russo-Polish border. Following the discovery in 1959 of the Groningen gas field in The Netherlands, and the spread of exploration activity to the southern North Sea, it was confirmed that this basin also extended westwards to eastern England. This extension is bounded to the north by the fragmented Mid North Sea–Ringkøbing–Fyn system of highs and to the south by the London–Braabant Platform (Fig. 1).

A study of the Upper Rotliegendes of this basin, based largely on core data from these early wells, established that it comprised four main depositional facies; fluvial (wadi), aeolian (Fig. 2), inland sabkha and desert lake (Glennie, 1972). This Rotliegendes basin was fairly large. At its fullest extent, the basin-centre lacustrine facies alone covered an area some 1200 km E–W by 100–200 km N–S, and accumulated to a thickness exceeding 1000 m. The Rotliegendes desert was sufficiently arid for bedded halite to form important horizons especially within the earlier deposits of the desert lake, and for gypsum crystals (now mostly anhydrite) to grow within the sediments of the surrounding sabkhas.

As hydrocarbon exploration extended further north, another basin, containing Early Permian clastic sediments up to 500 m thick, was delineated beyond the Mid North Sea system of highs. It has become known as the Northern Permian Basin. And beyond the Grampian Spur, over which Rotliegendes sediments seem to be absent, the smaller Moray Firth Basin, with up to 600 m of Rotliegendes, is possibly continuous with the shallower Dutch Bank Basin. Because of considerable local subsidence, especially within the area of the Central Graben, the drill often did not reach the Rotliegendes. Even so, there are sufficient penetrations to confirm the presence of Early Permian rock sequences similar to those of the Southern Permian Basin, although bedded halite has yet to be recognised in their lacustrine facies.

A simple pattern of palaeowind directions had already been deduced for the Southern Permian Basin from cores and continuous dip-meter logs of North Sea wells (Glennie, 1972). It indicated that the Rotliegendes winds match those of the southern part of a modern Northern Hemisphere tradewind desert such as the Sahara. Additional palaeowind data supporting this interpretation has been published by Van Wijhe et al. (1980) from onshore and offshore areas of the Nether-
Fig. 1. Pattern of Early Permian (Rotliegendes) palaeowinds in the North Sea area. Areas of dune sand are stippled. 10°N palaeolatitude is indicated.
Fig. 2. Typical Rotliegendes dune bedding displayed in core from Shell/Esso well 49/26-4. Note the truncation of inclined bedding by sub-horizontal laminae just above porosity/permeability plug 246. The maximum dip angle in this photo is about 25°.
lands, and by Ellenburg et al. (1976) from Germany.

The latitudinal position of the basin is derived from regional palaeomagnetic data (e.g. Van der Voo and French, 1974; Habicht, 1979), which also indicate that NW Europe has rotated slightly in a clockwise direction since the Permian. South of the Variscan Highlands, Early Permian continental sediments are of a non-arid nature.

Early Permian wind directions have also been deduced from the dip-meter logs of some wells drilled in the northern basins, and are presented here (Fig. 1). These directions amplify the pattern already recognised in the south and give a clearer idea of the location of a desert centre of high barometric pressure.

Data from wells drilled prior to 1975 are now available to the public through the U.K. Department of Energy, and such wells have been used extensively in this study.

**DIP-METER LOGS**

The modern continuous dip-meter log measures, on a micro scale, the electrical resistivity of the rock adjacent to each of its four arms as the tool is pulled slowly towards the surface. The best planar correlation of the resistivity values measured by each of the arms is then computed in terms of dip angle and direction of dip relative to the North. With constantly dipping well-laminated strata of uniform lithology adjacent to the tool, the attitude of the bedding can often be measured with a surprising degree of accuracy. This has been confirmed in a number of cases where the Rotliegendes has been cored throughout the logged sequence (see Fig. 3, and also Van Veen, 1975; Van Wijhe et al. 1980).

Unfortunately, not all the dips and directions on a dip-meter log are of the bedding. The most obvious alternative is that of fractures in the rock, either joints or fault planes. This is patently not the case, however, with the 34° dip recorded at a depth of about 6858 ft in Fig. 3; as can be seen on Fig. 2, the dip angle at this depth is about 20° and no fractures are visible. An explanation might be found in the effects of a slight jerkiness as the tool is pulled up through the hole.

For reasons such as this, the dip-meter log is distrusted by some workers because it fails to give such a simple geometric record of the bedding attitudes as that seen in cores. Nevertheless, there is sufficient uniformity in the apparent dips recorded in most wells for them to have statistical significance even though the presence of every data point cannot be defended. Confidence is generated when the interpretation of the dip-meter data of each well conforms to a geologically plausible pattern.

In areas such as the Northern Permian Basin, where the sparse deep-well information indicates locally rapid facies changes but where the precise distribution of these facies is not yet clear, the differentiation between dune and intercalated fluvial sands can only be fully trusted when determined in cores. Unfortunately the Rotliegendes is not hydrocarbon bearing in many of the well penetrations in the Northern Permian Basin, so that cores were rarely taken in these cases. Instead, if confirmation is needed that the dip-meter data concerns dune sands and not some
Fig. 3. Idealised lithological column, gamma-ray, sonic and dip-meter logs of the top 300 ft (90 m) of the Rotliegendes in Shell/Esso well 49/26-4. The uppermost 90 ft (27 m) of the sands were homogenised and deformed during the Zechstein transgression and only locally display preserved aeolian bedding. Thus the dip-meter data for this upper sequence is rejected for palaeowind determinations.
other depositional facies, recourse must be made to other wireline logs knowing that there is regional evidence from other wells (such as cores) that dune sands are to be expected in the area. Calibration of the logs of non-cored wells is provided by those that were cored.

Sands and shales may be differentiated by means of the gamma-ray log. Within a sequence containing both fluvial and aeolian sands, the aeolian units may be determined from a combination of a Sonic (lower sonic velocity) or Formation Density log (lower density) and the Dip-meter log. With an ideal sequence of dune sands, the dip-meter log generally indicates a series of beds whose inclination steepens upwards until it is truncated by the low-angle beds of the overlying set (see Figs. 2 and 3, and Van Veen, 1975).

In practice, the ideal association of data is rarely found. Most drilled sequences of the Rotliegendes were not cored and the wireline logs are sometimes inconclusive, perhaps because of logging problems or because the local history of sediment-transporting wind directions was too complex for a simple interpretation to be possible (cf. well 29/25-1 in Figs. 5 and 6). Yet we are dealing with an area in which desert sediments are known to have been deposited, and there is sufficient core data to confirm the presence locally of very thick continuous sequences of dune sands. By a process of elimination, the clearly non-aeolian facies can be discarded. In some wells, this comprises virtually the total Rotliegendes sequence because only non-aeolian facies are represented.

In the Northern Permian Basin, the best dune bedding is preserved in the lower half of the Rotliegendes sequence. In many localities, this seems to be overlain by a considerable thickness of horizontal to low-angle, largely bimodal sheet sands that have no orientational value.

The upper 90 ft (27 m) or so of the Rotliegendes in Fig. 3 was homogenised and deformed by liquefaction during the Zechstein marine transgression (see Glennie and Buller, 1983). The dip-meter tool found few satisfactory points of correlation, and certainly none bear any relationship to depositional bedding attitudes. The higher sonic velocities seen in the top 30 ft. (9 m) of the Rotliegendes result from an increasing degree of post-depositional calcite (and anhydrite) cementation as the overlying carbonates and evaporites of the Zechstein are approached. By filling in the pore space, the cement alters the primary character of the sands and thus weakens its response to the dip-meter tool.

**BEDDING ATTITUDES AND ROTLIEGENDES PALAEOWIND DIRECTIONS**

The orientations of dip planes may conveniently be plotted on the upper hemisphere of a polar net (Fig. 4). The perpendicular (pole) to the bedding plane plots as a point, which indicates the slope and direction in which the bed faces. A horizontal bed plots in the middle of the net and thus has no directional value. The directional importance attached to the higher dip angles, on the other hand, is
Accretion laminae

Maximum angle of slip face is 34°

Dips only reach 34° in locally developed slip faces

Horn of dune
emphasised by their distribution nearer to the circumference of the net. With simple
dunes, the distribution of bedding attitudes on a polar net can be used to infer the
dune type and to deduce the direction of the wind that caused its construction.

In deducing palaeowind directions from well data, the following assumptions are
made:

(1) Most preserved bedding attitudes in a sequence of dune bedding results from
aeolian activity, and thus can be used to derive the directions of that wind.

(2) At the time of deposition in an arid desert, most dunes are of two relatively
simple types: (a) transverse dunes (or barchans where the supply of sand is limited),
the bulk of whose bedding dips downwind; or (b) if the wind strengths were too
great for transverse forms to be stable, the longitudinal seif type, whose bedding is
oriented almost at right angles to the regional sand-transporting wind but with a
down-wind component to it.

Thus, in an ideal case, the distribution of dune-bedding attitudes on a polar net
should be characteristic of the dune of which it forms a part. For example, if the
spread of points covers a relatively limited area (Fig. 4b), then they probably
represent the avalanche-slope bedding typically found on the leeward slopes of a
transverse dune. A radius bisecting the spread of points or, perhaps better, bisecting
the most dense cluster of points, indicates the down-wind direction (for convenience,
the arrow giving the wind direction is shown in the opposite quadrant). Where,
however, the spread of points forms two clusters that are almost diametrically
opposed to each other, it is indicative of deposition on the flanks of a seif dune (or
the horn of a barchanoid dune) whose axis is parallel to the regional sand-transport-
ing wind (Fig. 4g). Similar observations have also been made by Nurmi and Hepp
(1979). Outlines of the spread of bedding attitudes that the writer considers to be
characteristic respectively of transverse and seif dunes (Glennie, 1970, fig. 68) are
given on the polar nets of Figs. 4, 5 and 6.

The wind direction derived from a polar net ought to be the same as that given by
a rose diagram. In the case of an ideal transverse dune, this should be the statistical
mean (or mode) of the data distribution. The polar net has the advantage over the
rose diagram, however, in that the directionally more indicative higher dips are
emphasised, which permits a degree of subjective interpretation. This can be carried

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Fig. 4. Comparison of the bedding attitudes of ideal barchan (and transverse) dunes (a) and seif dunes (f)
with the dominant dune-constructing wind. The bedding attitudes of each of these dune types were
measured on peneplaned cemented Pleistocene dunes in Arabia, and are plotted on fig. 4b and g. The
dashed outlines on e indicate the ideal limits within which the distribution of points on a polar net is
characteristic of these basic dune types. The plots on 4c and d represent the bedding attitudes met when
only one side of a barchan dune is drilled through and the dashed arrow indicates its apparent rather than
its true derived wind direction. Similarly, 4h and i represent the opposing flanks of a seif dune with even
greater differences between the true wind direction and the apparent directions indicated by the bedding
attitudes.
Fig. 5. The basic data of this paper. Polar nets of dip-meter data from the Rotliegendes of 20 North Sea wells. Dashed outlines indicate areas of typical transverse (a) and self-dune bedding attitudes (b) and the arrow indicates the derived wind direction.

to its logical conclusion (as discussed in greater detail below) when the plotted dip data are all derived from the same flank of a self-dune; the true wind direction could be up to some 70° or more oblique to that indicated by the bisector of the scatter of points (Fig. 4h and i).
Using the above geological rather than strictly statistical reasoning, palaeowind directions were derived from dip-meter data of Rotliegendes dune sequences of many wells (Fig. 5). From these, a map was constructed, which gave the apparent
direction in which the wind blew at each well location (Fig. 1). In general, the arrows at each well location indicate a plausible and geologically (meteorologically?) acceptable pattern of early Permian winds, which blew in a clockwise direction around a barometric High located in the vicinity of the Mid North Sea structural high.

It must be expected, however, that the effects of occasional winds of a non-anticyclonic origin will provide a scatter of points that defy interpretation. Apart from these, the general spread of dip attitudes on the polar nets of some wells still seem to be anomalous. Since the regional palaeowind pattern is so relatively simple, further analysis of this anomalous data is warranted; but first, what should we expect to find in simple transverse and seif dunes?

**Simple transverse and seif dunes**

Transverse and barchan dunes form with winds of moderate velocity. The dunes migrate down-wind with no gain or loss in size when the sand grains that were deflated from the windward slopes are redeposited on the steeper avalanche slopes. The bulk of the latter face in a general down-wind direction at dips of up to 34°, the maximum angle of repose for dry sand, and are thus strongly indicative of the prevailing wind direction (Fig. 4a).

Barchans form where the supply of sand is too limited for the more laterally continuous transverse forms to survive. Because sand can saltate more rapidly over the hard, sand-free surface of a desert floor than over the sand-covered dune (Bagnold, 1941, p. 72) a transverse dune form is no longer stable on the flanks of a barchan, and the sand is drawn out into "horns" that point down-wind. Small, volumetrically unimportant, slip-faces can develop on the inner edges of the horns, which thus face at angles that are highly oblique to the main dune-forming wind direction.

If one studies satellite photos of the broad belt of parallel rows of uniform seif dunes that extend for hundreds of kilometers across the Rub al Khali in Arabia, it is clear that, collectively, they cannot be the vector resultant of two different winds that blew alternately from different quarters, as first suggested by Bagnold (1941) for the origin of seifs. In a little known paper, however, he later (Bagnold, 1953) advocated a change to this hypothesis to one involving an origin by helical wind vortices. This is interwoven below with some of the writer's own modifications.

Like the horns of a barchan, longitudinal seif dunes form in areas where the wind velocities are generally too high for transverse dunes to be stable. Seif dunes are formed by strong winds that flow in helical vortices whose axes are parallel to those of the dunes. The winds of two adjacent vortices diverge at ground level from the centre of the interdune corridor and carry sand grains obliquely up the sides of the flanking dunes (Glennie, 1970, fig. 73). The upper half of each vortex carries no sand as it rotates back down to ground level again over the interdune area. A slight imbalance between the vortices on opposite sides of a dune's axis can result in
avalanche slopes developing crestally. They form oblique to the axes of both dune and helical vortex and, along the length of the same dune, may alternately face in opposing directions. Thus each member of the parallel rows of seif dunes is the resultant of winds that blew in two converging directions at ground level, but these winds are part of a system of vortices whose own axes are parallel both to each other and to the resulting dunes. If, for any reason, there is a general reduction in windstrengths, then the surface of a seif dune may be modified by the development of parasitic transverse forms. When this is of short duration, evidence of their temporary existence may not be preserved. In a subsiding basin, however, their existence may be recorded in a pattern of bedding attitudes that have the characteristics of both seif and transform dunes. Using as a guide the outlines that characterise the two basic dune types, this combination of data points can be seen in many of the wells depicted in Fig. 5. In most cases, however, there is little doubt as to the direction of a dune-forming wind.

*Apparently anomalous dip data*

When a well is drilled through an ancient dune sequence, the distribution of bedding attitudes may represent the deposition of a series of migrating dunes. One such sequence of sand, deposited as part of an individual dune, is seen in the core in Fig. 2 from just below 6872 ft to the porosity plug number 246, a thickness of some 16 ft (5 m). Only the lower part of the original dune sequence is preserved and the bedding reaches a maximum dip angle of 25°. This relatively low maximum dip angle is partly accounted for by truncation, but in general, the dips in Rotliegendes dune sands rarely seem to exceed about 27°. The explanation may in part also be found in the effects of compaction and pressure solution resulting from deep burial. Locally, this has resulted in the loss of perhaps half the original porosity of the sands and thus also in a reduction of dip values (Glennie et al., 1978).

In areas of later tectonic tilt, dips may, of course, be considerably distorted and, if large, should be corrected before the palaeowind direction is deduced. In the wells 9/17-1 and 9/23-1 (Fig. 5), for instance, corrections had to be made for 30° tectonic tilts to, respectively, the northeast and the west. With well 29/25-1, the dip-meter log indicates that the overlying Zechstein carbonates are tilted to the south at an angle of 10°. On the other hand, in 30/16-10, most dips in the eastern sector of the net are concentrated between 20° and 25°. Perhaps here, there is a slight tectonic tilt to the west; and in the nearby well 30/16-8, the distribution of points strongly suggest the effect of a greater westward tilt. No corrections were made for these latter wells.

Corrections were made, however, for the wells 22/18-1 and 30/12-1, where a 10° tilt to respectively the south and the northwest were applied (30/12-1 is also shown uncorrected in Fig. 6). Where present, such tilts can be gauged very accurately if strata of known or probable horizontal habit have been cored (e.g. fluvial clays,
adhesion-ripple horizons; see, e.g., Glennie, 1972, figs. 10 and 14). In this respect, however, the attitude of the overlying marine Kupferschiefer at the base of the Zechstein need not always be horizontal. In the sand pits of Northeast England, its lateral equivalent, the Marl Slate, clearly drapes the Rotliegendes ("Yellow Sands") seif dunes and thus outlines the preserved relief of some 50 m (see Smith and Francis, 1968, Fig. 18). Surface slopes of around 8° are still preserved and there is no sand between the Marl Slate and the Carboniferous in the interdune areas.

In most wells, as seen from the vertical sequence of data points in say 22/10-1 or 30/16-1 (not illustrated), there seems to be little change in the deduced palaeowind direction with advancing time. In some wells, however, such as 29/20-1 or 29/25-1 (Fig. 6) the sands in the lower and upper halves of the Rotliegendes sequence seem to have been derived from directions some 90° or more apart. Since there is no regional evidence to support major changes in wind patterns, an alternative interpretation must be sought.

If the dip-meter data from both the upper and lower parts of the Rotliegendes sequence in 29/20-1 are plotted on the same polar net, the distribution of points

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Fig. 7. Palaeowind directions on the north flank of the Mid North Sea High. The rectangles correspond to UK offshore concession blocks, and the numbers to the respective wells. AuK and Argyll are oil fields with some production from Rotliegendes dune sands.
indicates a wind that conforms to the regional pattern (Fig. 7). This suggests that they were possibly derived from the opposing sides of a seif dune of long duration; if correct, then the axis of the dune must have migrated with advancing time, but the wind direction remained constantly from the west. This possibility is illustrated schematically in Fig. 8. The data from well 29/25-1 is more enigmatic because of its much greater spread, and currently defies an interpretation that is conclusive. A wind from the north is tentatively accepted for the lower half of the sequence and is indicated on the maps as such. The spread of points in the upper half of the sequence has no clear parallel elsewhere in the area.

A similar anomalous pattern occurs in the vicinity of the Argyll oil field in block 30/24 where, at first sight, different wind directions seem to have been responsible for dune formation in three separate wells (Figs. 5 and 6). Pennington (1975) suggests sediment transport from the southwest. If two of the wells (30/24-3 and 30/24-4) are plotted on the same net, however, the resulting apparent wind direction from the northwest is about the same as that found in the third well (30/24-2), which again matches the regional pattern (Fig. 7).

The location of these wells suggests the possibility that in the Argyll area, construction of dune bedding is controlled in part by pre-existing relief. On a relatively small scale, this could comprise the channel of a major wadi on the flank of the Mid North Sea High (Figs. 1 and 8), which was uplifted during the latest Carboniferous and Early Permian. On a slightly larger scale, the wells may have

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**Fig. 8.** Conceptual sketches indicating how pre-existing relief could control the bedding attitudes of aeolian sands. Strong, convection-induced winds in wadi channels commonly blow upstream and sand is first deposited at the wadi margins because that is where the wind experiences the greatest drag. Thus the deepest bedding dips towards the middle of the channel. With continued sedimentation, a dune will rise above the level of the surrounding country. If a seif, and the wind is still in the same direction, the bedding will dip away from its cresteal axis. If a barchanoid dune, then the bedding will be oriented in a down-wind direction. Polar plots a to e indicate the sort of bedding distribution to be expected from corresponding well locations. A. Wadi channel. B. Small half graben. No scale implied.
been located over different flanks of a small half graben such as those associated with the creation of the Central Graben and illustrated by Brennand and Van Veen (1975). As has been argued elsewhere (Glennie and Buller, 1983), there are reasons for believing that the Central Graben was already forming as a rift during the earlier Permian.

In a similar way, most of the dip-meter data from the well 30/12-1 plots in the NE quadrant of the polar net (Fig. 6). If these bedding attitudes are believed to represent a transverse dune, then a wind from about N220°E must be inferred. If, on the other hand, the possibility is accepted that the bulk of the bedding could have been deposited on one flank of a seif dune, then a wind from the northwest becomes probable. A correction for a tectonic tilt of about 10° to the northwest (compare the plots on Fig. 6) adds to the appearance of a seif-flank distribution of points with a wind from about N330°E. This latter direction conforms well with the regional palaeowind pattern (Fig. 7). Similar reasoning has been applied to the well 12/23-1 in the Moray Firth Basin.

Still somewhat anomalous are the uniform southerly dip directions exposed along several kilometers of the southern coast of the Moray Firth near Elgin (Fig. 1). Contrary to the younger age advocated by some workers (e.g. Walker, 1973), the lower part of this dune sequence underlies a zone of deformed and homogenised sands (cf. upper part of Fig. 3) whose origin has been equated with the Zechstein transgression (Glennie and Buller, 1983); thus these lower dune sands are of probable Rotliegendes age. The small spread of bedding attitudes over such a wide area of exposure must indicate dunes of transverse type, which were formed by winds that blew towards the south. Perhaps here, dune deposition was influenced by strong convection over the Scottish Highlands in much the same way that the ENE–WSW-oriented seif dunes at the southern margin of the Rub al Khali have slip faces directed southward towards the barren Hadramaut plateau (Glennie, 1970, p. 148).

REGIONAL SIGNIFICANCE OF ROTLIEGENDES PALAEOWINDS

It is clear from the foregoing analysis that the pattern of Rotliegendes palaeowinds deduced for the North Sea is essentially very simple; and to judge from the studies of modern sand seas by Breed et al. (1979), probably far simpler than the wind patterns of today. The very simplicity of the Early Permian atmospheric circulatory pattern was possibly the result of a polar glaciation when, because of the increased size and intensity of the barometric highs over the ice caps, all other air-pressure belts would have been squeezed towards the equator. Under these circumstances, the winds were probably stronger, colder and much more uniform in direction than is the case today. Frakes (1979) indicates that glacial conditions occurred over Gondwanaland from the middle Carboniferous (Namurian) until the late Permian (Kazanian). Another ice cap was possibly centred in eastern Siberia
(Ustrisky, 1973) or on a microcontinent in a polar ocean to the east of Siberia (cf. Habicht, 1979); it seems to have been much smaller than that over Gondwanaland.

In the Rotliegendes of the Southern Basin, halite deposition in the lacustrine facies seems to correlate with the greatest extension of the dune sands, which roughly coincided with the maximum development of the Australian ice sheet during the earliest Permian (Frakes, 1979). This suggests that aridity may be as much a function of the dessicating ability of strong, possibly even relatively cold, dry winds, as the occurrence of high temperatures. The waning of the Gondwanaland ice sheets was matched in the Rotliegendes basin by an extension of the lacustrine facies at the expense of dune sands. Presumably wind strengths also diminished somewhat (therefore a reduced rate of evaporation) and there was probably also some related shift in direction as the centres of tropical high pressure migrated farther from the equator. If correct, the rate of migration of dune sands would have been reduced, and the surfaces of seif dunes would have become covered by parasitic dunes of transverse type, as can be seen in many modern deserts of today. Evidence of this would have been largely destroyed, however, as much of the upper sands were homogenised and liquefied during the Zechstein transgression (Fig. 3).

The most striking feature of the distribution of Rotliegendes palaeowinds in the North Sea area is the reversal seen on either side of the Mid North Sea High, a distance apart of barely 300 km. It is similar to what Opdyke (1961) called the wheel-round when referring to the change in alignment of linear dunes in North Africa from NW–SE to NE–SW around an axis (the high-pressure Horse Latitudes) that seems to trend obliquely from about 24°N in Egypt to 32°N in NW Algeria (Fig. 9). A similar pattern has been described in Arabia by Holm (1960) and Brown (1960), where the arc of dunes known as the Dahna locally has a radius of curvature of about 400 km, the centre of which lies SW of Riyadh and about 23° from the equator.

Perhaps more directly comparable with the Rotliegendes in terms of the sharpness of the reversal in wind direction, are the dunes of central Australia (Fig. 9). They were formed by winds that rotated in a counterclockwise direction about an axis coinciding with the 26°S parallel (Jennings, 1968). Here, linear dunes were formed by strong Pleistocene winds that blew in opposite directions at points only some 400 km apart. Galloway (1965) and Williams (1973) correlate much of this aeolian activity in Australia with the last (Würm–Weichsel) glaciation, and note that temperatures were several degrees lower than today.

If the Permian and Triassic palaeolatitudes determined by Van der Voo and French (1974) and by Habicht (1979) are correct, then the axis of the Rotliegendes wheel-round in the North Sea area must have been in the vicinity of 10° north of the Permian equator. This is much closer to the equator than any comparable feature seen today or inferred for the late Pleistocene deserts.

The Jurassic–Cretaceous dunes of the Sambaiba Sandstone in South America have been reported as close as about 4° to the present equator (Bigarella, 1979). This
area had a palaeolatitude at the time of deposition that was not much different to now (Habicht, 1979). Together with its more southerly equivalent, the Botucatú Sandstone, this desert had a latitudinal extent similar to that of the modern Sahara. Palaeowind data suggest that, at least seasonally, the dunes of this desert were transported by a Southern Hemisphere anticyclonic wind system centred over the southern Amazon Basin, or some 15° to 20° from the equator.

Dunes of probable Pleistocene age have also been reported within 5° of the equator in both the northern and southern hemispheres (Fairbridge, 1964). The presence of a centre of high barometric pressure, as represented by the Rotliegendes wheel-round, so close to the equator, however, does not conform to modern meteorological patterns. Possibly the Permian distribution of land and water resulted in a shift of the global centres of high and low pressure in much the same way as a major barometric high occurs over central Asia today some 50° from the equator instead of the 25°–30° of Afro-Arabia.

This line of reasoning suggests a comparison with the little known intermontane Takla Makan Desert in the Tarim Basin of Central Asia (Fig. 9, and see Breed et al., 1979, figs. 260 and 266), which is only slightly smaller than the Southern Rotliegendes basin, rather than with the Sahara as earlier suggested by the writer (Glennie, 1972). The Lop Nor playa lake is at the lower east end of the basin. Dunes occupy the basin centre, and most are moulded by winds from the northeast. There is some suggestion of a local wheel-round over the northern edge of the basin, but the major Central Asian barometric high to its north is probably too strong for local centres to
develop fully. Although comparable in size and shape, one striking difference between these two desert areas, apart from latitude, is that whereas the floor of the Takla Makan Desert averages 1 km above sea level, the floor of the Rotliegendes desert may have been up to 1 km below the level of the Permian oceans (Glennie and Buller, 1983).

The deduced anticyclonic Early Permian wind pattern implies a different provenance for the Rotliegendes dune sands of the Northern and Southern basins. De Booy (1968) has shown that the heavy-mineral assemblage of the Southern Basin was derived from Precambrian sources in the Baltic Shield. In the Moray Firth and Northern Permian basins, however, an assemblage from Caledonian (and second cycle Old Red Sandstone) sources is to be expected.

Because of their considerable depth of burial, our knowledge of the Rotliegendes facies distribution in both the Egersund sub-basin of the Northern Permian Basin and in the Central Graben is very limited (Fig. 1). Whether or not the northern Central Graben acted as a barrier to dune-sand migration from the west, will have important implications for the sedimentary facies to be expected farther east.

CONCLUSIONS

Data from the dip-meter logs of North Sea wells can be used to deduce the bedding attitudes of Rotliegendes sand dunes. They plot on polar nets with distribution patterns typical for either transverse or seif dunes or, quite commonly, with elements from both dune types. From these data, the palaeowind direction at each well location can be derived. Apparently anomalous directions can locally be corrected if deposition on one flank of a seif dune is assumed, and especially if the resulting new direction then conforms to the regional palaeowind pattern.

A map of the opposing palaeowind directions on either side of the Mid North Sea High indicates that this area was the site of high atmospheric pressure, possibly the Early Permian Horse Latitudes, around which the wind rotated.

The size of the Rotliegendes desert basins and their location in the middle of a continent between two mountain ranges suggests an analogy with modern deserts in central Asia such as the Takla Makan desert.

The coincidence of strong aeolian activity with halite deposition in the desert-lake facies of the Rotliegendes and also with a major glaciation in Gondwanaland, suggests that aridity in deserts may be as much a function of the dessicating nature of strong, dry, even cold winds, rather than solely of high temperatures.

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THE PERMIAN WEISSLIEGEND OF NW EUROPE: THE PARTIAL DEFORMATION OF AEOLIAN DUNE SANDS CAUSED BY THE ZECHSTEIN TRANSGRESSION

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ABSTRACT


The Weissliegend is a European sandstone unit of largely late Early Permian age. It is underlain by the Early Permian Rotliegend red desert sandstones and is overlain by the conventionally accepted basal bed of the Zechstein—the bituminous marine shales of the Kupferschiefer. The Weissliegend sandstones are characteristically white or grey in colour and have been recognised beneath the North Sea, in Germany and in Poland. Equivalents, which are red or yellow in colour, occur in NE England and at the southern edge of the Moray Firth Basin in Scotland.

From an examination of cliff and quarry exposures in Britain, and of drill cores from southern North Sea gas wells, it is now believed that the bulk of the Weissliegend sandstones (and their equivalents) were originally deposited as aeolian dunes. These dune sands, however, were later modified by a widespread event, the Zechstein transgression, which caused their partial homogenisation, the creation of large-scale soft-sediment deformation structures, and the local and minor reworking of some of the dune flanks.

The preferred mechanism of deformation is interpreted as: (1) entrapment of large pockets of air within the bodies of the dunes by flanking and overlying wetted dune sands; (2) venting of the air pockets when the rising internal air pressures overcame the weight of the hydrostatic head of water and the capillary (cohesive) strength of the overlying wetted sands; (3) the rapid replacement of air by water, which caused liquidisation of the original dune laminae; and (4) the associated collapse and final consolidation of the sands into a tighter packing configuration.

Deformations seem to be more developed in former transverse dunes than in seif dunes. The reason may be that the relatively tightly packed low-angle accretion bedding common on the flanks of seif dunes is more resistant to deformation than the looser avalanche sands that form a major part of transverse dunes. Limited reworking of former dune sands was probably best developed on the steep lee slopes of transverse dunes and the steeper upper slopes of seif dunes.

The lack of reddening of the Weissliegend sandstones-proper is attributed to a combination of their

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accumulation above the Rotliegend water table, to the rapidity of the Zechstein transgression, and to the
anoxic state of the early Zechstein sea floor. The Weissliegend sands, unlike the underlying Rotliegend
into which they grade, were thus never in a diagenetic environment that was conducive to reddening.

Finally, it is recommended that the term Weissliegend be dropped in any formational sense. It should
only be retained for the Weissliegend proper, and their equivalents, to denote a complex facies association
dominated by (1) the uppermost Early Permian Rotliegend dune sands (now partly deformed) that lay
above the water table just prior to the Zechstein transgression, together with (2) the minor erosional
marine products caused by that transgression. The latter, sensu stricto, are Zechstein sandstones of
earliest Late Permian age.

INTRODUCTION

Following the Variscan orogeny, the foreland areas of NW Europe became the
sites of developing basins, horsts and grabens (Fig. 1). The largest of the basins
extended 1500 km from eastern England to the Russo-Polish border, and was the
site of deposition of the Early Permian “Rotliegendes” (an old German miner’s term
for the red beds beneath the Zechstein). These red clastics were largely derived from

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**Fig. 1.** Rotliegend facies and palaeogeography, and limit of Zechstein transgression. Facies distribution
poorly known in Northern Permian Basin. Modified after Glennie (1972) and Ziegler (1978). D = Durham;
E = Elgin; J = well 49/24-J1000 (see Fig. 3); 4 = well 49/26-1 (see Fig. 4).
the rapidly degrading Variscan Mountains to the south, and were deposited in the fluvial (wadi), aeolian, sabkha and lacustrine environments of a tropical desert of internal drainage (Glennie, 1972).

The north-central part of the Rotliegend basin was occupied by a large desert lake, which may have been in communication with smaller lakes in the Northern Permian Basin via the Central or Horn Grabens (Figs. 1 and 2); its extent fluctuated across an encircling inland sabkha in response to variations in the rate of evaporation from its surface or in the supply of wadi water. The lake was flanked on its southern side by a broad belt of sand dunes whose continuity was occasionally disrupted by the wadi channels of northward-flowing ephemeral streams. The dune sequence locally reaches thicknesses approaching 300 m.

Lacustrine and subaerial sedimentation was brought to an end when the basin was flooded by marine waters of the Zechstein Sea. The base of the Zechstein sequence is conventionally taken to be marked by the thin (30–100 cm) black, bituminous Kupferschiefer (Copper Shale) or its English equivalent, the so-called Marl Slate.

At many localities from the North Sea to Poland, up to 50 to 60 m of the uppermost part of the Rotliegend sandstone facies is grey-coloured or even white, rather than red. In Germany, this led to the non-reddened sequences being named the “Weissliegend”. The origin of the Weissliegend has been a controversial subject for many decades (for references see Pryor, 1971a, b; Nemec and Porebski, 1977a). The two most favoured interpretations invoke an origin either as aeolian dunes

![Fig. 2. Conceptual block diagram of the Zechstein transgression of the Southern Permian Basin via the Central and/or Horn Grabens. Note suggested change in dune style from transverse in centre of block to seif in western basin-margin location.](image)
marginal to the transgressing Zechstein Sea, or as the deposits of a shallow-marine environment beneath that sea.

From a study of: (1) Weissliegend intervals seen in core samples taken from southern North Sea gas wells; (2) illustrations of Weissliegend strata seen in publications (especially those of Polish examples by Jerzykiewicz et al., 1976, and Nemec and Porebski, 1977a); and (3) what are now believed to be Weissliegend-equivalent sandstones of red and yellow colouration that occur in cliff and quarry exposures on the U.K. mainland, we have come to the conclusion that neither of these interpretations is correct. Instead, we believe that the Weissliegend-proper comprises essentially three separate elements:

(1) Non-reddened dune sands that are otherwise identical to and grade into red dune sands of the underlying Rotliegend.

(2) Non-reddened dune sands that have been locally deformed and homogenised. Deformation probably resulted from the inundation of the dunes by the rapidly rising waters of the Zechstein transgression.

(3) Non-reddened sands of limited extent that were reworked from the surfaces of the dunes, especially their steeper flanks, by the waters of the transgressing Zechstein Sea.

Clearly, the strict usage of the term Weissliegend should be limited to those sands of the Rotliegend basin that were never diagenetically reddened. Nevertheless, we now advocate that the term be extended to include the Weissliegend-equivalent sandstones seen on the U.K. mainland. The fact that these sandstones are red or yellow in colour, rather than the traditional white or grey, will be the subject of further discussion. We also recommend that the Weissliegend be discarded as a stratigraphic or formational unit, and be retained only to indicate a complex facies association of the uppermost Rotliegend (as outlined in elements 1 and 2 above), and the earliest Zechstein sediments (element 3).

In the following sections of this paper, we intend to outline the history of events that gave rise to the super-position of non-reddened over reddened dune sands, and to the deformation structures within the sands. The lines of evidence and argument leading up to this historical assessment include a description of cores and exposures, an explanation of why the Zechstein transgression is believed to have been an extremely rapid event, and an attempt to explain how the inundation of a dune field could have resulted in the creation of widespread deformation structures.

SEDIMENTARY AND DEFORMATION STRUCTURES

North Sea cores

The Early Permian reservoir rocks of the southern North Sea gas fields consist largely of Rotliegend/Weissliegend sandstones. The upper 15 m or so of these sandstones is increasingly cemented (mainly dolomite and anhydrite) as the boundary
Fig. 3. Cored section from Shell/Esso well 49/24-J1000 (southern North Sea) showing a thin interval (65 cm) of blurred and mildly deformed laminae with an horizon of structureless sandstone. This interval is overlain sharply by the Kupferschiefer (K), and underlain by laminated and cross stratified dune sandstone (A). R = remnant aeolian bedding; E = small (cm-scale) fluid escape structures; H = a thin horizon of almost structureless (homogenised) sandstone; D = blurred laminae which have suffered a degree of dilatency.
Fig. 4. Core photographs selected from the uppermost 120 ft (37 m) of sandstone beneath the Kupferschiefer in Shell/Esso well 49/26-1 (southern North Sea). The Kupferschiefer was penetrated at about 6000 ft, but was not cored. $D =$ parts of larger-scale soft-sediment deformation structures;
$E$ = small (cm-scale) fluid-escape structures; $H$ = structureless (homogenised) sandstone; $R$ = remnant aeolian bedding; $C$ = horizons of apparent clay enrichment; $A$ = laminated dune sandstone. The sandstone is enriched in pyrite between porosity/permeability coreholes 29 and 30.
with the overlying marine Kupferschiefer and Zechstein carbonates and evaporites is approached. This cementation, which has a deleterious effect on gas production, is thought to have arisen from the downward percolation of early Zechstein ground waters (Glennie et al., 1978).

Cementation apart, there are now sufficient cores of Rotliegend sequences to confirm the early interpretation of Glennie (1972) that a widespread sandstone facies represents aeolian dune deposits. Furthermore, the uniformity and unimodality of foreset bedding attitudes, as measured from dipmeter surveys and cores (Glennie, 1983), suggest that the sequence was deposited by dunes of transverse type (Fig. 2).

Towards the top of many of the cored intervals, however, the characteristically reddish (Rotliegend) sandstone commonly becomes pinkish, grey or whitish in colour (Weissliegend), and the typical laminated aeolian bedding tends to break down into an association of indistinct facies whose characteristics are, at first sight, far from clear. This association typically extends right up to the sharp and planar boundary of the overlying Kupferschiefer, and has a variable thickness ranging from less than one metre to a maximum observed thickness of over 50 m.

Locally, the critical sections (e.g. Fig. 3) contain relatively sudden, yet gradational, changes from laminated and cross-bedded dune sandstone, into a thin (65 cm in Fig. 3) interval of blurred and mildly deformed laminations with horizons of indistinct or structureless sandstone. More commonly, however, there are much thicker and more complicated zones of structureless horizons. These are often associated with “ghostly” contortions and areas of clay enrichment which alternate with, or grade into, horizons of blurred and indistinct lamination (e.g. Fig. 4). In other examples (not illustrated) the loss of structure tends to intensify upwards, yet the lower boundary between the indistinct bedding and laminated dune sandstone is always relatively sharp. In many cases the blurred and mildly deformed laminations are reminiscent of original aeolian bedding.

Early and unsubstantiated interpretations of these deposits (e.g. Glennie, 1972; Van Veen, 1975) have appealed to some type of marine reworking of original dune sands caused by the waters of the transgressing Zechstein Sea. We now believe that these intervals represent the internal structures of deformed aeolian cross bedding. The cores described above contain features and sequences very similar to those recorded by Doe and Dott (1980) from contorted zones in the Weber and Navajo sandstones of Utah. The stratigraphic position of our zones leads us to believe that they were generated either by events: (1) leading up to the Zechstein transgression; (2) related to the Zechstein transgression; or (3) coincident with the transgression.

In order to test these hypotheses we examined exposures of the “Yellow Sands” of Durham, and the Hopeman Sandstone of Morayshire (D and E, respectively, on Fig. 1).
The "Yellow Sands" of Durham

In County Durham, at the western edge of the Southern Permian Basin, an aeolian facies of the Rotliegend ("Yellow Sands") is wonderfully displayed in a series of quarry exposures. These unconsolidated aeolian sands have a preserved thickness beneath the Marl Slate of up to 60 m.

In contrast to the transverse dune bedding recognised in the southern North Sea gas wells, the Yellow Sands are thought to have been deposited over a Carboniferous substrate as a series of large and essentially simple seif dunes with a NE–SW alignment. These are shown conceptually in Fig. 2. The factors that led to this interpretation are: (1) observations of interdigitating foresets in axial zones; (2) bi-modal cross-bedding azimuth distributions which are symmetrical about the dunes' axes and presumed dominant wind direction (see Glennie, 1970, figs. 79 and 80; Glennie, 1972, fig. 13); (3) outlines of isopachs (Smith and Francis, 1967, fig. 11), which clearly show the presence of parallel linear features whose axes of elongation correspond to the dominant wind direction deduced from (2) above; and (4) the presence of symmetrical, if somewhat rounded, profiles of individual dunes preserved beneath the Marl Slate (Fig. 5).

Deformation structures are occasionally found within these seif dunes, but they are neither common nor particularly large (one of the largest is shown in Fig. 6). They are always located at the very top of the aeolian sequence and seem to be limited to the more steeply inclined bedding near the axial parts of the dunes.

Perhaps one of the most striking aspects of these quarry exposures is the absence of a thick transgressive sequence, and the absence of a planar transgression surface. As mentioned in (4) above, the former dune surfaces have simply been truncated to rounded profiles which, in crestal areas, commonly have little more than a few centimetres of reworked sands between the dune bedding and the Marl Slate. Apart from the absence of their former crest lines, the rounded profiles probably give a close approximation to the original sizes and shapes of the Permian seif dunes.

Access to the transgression surface itself, and to the sequence lying immediately below the Marl Slate, is difficult because of active sand extraction in these Durham quarries. The critical exposures are invariably at the top of a vertical face of unconsolidated sand some 30 m or more high; and because of similarity in grain size, the contrast between the aeolian and reworked sand is difficult to discern from a distance. Nevertheless, there are a few accessible exposures in which the reworked sands can be seen to locally reach thicknesses of 4–5 m on the dune flanks, but generally seem to be much thinner. These sands grade up into the base of the Marl Slate and locally contain disarticulated valves of *Lingula credneri* in the top 2–3 cm below the contact (Bell et al., 1979).

At one locality (Fig. 7) the reworked sands can be seen to have been derived from up-slope avalanching of the original dune surface. These avalanche sands are either structureless, vaguely layered, or contain complex wavy laminations. Elsewhere,
Fig. 5. Panorama showing the curved surface of a former seif dune after rounding during the Zechstein transgression. The surface conforms to the base of the Marl Slate, which was largely bulldozed away prior to quarrying the underlying almost unconsolidated Permian “Yellow Sands” (Weissliegend equivalent). Crime Riggs East Sand Pit, Country Durham, England.
Fig. 6. Typical interdigitating style of cross-stratification in a near-axial location of a former seif dune. Large-scale deformation (D) is limited to a near-crestal position where over-steepened bedding dips obliquely towards the viewer. Permian “Yellow Sands”, Crime Riggs West Sand Pit, Durham, England.

Some of the reworked sands contain examples of very clear surface-parallel lamina- 
tions and low-angle discordances. These latter horizons may represent local occur-
rences of beach or near-shore sands of very limited thickness and extent (Fig. 8). 
Their development seems to be confined to only part of the upper dune slopes and, 
over a short distance, may range in thickness from the total reworked sequence 
(some 60 cm in Fig. 8) to only the top 10 cm.

Another feature of the locality illustrated in Fig. 8 is that the marine reworked 
sands truncate both relatively undeformed dune sands and an almost structureless 
sequence that grades down into normal undeformed laminated dune sands. Clearly 
the loss of structure of the dune sands preceded the marine reworking of the 
overlying sands. These structureless sands are very similar in appearance to some of 
those seen in well cores, including the gradational relationship into adjacent laminated 
sands. The development of deformation structures, however, is much too limited in 
these seifs for a detailed comparison to be made with North Sea cores. As will be 
explained later, we suspect that the difference in degree of deformation is directly
Fig. 7. The scalloped surface of a former seif dune overlain by several metres of low-angle, weakly layered, wavy-laminated or structureless sands. The latter are interpreted as the avalanche products of the upper slope of the original dune surfaces, that were perhaps undercut by the rising floodwaters of the Zechstein Sea. These reworked sands grade up into the Marl Slate in the upper right-hand corner. Carbonates of the Zechstein Cycle 1 form the background. Permian “Yellow Sands”, Eppleton Sand Pit, Hetton Downs, Durham, England.

related to dune type; transverse in the centre, and seif at the western edge of the basin.

What is of greater importance here is that the quarry exposures in County Durham have given us an idea concerning the rapidity of the Zechstein transgression. We argue that the thinness of the reworked sands, together with the essential preservation of the dune forms beneath the Marl Slate, could only have resulted from an extremely rapid transgression. Had the transgression taken thousands, or even hundreds of years to complete, it seems unlikely that 50 m-high dunes would have survived with only a few tens of centimetres of reworking.

The Hopeman Sandstones of Morayshire

In striking contrast to the limited extent of deformed sandstones in Durham, is another sequence of aeolian sandstones on the Moray coast of Scotland, north of
Fig. 8. The almost structureless sand (H), sandwiched between two sets of dune bedding, grades down to the left into well-laminated dune sands. To the right, they pass into weak vertical laminae, which like the upper dune sands in the photo, are truncated and overlain by the low-angle basal Zechstein marine sands (Ze), here thought to have the characteristics of local beach development. The overlying Marl Slate was removed by quarrying. Permian "Yellow Sands", Crime Riggs East Sand Pit, Durham, England.

Elgin (Peacock, 1966; Peacock et al., 1968), where deformation has occurred on a scale similar to that inferred from North Sea cores. Both undeformed dune bedding and deformation structures are exposed along some 10 km of cliffs and foreshore (Fig. 9A) where they are collectively known as the Hopeman Sandstones (Peacock et al., 1968). The relative uniformity of the bedding attitudes of undeformed sandstones (Figs. 9B and 10) indicates that they were deposited as a series of transverse dunes. The Hopeman Sandstones can be divided into two units, the lower unit having the simpler dunes of the two (Fig. 9B). The lower unit, which is mostly exposed along the foreshore, repeatedly grades both laterally and upwards from
Fig. 9. (A) Map of part of the Moray coast, Scotland, showing areas where large-scale deformation occurs in the lower unit of the Hopeman Sandstone; (B) polar nets of the bedding attitudes of dune sands of presumed Early Permian and younger ages. (C) listing of some key localities.
Fig. 10. South-dipping sequence of sandstones of a transverse dune that probably correlates with the Weissliegend. The sandstones grade up into a large deformation structure, which does not extend south of this point. Hopeman Sandstone, Locality 11, East Covesea, Morayshire, Scotland.

dune-bedded sandstone into areas of large deformation structures several hundred square metres in extent and up to about 20 m in thickness. The upper unit, on the other hand, comprises undeformed dune-bedded sandstone that has been extensively quarried for building stone. The contact between the deformed sequence and the overlying dune sandstones is either poorly exposed or is difficult of access.

Bedding attitudes in the undeformed strata of the lower unit indicate that the dominant dune-forming wind blew towards the south (Figs. 9B and 10). In the upper sandstones, however, the palaeowinds seem to have veered to the southwest, with non-conforming dips suggesting that the wind pattern was not so stable as during deposition of the lower sequence (Fig. 9B).

In isolated outcrop areas some 6 km to the south, about 45 m (150 ft) of dune sandstone have been dated by fossil reptiles from near the base of the sequence as Late Permian or Early Triassic (Walker, 1973). We believe that these dated sands, which at this locality lack deformation structures and overlie the Devonian Upper Old Red Sandstone, probably correlate with the undeformed upper dune unit of the coastal sequence (Fig. 12). We suggest that the undated (only reptilian foot prints)
sandstones of the lower coastal unit, which had a more basinward palaeolocation than the inland exposures, are of Early Permian age, and are therefore time equivalents of the Rotliegend. The complete Hopeman Sandstone sequence has a thickness of 60 m in an IGS boring at Clarkly Hill, SE of Burghead (Fig. 9A; Peacock et al., 1968). Offshore, however, the Rotliegend sandstones alone (i.e. the lower unit) have a thickness of up to 600 m (Ziegler, 1982).

At locality 9 on the coast (Fig. 9A), where the cliff is about 20 m high, north-dipping dune bedding of the lower unit rapidly curves up to form the flanks of a large fluid-escape structure whose base is not seen (Fig. 11). The uplift seems to be linear rather than circular, with its axis almost perpendicular to the strike of the dune bedding. Definition of the deformed dune bedding weakens towards the core of the uplift where the sands take on an almost structureless appearance. The flanking laminae, on the other hand, are fairly well preserved and seem to have been foreshortened by crumpling. This suggests that the structure is in some way also connected with the vertical effects of sand compaction. Deformation structures
Fig. 12. Suggested stratigraphic relationship between the dune sandstones observed along the Morayshire coastline (Scotland) and those penetrated during offshore drilling in the Moray Firth. If our belief that the deformation structures were formed in response to the Zechstein transgression is correct, then the lower dune sandstone unit must equate with the Rotliegend, and the deformed sandstones with the Weissliegend. The overlying dune sandstones of the upper unit might be of Late Permian or Triassic age (for further explanation, see text). (Figure is not to scale).
associated with another more complex fluid-escape structure some 10 m high, can be seen in the cliff just below Covesea lighthouse.

An another locality (6 on Fig. 9A), a series of shallow concave structures (Fig. 13), some several metres across and with well-preserved concordant laminae, are separated by narrow vertical zones of structureless sandstone (Fig. 14). These saucer-like structures occur in different sizes and with varying degrees of distortion, and are fairly common features of the foreshore. They are interpreted as comprising areas of formerly continuous sand with surface-parallel lamination, such as is commonly found on the upper windward slopes of transverse dunes; patches of these sands now seem to have sagged into an underlying quicksand.

This deformation appears to have taken place in a sub-aquatic environment, as is evidenced by the upward gradation of the deformation sequence, with very little angular discordance, into a thin bed capped by shallow-water ripple marks that have been partly erased by streaming lineations. Thus the water in which this event took place is likely to have been very shallow. Other smaller shallow-water ripples are overlain by some 15 cm of structureless sand, which is thought to have been
extruded over the submerged surface of the dune from a nearby sand volcano, or oozed over the subsiding saucers from marginal linear fractures such as that depicted in Fig. 14.

At a nearby third locality (5 on Fig. 9A), some 3 m of deformed sandstone is clearly inverted over an area of some 100 m$^2$ or more above a fold plane that dips gently to the north-west (illustrated by Peacock, 1966, PL.IX, fig. 3). Such an unexpected structure in what is, or was, still essentially a sequence of dune sands, is interpreted by us as having been formed in moist sands in a sub-aerial environment just above water level, where the surface beds of the sand dune were locally strongly deformed. This was possibly accomplished by a combination of subsidence below the future location of the inverted bed, and an adjacent linear uplift formed by a fluid-escape structure upslope (downwind) from the axis of inversion. This interpretation is shown schematically in Fig. 15.

Thus, we have strong evidence that both large-scale deformation and related loss of bedding definition of a sequence of dune sands took place in the presence of water; and, in some cases, we have clearly demonstrated that the deformation must have taken place below a shallow cover of water, presumably shortly after submer-
Fig. 15. Conceptual cross-section of transverse dune to illustrate how the dune bedding originally illustrated by Peacock (1966; PL IX, fig. 3) may have become overturned.

gence. The similarity between the deformation structures seen on the Moray coast and the features seen in North Sea cores is so striking that we have no hesitation in suggesting that the latter formed under broadly similar aquatic conditions. Indeed, with the knowledge that Late Permian (Zechstein) sandstones occur only a short distance off the Moray coast (Chesher et al., 1972), and that thick Rotliegend sandstones have been drilled through farther offshore (Fig. 12), we feel that the origin of the deformation structures described above was probably also intimately related to the time of the Zechstein transgression.

The dune sandstones that clearly underlie the deformation structures east of Covesea (Fig. 10) can be traced southwards along the line of a former (sub-Recent?) Clifford coastline for another 200–300 m (Fig. 9A). However, no more deformation structures can be seen. Deformation structures are also lacking from the boring at Clarkly Hill (Fig. 9A; see also Peacock et al., 1968). This leads to the conclusion that the present coast between Burghead and Lossiemouth may approximate to the maximum extent of the Zechstein Sea (Fig. 1).

Thus, deposition of the lower dune sequence of the Moray coast was probably time equivalent with the Rotliegend of other localities, whereas the overlying undeformed dune sands are of Late Permian to Early Triassic age. The lack of any permanent Zechstein deposition in this area is ascribed to its extreme basin margin location, but evidence that the transgression reached the Moray coast is deduced from the occurrence of water-laid sandstones (ripple marks, streaming lineations) in association with the massive deformation structures. The validity of this correlation would also seem to be supported by the common occurrence within the Hopeman Sandstone deformation series of small radiating clusters of barite crystals (Peacock et al., 1968) and the recognition of sulphide minerals in the Weissliegend of North Sea cores (Fig. 4) and in Poland (Jerzykiewicz et al., 1976). Their origin seems to be
related to the interplay between the oxidising ground waters that prevailed during Rotliegend deposition and the strongly reducing conditions of Kupferschiefer deposition.

*The Fore-Sudetic Monocline of Poland*

We have already mentioned Poland as being a country where deformation structures in the Rotliegend/Weissliegend sequences are similar to those described above. It is pertinent, therefore, to outline the salient features of these Polish examples. The following descriptions are derived largely from three papers (in English): one, by Jerzykiewicz et al. (1976), concerns exposures in copper mines together with related borings; and the other two, by Nemec and Porebski (1977a, b), are based on cores from 28 borings scattered over an area of some 2000 km². The regional distribution of Rotliegend facies and the extent of the Zechstein transgression is described by Pokorski and Wagner (1975) and Depowski (1978).

In the Fore-Sudetic Monocline of SW Poland, the complete Rotliegend/Weissliegend sandstone sequence has a thickness that reaches some 300 m, the lowest 60 m of which is largely of fluvial origin, and the remainder is aeolian. The uppermost 1–60 m of this latter sequence is light grey to pale orange in colour, and is thus classified as Weissliegend. The change from orange to grey, however, is gradational and does not coincide with any sedimentological break (Nemec and Porebski, 1977b). The top 0–20 m of the Weissliegend comprises sandstones that have been subjected to local, possibly post-depositional, clay enrichment, to various types of deformation similar in style to those seen in North Sea cores, and to reworking in a marine environment. A bioturbated horizon is locally recognised just beneath the Kupferschiefer.

Jerzykiewicz et al. (1976) divide the Weissliegend into two units, alpha and beta. The lower alpha unit comprises planar cross-bedded sandstones which, following Pryor (1971b), these Polish writers ascribe to deposition in a shallow-marine environment similar to that described by Houbolt (1968) from the present strongly tidal North Sea. Nemec and Porebski (1977a), on the other hand, considered these sands to be aeolian, and probably deposited as coastal dunes marginal to the transgressing Zechstein Sea.

The upper beta unit, which includes the structureless sands and deformation structures, is interpreted by both these groups of writers as having been deposited in a shallow-marine environment. Much of the deformation is ascribed by Jerzykiewicz et al. (1976) to liquefaction by dewatering. In the light of our experience elsewhere, this is an interpretation for which we have sympathy, although we propose different reasons for its occurrence. The Polish upper Weissliegend sequence is very similar to that seen in the North Sea; like the North Sea, however, we can discern no sedimentary structures in the photographic illustrations from Poland that are diagnostic of a shallow-marine environment of deposition.
The occurrence of sulphide minerals in the Weissliegend of Poland, and the support it gives to regional correlations such as the timing of the Hopeman Sandstone deformation, has already been mentioned. The distribution of these sulphur-based minerals is evidence of the very large area that was subjected to a remarkably similar diagenetic process.

**ZECHSTEIN TRANSGRESSION**

In the previous sections we have tried to demonstrate that the preservation, deformation and reworking of the original aeolian dune sands seem to have occurred in the presence of water, and that this water was clearly supplied by the transgression of the Zechstein Sea. Furthermore, the evidence indicates that this transgression must have been geologically extremely rapid (Smith, 1970, 1979), since it failed to peneplain the dune-covered surface of the Rotliegend basin, and the small volume of marine-reworked sands lack structures indicative of more than very limited beach or shallow-marine development.

The amount of aquatic reworking of these Permian dunes is far less than that found by Tanner (1970) in the upper part of the Jurassic Entrada Sandstone of New Mexico. There, truncation of aeolian dunes by the action of water of the succeeding "Todilto Lake" resulted in the interdune hollows being infilled with up to 30 m of water-laid sandstone.

Another feature supporting the idea of a rapid transgression is the paucity of fauna in the marine-reworked sands. Fossils are known only from the uppermost 10–20 cm or less of sand, and even then seem to be confined to basin-margin locations such as NE England where they have clearly been reworked prior to deposition of the Marl Slate. There is no evidence of bioturbation. Pelagic foraminifera and fish are known from just below the Kupferschiefer in a basin-centre environment in NW Germany (Plumhoff, 1966). The implication is that the basin centres became the sites of deep water too rapidly for normal shallow-marine forms to become established and, because of a lack of light, no others colonised the area prior to deposition of the Kupferschiefer.

Facies variations within the early Zechstein sequences indicate that these rocks were deposited over an area of considerable relief (see, e.g. Smith, 1970; Schlager and Bolz, 1977; Taylor, 1981). Shallow-marine limestones and sabkha anhydrites covered the structural highs and basin margins, whereas halites accumulated in great thicknesses over the more basinal areas of former Rotliegend deposition such as the basin-centre desert lake. Thus, prior to the transgression, the surface of this Rotliegend continental basin was probably considerably below that of the Permian open ocean, which lay almost 1500 km away between North Greenland and Finnmark (Forbes et al., 1958; Calloman et al., 1972).

For the Zechstein transgression to have taken place rapidly, both the Rotliegend
depositional basins and the route along which the transgressing floodwaters travelled must have been below the level of the Permian oceans. The route followed by the floodwaters probably comprised an essentially tensional proto-Atlantic rift (Russel, 1976) and the Viking/Central graben system (Fig. 16). With a route already constructed, the transgression possibly became effective because of a eustatic rise in sea level caused by the waning phase of a Permian polar glaciation (cf. Smith, 1964, 1979); later glacial and interglacial cycles were possibly responsible for the cyclic nature of Zechstein deposition as the ocean level rose and fell. This interpretation is broadly in keeping with the findings of Ustrisky (1973) regarding variations in the Permian peri-polar climate in Siberia.

Fig. 16. The post Variscan mega-fracture pattern in the North Atlantic (simplified after Russel, 1976; Ziegler, 1978) illustrates the route along which the marine waters of the Zechstein Sea may have flowed from the Early Permian open ocean to the Rotliegend basins. \( M = \) Moray Firth Basin; \( NPB = \) Northern Permian Basin; \( SPB = \) Southern Permian Basin; \( MNS = \) Mid North Sea High; \( RF = \) Ringkøbing-Fyn High; \( VG = \) Viking Graben; \( CG = \) Central Graben; \( HG = \) Horn Graben; \( OG = \) Oslo Graben.
Another critical factor needed for rapid flooding of the Permian basins would be the easy and rapid removal of what may have been no more than a valley-bottom sedimentary barrier separating the boreal ocean from the water-free channel to the south. By contrast, the influx of Tethyan water via SE Poland seems to have been limited (Ziegler, 1982). If the barrier here was more durable than that in the north, local influx could have been related directly to the rate of eustatic rise in sea level, which would have been relatively slow.

From the evidence of Zechstein Cycle I carbonates cropping out in NE England, Smith (1970, 1979) showed that the Zechstein Sea in this basin-margin location must have had a depth of at least 60 m; for the basin centre, he hazarded a depth of 250 m. Using different criteria, Ziegler (1982) calculated that, in order to accommodate the 1500 m of Zechstein Cycle II halite found in the centre of the Southern Permian Basin, the surface of the Rotliegend desert lake must have been in the region of 200–300 m below mean sea level prior to Zechstein flooding.

If we accept a depth of 250 m as being realistic, then the volume of water initially required to fill the basin would be in the order of 75,000 km$^3$. Up to another 35,000 km$^3$ would also be needed to fill the Northern Permian Basin, making some 110,000 km$^3$ of water that would have to be transported along the channel from the area of the present Arctic Ocean (Fig. 16). The Moray Firth Basin was probably too shallow to add much to this figure.

Ignoring the effects of seepage and evaporation, the basins would have been completely filled in about 6 years if the water was supplied at a rate of 50 km$^3$ day$^{-1}$. This volume could have been achieved if the rift channel was 10 km wide, the depth of water in the channel was 20 m, and it flowed with a mean velocity of 3 m s$^{-1}$. The suggested parameters controlling the rate of water influx are possibly conservative, and conceivably the rate could have been doubled. For instance, the minimum width of Permian strata in the Viking and Central Grabens seems to be at least 20 km, and mostly is much more.

More pertinent to our thesis is that at the suggested rate of influx, the daily rise of water level in the basins could have been measured initially in decimetres per day. Thus lake-margin dunes 50 m high could have been completely inundated in, say, 8 months.

The limited extent of reworking of dune sands possibly reflects the very rapid Zechstein transgression, there being little time during which individual dunes are likely to have been affected by storm-induced waves. Furthermore, on a gently sloping basin floor, most dunes will have been protected from strong wave action by the shoal system built of those already submerged dunes nearest to open water. Any turbulence associated with the entrance of the Zechstein floodwaters to the basin must have been dissipated in the waters of the pre-existing desert lake long before the dunes were reached some 100 km or more from the point of entry (Fig. 2) although a metre of presumably reworked sandstone overlies the desert-lake facies in some North Sea wells (e.g. Rhys, 1974, fig. 2b). A modern example of the survival of
a pattern of dunes that were inundated in a temporary desert lake is illustrated by Glennie (1970, fig. 49) from the Djofra Graben, Libya.

Recently formed seif dunes rarely seem to exceed a height of 10–20 m. To judge from the Durham exposures, and from the depth of deformation in North Sea wells, the Rotliegend dunes were relatively large and probably reached a height in excess of 50 m. This height difference can be explained if it is assumed that the larger Rotliegend dunes resulted from a wind system that had been strengthened by the effect of an intense Permian glaciation (cf. the probable Pleistocene age of large Arabian and Saharan dunes; Glennie, 1970, p. 94, 1983). The amelioration in the climate that is presumed to have led to a eustatic rise in sea level and to the Zechstein transgression, would have resulted in a much weaker wind system as the polar area of high pressure contracted. The lack of strong desert winds, in turn, will have meant reduced evaporation and an expansion of the desert lake (Fig. 2; Glennie, 1983). It may also have resulted in minimal wave action during the critical period of flooding.

TOWARDS A MODEL

So far, our thesis hinges on two main lines of argument: first, that the partial preservation and limited reworking of the dune field resulted from the extremely rapid transgression of the Zechstein Sea; and second, that the deformation took place during the flooding, and perhaps shortly after individual dunes had been submerged. The outstanding points left for consideration are the mechanism of deformation, and the event (or events) that triggered that mechanism. Can, for example, the effects of the transgression be appealed to as having caused the generation of the structures, or does an additional and coincident event have to be sought?

The mechanism of failure

In recent works there is agreement that many large-scale deformation structures resulted from liquefaction of originally unconsolidated, yet water-saturated, aeolian dune sands (e.g. Doe and Dott, 1980). Some of the features which support this hypothesis are complexly folded and contorted zones whose relatively sharp lower boundaries grade quickly downward into undisturbed dune bedding. Internally, some of the original dune laminae tend to be preserved in spite of the deformation, and gradational changes from undisturbed laminae, through blurred laminae, to indistinct (structureless) horizons, are thought to be indicative of increasing grain dilation and flow as the sands were held in a liquid-like state (Doe and Dott, 1980). Each of these features has been observed in the Weissliegend sandstones (e.g. Figs. 3, 4 and 10), together with more dramatic flow phenomena such as sandstone dykes and large-scale fluid-escape structures (Figs. 10 and 14). Liquefaction, therefore, seems an appropriate mechanism for the Weissliegend.
Possible triggering mechanisms

Various triggering mechanisms have been put forward to explain the generation of large-scale deformation structures both in aeolian and subaqueous sandstones (see review passages in Doe and Dott, 1980; Horowitz, 1982). Of the variety on offer, only three of them seem conceivable for the Weissliegend. These are: (1) liquefaction induced by earthquakes (e.g. Horowitz, 1982); (2) liquefaction resulting from storm waves (Seed and Rahman, 1977); and (3) liquefaction caused by rapid fluctuations of ground-water level (Doe and Dott, 1980).

A fourth possibility is deformation resulting from the entrapment and expulsion of large quantities of air. (This is of our own making.)

Earthquake-induced liquefaction

In order to adopt the earthquake-induced hypothesis, it would be necessary to envisage intense seismicity on a regional scale. There is, however, no evidence to substantiate regional seismicity at this time. Furthermore, the period of seismicity would have had to have corresponded exactly to the period of transgression, and such a condition seems unnecessarily fortuitous. We therefore eliminate this triggering mechanism for the Weissliegend, and turn instead to those involving the role of water, and thus the role of the Zechstein Sea itself.

Storm wave-induced liquefaction

Seed and Rahman (1977) have demonstrated that cyclic pore-pressure changes due to oceanic storm waves can cause liquefaction of shallow subaqueous sands to depths of several tens of metres. These depths of potential failure are appropriate for the Weissliegend, but the credibility of such an idea has to be tested against the likely hydrographic environment developed during the sudden inundation of an intracontinental dune field.

Clearly the dimensions of the Permian basins were not comparable to those of oceanic settings, even when fully flooded. A maximum depth below sea level of 250 m has been suggested for the level of the desert lake prior to flooding, and the widest dimension of the southern basin is approximately 1500 km (Fig. 1). During the flood, the intracontinental sea floor would thus have had extremely low slopes and (initially) very shallow water depths throughout. Furthermore, the submergence of dunes locally 50 m in height would have resulted in extensive zones of essentially shoal-like sands with extremely hummocky bathymetries. Such conditions are hardly conducive to the generation, procession and shoreward penetration of major storm waves.

Other adverse factors relate to the prevailing meteorological conditions. Earlier we have mentioned that the eustatic rise in global sea levels (which led to the transgression) was probably accompanied by an amelioration in climate and the development of a weaker wind system. If correct, the intensity and frequency of
storms would have lessened during this time; also, the period of flooding itself is thought to have been extremely short. Moreover, the waves would have had to have attacked the retreating shore lines from virtually all quarters in order to meet the requirement of regional, penecontemporaneous deformation; and this is an unlikely situation.

Another objection concerns the generally accepted precondition for liquefaction that sands must be saturated (e.g. Horowitz, 1982, p. 155). Many of the dunes under attack would still have been partly above the floodwater level, and even those that had been recently submerged would have contained some entrapped air (see the following sections). Much of the effects of shocks induced by any substantial wave trains would have been quickly lost due to the low viscosity and density of air (Doe and Dott, 1980, p. 799).

Taking all of these limitations into account, the likelihood of substantial storm waves ever having been generated, yet alone having caused widespread liquefaction and deformation, seems very slight indeed. We therefore reject this triggering mechanism for the Weissliegend.

*Liquefaction caused by changes in water-table level*

Doe and Dott (1980) have argued that steeply dipping avalanche deposits are so susceptible to granular collapse and liquefaction that even a subtle triggering mechanism, such as rapid fluctuations in water-table level, may be sufficient to cause failure. The Zechstein flooding would certainly have caused the ground-water table to have risen quite suddenly within the dune field, and there was clearly no shortage of avalanche deposits available for attack as most of the Weissliegend dunes are believed to have been of transverse type.

The major problem, however, is that the postulated rapidity of flooding would not have allowed sufficient time for this triggering mechanism to have affected the entire bodies of the dunes before other processes may have taken over. The rise in floodwater level would have greatly outpaced the rise in the level of the ground water. Whilst the upward movement of the water table would still have been affecting only the lowermost bodies of the dunes, the external dune surfaces would have been undergoing progressive, and finally total, submergence.

During these stages, air is likely to have become trapped between the rising level of the water table and the sediment/floodwater interface. The inward penetration of complex fronts of capillary water would have been assisted by the increasing weight of the floodwater column, and the fronts would have moved preferentially along the finer-grained laminae because capillary forces are greater in these and in sands that are less well sorted (De Boer, 1979). This preferential movement would have resulted in the by-passing of much air on two scales: (1) within the intergranular pore spaces of especially the laminae of coarser sand (unpublished experimental observation by K.W.G.); and (2) larger pockets of air deep within the body of the dune. Later, as flooding proceeded, the deeply entrapped pockets of increasingly compressed air
would have tried to force their way upwards, perhaps coalescing with other pockets. The escape of the air in these pockets (see following section) may initially have been resisted by the cohesive strength of the newly flooded and more crestal dune surfaces, which would still have contained a three-phase sand/water/air mixture (cf. De Boer, 1979). In the totally saturated sections of the dunes, the water table would thus have been eliminated; in the parts containing air, any further effects of the rising water table would have become intricately associated with the entrapment of the air pockets, and the possibility of further liquefaction being induced by the rising water table per se, would have been lost.

It follows, then, that any deformation that resulted *solely* from rapid fluctuations in the level of the water table, could only have been centred in the lower bodies of the dunes. The influence of such deformation events may have occasionally extended to the tops of the dunes, perhaps aided by the collapse of overlying sand (Horowitz, 1982), but it is unlikely to have occurred in every case. Yet a feature of the Weissliegend deformation is that it invariably extends right to the very tops of the dune sandstone sequences (e.g. Figs. 3, 4 and 6 and perhaps 10 and 11). Furthermore, some of the deformation is thought to have occurred after submergence, and this, clearly, could not have been triggered by fluctuations in water-table level.

On these grounds we exclude the possibility that fluctuations in the level of the water table acted as the sole triggering mechanism for the Weissliegend deformation. Indeed, we are sceptical whether such a delicate mechanism could ever be capable of producing large-scale liquefaction and deformation in dune sands.

**Deformation resulting from air entrapment and expulsion**

In the previous section we introduced the concept of air entrapment. This concept, when related to the generation of deformation structures, is not new (e.g. De Boer, 1979); but the magnitude which will be invoked below is on an hitherto unimagined scale.

Because of the inherent high porosities (Hunter, 1977) and permeabilities of dune sands, hydrostatic pressures in water-saturated sands of a dune will increase in concert with flooding at a rate of one atmosphere for every 10 m depth below the water surface. That same high permeability permits the total pocket of air-filled dune sand to have the same pore pressure throughout, so that for every 10 m column of air trapped within a dune, there will be a pressure differential at its contact with the overlying wetted dune sand of one atmosphere. This pressure differential will be potentially greatest for large (high) dunes that are only just covered by water, and values of several atmospheres are not inconceivable if the dunes contain columns of entrapped air some tens of metres thick.

These sorts of pressure differences (from one to perhaps five atmospheres) are considered adequate to allow air to eventually break through the permeability barriers near the tops of dunes. Experimental observation again indicates that the barriers themselves are likely to weaken with time as air percolates out of the
near-surface pore spaces and is replaced by water; the former capillary strength of
these lamellae when moist will then be lost, and the newly water-saturated sands will
no longer be able to withstand the pressure of the underlying entrapped air. Once a
breach was initiated, it would have grown rapidly as large volumes of pressured air
escaped upwards, expanding as they flowed, and were replaced by water from below
and the sides under the influence of a considerable hydrostatic pressure. The chain
of events accompanying the entrapment, and particularly the venting of large air
pockets, could well result in profound deformation of the internal dune structure,
particularly if the dune sands were in a highly metastable condition.

The appeal of this hypothesis is that it readily satisfies the temporal, areal,
stratigraphic and dynamic demands of the Weissliegend. Deformation would have
started in the centres of the Permian basins, and would have proceeded outwards (in
a concentric fashion) concomitant with the flooding. The extension of the deforma-
tion zones to the tops of the dune sandstone sequences would simply reflect that the
tops of the air pockets were lodged just below the dune surfaces, and that their
entrapment and venting were surface-related phenomena. The extreme thicknesses of
some of the deformation zones (over 50 m in some North Sea cores) would be
explained by the occasional entrapment of huge columns of air in large dunes.

Unfortunately, the quality of our field exposures does not allow us to reconstruct
the chain of events leading to the deformation of any particular dune. We can,
however, offer an hypothetical reconstruction for an imaginary Weissliegend dune of
transverse type by piecing together a number of features observed at various
localities, and in core samples. The visual framework for this reconstruction is given
in Fig. 17: the upper diagram (17A) shows the elements of air entrapment; the lower
diagram (17C), the suggested outline of the resultant deformation, and the distribu-
tion of specific deformation structures.

Initially, there may have been some deformational effects which arose from the
unification of separate air pockets, or by air pockets trying to reach higher levels
within the body of the dune. The unification would have occurred through the
breaching of intervening zones of moist laminae; the upward movements would have
required the displacement of overlying pore water, and the replacement of air-filled
pores by underlying pore water. Accompanying, or subsequent to, these events, the
pockets of increasingly compressed air may also have been capable of lifting, or
doming, the overlying sands. This process, too, could have resulted in deformation,
aided by density instabilities set up between the overlying saturated sands and the
underlying air-filled sands (cf. De Boer, 1979).

On venting, the release of pressure would have caused further upsurgence of pore
water into the previously air-filled body of the dune, as well as the inward collapse
of any up-domed cavities. Such movements could have led to widespread failure in
the dune structure triggered (perhaps) by an interdependent combination of lique-
faction, fluidisation (?) and foundering. The internal expression of this failure would
have been in the forms of complexly folded and contorted structures (e.g. Fig. 10),
A PERMEABILITY BARRIER AT THIS BEDDING CONTACT IF DAMP
30°-34° AVALANCHE SLOPE SANDS UNSTABLE IN RISING WATER
ACCRETION BEDDING
ENTRAPPED AIR POCKET

PRE-INGRESSION WATER TABLE

MODIFIED DUNE FLANK WITH PRISM OF MARINE-REWORKED DUNE SAND AVALANCHES/SLIDES

ORIGINAL DUNE PROFILE
RELOVENTY UNMODIFIED PROFILE

PRE-INGRESSION WATER TABLE

TRUNCATION SURFACE OF OLDER DUNE SEQUENCE

K - KUPPERSCHIEFER (COPPER SHALE (DEPOSITED BELOW WAVE BASE AFTER DEFORMATION))
WM - WEISSLEGEND - MARINE REWORKED AND SUBSEQUENTLY DEFORMED DUNE SAND (WHITE, GREY)
WM - WEISSLEGEND - CONTORTED AND DEFORMED DUNE LAMINATE-HOMOGENEOUS SANDS (WHITE, GREY)
WD - WEISSLEGEND - WHITE, GREY SANDS WITH DUNE CROSS-STRATIFICATION
Rh - ROTLEGEND RED SAND, SLIGHTLY DEFORMED/FADED DUNE LAMINATE, PARTLY HOMOGENEOUS
RD - ROTLEGEND RED SAND WITH UNDISTURBED DUNE CROSS-STRATIFICATION
NOTE: THERE IS OFTEN A COLOUR TRANSITION BETWEEN R.B.W.

W = WM + WH + WD, R = RH + RD

SAND VOLCANO / DYKE
LAMINATION
CROSS LAMINATION
WAVY LAMINATION (IN WM)
HOMOGENISED, SWIRLED, INDISTINCT
and the production of homogenised sands (e.g. Fig. 4). Flow or collapse may also have caused brecciation of damp, cohesive sands, and their incorporation as xenolithic fragments within an essentially structureless matrix (see Peacock, 1966, fig. 2).

The small percentage of clay-size particles already within the dune sequence may also have been concentrated by the sieve effect of fine-sand laminae as the particles are carried upward (elutriated) by the rapidly rising ground water that replaced the air. These phenomena might have resulted in the local development of dirty sandstones and anomalous horizons of clay enrichment (e.g. Fig. 4) which otherwise cannot be explained as the products of normal processes of aeolian sedimentation.

The external expression of these events may have included eruptions of homogenised sand over the dune surfaces, impelled, possibly, by mixtures of air and water (sand dykes/volcanoes; e.g. Figs. 11 and 14). These eruptions might also have been accompanied (or preceded) by shallow-marine currents capable of producing ripples and streaming lineations.

As the general activity subsided, late flushes of escaping fluid might have caused further dislocation to the dune surface (e.g. Fig. 15). This, together with the development of subsidence features (Fig. 13), could have occurred in response to the underlying disturbed sands consolidating into a final and tighter packing configuration.

From start to finish, the entire sequence of events accompanying the expulsion of air from any one dune may have taken only a matter of minutes—if that.

**Distribution of deformed and reworked sands related to dune type**

Certain features of the above reconstruction, particularly the internal events, may also be applied to seif dunes; but it is a characteristic of our Weissliegend exposures.

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Fig. 17. Conceptual model to illustrate the effects that the Zechstein transgression may have had on an hypothetical dune of transverse type. A. Principle of air entrapment. Also shown is the original (asymmetric) dune profile and attitude of the avalanche and accretion bedding, together with associated permeability and capillarity trends. B. Inherent instability and reworking of the leeward avalanche slope during inundation by the Zechstein Sea may have resulted in the development of a more symmetrical outline. The prism of marine-reworked sand may contain local beach-like (?) and shoal facies, together with avalanche deposits which may have been derived from upslope by undercutting of the dune avalanche surface sands by surges, lapping and limited wave activity, and/or the collapse, slumping and avalanching of submerged sands rendered unstable within their newly found subaqueous environment. C. Suggested outline of deformation resulting from liquidisation and collapse of the dune sands. The possible locations of specific deformation and other sedimentary structures are also indicated. The colour distribution of the Weissliegend/Rotliegend sandstones is given in the key. Unlike the underlying Rotliegend, the Weissliegend sandstones were never reddened because they were never in an environment conducive to reddening; at the right-hand edge, however, the lower parts of the avalanche slopes may have been buried beneath the water table long enough for the sands to acquire sufficient ferric ions to be stained pink.
that deformation is preferentially developed in dunes of transverse type (cf. Figs. 17C and 18). The explanation for this may be related to bedding styles. For example, transverse dunes contain a relatively simple internal cross-stratification consisting of steeply dipping avalanche deposits (shown schematically in Fig. 17A). As previously mentioned, avalanche deposits are generally undercompacted and metastable, and are thus particularly susceptible to failure. By contrast, seif dunes contain a more complex and interdigitating style of internal cross stratification formed largely by lower angle and more tightly packed bedding deposited from traction (Fig. 18). These beds, with the exception of local avalanche slopes developed in near-crestal positions, are thus more stable. Deformation in seif dunes is therefore far less likely than in dunes of transverse type. When deformation does occur, it should be located only in the near-crestal positions (e.g. Fig. 6, and diagrammatically in Fig. 18).

Similar criteria may be applied to the distribution of sandstones reworked by marine processes. In Fig. 17B, the modification of an original transverse dune outline to a more symmetrical form is achieved by an asymmetrical development of reworked sandstone. This, again, relates to bedding style in that the avalanche slope is believed to be more susceptible to reworking than the more gently dipping and tighter packed accretion bedding of the windward slope. This relationship, however, must remain conceptual. Reworked sandstones that could be ascribed to an origin on the former avalanche slopes were not recognised on the Moray coast. Sandstones of this type are either still preserved beneath the present exposure surfaces, or they became involved in the deformation that affected the adjacent non-reworked avalanche deposits (Fig. 17C), and their temporary marine origins are no longer recognisable.

In the case of seif dunes, the prisms of reworked sandstones are thought to be symmetrically disposed about the upper and middle flanks. This symmetrical development probably formed in response to two controlling factors: (1) that the original dunes were symmetrical; and (2) that the more inclined near-crestal sands were more susceptible to erosion than sands on the lower slopes. This relationship is shown in Fig. 18.

**Stratigraphic implications of the model**

Finally, some of the stratigraphic implications of our model will be discussed. The main topics are: (1) an explanation of the colour distinction between the Rotliegend and Weissliegend sandstones; (2) an assessment of whether this colour distinction can be used as a precise stratigraphic tool; and (3) a comment on the inadvisability of using the term “Weissliegend” in a stratigraphic/formational context.

**Colour**

The classic distinction between the Rotliegend and the Weissliegend has been made on the basis of colour (see Fig. 17C). In its simplest expression, the Rotliegend
Fig. 18. Conceptual model of an hypothetical Weissliegend dune of seif type submerged during the Zechstein transgression. Both deformation structures and prisms of reworked sands are shown in near-crestal positions only. Here, the bedding is composed of avalanche deposits, which are thought to be more susceptible to failure and reworking than the more resistant, low-angle accretion deposits of the lower slopes.
comprises those dune sandstones that were reddened in an oxidising and alkaline environment beneath a slowly rising water table (for the origin of red beds see Walker, 1967, 1976, and Turner, 1980). The reddening took place before the Zechstein transgression. The Weissliegend, on the other hand, comprises those dune sandstones that were above the Lower Permian water table prior to the transgression; and because their pore spaces were essentially air-filled, no reddening took place.

This fundamental colour distinction was also maintained subsequent to the transgression. The rapid rise in floodwater ensured that the pre-ingression water table rose very little before the overlying air-filled sands were saturated with sea water; the acidic and reducing environment that quickly developed on the sea floor (as evidenced by the Kupferschiefer and by the presence of pyrite in the upper Weissliegend sandstones) would have been the very antithesis of that required for reddening. The reworked sandstones, the deformed sandstones, and the dune-bedded sandstones into which the deformation structures grade, thus remained whitish or grey. In Germany, some Weissliegend strata are of fluvial origin (e.g. Pryor, 1971b); like their aeolian equivalents, their lack of reddening can be attributed to deposition above the permanent Rotliegend water table.

**Precise subdivision on the basis of colour**

In an ideal situation the colour distinction would serve well as the basis for a precise subdivision between the two sandstone sequences. The perversions of geology, however, seldom allow such a naive tool to be used, and there are a number of exceptions which undermine its application.

For example, the Weissliegend sandstones are locally orange or pinkish, especially near their base. These colourations may have resulted from temporary periods of submergence beneath the water table. The creation of the pink colour may also have occurred above the water table following heavy dew, by partial wetting by rainwater (perched water tables) or from the local reworking of already reddened sand grains without the complete removal by abrasion of their pellicle of iron-oxide-impregnated clay (Norris, 1969; Glennie, 1970, p. 184; Walker, 1979, p. 72). Furthermore, the main colour boundary between the Rotliegend and Weissliegend is commonly gradational, and probably represents the upper fluctuating limit of ground water-induced reddening prior to Zechstein flooding. The Rotliegend sandstones can also grade into greenish and whitish zones (e.g. in North Sea cores), particularly in close proximity to post-depositional fractures, and this may occur at any stage from the surface of the Rotliegend right down to its basal contact with the Carboniferous (Glennie et al., 1978). And finally, there is the case of the Weissliegend-equivalent sandstones of Morayshire, which are red. This anomaly would be explained if the sandstones never had a permanent protective covering of marine Zechstein strata. Their reddening, together with that of their cover of younger dune sandstones, would presumably have occurred during the Late Permian and Triassic.
Other examples could be added to the list, but the implication is already clear: a simple stratigraphic subdivision on the basis of colour is no longer acceptable. Colour distinction may certainly serve as a guide in some instances, but it will be neither precise, nor infallible.

**Precise subdivision on the basis of sedimentary features**

Perhaps it is safer to ignore colour and to rely on sedimentary features as a basis for subdivision. With reference to Fig. 17C, the Weissliegend sandstones could well be regarded as a complex and hybrid association of dune, deformation, and marine-reworked facies. The Weissliegend could thus be defined on the basis of this association.

Unfortunately this cannot be so as, without the suggestion of where the palaeo-water table lay, as given by colour changes, there is no way of knowing which dune beds lay above the water table and which were below. In Fig. 17C, for example, it is possible to drill through a complete sequence of undeformed non-reddened dune-sand (WD) and the same applies to most of the seif dune in Fig. 18. With the limited data available from wells, it cannot be known if an undeformed sequence of bedding grades laterally into deformation structures. Subdivision purely on the basis of sedimentary features is also not infallible.

**Time/rock relationships (Fig. 19)**

Finally there is the question of whether the term "Weissliegend" should be maintained in a stratigraphic/formational context.

We have demonstrated that the deformation structures were formed as a direct result of the Zechstein transgression. Their creation (from a temporal viewpoint) was an Upper Permian event. These structures, however, occur within the bodies of
Lower Permian dunes and grade into undisturbed dune-bedded sandstones. Their formation was therefore diagenetic (i.e. mechanical diagenesis). On the other hand, the sands reworked from the dune flanks are the erosional products of the transgression, and thus form the basal beds of the Upper Permian Zechstein. The sandstones grouped together under the term “Weissliegend” therefore straddle a major stratigraphic boundary.

In the light of this, we recommend that the term “Weissliegend” be dropped in any formational sense. It is perhaps best retained to denote only a complex and hybrid facies association dominated by: (1) the uppermost Lower Permian Rotliegend dune sandstones (now partly deformed) that originally lay above the water table prior to the Zechstein transgression; together with (2) the sandstones of earliest Upper Permian Zechstein age, which represent the reworking of the dune flanks during that transgression. The colour connotation inherent within the term “Weissliegend” should be retained only as a reminder that probably none of these deformed and reworked sands were red at the time of the Zechstein transgression.

CONCLUSIONS

Some of the most striking results to come from this study are summarised below:

(1) The Weissliegend sandstones that we have studied were originally and entirely deposited as aeolian dunes that lay above the water table just prior to the Zechstein transgression.

(2) The extremely rapid inundation of the dune field by the Zechstein Sea resulted in: (1) very slight reworking of the dune flanks by marine processes; and (2) partial deformation of dune cross stratification within the bodies of the dunes. The reworked sandstones consist of avalanche deposits, shoal sands, and limited beach (?) developments. The deformed dune sandstones contain complexly folded and contorted structures, together with structureless horizons.

(3) The deformation of the dune sands is believed to have been triggered by liquefaction as many of the deformation structures grade laterally and downward into undeformed, cross-stratified dune sandstone.

(4) Liquefaction may have resulted, in part, from rapid fluctuations in the water-table level (the level being driven upwards by the progressive flooding), and/or from the entrapment and expulsion of large quantities of air: the air would have been entrapped between the water table and the inundated and saturated surfaces of the dunes. Of the two mechanisms, air entrapment is favoured.

(5) Deformation by this mechanism would have started nearest the centres of the Permian basins, and would have proceeded outwards in concert with the flooding.

(6) Deformation of the original dune sands seems to have been preferentially developed in dunes of transverse type. The explanation for this is that the avalanche deposits of transverse dunes are often undercompacted and metastable. They are therefore more susceptible to liquefaction and collapse than the generally lower-an-
gle and more tightly packed traction deposits which make up the greater part of seif dunes.

(7) Finally, there is a stratigraphic problem which arises from the hybrid nature of the Weissliegend. It is mostly an Upper Rotliegend (conventionally Early Permian) rock unit, but with marine-rewired sandstones forming the base of the Zechstein (and thus earliest Late Permian). We therefore recommend that the term Weissliegend be dropped in any stratigraphic/formational sense. It should only be retained to denote those dune sandstones of the uppermost Rotliegend (now partly deformed) that lay originally above the water table prior to the Zechstein transgression, together with the minor erosional marine products of that transgression.

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LOWER PERMIAN ROTLIEGEND DESERT SEDIMENTATION IN THE NORTH SEA AREA
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INTRODUCTION

During the past two decades, our knowledge of the distribution and environment of deposition of the Rotliegend has been added to greatly by the drilling of many wells in the North Sea area; first, in the southern North Sea where Rotliegend sandstones form the main reservoir for gas (e.g. Van Veen, 1975), and later in the central North Sea where these sandstones (in addition to Zechstein carbonates) are oil reservoirs in two producing fields, Auk (Brennand and Van Veen, 1975) and Argyll (Pennington, 1975).

The "Rotliegendes" have long been recognised by miners of the Kupferschiefer in Germany and Poland as the red beds that underlie the Late Permian Zechstein sequence of N.W. Europe. The Rotliegend sequences overlie Carboniferous or older strata. The origin of the Rotliegend sediments, and of their equivalents in Britain, as the deposits of an arid continental environment, has been advocated for many decades (e.g. Born, 1921). An analysis of the lithologies and inferred depositional environments of the Rotliegend desert sequence of the southern North Sea was given by Glennie (1972) following extensive studies of modern desert sediments (Glennie, 1970), outcrops of Permian strata, and of cores and wireline logs from wells. Since then, more has become known of the Rotliegend sedimentary rocks, especially in the UK part of the central North Sea (Glennie, 1983), as additional well data is released to the public each year by the UK Department of Energy.

Rotliegend deposition took place in the environment of a Northern Hemisphere Trade Wind desert during the later phases of the major Southern Hemisphere glaciation over Gondwana. I intended to show that there probably was a causal relationship between Rotliegend deposition and polar glaciation, and that, despite probable lowered global air temperatures, the marked aridity of the Rotliegend deserts was the result of winds that either were stronger or, more likely, that blew at sand-transporting velocities for much longer periods of the year than is now the case. If correct, then the late Carboniferous-early Permian peak to the Gondwana glaciation also implies a stronger coeval wind regime outside the polar and equatorial areas of the world.

Rotliegend deposition was brought to an abrupt end early in the Late Permian as the result of the rapid flooding of the continental basins by the marine waters of the Zechstein Sea (see e.g. Smith, 1970, 1979).

THE ROTLIEGEND BASINS

The 'early' Permian Rotliegend sedimentary sequences of the North Sea area were
Fig. 1 - Upper Rotliegend Facies and Palaeowind Map of North Sea Area
deposited in three, essentially east-west trending basins (Fig. 1). The largest of these basins, now known as the Southern Permian Basin (Ziegler, 1978, 1982) extends almost 1500 km from eastern England to about the Polish-Soviet border. The preserved-width of the basin locally approaches 400 km, and sediments accumulated within it to a thickness of over 1500 m (Gill, 1967). In the area of the southern North Sea, this sedimentary sequence has been penetrated by many wells and is now becoming increasingly recognisable at depth on modern high-resolution seismic lines.

The smaller Northern Permian Basin is separated from its southern neighbour by the east-west trending Mid North Sea-Ringkøbing-Fyn system of highs and intervening grabens (Central and Horn Grabens). The basin is now divided into two parts by the NNW-SSE trending northern extension of the Central Graben. Its eastern limit seems to coincide with the southerly extension of the NNE-SSW trending Oslo Graben, whereas little more than 500 km to the west, the basin character is lost in the vicinity of the Forth Approaches. Rotliegend sequences up to 600 m thick have been penetrated by the drill. Rotliegend sandstones exceeding 100 m in thickness, have also been penetrated in several wells up to 150 km northwards into the Viking Graben (Fig. 1).

The relatively small Moray Firth Basin is separated from the Northern Permian Basin by the SW-NE trending Grampian Spur, over which Rotliegend strata are not found. Rotliegend sandstones accumulated to a thickness of 600 m in the centre of the basin. The western margin of the basin coincides with the Great Glen Fault.

Although not strictly Rotliegend, because they are not overlain by Zechstein strata, sedimentary sequences of similar age and facies also occur in a series of half grabens that stretch in an arc from the south of England, through the west Midlands of England, the Cheshire, Manx-Furness and Ulster Basins to the Minch Basins of N.W. Scotland (Smith et al., 1974). A shorter but parallel arc of small basins curves from the Vale of Eden, through southwest Scotland, to the Isle of Arran (Smith et al., 1974; Brookfield, 1978; Lovell, 1983).

The development of these half grabens seems to be matched by similar depressions in S.W. Germany and eastern France, and thus reflects the early collapse of the Variscan Highlands and its northern foreland under an E-W tensional regime.

Sedimentary sequences of desert origin that extend back into the Pennsylvanian have been described from the west-central United States (see McKee, 1979a). Making allowance for the later opening of the Atlantic, these American sequences probably formed part of the same Northern Hemisphere belt of Trade Wind deserts as the Rotliegend of N.W. Europe (see Fig. 4).

**ROTLIEGEND FACIES AND DEPOSITIONAL ENVIRONMENTS**

The Rotliegend is divided into two distinct rock-stratigraphic units, the Lower and Upper Rotliegend, on the presence or absence of volcanic rocks associated with the sedimentary sequence. Although the Lower Rotliegend is generally overlain by the Upper, the latter may locally be coeval with younger parts of the Lower Rotliegend in areas where there was no Early Permian volcanic activity (Fig. 2).
a) Lower Rotliegend

The range in composition of the basic to intermediate volcanics of the Lower Rotliegend, and their distribution adjacent to known or inferred faults, suggests that their origin was probably related to the earliest tensional movements connected not only with the Permian basins, but also with the North Sea graben systems. Katzung (1975) considered that these faults were initiated during the Stephanian-Autunian.

The Lower Rotliegend is best developed in northern Germany, where it forms a continuous sequence up to 2000 m thick (Plein, 1978) that is inferred to range in age from the Stephanian into the Autunian, and in the area of the Oslo Graben-Bamble Trough. Similar rock associations of about the same age, but covering generally smaller areas, are known in France, S.W. England, S.W. Scotland, and in some of the flank areas of the Mid North Sea-Ringkøbing-Fyn Highs (see Ziegler, 1982). No volcanic activity of Saxonian age is known on continental Europe (Falke, 1976).
b) Upper Rotliegend

In the Southern Permian Basin, the Upper Rotliegend is made up of four distinctive facies associations which have been interpreted as the products of sedimentation in fluvial (wadi) aeolian, sabkha and lacustrine (desert lake) environments; the latter includes deposits of bedded halite. Their main sedimentary characters as recognised in cores are illustrated by Glennie (1972) and Plein (1978). These facies are distributed from south to north across the basin in essentially the same order as given above, with the desert-lake facies coinciding with the main axis of subsidence. Only a narrow strip of rarely drilled Rotliegend occupies the area between the lake sediments and the area of the highs to their north. On present evidence, the same sedimentary facies (wadi, aeolian, sabkha, lacustrine) also occur in the Northern Permian and the Moray Firth basins, although the presence of a bedded lacustrine halite has yet to be demonstrated. It should be realised, however, that the Permian geology of these two basins is still incompletely known; the Rotliegend in the deeper parts of the basins is rarely reached by the drill and, except on the Moray coast near Elgin (Fig. 1), it is not seen in outcrop.

i) The fluvial sequences of the Southern Permian Basin comprise locally conglomeratic sandstones that are characterised by the occurrence of curled clay flakes, denoting subaerial desiccation. Conglomerates form a greater part of the sequence in The Netherlands and in Germany than in British waters, and in the former two lands they are generally rich in volcanic clasts. The fluvial sandstones commonly alternate with well laminated, horizontally to relatively steeply inclined (20-30°) sandstones that are interpreted to be aeolian, and with structureless sandstones, which are thought to have been homogenised by liquefaction of previously deposited sands during local fluvial flooding (cf. Selley, 1969). Thus, together, these beds indicate the ephemeral nature of the stream flow with intervening periods of sub-aerial exposure, desiccation and wind activity. Some thicker mud-cracked clays seem to have had their cracks infilled with sand from above, whereas others were recognisably injected from below by a slurry of sand and water to form sandstone dykes (Glennie, 1970; 1972).

The fluvial sequences are more common along the southern margin of the basin and in the lower part of the Rotliegend; many smaller wadi channels probably did not extend as far as the desert lake, but terminated in small interdune sabkhas. Intermittently along the southern margin of the basin, major wadi channels cut through the belt of dune sands, but even here, the fluvial sands commonly alternate with those of aeolian origin (Fig. 1) thus emphasising the sporadic nature of rainfall in the area. Ellenberg et al. (1976) describe local deltaic sedimentation along the southern margin of the desert lake in E. Germany, and in Poland, Pokorski (1978) described widespread braided river channels and ephemeral fluvial activity.
The character of wireline logs indicates that fluvial sands also occur in some parts of the Northern Permian and Moray Firth basins. The main directions of fluvial transport have not yet been determined, however.

ii) The aeolian sequences penetrated in North Sea wells generally comprise sets of moderately to well-laminated sandstones that are horizontally bedded at the base, and upwards become inclined at angles that increase to a maximum of 25° or 30°. The set is then generally overlain by horizontally bedded sandstones forming the base of the succeeding set (see cores in Glennie, 1972; 1983). Locally, low-angle sandstones may be overlain by the steeply-dipping laminae of a sequence of prograding sandstones.

The well laminated steeply-dipping sandstones are interpreted to have resulted from deposition by grainfall (Hunter, 1977), and the decimetre-thick, poorly to non-laminated sets from sandflow. Both types are common in North Sea wells. Ripples are rarely well displayed. In sub-horizontal bedding, however, slight changes in lamina thickness within the width of a core (10 cm) possibly indicate an origin by ripple migration in an interdune area as described by Kocurek (1981). Irregular adhesion-ripple laminae have been described by Glennie (1972, Fig. 14). They are possibly more common in areas where the rate of subsidence matched that of deposition, as locally seems to have been the case in the Sole Pit Basin (Glennie, Mudd and Nagtegaal, 1978).

The sets of laminae seen in North Sea cores between deflation surfaces generally range between about one and seven metres in thickness. Locally, however, sets have been measured that comprise up to 20 m or so of continuous bedding. Since most of the thinner bedding sets probably represent only the preserved lowermost portions of migrating dunes, the very thick sets probably represent an unusual event in which most of the lee-slope bedding of a dune is preserved. This could happen in a sand sea, for instance, where a temporary wadi channel becomes permanently infilled by a single phase of migrating dune. It could also be achieved by the infill of interdune hollows in a growing sand sea such as those illustrated by Glennie (1970, Fig. 72) from the Al Liwa, eastern Arabia. For example, the Rotliegend sand sea reached a total thickness of some 300 m in the Sole Pit Basin (Glennie and Boegner, 1981). Thus, although we have no idea of how much of the upper part of such large dunes was removed during later dune migration, the 20 m thick sets of bedding can be taken as a crude approximation to the minimum height of the dune on which it was formed. In this context, Glennie and Buller (1983) have presented other evidence suggesting that many Rotliegend transverse dunes of the Southern Permian Basin were probably over 50 m high at the time of the Zechstein transgression.

The bedding attitudes of the dune sands measured on cores and deduced from dip-meter logs of wells in the Southern Permian Basin indicate that the Mid Permian wind blew over the area from essentially east to west (Glennie, 1972; 1983; Ellenberg et al., 1976; Van Wijhe et al., 1980). At least in the western half of the Northern
Permian Basin, however, the winds blew from the northwest (Fig. 1). Thus, a centre of barometric high pressure must have been located over the Mid North Sea High (Glennie, in press). A similar centre of high pressure to the west of Wyoming and North Colorado has been proposed by Fryberger (1979) to account for southerly-directed foresets found in the Pennsylvanian-Permian Weber Sandstone.

In the Rotliegend sandstones of many North Sea wells, the relatively narrow spread of bedding attitudes indicates that these sands were deposited largely on dunes of the transverse type (Fig. 3). In N.E. England (D in Fig. 1), however, wind strengths seem to have been too strong for transverse dunes to be stable, and longitudinal (seif) dunes were formed. Because of their almost complete preservation beneath the marine Marl Slate (Kupferschiefer equivalent) at the base of the succeeding Zechstein sequence, much of their original shapes can be followed in a series of large quarries (sand pits), mine shafts and borings. These dunes are aligned along an ENE-WSW axis (N60°E) and have been preserved to a height of up to 60 m above the Carboniferous surface (Smith and Francis, 1967, Fig. 18 and PL 15; Glennie and Buller, 1983). In the interdune areas, the Marl Slate generally rests directly on Carboniferous strata with no Rotliegend sandstone intervening. The thickness of sand visibly reworked from the dunes by the transgressing Zechstein Sea ranges between some 10 cm in what are believed to be former dune-crest locations, to a maximum of around 4 m on the dune flanks (Glennie and Buller, 1983). Marine faunas with Zechstein affinities occur in the 2-3 cm immediately below the Marl Slate (Bell et al., 1979).

![Polar Nets](image)

**Fig. 3** - The poles of dune bedding (derived from dipmeter logs) plotted on Polar Nets (upper hemisphere) for three different localities in the Southern Permian Basin. The distribution of bedding attitudes suggests the type of dune of which they form a part, and arrows give deduced wind directions:

**A** Limited distribution - simple transverse dune

**B** Most dips concentrated in areas (b) on either side of the dune axis - seif dune

**C** Most bedding attitudes concentrated in a general down-wind location on the net but with some dips almost at right angles to the deduced wind direction, and so suggestive of limited transverse instability and the possible presence of barchan-like horns - barchanoid dune.

Note: Bedding attitudes in A originally published by Van Wijhe et al. (1980) as a rose diagram.
This limited amount of reworking is in marked contrast to the much greater thickness of reworked sandstone locally present in the Triassic-Jurassic Entrada Sandstone of New Mexico (Tanner, 1970). There, dunes, with an original relief of over 45 m, were truncated and had their interdune areas infilled with water-laid sandstones some 30 m thick in the aquatic environment of 'Lake Todilto'.

At outcrop, the seif dunes of Durham are not red but yellow in colour, from which they derive their name of 'Yellow Sands'. In North Sea wells, in Germany and in Poland, however, the uppermost sands of the Rotliegend sequence are commonly grey or white for up to some 50 m beneath the Kupferschiefer, and are referred to as the Weissliegend. The origin of these sands has been the subject of dispute over the decades, with both aeolian and marine derivations being attributed to them (see references in Pryor, 1971a, b; and Nemec and Porebski, 1977). In both the southern North Sea and in Poland, the Weissliegend dune bedding grades into large-scale soft-sediment deformation structures. Similar structures occur in the middle of the Hopeman Sandstones (Peacock, 1966) on the southern coast of the Moray Firth near Elgin (Fig. 1). Here, a zone of deformation divides the dune sequence of the area into two units: a lower unit of southward-dipping dune sands for which correlation with the Rotliegend is suggested, and an upper unit of southwest-dipping dune sands of probable Late Permian or Early Triassic age (Walker, 1973). The creation of the deformation at the top of the lower dune unit, and also of similar structures in the Weissliegend of the North Sea and Poland, are attributed to the effects of the Zechstein transgression (Glennie and Buller, 1983). This event will be referred to again.

iii) The Lacustrine Sequence in the Southern Permian Basin (Fig. 1 and 2) consists primarily of red-brown mudstone with minor siltstone. Several horizons of halite, with a cumulative thickness in excess of 100 m (Plein, 1978), are developed over much of the axial part of the basin. The halite is concentrated in the middle part of the Upper Rotliegend sequence in the axial part of the basin in Germany, but wedges out towards the basin margins so that in UK waters, for example, it is limited to the lower part of the section. In northern Germany, where the lacustrine sequence reaches a thickness of around 1500 m, the halites attain a sufficient thickness to react diapirically; the sequence is then referred to as the Haselgebirge facies. Carbonate and both bedded and nodular anhydrite are absent from the lacustrine facies.

The lacustrine strata seem to be devoid of fossils except in the metre or so beneath the Kupferschiefer (Plumhoff, 1966). As these fossils have strong Zechstein affinities, they should, perhaps more correctly, also be attributed to the Zechstein marine transgression (Falke, 1976).

The lacustrine environment of deposition of this sequence was deduced from the lack of fossils, from the absence of carbonates, and from the non-marine composition of the evaporites (Glennie, 1972). This interpretation is supported by the low
bromide content of the Rotliegend halite and by the sulphur-isotope analysis of the associated anhydrites (Holser, 1979). Holser (1979) suggests that because of their relatively great volume, these salts must have been derived from older deposits of marine origin, and suggests exposed Devonian and Early Permian halites of European USSR as possible sources. The writer's experience in modern deserts convinces him, however, that the required volume of halite can be concentrated and precipitated from lake water supplied by a normal system of seasonally (?) flowing rivers, provided the rate of evaporation is great enough and of long duration. As will be shown later, both these factors were probably operative in the early Permian Rotliegend desert.

There are no known sedimentary structures within the lacustrine sequence that are indicative of sub-aerial desiccation or erosion, although such structures are known in the anhydritic mudstones of the sabkha facies, which encircles this desert lake. The lake is therefore believed to have been a permanent feature of the Southern Permian Basin throughout the earlier Permian. At its fullest extent, the lake must have covered an area of some 1000 km east-west, by up to 200 km north-south.

The evidence of fluvial transport directions in the Rotliegend sequences of the Southern Permian Basin indicates that most of the water supplied to the desert lake was derived from the south, from the Variscan Mountains, with only minor quantities from other sources. Fluvial activity seems to have been more important in the east than in the west. The sources of water for the northern two basins are unknown.

iv) As we have seen, a sequence of Sabkha deposits occupied a broad area encircling the desert lake in the Southern Permian Basin, and deposits of this type are found in smaller areas of local subsidence in the Northern Permian and Moray Firth Basins (Deegan and Scull, 1975; Smith, 1976, Fig. 1). Using tetrapod footprints for dating, Haubold and Katzung (1978) indicate that playa and sabkha sediments already existed in Germany during the Late Autunian.

These sedimentary rocks comprise poorly-bedded clays with minor silts and sands that display many features indicative of a sabkha (e.g. desiccation cracks, sandstone dykes, adhesion ripples, anhydrite nodules). These features collectively indicate an aquatic depositional area that was subject to extensive subaerial desiccation in an arid environment (Glennie, 1970). Thus the sabkha sediments represent the area that was covered by water only during the maximum extensions of the desert lake. Although there probably were annual fluctuations in the extent of the lake and associated sabkha, this is difficult to illustrate with a limited number of well borings. Fluctuations on a much longer time scale are clearly apparent, however, and are indicated schematically on the rock-stratigraphic facies diagram (Fig. 2), and in the well-log correlations illustrated by Adrichem Boogaert (1976) for the Netherlands. Another important point that these figures illustrate is that halite precipitation in the Southern Permian Basin seems to have been coeval with the maximum development of the aeolian facies. This point will be referred to again.
Fig. 4 – Palaeo-relationships of Rotliegend and North American Deserts. Plate reconstruction after Scotese et al. (1979): Late Carboniferous.

HISTORICAL DEVELOPMENT OF ROTLIEGEND CLIMATE

The change from the humid equatorial conditions under which the Carboniferous Coal Measures were deposited, to the arid climate of Rotliegend deposition, was probably mostly a result of the northward drift of Laurasia. The Rotliegend basins came to occupy a latitudinal position north of the equator (Fig. 4) similar to that of the present North African-Arabian deserts. (During the past decade, the published palaeomagnetic location of NW Europe has varied by some of 10° of latitude - see e.g. different latitudes used by Glennie, 1972 and 1983; the latitudes illustrated by Scotese et al., 1979, are adopted for this paper, and are similar to those used by Glennie, 1972).

During the Late Carboniferous, the newly created Variscan Highlands in western Europe occupied an equatorial climatic niche and, to judge from coals of Stephanian and Autunian ages in Saarland and Bohemia, had a fairly high rainfall. A time-related reduction in the volume of Rotliegend fluvial sediments in the Southern
Permian Basin, and an associated spread of dune sediments, may indicate that the main sources of river water in the Variscan Mountains were also moving away from the equatorial rain belt.

These climatic changes seem to have been heralded in Britain and north Germany by a northward migrating alteration with time from grey coal measures during the earlier Westphalian (A + B) to a "Barren" red-bed facies of upper Westphalian C to Stephanian age.

Over much of the area, there was a major hiatus between Carboniferous and Upper Rotliegend strata (see e.g. Van Wijhe et al., 1980; Fig. 10), so that the detailed history of climatic change during that time span has to be deduced from areas beyond the desert where sedimentation was continuous. In the southern Pyrenees, for example, Nagtegaal (1969) has shown that there was a progressive change from a humid Westphalian D climate to one that was semi-arid (alluvial fan/steppe) at about the Autunian-Saxonian time boundary. A similar climatic change has been deduced from the late Carboniferous and early Permian strata of France (e.g. Feys, 1976) and Bohemia (Holub, 1976). Holub, however, stresses that superimposed on this gradual increase in aridity were alternations of more humid and more arid periods, with culminations of aridity during the Saxonian (U. Rotliegend) and the Thuringian (Zechstein).

These two phases of increased aridity seem to be matched in the Southern Permian Basin by the two major periods of evaporite precipitation - the evaporites of the Rotliegend desert lake and of the later Zechstein Sea; and as may be expected, the earlier of the two seems to coincide with maximum dune activity. The depositional history of the later Rotliegend is one of extension of the desert lake and sabkha facies at the expense of the dune sands (Fig. 2).

Deposition of the continental Rotliegend desert sequence was brought abruptly to an end by the Zechstein transgression. This transgression seems to have been a very swift event. It flooded Rotliegend basins whose surfaces may have been some 250 to 300 m below sea level (Smith, 1970, 1979; Ziegler, 1982; Glennie and Buller, 1983). A global rise in sea level seems to have caused approximately simultaneous marine transgressions over much of the Russian platform, many of the Arctic islands, central-east Greenland and central North America, as well as large parts of western Europe (Smith, 1964). This eustatic rise was probably associated with the decay of the Gondwana ice cap, and seems to have been great enough to breach what was possibly no more than a valley-bottom sedimentary barrier somewhere between NE Greenland and the North Cape of Norway. Once breached, water will have poured south along the proto-Atlantic rift (Russel, 1976; Ziegler, 1978), which also must have been below sea level along its entire length. The ensuing flooding of the Rotliegend basins is inferred to have been fast enough to have caused widespread liquefaction within the Rotliegend dune sands with associated soft-sediment deformation. Deformation seems to have been confined largely to dunes of transverse type, which have a high percentage of under-compacted avalanche-slope sands. The seif dunes of the Durham area, on the
other hand, with their lower-angle style of cross bedding and inherently tighter grain packing, seem largely to have resisted liquefaction (Glennie and Buller, 1983).

IMPACT OF GONDWANA GLACIATION ON ROTLIEGEND WINDS

Because of an associated eustatic rise in sea level, the demise of the Gondwana glaciations seems to have resulted in the end of continental desert conditions over many low-lying areas of N.W. Europe. It is suggested here, that the periods of more intense aridity in the Rotliegend desert areas were caused by stronger global wind systems, which in turn, were driven by polar areas of glacially-induced high barometric pressure over Gondwana. This general premise, which is illustrated conceptually in Fig. 5, has been advocated by many workers (e.g. Lamb, 1961; Glennie, 1970; Manabe and Hahn, 1977) and is supported by the evidence from Pleistocene deserts in

Fig. 5 - Conceptual Differences in Width, Location of the Earth's Air-pressure Belts, and the Intensity of the Associated Wind Systems in Relation to the Size of the Polar Ice Caps.
Australia (Galloway, 1965; Williams, 1973; Bowler, 1976), South Africa (Lancaster, 1981), the Persian Gulf (Sarnthein, 1972, 1978) and offshore North-West Africa (Sarnthein et al., 1981).

Several pertinent analogies can be made concerning wind activity in the Permian palaeodesert of Northwest Europe and the Pleistocene-Recent deserts of the world.

In arid deserts, there seem to be two simple basic types of dune; the transverse dune (or barchan where there is a shortage of sediment) which is formed by winds of moderate strength and has its axis transverse to the dominant sand-transporting wind; and the linear or seif dune, which forms in winds that are too strong for transverse stability and has its axis parallel to the dominant sand-transporting wind (Glennie, 1970).

The writer believes that the bulk of the existing major dunes of North and South Africa, Arabia, India and Australia originated with such basically simple forms during the last (Weichsel/Wisconsin) glaciation. Many of these giant dunes have heights that exceed 100 m and a width of 1 km or more. Since then, however, wind intensities have decreased, and the maximum size of seif dunes built today seem to have an average height of little more than 5 or 10 m. Perhaps more important in terms of the building and preservation of Pleistocene dune forms, is that at the time of their formation, winds possibly blew with sand-transporting velocities for much of the glacial winter; according to Wilson (1971, p. 190; using data from Dubief, 1952) modern sand-transporting winds in an area of Algeria blow for a cumulative total of only some 52 hours per year (mean deviation 39 hours). Bagnold (1941, p. 69) states that sand flow is proportional to the cube of the excess velocity above that at which sand begins to move. This fact lays great stress on the transporting power of strong winds; coupled with the probable much greater annual duration of strong winds during high latitude glaciations, the annual amount of sediment moved must then have been immeasurably greater than today.

The giant Pleistocene dunes include a high proportion of linear forms. In the present interglacial conditions the winds are generally weaker and the surfaces of the Pleistocene dunes are now commonly covered by much smaller parasitic dune forms whose axes are transverse to the prevailing winds (see e.g. Cooke and Warren, 1973, Pl 4.1, 4.2). In many places, the direction of today's winds is not in equilibrium with the former giants across which they blow, and the latter suffer a greater or lesser degree of morphological change. As a result, the axes of younger dunes may be seen to cut across those of the older dune systems (see Breed et al., 1979, Fig. 202, for example). Where the old and new wind directions are similar, erosion of the older system of giants is less obvious, but the small size of the resulting new dunes is plain to see (e.g. Glennie, 1970, Fig. 74).

In the Rotliegend, the bulk of the dunes seem to be of the transverse type, with some suggestion of minor transverse instability. Seif dunes are clearly present only at the western basin margin in County Durham (Fig. 1), where wind strengths presum-
Fig. 6 - Polar Nets of Rotliegend Dune-bedding Attitudes derived from Dipmeter Logs, and Changes in the Deduced Palaeowind Directions with Depth. UK North Sea Wells 49/26-2 (Southern North Sea) and 9/28-2 (Viking Graben).
ably were greater than in the basin centre. Seifs may also be present locally in the Northern Permian Basin (Glennie, 1983). The distribution of their bedding attitudes in comparison with that of an inferred transverse dune is illustrated in Fig. 3.

The study of Pleistocene-Recent dunes shows that there have been changes of both wind strength and wind direction with time. Similar variations with time can be seen from a study of the bedding attitudes in Rotliegend dune sands as derived from dipmeter logs (Fig. 6). The derivation of the palaeowind directions is subjective rather than strictly statistical, so that some of the inferred directions could be questioned, although arguments in their support have been given at some length by Glennie (1983). There can be little doubt about the general wind direction at most well localities, however.

In both the illustrated wells, the dunes are probably of the transverse type. Especially in well 49/26-2, the individual depth sequences, except perhaps for the deepest, seem to conform closely to the limited spread of bedding attitudes typical of transverse dunes. Because of slight changes in wind direction with time, the added spread in dip attitudes seen in the combined data for the total Rotliegend dune sequence gives a false impression of deposition on a barchanoid dune, with its element of transverse instability.

Simple transverse dunes seem to be less likely in the case of well 9/28-2, especially at the deeper levels, which have a more barchanoid distribution of bedding attitudes. Here too, however, it may be a case of sampling at too coarse an interval. What is more remarkable, perhaps, is the relatively small range of palaeowind directions inferred for a time span of deposition that runs into millions of years. These observations may reflect a situation involving a stronger glacial control of desert winds earlier in the Permian, and a weakening wind system as the Gondwana ice cap declined. Similarly, the winds of the last Pleistocene glaciation probably had much less variability in direction than those of today. Certainly, the depositional results of weaker Permian interglacial winds are likely to have been largely removed by the stronger winds related to each succeeding glaciation over Gondwana.

A weakening wind system is likely to be the reason why, with advancing time, the area of the Rotliegend desert lake increased at the expense of the dune sands. The intensity of evaporation that gave rise to halite precipitation earlier during Rotliegend deposition is thought to have been more closely related to dry high-velocity trade winds than to high temperatures. Indeed, evidence from the Pleistocene indicates that, although the snowline in Ethiopia was lower by some 600-1000 m during the last glaciation (Adamson et al., 1980), the level of Lake Victoria was also much lower than now, and in the Sudan, linear dunes were active prior to late Glacial flooding of the Nile valley (Williams, 1975). Similarly, Galloway (1965) reports that, although Australia was considerably colder during the Late Pleistocene than now, it was also more arid with associated strong winds capable of building seif dunes.
The oldest known glaciation in Gondwana is found in Early Carboniferous strata in the Andean belt of Argentina; the youngest (icerafted) glacial deposits were recorded in Tasmania and the Sydney Basin of Australia and are of mid-Permian age. Between these extremes, glacial conditions continued intermittently for some 80 million years, with at least 12 advances and retreats being recognised in the Paraná Basin of Peru (Martin, 1981).

The Gondwana glaciation seems to have been at its maximum extent at about the Carboniferous-Permian time boundary (Stephanian- Early Autunian). In Western Europe, local intermontane coals of these ages testify to the existence of rainfall within the Variscan Highlands, but on their northern foreland there were increasing signs of aridity. Dune sands associated with earliest Permian volcanics occur both in S.W. England (near Exeter) and S. W. Scotland (Ayrshire). At about this time in North America, the era of Pennsylvanian coal deposition was drawing to a close in the Appalachians, but farther west the Weber (Fryberger, 1979) and de Chelly (Weber, 1979) dune sands were being deposited. In both formations, the dunes seem to be of the transform type, Weber (1979) making a comparison between the bedding styles of the de Chelly and the Rotliegend dunes of the North Sea gas fields. Thus, although wind strengths were not extreme, dune activity coincided with the time of a major high latitude ice cap. Had Permian ice caps developed coevally in both hemispheres, as in the Pleistocene, Rotliegend wind strengths may well have been greater.

The major hiatus between the Carboniferous and the Upper Rotliegend of the Southern Permian Basin contains an anomaly. A greater volume of rock seems to have been eroded at this time than can be accounted for in the Rotliegend sequence. Inversion along the Sole Pit axis of the Southern North Sea (Glennie and Boegner, 1981) resulted locally in erosion of the total Westphalian sequence, which probably had a depositional thickness approaching 1000 m (Fig. 7), before deposition of the succeeding Rotliegend sequence. Also, prior to Zechstein deposition, early Permian erosion over the uplifted Mid North Sea High penetrated through the former Carboniferous cover deep into the Devonian. This anomaly can be explained if large-scale removal of the finer fraction by deflation is accepted as plausible, the products of deflation being deposited elsewhere in the oceans or on land as loess. On a similar large scale, deflation of much of the Tertiary products of erosion from the west side of the Oman Mountains in Arabia has been proposed by Glennie et al. (1974).

In the context of deflation, the time of origin of the Rotliegend desert lake is not known. Its position overlying sequences of Westphalian C to D (Fig. 7) and even Stephanian strata (Van Wijhe et al., 1980) suggests that in these areas, deposition may have been locally continuous across the Carboniferous-Permian time boundary. If correct, then the lake itself, or moisture from near-surface groundwater, will have prevented further erosion by deflation; to the contrary, these wet areas may well have trapped wind-blown clay and silt particles to form some of the earliest sediments of the desert lake.
Because of the equatorial location of their depositional area, the Westphalian Coal Measures of N.W. Europe were probably largely unaffected by the Gondwana glaciations except in terms of a relatively cool climate and occasional glacio-eustatic rises in sea level to give the distinctive marine marker horizons found in the Coal Measures. As soon as this area began to drift into the belt of Trade Winds, however, increased aridity coupled with strong winds resulted in severe deflation, which was more marked in the north than to the south.

CONCLUSIONS

The Rotliegend sedimentary sequence was deposited in several continental basins of a northern hemisphere tropical desert, the largest of which were the Southern and Northern Permian basins. Desert deposition succeeded an early Permian period that involved considerable erosion of pre-Permian strata, which was possibly caused in part by strong deflation. This northern hemisphere wind activity seems to have coincided with the early waning stages of a major polar ice cap over Gondwana, which, like the Pleistocene glaciations, may have affected the patterns of temperature and barometric pressure on a global scale.

The change from a humid Carboniferous climate to the arid conditions of the Rotliegend desert was probably largely the result of the northward drift of N.W. Europe from the vicinity of the Late Palaeozoic equator into the zone of prevailing northeast Trade Winds; thus the Rotliegend desert would also be within the rain shadow of the Variscan-Ural system of mountains.

The axial part of the Southern Permian Basin was occupied by a major desert lake whose waters were fed especially from the eastern Variscan Mountains. The lake contained bedded halite in the middle part of its sedimentary sequence. Halite precipitation seems to have coincided with the maximum extension of the belt of dune sands to its south, thereby emphasizing the relationship between aridity and wind activity in this desert area.
A relatively strong glacially-induced wind system is thought to have resulted in a fairly constant anticyclonic pattern of mid-Permian winds over the present North Sea area; essentially from the east over the Southern Permian Basin and from the west over the Northern Permian Basin. At any one locality, the direction of the mid-Permian palaeowinds shifted slightly with time, but the overall direction seems to have been remarkably constant. These minor changes are thought to reflect differences in the global wind pattern, which probably became less uniform and less intense as the Gondwana ice cap dwindled, and resulted in much lower aeolian transport rates in the Rotliegend desert.

A glacio-eustatic rise in sea level is thought to have resulted in the Zechstein marine transgression. At that time, the surface of the Rotliegend desert lake is estimated to have been some 250 m below sea level and, indeed, the entire route along which the Zechstein marine water must have flowed from the vicinity of the present Arctic Ocean must also have been below sea level. The Rotliegend basins were flooded so rapidly that there was no extensive marine reworking of the exposed unconsolidated Rotliegend dune sands, and much of their former sub-aerial relief is now preserved beneath the draping basal Kupferschiefer. Internally, however, especially with transverse dune forms that were exposed above the water table, the former avalanche bedding was subjected to widespread homogenisation and to large-scale soft-sediment deformation as the dunes were inundated beneath the rising Zechstein Sea.

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Chapter 3

Early Permian—Rotliegend

K.W. GLENNIE

3.1 Introduction

The ‘Rotliegendes’* is an old German miner’s term for the red beds that underlie the Zechstein. The classical Rotliegend sedimentary sequence was deposited in a post-Variscan basin that extended some 1500 km from eastern England to the Russo-Polish border and has been referred to as the ‘Southern Permian Basin’ (Figs. 3.1, 3.2). Seismic surveys and offshore drilling have shown that another, much smaller, Rotliegend basin occurs between the fragmented Mid North Sea—Ringkøbing Fyn High and the Shetland and Egersund platforms, and is known as the ‘Northern Permian Basin’. A third area of Rotliegend deposition is limited to the Moray Firth Basin. Of the same approximate age of creation is a series of small half grabens, which stretch from SW England to SW and W Scotland; their fill of Permian sediment has been correlated with North Sea sequences by Smith et al. (1974), and Lovell (1983); see also Smith (1972).

In the Southern Permian Basin, the Rotliegend can be divided into two distinct units, the Upper and the Lower Rotliegend. The Lower Rotliegend is characterised by the presence of volcanic rocks, which are not found in the Upper Rotliegend; the two units are possibly partly coeval (Fig. 3.4).

Sandstones of the Upper Rotliegend form a most important reservoir rock for gas in the Southern Permian Basin. They contain some \(4.1 \times 10^{12} \text{ m}^3 (145 \times 10^{12} \text{ ft}^3)\) of proven recoverable reserves, of which \(1 \times 10^{12} \text{ m}^3 (35 \times 10^{12} \text{ ft}^3)\) are in offshore fields of the Southern North Sea (Ziegler, 1980a) and \(2.4 \times 10^{12} \text{ m}^3 (86 \times 10^{12} \text{ ft}^3)\) are in the giant Groningen gas field in The Netherlands (Fig. 3.12). The source for all this gas is the Coal Measures of the underlying Carboniferous, which, depending on the temperature gradient, gave up its gas when buried at depths of between 4000 and 6000 m (Lutz et al., 1975; Van Wijhe et al., 1980). The seal is provided by the overlying Zechstein sequence. The Zechstein cycle II (Stassfurt) halite (see the following chapter by Taylor) is the most important individual seal because it is regionally thick and is able to flow and thus heal any fault-induced fracture.

In the Northern Permian Basin, Rotliegend sandstones are oil bearing in both the Auk and Argyll fields, the source rock being the Upper Jurassic Kimmeridge Clay, which matured deep in the adjacent Central Graben.

The colour of the Rotliegend sedimentary rocks resulted from post-depositional diagenetic reddening

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* Rotliegendes: I am informed by my more linguistic colleagues that the term ‘Rotliegend’ is both simpler and more correct in English.
when ferrous ions in the ground water were oxidised to the ferric state. As shown by Walker (1967), a diagenetic environment conducive to such oxidation is commonly present beneath the surfaces of tropical deserts. The palaeogeographic, climatic and diagenetic significance of red beds formed in desert and other depositional environments is discussed in considerable detail by Turner (1980).

The uppermost part of the Rotliegend sandstone sequence is commonly grey or white in colour, which has given rise to the German name ‘Weissliegend’. It is thought that the Weissliegend sands were above the water table at the time of the mid Permian Zechstein marine transgression and thus were never in a diagenetic environment in which they could become reddened (Glennie and Buller, 1983).

### 3.2. Lower Rotliegend

The Lower Rotliegend comprises an association of rocks that is predominantly volcanic in character but includes some sedimentary sequences, especially in Germany (Falke, 1972; Plein, 1978), that were deposited largely in fluvial and lacustrine environments under a climate that alternated between humid and arid. Aeolian sandstones also occur, and locally have good reservoir potential (Schneverdingen Sandstone; Drong et al., 1982). The distribution of the Lower Rotliegend is very limited in comparison to the Upper Rotliegend (cf. Figs. 3.1, 3.3). The Lower Rotliegend volcanics are best developed in northern Germany and the Oslo Graben-Bamble Trough areas. Similar associations of sediments and volcanic rocks of about the same age (Late Stephanian-Early Autunian) but covering much smaller areas, are known in France, S.W. England, S.W. Scotland and some of the flank areas of the Mid North Sea-Ringkøbing-Fyn Highs.

The range from basic to intermediate volcanic rocks in the Lower Rotliegend, and their distribution adjacent to known or inferred faults, suggests that their origin was possibly related to the earliest tectonic movements connected with the creation not only of the Permian basins, but also with the graben systems of the North Sea (Viking-Central grabens, Oslo-Horn grabens) and Germany (e.g. Schneverdingen Graben).

If this interpretation is correct, then the North Sea graben system could have started to form during or shortly after the final stages of the Variscan Orogeny; this, in turn, coincides with a Late Westphalian age for uplift of the London-Brabant Platform and a Stephanian date for the earliest positive movements of the Mid North Sea High.

Some idea of the scale of Latest Carboniferous to Early Permian differential vertical movements can be gained when one considers that, prior to the Zechstein transgression, erosion removed all the previously deposited Carboniferous strata over large parts of the Mid North Sea High; and adjacent to the Central Graben, Zechstein erosion additionally cut deep into the Devonian Old Red Sandstone and even older strata, thus implying even greater uplift. Within the Southern Permian Basin, Late Carboniferous to Early Permian inversion along the Sole Pit axis (Fig. 2.13) resulted in erosion, locally, of the complete Westphalian sequence prior to deposition of the Upper Rotliegend.

### 3.3 Upper Rotliegend

#### 3.3.1 Southern Permian Basin

The Upper Rotliegend is made up of four distinctive facies associations, which have been interpreted as the products of deposition in fluvial (wadi), aeolian, sabkha and lacustrine environments (Figs. 3.4, 3.6), cores of three of which have been illustrated by Glennie (1972). All four facies are widespread in the asymmetric Southern Permian Basin, whose floor sloped from south to north over much of its area (see Fig. 3.11). In the Southern North Sea area, the sandy facies is referred to as the Leman Sandstone in U.K. waters and

![Fig. 3.2. Upper Rotliegend isopachs and isolated thicknesses. Modified from Ziegler (1980).](image-url)
as the Slochteren Sandstone in The Netherlands (Figs. 3.4, 3.5). The clayey lacustrine facies of both areas is named the Silverpit Formation (Rhys, 1974; NAM, 1980). The characteristic lithologies that make up these formational units have been studied in cores and can be deduced from the wireline logs, for instance, of the type and reference sections depicted in figure 3.5 for different parts of the North Sea area.

Fluvial facies

The fluvial sequences are characterised by the occurrence of curled clay flakes, indicating frequent subaerial exposure and desiccation. Some thicker clays seem to have had their cracks infilled with sand from above; others were injected with a slurry of sand and water from below to form sandstone dykes (Glennie, 1970). The sandstones directly below clay beds display centimetre to decimetre foresets with low-dip and commonly discontinuous laminae, which are interpreted as having been deposited by flowing water. These sandstones are locally conglomeratic, with some of the contained pebbles consisting of red clay similar in character to the bedded claystones. Laminated sandstones locally grade up into apparently homogenous sandstones that lack sedimentary structures. Some homogenous sands may contain large pebbles and are obviously of fluvial origin; others, however, grade down into sandstones with well-defined laminae typical of the aeolian sands that are commonly interbedded with those of fluvial origin. The origin of these structureless sands will be discussed below (Section 3.3.2).

Fluvial sands tend to be well cemented with dolomite (originally calcite?) and only locally form good reservoirs for hydrocarbons (e.g. in the Groningen Field). Many of the above features indicate that these essentially fluvial sequences should be interpreted as having been deposited by ephemeral streams in an arid or semi-arid environment; and may thus be referred to as wadi deposits.

Wadi sandstones are common along the southern margin of the Rotliegend basin (Fig. 3.6) and especially in Holland, Germany and Poland (see Ziegler, 1982, Encl. 13). In the U.K. sector of the North Sea, wadi sandstones are generally poorly developed, the Rotliegend sequence being dominated by sandstones of aeolian origin (see e.g. Nagtegaal, 1979). Such wadi sandstones are better displayed in the East Netherlands (Bungener, 1969) and in other basins with areas of greater relief, as in Arran, Scotland (Clemmensen and
Fig. 3.5. Rotliegend type and reference well sections.
Abrahamsen, 1983), SW Scotland (Brookfield, 1980), and Devon, England (Laming, 1965). A sequence of cores from the southern North Sea that include wadi sands and conglomerates are illustrated by Glennie (1972, Fig. 10).

**Aeolian facies**

The aeolian sands are most readily recognised when they conform to a series of criteria which include:

(a) well defined planar or trough-bedded strata in which adjacent laminae commonly show sharp grain-size differences between that of very fine sand and some 1 or 2 mm; (b) many intra-formational unconformities, above which the laminae are commonly horizontal; there is an upwards increase in the inclination of the laminae to an angle of some 20° to 25° before being terminated by the next low-angle truncation, and (c) a lack of mica flakes. Sequences of almost continuous aeolian bedding of this type locally reach thicknesses of one or two hundred metres or more. Analysis of the orientation of the aeolian bedding both in outcrop (e.g. Durham, England) and in wells (dip-meter logs) indicate that both transverse and seif dunes were formed in the Southern Permian Basin (Fig. 3.11).

Within the Southern Permian Basin, the Early Permian winds blew from roughly east to west (Glennie, 1972, 1983a; Van Wijhe et al., 1980), which, after correcting for the rotation of N.W. Europe since the Permian, suggests that the Rotliegend was deposited in a 'Trade Wind' desert of the Northern Hemisphere similar to the Sahara of today (Figs. 3.6, 3.8, 3.9).

In some North Sea wells, typical dune bedding grades into intervals of irregular wavy laminations, which are interpreted to be adhesion ripples, formed when wind-blown sand adheres to the damp surface of a sabkha. These adhesion ripples are believed to have been deposited over interdune areas whose surfaces coincided with the water table. The presence of small blebs of anhydrite within the adhesion ripples testifies to the hot and arid climate which caused gypsum to crystalise within the sands.

Dune sands form the main reservoir rock for gas in the Southern Permian Basin. At the time of deposition their...
porosities probably averaged around 42% (Hunter, 1977). With increasing depths of burial the porosity was progressively reduced, first by compaction and pressure solution and then by the growth of authigenic minerals including chlorite and illite (Glennie et al., 1978; also Fig. 3.15). At greater depths, especially after the porosity has already been reduced to around 10%, illite develops in a fibrous form, which has a very deleterious effect upon permeability without significantly reducing porosity.

Lacustrine facies

The lacustrine facies consists primarily of red-brown mudstone with minor siltstone. Several halite horizons occur in the lower half of the sequence in U.K. waters (Figs. 3.4, 3.5, 3.11), but in North Germany, halite is best developed in the middle of the Upper Rotliegend sequence, thus implying a longer local history of sedimentation. The salts attain a sufficient thickness to react diapirically in northern Germany (see e.g. Plein, 1978, Fig. 1), and the deformed lacustrine sequence is then known as the Haselgebirge facies. In the Glückstadt Graben at the southern end of the Danish peninsula, Rotliegend halite occurs in the same diapiric structures as halite of Zechstein age (Best et al., 1983, Fig. 4), mobilisation probably being triggered in the Triassic by earth movements associated with the Hardegsen disconformity (see Fisher, Section 5.2.1). Although the lacustrine facies achieves a thickness of some 1500 m in northern Germany, it seems to be devoid of fossils apart from the top metre or so (the fossils have strong Zechstein affinities and are probably attributed more correctly to the Zechstein marine transgression).

The basin-centre parts of this sequence also contain no known sedimentary structures indicative of subaerial desiccation or erosion, although this may merely reflect the lack of coring in commercially unproductive rocks. The desert lake is therefore believed to have been a constant feature throughout the early Permian. At its fullest extent, it must have covered an area of some 1200 km E-W by over 200 km N-S (Fig. 3.1).

Sabkha facies

Between the deposits of the desert lake and the more southerly depositional areas of the wadi and aeolian sands (Figs. 3.1, 3.6, 3.11), is a broad band of poorly bedded clays, silts and sands that display many features indicative of a sabkha (e.g. mud cracks, sandstone dykes, adhesion ripples, anhydrite nodules—see Glennie, 1970; Nagtegaal, 1973). These features collectively indicate a largely aquatic depositional area that was subject to limited aeolian deposition and to sub-aerial desiccation in an arid climate. The sabkha sediments represent the area that was covered by water only during the maximum extensions of the desert lake (Fig. 3.1).

The lacustrine and sabkha facies lack reservoir development. To the contrary, they are much more likely to act as a seal for underlying hydrocarbon accumulations.

In the Southern North Sea and The Netherlands, the Rotliegend sequence is developed, from south to north and from base to top, in the following generalised facies (Figs. 3.4, 3.11): (a) mixed wadi and aeolian, (b) mainly aeolian, (c) sabkha, (d) desert lake. The whole sequence is capped by the Kupferschiefer with only a few centimetres to a maximum of three or four metres of marine-reworked sands between (Fig. 3.4).

The 'Weissliegend'

At many localities in the belt of dune sands, and for a distance beneath the Kupferschiefer of up to 50 m (150 ft) or more, a sequence of uncoloured ('Weissliegend') structureless sands alternate with, and grade into, highly deformed strata as well as into beds in which the original aeolian bedding is only weakly preserved. These non-depositional features are collectively believed to have resulted from a very rapid Zechstein transgression, and the escape of air trapped beneath the wetted surface of aeolian sand dunes during the rapid rise of water level (Glennie and Buller, 1983). Such a rapid transgression has already been invoked by Smith (1979) to explain other features in the overlying Zechstein of NE England.

3.3.2 Moray Firth and Northern Permian Basins

The facies distribution of the Rotliegend in these two basins is still incompletely known (Fig. 3.6). In the deeper parts of the basins the Rotliegend is rarely reached by the drill and, except near Elgin on the Moray coast, it is not seen in outcrop.

On present evidence, both basins probably contain rocks of the same general sedimentary facies already recognised in the southern basin (fluvial, aeolian and sabkha, although the presence of a bedded lacustrine halite has yet to be demonstrated). The sand-dominated sequences in U.K. waters have been designated the Auk Formation and the shaly sequence the Fraserburgh Formation (Deegan and Scull, 1975).

Much of the Auk sequence in the type locality has been interpreted as aeolian dunes. Dip-meter data indicate that here the winds blew in a direction opposed to that of the Southern Permian Basin (i.e. towards the E and SE; see Fig. 3.6); a barometric high must therefore have existed in the vicinity of the Mid North Sea structural high. The basal 14 m of the Auk Formation at the type locality (Fig. 3.5) is conglomeratic and contains clasts of quartz and schist. The shales of the Fraserburgh Formation in Shell/Esso well 21/11-1 contains dolomitie and micaceous sandstone stringers, which are anhydritic, and also adhesion ripples. The depositional environment is here interpreted as a dune–bordered sabkha (Deegan and Scull, 1975). Thus the Auk and Fraserburgh formations are broadly similar to the Leman Sandstone and Silverpit Claystone formational sequences of the Southern Permian Basin.
Similar sequences have been recognised in released Norwegian wells (Fig. 3.5) but as yet are unnamed.

In the U.K. part of the Northern Permian Basin the thickness of the Rotliegend changes rapidly from place to place, making correlation between wells very difficult. The differences may reflect deposition in small rotated half grabens such as formed on the flank of the developing Central Graben (cf. Fig. 3.11). Such an interpretation could explain the derivation from adjacent fault scarps of the clasts of quartz and schist in the Shell/Essco well 30/16-I mentioned by Deegan and Scull (1975).

It seems likely that in the Moray Firth Basin, Rotliegend sedimentation kept up with subsidence, as the overlying Zechstein is entirely in a shallow-marine facies. In the centre of the Northern Permian Basin, on the other hand, subsidence probably greatly exceeded sedimentation and the succeeding Zechstein is dominantly in a basinal halite facies (see Taylor, Chapter 4, Fig. 4.9; Taylor, 1981). Like the central parts of the Southern Permian Basin, the halite reached thicknesses over much of the northern basin that were great enough to permit diapirism (Taylor, Fig. 4.11).

On the Moray coast of Scotland, north of Figgis, around 200 feet (60 m) of dune sands, known locally as the Hopeman Sandstone (Peacock et al., 1968), contain some reptile foot prints, which have been tentatively dated as Early Triassic or Late Permian. Bedding attitudes indicate that the contemporary winds blew in both southerly and south-westerly directions. Intermittently along 10 km of this coastline, and coinciding roughly with the deepest sands exposed, the dune bedding grades both laterally and upwards into structureless and distorted sequences up to 20 m or more thick (Fig. 3.7). As with the Southern North Sea examples, the origin of these deformation structures also is ascribed to the escape of air through the wet surface of aeolian sand dunes prior to deposition of the basal Zechstein Kupferschiefer (Glennie and Buller, 1983). The structureless sands may result from the rapid upward replacement of escaping air by water, which causes slight differential grain movement that obliterates the bedding. The entrapment of sufficient air to deform these sands on such a large scale must have resulted from the differential capillary penetration of water into the dunes related to a very rapid rise in water level. The widespread occurrence of these deformation structures at the same sub-Kupferschiefer stratigraphic level in all three North Sea Permian basins implies that it must have been the Zechstein transgression that caused the rise in water level. For such a rapid rise in water level to be possible, the surface of the Rotliegend desert must have been below global sea level.

Although Zechstein strata are not recognised on land in the Moray Firth Basin, apart from the above described effect, they are known, from both bore hole and seismic data, to be present in the subsurface only few kilometres offshore (Fig. 3.7). Thus these basin-margin Moray dune sands probably cover a greater area than formerly believed, from Early Permian (Rotliegend) to Early Triassic. Offshore, Rotliegend sandstones alone attain a thickness of up to 600 m (Fig. 3.2).

Unlike the 'Weissliegend' sandstones of the North Sea basins, the Hopeman Sandstones of the Moray coast are stained red. This probably reflects a subsurface diagenetic environment induced by overlying desert conditions during the later Permian or early Triassic. The Hopeman sands were probably at the limit of the Zechstein Sea and were only temporarily covered by its waters.

![Fig. 3.7. Dune sand relationship—southern Moray Firth.](image-url)
3.4 Historical development

3.4.1 Climate

The change from the humid equatorial conditions under which the Carboniferous Coal Measures were deposited to the arid climate of Rotliegend deposition, was probably mostly an effect of the passive northerly drift of Laurasia. The Southern Permian Rotliegend Basin came to occupy a latitudinal position north of the equator similar to that of the present North African-Arabian deserts (Fig. 3.8). Lower Rotliegend aeolian sandstones in Germany (Drong et al., 1982) and evidence of strong pre-Saxonian deflation in the southern North Sea area (Glennie, 1983b), indicates that the Southern Permian Basin had already entered the Trade Wind desert belt early in the Permian.

The newly created E-W trending Variscan Highlands occupied a near equatorial location and will have had at least some tropical rainfall to judge from the Stephanian to Autunian coals in Saarland and Central France. A time-related reduction in the volume of fluvial sediments within the Southern Permian Basin may indicate that during the earlier part of the Rotliegend depositional history, the sources of fluvial activity were still just within the equatorial zone of higher rainfall; later, these source areas may also have entered the region of desert climate, resulting in an extension of the area of dune sands and some reduction in the size of the desert lake.

Well data, coupled with excellent quarry exposures of aeolian sands in Durham, England, indicate that many of the Rotliegend dunes attained a height of up to 50 m or more.

Today's winds are capable of constructing only small seif dunes with an average maximum height of 5 to 10 m. The large modern seif dunes of Arabia and North Africa, possessing heights of some 100 m and wave lengths of 1 or 2 kilometres, were probably constructed during the Pleistocene. The most likely reason for these size differences is that the large areas of high barometric pressure associated with major glaciations (Permio-Carboniferous as well as Pleistocene) will have caused a concentration of the world's air pressure belts towards the equator and thus have created a shorter distance between the zones of high and low pressure than is now the case (Fig. 3.9). The resulting wind systems probably had higher average velocities than now, and were also colder than now, interpretations that are gaining support among many workers (e.g. Galloway, 1965; Bowler, 1976; Krinsley and Smith, 1981; Rea and Janecck, 1982).

These Pleistocene winds possibly persisted for a much greater part of the year instead of blowing for only a few hours or days at a time as is now the case. With ice caps the size of those found in Gondwana, the Permian dunes, like those of the Pleistocene glaciations, are likely to have been built on a scale that is impossible today (see also Glennie, 1983a, b).

If these deductions are correct, the Early Permian desert winds are likely to have been strong and, because of their long continental route before reaching the western part of the Southern Permian Basin, also very dry. Thus the area of dune activity will have extended during Gondwana glaciations, and strong evaporation will have reduced the size of the desert lake to a minimum. The dying stages of the Early Permian glaciations, on the other hand, should have coincided with a generally weaker wind system with a resulting combination of a higher convection-induced rainfall and less evaporation. This may explain the lateral extension of the lacustrine and sabkha facies for a considerable period prior to the Zechstein transgression. These points are suggested schematically in Figures 3.4 and 3.11.

3.4.2 Basin formation

Late Westphalian N-S compression brought about the creation of the Variscan Highlands to the south of Britain, and associated right-lateral shearing in the U.K. part of the Southern North Sea seems to have resulted in inversion of the Sole Pit area (Glennie and Boegner, 1981). The early collapse of the Variscan Highlands in the west, and the development of a horst and graben system that was possibly allied to the creation of a proto-Atlantic fracture system, gave rise to E-W tension and to right-lateral extension movements in the Southern North Sea area (Fig. 3.10; Glennie and Boegner, 1981) and in Germany (Drong, et al., 1982). These movements resulted in subsidence of the Southern and Northern Permian Basins, leaving the Mid North Sea structural high as a relic of relative stability. It also resulted in the en-echelon development of NW-SE
trending sub-basins, such as the Sole Pit and Broad Fourteens basins, which continued to subside until the late Mesozoic, and possibly of other sub-basins as far east as Poland.

The same E-W regional tension is suspected of causing the development of the N-S and NW-SE oriented fracture systems of the North Sea (e.g. Oslo and Horn Grabens, Viking and Central Grabens) already during the Early Permian and perhaps even in the latest Carboniferous (Figs. 3.3, 3.10). Lower Rotliegend volcanics occupy the Horn Graben (Figs. 3.3, 3.11) and the Mid North Sea flanks of the Central Graben. And some of these subsiding grabens were deep enough by Zechstein time to allow the accumulation of salt that was sufficiently thick to move later diapirically (Southern Viking Graben and Central Graben). This contrasts with the areas flanking the grabens where early Zechstein strata are locally absent and the later Zechstein was deposited entirely in a shallow-marine carbonate/anhydrite facies (Fig. 3.11; see also Taylor, 1981).

3.4.3 Zechstein transgression
As shown earlier, the ‘Weissliegend’ sediments of the uppermost Rotliegend contain important evidence concerning the rapidity of the Zechstein transgression, which is why the transgression is discussed in this rather than in the succeeding chapter.

It is along a combination of the proto-Atlantic and North Sea fracture systems that the waters of the Zechstein transgression are presumed to have been transported from the Permian open ocean somewhere between the northern coasts of Greenland and Norway (Fig. 3.10).

Future evidence from the offshore areas between East Greenland and Norway may indicate whether the mid-Permian proto-Atlantic rift was a narrow graben, as implied in Figure 3.10, or was a broader arm of the ocean as suggested by Callomon et al. (1972), and depicted by Taylor in Figure 4.1.

The Zechstein transgression probably started because a world wide rise in sea level, coinciding with the end of a phase of Permian glaciation, permitted oceanic water to flow along a pre-existing tensional fracture system. The surfaces of both the northern and southern Permian basins were probably well below the level of the open ocean, so that once the water began to flow south along the fracture the transgression continued until the level of the Zechstein Sea matched that of the ocean (Figs. 3.10, 3.11).

If the surface of the Rotliegend desert lake lay about
Fig. 3.10. Post-Variscan fault patterns. Modified after Russell (1976) and Ziegler (1978). CG Central Graben; HG Horn Graben; M Moray Firth; MNS Mid North Sea High; NPB Northern Permian Basin; OG Oslo Graben; RF Ringkøbing-Fyn High; SPB Southern Permian Basin; VG Viking Graben.

Fig. 3.11. Conceptual block diagram of the Southern Permian Basin and Central North Sea system of highs at the time of the Zechstein transgression. Zechstein Sea is presumed to have flowed into basin via Central and/or Horn Graben. Note suggested change in dune style from transverse in centre of basin to seif in western basin-margin location. From Glennie and Buller (1983).
250 m below the level of the open ocean, it would have required some 75000 km$^3$ of water to fill the Southern Permian Basin (Fig. 3.1), and another 35000 km$^3$ of water to fill the Northern Permian Basin. Ignoring seepage and evaporation, these basins could have been filled in about 6 years if they were jointly flooded at the rate of say 50 km$^2$/day (e.g. channel of water 10 km wide, 20 m deep and average velocity of 3 m/sec).

Such a flood of water into the Southern Permian Basin may have caused some initial scouring of the desert lake sediments where it debouched into the basin, but on the other side of the lake, 100 km or more away (Fig. 3.11), erosion will have been minimal. If the proposed flow rates are correct, then initially the rise in water level will have been around 30 cm/day, and lakeside dunes, 50 m high, will have been covered with water in just over 150 days. It is this relatively rapid and continuous rise in water level which is thought to have been responsible for the in-situ deformation of the 'Weissliegend' upper part of the Rotliegend sedimentary sequence. The surfaces of these dunes were reworked by wave action only to a limited extent. Thus the original shapes of the dunes were only slightly modified and considerable relief was preserved, as can be seen from exposures in NE England. The succeeding Kupferschiefer draped this dune relief. It is unrealistic, therefore, to attempt a detailed correlation of Rotliegend reservoir sequences in the belief that the Kupferschiefer formed a horizontal datum plane.

3.5 Hydrocarbon occurrences

3.5.1 Gas fields

Exploration in the hostile environment of the North Sea was triggered by the realisation that there were sufficient reserves in the Rotliegend reservoirs of the giant Dutch Groningen field to alter the fuel economy of much of NW Europe from a reliance on coal and oil to one based more extensively on gas. Although discovered in 1959, it was not realised until several more wells had been drilled that the continued discoveries all belonged to the same field (Te Groen and Steenken, 1968). With a surface area of almost 800 km$^2$, and a porosity that ranged from 10-25% (permeability 0.1-1000 mD), the field was conservatively estimated in 1968 to have ultimate recoverable reserves of some $1650 \times 10^3$ m$^3$ ($58 \times 10^{12}$ ft$^3$) of gas (now believed to be nearer $2425 \times 10^3$ m$^3$ or $86 \times 10^{12}$ ft$^3$); the unexplored southern North Sea lay down the palaeowind, roughly due west from Groningen (Figs. 3.1, 3.6, 3.12). The basic details of this giant are given in a series of articles by Te Groen and Steenken (1968), van der Laan (1968), Bungener (1969) and Stäuble and Milius (1970).

All the producing Rotliegend gas fields of Germany, The Netherlands and the Southern North Sea are underlain by Westphalian Coal Measures and overlain by Zechstein salt. The desert lake and sabkha facies contain no reservoir rocks and, apart from the Groningen area and the small Rough field offshore Yorkshire (see Robertson, 1981, Fig. 6), the fluvial sands are generally too well cemented to form good reservoirs. The commercially productive reservoirs are therefore largely confined to the aeolian facies. These factors result in the southern North Sea Rotliegend gas fields being limited effectively to an E-W band some 100 km wide stretching from the North German Plain, through Groningen, to the east coast of England (Fig. 3.12).

Gas generation, migration and entrapment

Superimposed on these three basic requirements for an oil or gas field of source, reservoir and cap rocks, is the general need for structural deformation to create a trap.

It is axiomatic that a trap must be formed before migrating gas can be retained in a reservoir. In the Southern Permian Basin, some early traps formed in areas of slower subsidence, whereas others resulted from subsidence followed by inversion. In some of the latter cases, subsidence was continuous from the Early Permian until the Late Cretaceous, when uplift in the order of 1 to 4 km took place (Figs. 3.13, 3.14, 3.15, 3.16); this was followed by further differential subsidence during the Tertiary. In other areas, uplift began in the Mid Jurassic or Early Cretaceous, to be followed by more subsidence in the later Cretaceous. All these movements were probably coincident with widespread but minor transcurrent faulting related first to the development, and then to the final demise, of the major North Sea graben system (see Chapter 2.5).

Gas generation has been shown by Van Wijhe et al. (1980) to result from burial of the Coal Measures to depths of some 4000 m or more (Fig. 3.13). In the areas around Hamburg, Germany, Westphalian coals have reached the rank of anthracite (Bartenstein, 1979), and must by now have given up most, if not all, of their gas. This was an area of rapid subsidence throughout the Permian (e.g. 1500 m of Rotliegend desert-lake sediments) and Triassic (almost 9 km thick in the Glückstadt Graben—Best et al., 1983, Fig. 4). Much of the succeeding Jurassic was removed during the 'Late Cimmerian' phase of erosion around the Jurassic-Cretaceous time boundary, preserved sequences being confined largely to the rim synclines of Zechstein diapirs, which were already active during the later Triassic (Best et al., 1983, Fig. 4). Under these conditions of burial, gas generation in the Glückstadt Graben (Fig. 2.2) probably began already during the Triassic. With impervious desert-lake sediments immediately above, this gas must have migrated to the south to reach porous Rotliegend sandstones; to the north, the lack of a Rotliegend reservoir and the general absence of Mesozoic seals over the Ringkøbing-Fyn High will have resulted in the escape of much gas to the surface.

Gas generation in the Broad Fourteens and Sole Pit Basins probably took place during the later Jurassic and early Cretaceous, so that the Rotliegend reservoirs in contemporary highs flanking the basins were
Fig. 3.12. Distribution of some Rotliegend producing oil and gas fields.
probably charged with gas at that time. After inversion of the basin areas, however, gas remigrated into a reservoir that was already partly damaged by the growth of authigenic illite during the earlier deep burial (see Glennie et al., 1978 and Fig. 3.15). In the Groningen area, however, Early Cretaceous uplift separated two periods of gas generation (Fig. 3.13) and it is probable that the Groningen field is still being charged with gas.

With such a long history of vertical and horizontal movement, it is not surprising that the Rotliegend reservoirs of the southern North Sea are highly faulted (Fig. 3.16). These fault movements undoubtedly triggered diapirism in the overlying Zechstein salt; in turn, some idea of the times of fault movement can be deduced from the erosional and depositional history of the rim synclines flanking the diapirs (see Taylor, Chapter 4, Figs. 4.13-4.15). Because of the relatively high acoustic velocity of halite, the rapid changes in its thickness resulting from halokinesis considerably distorts the shape of the underlying seismic reflectors, and must be taken into account when converting time maps of the Rotliegend to depth (see e.g. Butler, 1975; Christian, 1969).

The search for Rotliegend gas below the North Sea began in U.K. waters in 1964, and soon resulted in the discovery of a series of important fields, West Sole—Viking—Leman—Indefatigable (Figs. 3.12, 1.5), of which the last two account for some $425 \times 10^9$ m$^3$ ($15 \times 10^{12}$ ft$^3$) of recoverable gas; these four fields have been described, respectively, by Butler, Gray, van Veen and France, all of which can be found in Woodland (1975).

Some gas fields of the Netherlands offshore that were strongly influenced by inversion movements have been described by Oele et al. (1981) for quadrants K and L (see Fig. 3.17) and by Roos and Smits (1983) for block K/13. The latter authors consider that Late Cretaceous inversion enabled the Triassic Bunter Sandstone in block K/13 to be charged from a gas-filled Rotliegend reservoir; the preservation of gas in two small Rotliegend fields (structures E and F in Figure 5.7) supports this interpretation. Similar fault movements probably caused the breakdown of the intervening Zechstein seal and permitted gas to transfer from the Rotliegend sandstone to both Zechstein (Plattendolomite) and Triassic reservoirs in the Bergen area of the Netherlands (van Lith, 1983) and to the Triassic reservoirs of the Hewett field on the margin of the Sole Pit Basin (see Fig. 5.5; and Cumming and Wyndham 1975); gas was still retained in the Rotliegend reservoir of the adjacent small fields, Deborah and Dotty (Fig. 3.12).
Fig. 3.14. Burial histories of the base Rotliegend at selected well locations, Sole Pit Basin, southern North Sea. From Glennie and Boegner (1981).
Fig. 3.15. Depth-related diagenesis in the Leman Bank and Sole Pit areas of the U.K. Southern North Sea. From Glennie et al. (1978).

Fig. 3.16. Geological cross-section, U.K. Southern North Sea. Note fault-bounded relief of Rotliegend sequence, thick cover of Zechstein salt over Indefatigable, and much thinner salt over Leman and adjacent to the Dowsing Fault.
Reservoir quality and diagenesis

In general, the best reservoir sands occur in the dominantly aeolian middle part of the Rotliegend sequence, with porosities in some fields ranging up to about 25%, and air permeabilities with values in excess of 100 mD where not damaged diagenetically. The porosity of the aeolian sands directly beneath the Kupferschiefer is commonly reduced for two different reasons: the original grain packing became tighter in the Weissliegend sands (the top 0-65 m) because of deformation associated with the Zechstein transgression; and the porosity of the upper 10 to 15 m of sands, whether deformed or not, is drastically reduced because of a dolomite cement believed to result from proximity to the overlying Zechstein carbonates. Robinson (1981) finds that the ratio between horizontal and vertical air permeability of most aeolian and wadi sands is between 1 and 100. This distinction is not seen in the Weissliegend sands because of either a lack of distinct bedding, or the presence of deformed bedding. Both these points are reflected by dip-meter logs, which generally show no dip data or, alternatively, only random dips.

The thick Rotliegend sequence found in the Leman Bank gas field (Figs. 3.12, 3.16) comprises mostly foresetted aeolian sandstones similar to those illustrated by Glennie (1972, Fig. 12) which are interpreted as having been deposited on the avalanche slopes of transverse dunes; these undercompacted sands had a naturally high primary porosity. In the Indefatigable area (Fig. 3.12, 3.16), on the other hand, the Rotliegend aeolian sequence is much thinner and has many horizons of silt and anhydrite-rich (former gypsum crystals) adhesion-ripple sands, which had an inherently poorer porosity and permeability than the foresetted dune sands. The burial-related growth of authigenic minerals, and especially of illite, in areas of former deep burial such as the Leman Bank field, has resulted in the porosity and permeability of the dune sandstones now being much lower on average than those of the Indefatigable field, which was never so deeply buried (compare well 49/24-1 in Fig. 3.14 with Fig. 3.15).

The capillary effects associated with smaller pore connections in the diagenetically more damaged reservoir of Leman Bank probably account for the overall difference in recovery factors between the two fields, from 80% of the gas in place for Indefatigable to 75% for Leman Bank. Similarly, Robinson (1981) attributes a relatively high water saturation in the Amoco well 47/15-2 to the reduction in the size of pore throats related to the precipitation of authigenic minerals.

As we have seen, because of the effects of diagenesis, the quality of most gas reservoirs becomes poorer with increased depth of burial. It comes as a pleasant surprise, therefore, to find an example where Lower Rotliegend aeolian sandstones have better porosity and permeability at a depth of over 5000 m than the dune sands of the overlying Upper Rotliegend. This rather unusual situation is found in the Schneverdingen Graben in West Germany, about 50 km east of Bremen (Drong et al., 1982). The almost 400 m thick Schneverdingen Sandstone was deposited in an actively subsiding narrow graben (Fig. 3.18) and has preserved porosities of up to 15% and permeabilities in the range of 1-10 mD, in contrast to the 1-5% porosity and < 0.1 mD permeability of the Upper Rotliegend sandstones. Drong and his colleagues suggest that the differences in reservoir quality might reflect differences in climate between...
the times of deposition of the two Rotliegend dune sequences (less calcite and anhydrite cement in the Schneverdingen Sandstone). It seems just as likely, however, that the porosity differences could also be related to differences in grain size, with stronger capillary retention of formation water (and hence stronger diagenesis on burial) in the finer grained (0.1-0.25 mm) Hauptsandstein than in the coarser (0.25-0.5 mm) deeper sands of the Schneverdingen Sandstein.

In the Northern Permian Basin, the Zechstein salt cap rock is present over the greater part of the area (Fig. 4.9), but the all important source rock for gas, the Carboniferous Coal Measures, is conspicuously absent (see Figs. 2.2, 2.8). Thus in this area, those Rotliegend reservoirs that have been penetrated by the drill are devoid of gas. Where the Rotliegend sandstones of this basin form reservoirs for hydrocarbons, they have been brought into the correct geometric relationship with a mature source rock for oil.

3.5.2 Oil fields

Although minor amounts of condensate and Natural Gas Liquids (NGL) are known from some Rotliegend structures in the Southern Permian Basin, they occur as the heavy fractions of gas of Carboniferous origin.

The Auk and Argyll oil fields in the Central North Sea produce much of their oil from the basal Zechstein carbonates. In the shallower parts of the fields, however, Rotliegend sandstones are also saturated with oil. These fields are situated close to the western flank of the Central Graben. Pennington (1975) suggests that the oil found in the Argyll field is derived from structurally adjacent Paleocene shales. These shales, however, probably have not been buried sufficiently deeply to be mature. It seems much more likely that the oil source rock for both the Auk and Argyll fields, like most of the fields of the Central and Northern North Sea, is the Kimmeridge Shale, which matured deep in the Central Graben sometime during the mid to late Tertiary. The oil probably migrated up graben-flank faults and became trapped beneath the general cover of a relatively impervious Upper Cretaceous chalk (Fig. 3.19).
Fig. 3.19. Schematic structural setting and cross-sections of the Auk and Argyll oil fields. Source Rock—probably Upper Jurassic Kimmeridge Shales. Reservoir Rock—Zechstein dolomites and Rotliegend sandstones. Modified from Brennand and van Veen (1975), and Pennington (1975).
3.6 Acknowledgements

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EXPERIMENTAL SOFT-SEDIMENT DEFORMATION

IN ARTIFICIAL DUNE SANDS

by

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ABSTRACT

Four simple experiments were carried out to see what happens when a small artificially-constructed sand dune is fairly rapidly inundated by non-scouring water, and especially to see whether soft-sediment deformation could be induced of a type known in the Permian Weissliegend dune sands of N.W. Europe.

The first two 'dunes' were built at one end of a fish tank by avalanching sand down a slope. The third and fourth were constructed in the middle of the tank with gentler 'windward' slopes; flooding then took place at both ends simultaneously. Sharper grain-size contrasts between adjacent laminae were achieved in the later experiments by using some sieved sands with a smaller range in size.

During flooding, water rose by capillary action faster along fine-grained laminae than along coarse-grained laminae, so that air became trapped in the pore spaces of the latter. With the more complex bedding geometry of later dunes, pockets of air-filled sand were trapped beneath laminae of wetted sand. The natural upward escape of this air caused limited deformation in the overlying sands.

To study further the process of deformation, low-pressure air was injected into the base of the third and fourth 'dunes' via a capillary needle. With time, some relatively large deformation structures, similar in style to those seen in the Weissliegend, were formed within and at the surface of the 'dunes' by the upward escape of air through laminae, both above and below the level of free water in the tank, that had originally been wetted by the capillary rise of water. The ever-changing location of air-filled vugs within the heart of the dune resulted in considerable homogenisation of the original laminae. Deformation was most readily achieved in the fourth dune, in which much of the sand was relatively coarse and clean.
EXPERIMENTAL SOFT-SEDIMENT DEFORMATION IN ARTIFICIAL DUNE SANDS

K.W. Glennie

INTRODUCTION

Four simple laboratory experiments were carried out to study the effects of the fairly rapid submergence of dune sands beneath water. These experiments stemmed from work by Glennie and Buller (1983), who have proposed that the soft-sediment deformation structures found in the Weissliegend (non-reddened upper part of the Early Permian Rotliegend) dune sands of N.W. Europe are the result of a rapid but essentially non-erosional inundation by the marine waters of the Zechstein Sea. They suggest that the surface of the Rotliegend desert lake in the centre of the Southern Permian Basin lay well below the level of the world's oceans, and when the Zechstein Sea began to flow into it from a boreal source somewhere between Greenland and Norway, the water did so via a pre-existing proto-Atlantic fracture system. They further suggest that, at least during the early stages of flooding, the level of the Zechstein Sea might have risen at a rate measured initially in decimetres per day, so that lake-side dunes, possibly some 50 m high, could have been covered by water within, say, 8 months or so.

Under these conditions of geologically rapid flooding, water would have penetrated some parts of the dunes more rapidly than others because of differential capillary flow along laminae of varying grain size; as a result, air could have become trapped beneath permeability barriers formed by laminae of fine-grained sand that had been wetted by the capillary rise of water.

Because of the inherent high porosity (about 40-45%; Hunter, 1977) and permeability (many Darcies) of dune sand, the hydrostatic pressure in the water-saturated sands of a dune will increase in concert with flooding at a rate of one atmosphere for every 10 m depth below the water surface. That high permeability also permits the total pocket of air-filled dune sand to have the same pore pressure
throughout, so that for every 10 m column of air trapped within a dune, there will be a pressure differential at its contact with the overlying wetted dune sand of one atmosphere (Fig. 1). This pressure differential will be potentially greatest in large (high) dunes that are not fully covered by water, and values of several atmospheres are conceivable if the dunes contain columns of entrapped air some tens of metres high.

The air is trapped in dune sand that is almost free of clay-size particles. The internal air pressure at the top of the air pocket has to be contained by a combination of the capillary pressure at the water/air interface created by the small effective radii of the pore throats of the containing sand laminae, and the hydrostatic pressure associated with the overlying column of free water, if any. Should the internal pressure exceed these combined forces, then air will escape upwards and deform the moist sands through which it passes. This air will be replaced from below by water which, if rising fast enough, could also cause some internal deformation.

The writer considered that it might be possible to reproduce in the laboratory some of the processes involved in the soft-sediment deformation of dune sands and, perhaps, to observe other, hitherto unsuspected, features.

One further aspect of the flooding of dune sands required confirmation. There is a belief among some geologists that when aeolian sands are submerged beneath water, the 32-34° maximum angle of repose of these sands readjusts to a lower figure; although, to the writer's knowledge, it has never been stated in print, a new angle of 25-27° has been verbally suggested as possible. This suggestion seems to lack both scientific observation and measurement, however, and the way in which a lower angle might be achieved is not known beyond a vague suggestion that the surface sediment will slump. Observation of changes in the angle of surface slope became another of the aims of the experiments outlined below.

The purpose of this paper is to report the observations made during the experiments, and to suggest explanations for their origins.
EXPERIMENTAL AIMS

Glennie and Buller (op. cit.) indicate that the Weissliegend soft-sediment deformation structures seem to be best developed in the undercompacted aeolian sands that were deposited on avalanche slopes. All experiments were therefore conducted on artificial dune bedding formed mostly by allowing it to accumulate by avalanching down a slope (Fig. 2). The limited aims of each succeeding experiment were, in part, the outcome of its predecessor.

These aims were:

1) To see whether there is any reduction in the angle of surface slope when aeolian slip-faces are submerged beneath water.

2) To observe the capillary movement of water along sands deposited on artificially constructed avalanche slopes.

3) To see whether the capillary flow of water within dune sands is likely to be capable of entrapping air, given a suitably complex cross-bedding geometry and a level of free water outside the dune that is rising fast enough.

4) To see whether soft-sediment deformation could occur in an artificially constructed small-scale dune by the escape of entrapped air.

Although not designed as a primary aim of the experiments, temptation resulted in the surface morphology of the first two dunes being severely modified by induced wave action (Fig. 5), followed, in the case of the first experiment, by a little seismicity (Fig. 6). These activities were pertinent to the origin of the Weissliegend deformation structures since several writers have suggested that the one or the other could cause soft-sediment deformation (e.g. Seed and Rahman, 1977; Horowitz, 1982); Glennie and Buller, however, had considered and rejected their relevance for the Weissliegend.
EQUIPMENT AND AVALANCHE-SLOPE CONSTRUCTION

All four experiments were conducted within the confines of a glass-sided fish tank whose internal dimensions are 198 x 38 x 39 cm. Because of strengthening across the top with additional strips of glass, the effective height of the tank is only 37 cm.

For the first two experiments, avalanche slopes were constructed at one end of the tank by pouring sand onto an inclined sheet of hardboard (Fig. 2). The falling sand was deflected onto the back wall of the tank and it accumulated by sliding over the previously formed slope at about the angle of repose for dry sand. No great care was taken with the 'dune's' construction, so that local overloading of the slope occasionally caused avalanches that were several grain diameters deep. It was noticeable that even over the short slopes that could be built within the confines of the tank, a distinct natural lamination was developed. This was an outcome of avalanching, the coarsest and roundest grains presumably travelling the farthest (Bagnold, 1941). This latter effect was most apparent in the fourth experiment (Fig. 20), in which the long axes of the coarsest grains (shell fragments) reached diameters of 3 mm or more. Coloured sands provided marker horizons whose angle of inclination could easily be measured. The highest dips measured were about 32½°, with many slopes in the 30-31° range.

The grain sizes used in the first experiment were all in the range 150-300 μ (fine to medium sand). In the second experiment, the bulk of the sand was in the range 75-150 μ, with sieved marker fractions (e.g. 150-177 μ) that varied within the limits of 150-300 μ. In the third experiment, the bulk of the sand was again in the grain size range of 150-300 μ but with the coloured marker horizons and some other sieved but uncoloured fractions that fell within the ranges 105-125-149-177-210-250—>250 μ. For the fourth experiment, the sand was taken from the landward side of coastal dunes in The Netherlands. Its grainsize varied from 88 to over 3000 μ in diameter, with the bulk (∼60%) falling in the range 177-350 μ.
During 'dune' construction it was apparent that an avalanche sometimes stopped moving if sedimentation ceased virtually at the time the avalanche was induced. The result was the creation of small 'compression ridges', a few grain diameters high, parallel to the strike of the bedding. A major avalanche, on the other hand, resulted in a noticeable thickening in the lower slopes and a lower angle of inclination of its surface. Such sands finally curved tangentially to the floor of the tank, coming to rest with a narrow horizontal component directed away from the 'dune' (see Fig. 20).

FLOODING DURING THE FIRST EXPERIMENT

In all experiments, one aim was to see what happened to the 'dune's' surface bedding when passively inundated with water; water therefore had to be introduced at such a rate and in such a manner that no scouring of the dune sands took place. As already mentioned, Glennie and Buller (1983) suggested that when the Weissliegend dunes were flooded by the Zechstein Sea, the rise in the water level may have been measured in decimetres per day. The first experiment (both building the dune and flooding it) took only one afternoon to carry out; the level of the water rose to 23 cm after about one hour, when flooding was stopped.

Water did not climb the dune laminae as fast as it penetrated across them, so that it rose within the dune with a fairly regular advancing 'front' that had an average angle to the horizontal of about 22° (Fig. 3). While the level of the water was still rising, the leading edge of damp sand was about 5 cm up the inclined surface beyond the limit of free water. Once the level of free water ceased to rise, capillary drive soon extended the moisture some 15 cm up the dune slope (Fig. 4) and the slope of the advancing 'capillary front' within the dune lessened until finally it became horizontal. These observations suggest that:

a) the rate of flooding was almost as fast as the rather poor natural capillary rise of water along the laminae, and

b) the resistance to water penetration across the laminae was also relatively poor.
In the succeeding experiments, therefore, the rate of flooding would have to be drastically reduced, and the capillary differential between the laminae would have to be improved if the third and fourth experimental aims were to be achieved.

As flooding proceeded, it was noticed that very small quantities of sand were avalanching sub-aquatically down the lower slopes. In time, small mounds, perhaps a few grain diameters high and a centimetre or more across, seemed to form over the surface of the submerged dune and acted as the centres from which much of the avalanching began; they were also the sites from which bubbles of air escaped from within the dune, lifting sand grains in the process. Other, apparently smaller, bubbles of air escaped with minimal disturbance of sand grains. About one hour after flooding had started, the slope of the submerged surface had already reduced to about 30.5°, with the slow, creeping, avalanche process still continuing (compare Figs. 3 and 4).

It was then realised that while water had been climbing the finest-grained laminae by capillary action, air was being by-passed in the pore spaces of the coarser-grained laminae. Given time, possibly all the entrapped air would escape upwards.

Here was a possible mechanism for the origin of the seals that permitted the entrapment of pockets of air within the Weissliegend dunes; it also implied their time-dependent breakdown, which could lead to deformation of the overlying unconsolidated sands. The combination of fine-grained laminae with water-filled pores adjacent to laminae with air-filled pores results in a very strong capillary bond that is capable of enabling unconsolidated sediment to have sub-vertical slopes when exposed subaerially (e.g. Fig. 5); this bond is induced by the surface, or interfacial, tension of the fluid surrounding the sand grains (i.e. the result of the fluid trying to reduce its surface area to a minimum). The bond also results in a permeability barrier that is capable of containing pockets of air under considerable internal pressure within submerged dunes. When air escapes from the pores of the coarser-grained laminae and is replaced by water, however, the 'capillary strength' of the fine-
grained laminae is largely destroyed. When this happened within the submerged Weissliegend dunes, the entrapped air was able to break through newly water-saturated sands, deforming them as it escaped.

Gentle oscillatory wave action resulted in the relatively slow erosion and partial slumping of the sands exposed above the water level, the creation of a shallow sloping submarine shelf from the re-sedimented sand grains, and the overall reduction of the submarine slope by down-slope movement of sand (Fig. 5). The slope of the shelf was about 13°, the upper submarine slope about 26.5°, the lower slope about 29°, and the foot of the slope, despite the overall addition of a considerable thickness of sediment during wave action, maintained an angle of around 31.5°. Apart from the subaerial slumping, wave action did not result in soft-sediment deformation.

Seismic activity, induced by tapping the side of the tank with a rubber hammer, resulted in the rapid collapse of the sediment remaining exposed above water to an angle approximating to that of the submarine shelf (Fig. 6); the slumped sand was redeposited mostly over the submarine shelf, resulting in a slight regression of the shoreline. The redeposited sediment on the submarine slope now took on a decided upward-thinning wedge shape. The angle of the old submarine-shelf marker sand increased by some 4-5° and the dune's internal markers developed a slight but definite convex-upward bend. Although internal warping had taken place as the result of seismic activity, it was certainly not of the type seen in outcrop in Scotland or in North Sea cores (see Glennie and Buller, 1983; Figs. 4, 10, 11).

THE SECOND EXPERIMENT

The results of the first experiment were so encouraging that a second was undertaken. The main aims of the second experiment were to induce a greater capillary differential, and hence to improve the chance of trapping pockets of air. This was to be achieved first, by using a finer overall grain size (75-150 μ, instead of 150-300 μ),
and second, by guaranteeing greater differences in the grain size of adjacent laminae by occasionally depositing sieved sands possessing a narrower range in grain diameter.

Flooding went according to plan and took 9 hours to complete. Water travelled rapidly along the finer-grained laminae by capillary action (Fig. 7) and air was trapped in the pores between laminae of coarser grains. A pocket of air was eventually trapped beneath a lamina wetted by capillary action (Fig. 8); the principle of air entrapment had been confirmed even though it was not possible to maintain a visible three-dimensional seal against the glass side of the tank.

An unexpected reaction to flooding with water was the creation of surface slip-scars parallel to the strike of the laminae. They were probably caused by a shortening of the finer-grained laminae in response to their greater water-induced capillary pressure, with slippage occurring within the much weaker coarser-grained laminae with their air-filled pores. 'Fault scarp's about 2-3 mm high were created across the surface of the dune (Fig. 7). These planes of slippage developed first in the near-surface laminae and were followed progressively by several others slightly deeper within the dune.

The penultimate set of laminae to be deposited comprised very fine grains (75-105 μ) and seemed to contain some silt and clay-size particles. It was overlain by a thin set of surface laminae in the grain-size range 105-150 μ and underlain by a blue marker sand ranging from 177-250 μ. The capillary 'strength' associated with this grain-size combination seemed to prevent the escape of larger air bubbles through the sloping surface of the dune, so that there was no obvious repetition of the small-scale avalanching that was so noticeable in the first experiment. Small bubbles of air did escape, however, as was evidenced by their concentration on the surface of the water, and over the course of several hours the subaquatic surface of the dune seemed to become smoother. Thus sand movement in the order of a grain diameter was probably still taking place.
THE THIRD EXPERIMENT

The success of the first two experiments led to a decision to make a third. This time, however, a geometrically more realistic 'dune' was constructed, which could be flooded simultaneously from both sides (Fig. 9). It again comprised sands deposited largely on avalanche slopes, but its geometry was complicated by an internal plane of truncation and by the deposition of 'topset' laminae over its 'windward' slope (deposited in the tank under the influence of a jet of air).

As in the first experiment, the grain-sizes used were mostly in the range 150-300 μ, but with the addition of sieved sands for some of the laminae in the ranges: 105-125-149-177-210-250->250 μ. The coloured marker sands fell within the same ranges. When complete, the dune had an overall 'downwind' length of about 140 cm and a height of 33 cm. A final leeward slope of about 32° (locally 33°) was achieved over a length of about 60 cm. The 'windward' slope varied between about 20 and 25°.

Flooding took 7 hours 20 minutes to complete, with water rising at an average rate of just over 4 cm/hour. The final water depth was 32 cm. As in the second experiment, the finer-grained laminae were wetted much faster than those of coarser grain size (Figs. 9, 10) and similar 'capillary-induced' bedding-plane slippage was noted. In reaction to the more complex bedding geometry there were many temporary pockets of air-filled sand adjacent to the glass side of the tank (Figs. 10, 11) but, because none were able to retain their lateral seal, all were eventually wetted. It was clear, however, that air was again retained in the pore spaces of the coarser-grained laminae.

Probably because of the lower capillary pressure associated with the pore throats of the generally coarser grain sizes used in the third experiment, air bubbles were again noticed escaping from the surface of the dune, with a greater number coming from the 'leeward' avalanche slope than from the windward slope; and once more the bubbles were associated with small avalanches of sand, especially down the
steep 'leeward' avalanche slope. Although the original leeward slope was around 32° at the time of deposition, it had reduced to about 30.5° some 7 hours after flooding began.

Air accumulated very slowly below the seal of surface bedding near the centre of the 'windward' slope and, within three days of the start of flooding, its escape had caused a slight local deformation of the surface laminae, and a small underlying vug remained for several days until destroyed by later deformation (Fig. 17). Thus it was confirmed that the escape of naturally entrapped air could deform the bedding of an inundated aeolian dune.

It was realised, however, that because of poorly developed seals, insufficient air had been trapped in the dune to cause more than very minor deformation. In a large dune, much more air might escape through the damp sands to the surface. Since the principle of deformation by air escape had been established, it was decided to further the study of the deformation process by injecting air into the heart of the dune. This was achieved by inserting a 70 cm long capillary needle (1 mm outside diam., 0.6 mm inside diam.) along the floor of the dune (reaching to the end of the long arrow in Fig. 17), which was attached to a supply of compressed air. The air pressure was measured with a water manometer. Control of the air pressure proved very difficult initially, and at first it certainly exceeded the equivalent of 200 cm of water (blew the water out of the manometer). Eventually an upward escape of air was achieved at a maximum rate of around one bubble every 10-30 seconds by applying a pressure equivalent to about 110 cm of water (0.11 bar). It is stressed here that the air was injected from a point source within the heart of the dune, and that its initial influence should therefore have been fairly localised.

The first effect of air escape on the bedding was locally dramatic. Almost within seconds of its application, air began to escape up-slope from the terminal point of the needle (Fig. 12) and from then on deformation of the dune bedding was a fairly continuous process. Deformation was not confined to one spot, but neither was the entire dune deformed. Rather, deformation seemed to be confined to one
small area until such time as the process caused a locally higher resistance; the air then found a path of lesser resistance, and began to escape elsewhere.

Most of the deformation process followed a fairly set pattern. First, an air-filled void was created (Fig. 13) by uplift of the overlying laminae of damp sand. The void could be oriented both parallel to the sand laminae (in which case it seemed to form preferentially in the coarser-grained laminae) and oblique or even at right-angles to them. In general, the void slowly increased in size causing the overlying strata to be deformed convex upwards.

All voids, however, eventually propagated upwards across the laminae towards the surface, sometimes vertically and sometimes perpendicular to the bedding. A point was eventually reached when the air pressure within the void overcame the combined strength ('capillarity bond') of the overlying wet sand and the hydrostatic load of the confining column of water; the wet sands sheared along one flank of the deformation structure, and the air escaped to the accompaniment of a quick 'trap-door' rise and fall of the overlying wet sand. This caused further deformation and, as sand may have been dislodged in the process, the 'lid' to the void did not always fall back into place and a small void was therefore retained (Fig. 13). If the internal air pressure again built up, the void became the centre from which a further air-escape deformation process was propagated. If, however, the local supply of air was cut off for any reason, or the air failed to achieve sufficient pressure to form another escape route, such a void became a permanent feature of the dune. Structures of this type were first described from beach sand by Reade (1884), and were also noted by Emery (1945). Stewart (1956) aptly suggests that they should be named 'air-heave' structures.

Apart from the purely mechanical air-induced movements of relatively rigid 'trap doors' (e.g. Fig. 13), the near-surface sand also seemed to be subjected to very local and short-lived fluidisation. When this occurred, the air-escape movements were usually very rapid and often involved the vertical displacement of sand in both upward (positive) and downward (negative) senses, probably as a result of
density instability. Where laminae were subjected to negative dis-
placement, the grains were commonly dilated or even 'swirled' as the
sands were temporarily fluidised, and the marker horizons became
very diffuse (see Figs. 14, 18).

The area of initial deformation took place around the middle of the
'windward' slope just below the up-dip end of the first long ava-
lanche slope. Slowly, a redistribution of sand took place. This
involved the uplift of grains by escaping bubbles of air, with many
sand grains being carried to the surface of the water, and their
resedimentation by fall-out. If the depositional slope was suffi-
ciently great, these 'fall-out' grains could be involved in short-
distance (several centimetres) avalanching. In time, considerable
local changes were made to the 'dune's' surface morphology (cf.
Figs. 15, 17, 18).

After a relatively short initial period of air escape and deforma-
tion in the middle of the 'windward' slope, the area of deformation
activity changed to near the foot of the 'leeward' avalanche slope.
The reason for this change is believed to derive from the creation
of a narrow zone of relatively high permeability when the air-injec-
tion capillary tube was inserted along the floor of the tank. The
initial deformation activity was again dramatic, with air escape
presumably taking place at the point of least resistance, which here
included an almost 30 cm hydrostatic head of water. It caused the
repeated creation and collapse of a large air-filled void, which
eventually remained permanently open (Fig. 16); it also resulted in
the resedimentation of almost un laminated sand on the surface and,
locally within the dune, some sets of laminae were partly to com-
pletely homogenised. On the other hand, the former location of other
laminae seemed to be emphasised by the creation of a few enlarged
air-filled pores (Fig. 16, to left of upper marker overlying the
void).

In the course of time, the site of air escape on the leeward slope
slowly moved stepwise to higher locations (see Fig. 17). This may
have been the result of two opposing effects:
1) Increased resistance to air flow across the deformed laminae of the lower slopes; this may have been caused, in part, by the increased rigidity of the laminae with each stage of deformation, but perhaps also by an increase in the strength of these sands by the creation of a net-work of oversized air-filled pores (Fig. 16) in sands that may earlier have been almost saturated with water, and

2) The creation of a continuous air-filled zone that propagated upwards along the coarser-grained laminae to a point where the internal air pressure overcame the confining surface-tension forces.

Occasionally, especially later during the process of air injection, the whole process of air escape was liable to cease for considerable periods until the internal air pressure was able to overcome the resistance and propagate a new escape route. With the passage of time, deformation activity slowed down unless the air pressure was increased.

Prior to the injection of air into the dune, about one half centimetre of sand at its crest was not covered by water. After some 6 hours of air injection the crest of the dune lay about one centimetre below the surface of the water. This was probably mostly the effect of an overall slight compaction of the sand resulting from interfacial tension effects during the deformation processes.

Within about 2 hours of the beginning of air injection, around one tenth of the dune adjacent to the glass side of the tank had been visibly deformed (Fig. 17); 9 days later, with a very slow rate of air injection, this proportion had increased to about one third and the overall height of the dune had been reduced by 5 cm (Fig. 18). There had been no intervening wave action or seismicity.
THE FOURTH EXPERIMENT

Because of the success of the third experiment, it was decided to conduct a fourth, whose progress could be recorded by video camera.

As has already been mentioned, the sands used in the fourth experiment were collected from the landward side of coastal dunes in The Netherlands, and contained a fairly high proportion of shell material. Those shells that would not pass through a 2.5 x 2.5 mm mesh were discarded; 42% of the sands fell within the grainsize range 250-500 μ. Thus the sands used in the fourth experiment were the coarsest of all.

The 'dune' was built in much the same style as in the third experiment, but with an extra horizon of wind-formed truncation (Fig. 19). In the earlier experiments, the bottom of the tank was covered with a thin 'blanket' of sand prior to dune construction. Because this sand was not laminated, it allowed water to spread rapidly along the base of the artificial dune, thereby increasing its rate of flooding and so reducing the chance of trapping air deep within the dune. This was obviated in the fourth experiment by allowing the sand to avalanche down directly onto the glass floor of the tank so that water tended to climb the fine laminae rather than pass through them. To avoid damaging the natural differences in permeability caused, in the third experiment, by inserting the fine capillary needle into the dune sand, the needle was placed into position before the dune was constructed. The final 'down-wind' length of the dune was 130 cm and its maximum height 31.5 cm.

Flooding proceeded as expected and reached a final water depth of 27.5 cm after 7¼ hours. The first 'faults' at the upper end of the avalanche slope were observed 40 minutes after flooding began, by which time the water in the fine-grained near-surface laminae had already reached a height of 18 cm above the level of free water. A steady stream of small bubbles of air was seen escaping from the submerged sands when the water had an average depth of around 7 cm. After only two hours of flooding (unbalanced water depth 7.2 cm 'down-wind' and 5.8 cm 'upwind') water had extended some 30 cm along
the 'down-wind' near-surface laminae by capillary action, and 29 cm along the 'upwind' laminae, and several pockets of air had been temporarily trapped within the dune adjacent to the glass side of the tank (cf. Fig. 19).

3 hours after flooding began, when the water depth was about 9.6 cm at both ends of the tank and only small pockets of sand remained dry, it was noticed that a row of fine sand grains was adhering to the side of the tank some 2 mm above the current level of the dune, and the overall height of the dune had decreased by a similar amount. This, presumably, was the effect of a shortening of laminae induced by the interfacial tension of the water surrounding the sand grains, combined with compaction caused by the weight of water now carried within the dune. At about the same time, a narrow linear vug (6.5 cm long, 0.5 mm wide) developed just above the topmost marker sand in about the middle of the avalanche slope; the overlying laminae presumably were adhering to the side of the tank rather than subsiding with the rest of the 'dune'. By the time water flooding ceased, the overall dune height had been reduced by some 6 mm.

4½ hours after flooding began, a dome, about 3 cm across, developed on the damp surface of the 'windward' slope just above the level of the free water, which had a depth of 16 cm; the dome fractured at its upslope edge, a few large bubbles of air escaped, and the dome collapsed to about half its former height. Other air-induced uplifts then formed both above and just below the level of free water, with air escaping from most of them. Thus again it was demonstrated on a laboratory scale that the escape of naturally entrapped air could deform the capillary-moistened surface laminae of a dune through which it passed. Uplift of the submerged air-escape structures resulted in small avalanches in those areas where the submarine slope was relatively steep. In more horizontal areas, air escape resulted in a pit surrounded by a small cone of 'fall-out' grains.

There was no obvious deformation of the avalanche slope except near its foot, where a slightly 'ruffled' appearance was interpreted as the result of a little downward creep of the surface sands assisted, probably, by the escape of small bubbles of air.
The artificial injection of air was not started until 22 hours after flooding ceased. The capillary needle and plastic tube leading to it had filled with water during flooding. With an air pressure equivalent to 61.5 cm of water, the water was expelled through the needle at the rate of about 1 cm$^3$/minute. At this rate, there was no deformation of the dune laminae for about one hour, when two linear vugs, one parallel to the dune laminae and the other roughly horizontal, developed up dip from the end of the air-injection needle about 5 cm below the submerged surface of the dune (Fig. 20). Deformation of the surface laminae by air escape was not to take place for another 20 minutes, however.

In the meantime, shortly after seeing the two linear vugs described above, it was realised that a fairly large submarine avalanche had occurred on the 'down-wind' slope. Its length was about 47 cm and its width around 16 cm. Because the sand was clean, the avalanche did not flow across the floor of the tank but built a small 'toe' at the foot of the dune. It had probably been triggered at about the same time as the creation of the two vugs described above.

As with the third experiment, once deformation of the surface sands began it became a fairly regular process, with air continuing to escape spasmodically from one spot for several minutes, and then all activity would switch to another location, as often as not in the middle of the dune, away from the glass sides of the tank. It proved difficult to maintain a constant air pressure, which fluctuated between about 60 and 90 cm of water.

After some 18 hours of air injection, the maximum height of the dune was 30.8 cm. 24 hours later, it was 30.4 cm. Because air escaped through the damp exposed part of the dune as well as through submerged areas, the former became highly pitted, and after 6 days, when the supply of air was disconnected, the average height of the dune had reduced to 27.8 cm (reduction of 3.7 cm) with a considerable volume of surface and expelled sand being redistributed down the 'upwind' slope). Deformation of the marker horizons showed how the internal dune laminae became homogenised by the ever-changing locations of air-filled vugs (Fig. 21, 22).
Eventually a fan of deformation spread through the dune from the vicinity of the air-injection needle, whose limits adjacent to the glass side of the tank intersected the laminae at the top of the avalanche slope and extended up dip from just beyond the end of the injection needle. The spread of deformation seen through the back side of the tank was about the same as at the front, thus indicating the good permeability of the dune sands parallel to their laminae.

One further item of interest was the generation of a second avalanche from about the middle of the first. From the shape of the ensuing surface relief, this second avalanche was almost certainly triggered by the uplift of sand associated with the escape of air through the 'down-wind' surface laminae. This seemed to be the only point at which air broke through these laminae of the 'leeward' slope.

MAIN EXPERIMENTAL RESULTS

The aims of the experiments, as outlined earlier, were all achieved.

1) There is a reduction in the angle of surface slope when aeolian slip-faces are rapidly submerged beneath water, but the amount of reduction is small (1-2°) and seems to be greater in coarser than in finer sands. In coarser sands, the initial slope can be reduced still further locally by submarine avalanching.

2) Observation of the hydrodynamic movement of water across artificially constructed slip-face laminae in the first experiment led to a much greater capillary differential in the later experiments by the greater use of some finer grain sizes and narrower grain-size ranges. This led directly to:

3) It was observed that pockets of air could be trapped within an aeolian dune provided the grain-size contrasts between laminae were great enough, that the bedding geometry was suitably complex, and that the level of free water outside the dune was rising at a rate in the order of decimetres rather than metres or millimetres per day.
Naturally-induced soft-sediment deformation was achieved in artificially constructed dunes; this was most readily obtained (even though still on a small scale) with the coarsest dune sands. The results of these experiments exceeded their limited aims in as much as observation indicated other effects of importance.

The seal that permits the entrapment of air in clean sand is a product of the higher capillary pressure at the water/air interface of the grains with smaller rather than larger diameter. The role of capillary pressure in the entrapment of migrating oil has been clearly described by Berg (1975) and Schowalter (1979); the same principles apply to the entrapment of air, and are illustrated in Fig. 23.

The capillary-pressure seal will be strengthened by the mechanical rigidity resulting from alternations of dry and water-wet laminations of differing grain-size (Fig. 24). The strength of the seal will be progressively destroyed as air escapes or is driven from the larger pores and is replaced by water.

Air pressure should build up within the submerged dune in the proportion of one atmosphere for every 10 m column of air-filled sand. Air will escape from a zone of higher pressure to one of lower pressure as soon as the pressure difference is sufficient to rupture the intervening permeability barrier. This results in the creation of relatively small fluid-escape structures and in minor homogenisation of sands in the core of a dune, and in larger escape structures especially in the surface-parallel laminae of the windward slopes.

In the third and fourth experiments, homogenisation of the sands occurred locally within the dune in response to the creation of air-filled vugs. This was especially the case when the vug, or irregular line of vugs sub-parallel to the former bedding, kept changing its location so that adjacent sand moved between overlying and underlying it, thereby homogenising a swath of laminae.
9) The whole deformation process involves a redistribution of sand both internally and externally, including some expulsion, which results in an overall lowering of the dune's crest line, and in a lessening of parts of the surface slopes.

10) The capillary forces engendered by the advance of water along the fine-grained sand laminae of the avalanche slope are sufficiently strong to cause a shortening of the laminae and associated slippage (shearing) along the coarser-grained air-filled laminae. This probably results in an overall tighter packing of the sand and a slight reduction in the height of the dune.

11) Soft-sediment deformation was not induced by wave action and, although a little warping of laminae did result from seismic activity, it bore no similarity to deformation caused by the escape of air.

POSSIBLE LIMITATIONS IN APPLICATION OF THE EXPERIMENTS

1) The artificial construction of the 'dunes' is believed to have resulted in bedding that is reasonably close to that found in nature. In nature, however, the natural sorting produced on the larger avalanche slopes is likely to be much better, with a sharper distinction between different grain sizes, and the internal bedding geometry will probably be more complex, especially with larger dunes. Furthermore, within the confines of a small tank, it was not easy to produce an even distribution of wind-blown laminae on the 'windward' slope. Thus, with the better sorting available in nature, their capillary-induced surface seals are likely to be more efficient than those formed in the tank.

2) In the third experiment, insertion of the air-injection tube greatly increased the natural cross-lamination permeability of the sands in its vicinity and permitted the back-flow of air towards the base of the 'leeward' avalanche sands, which then developed very clear deformation structures. The probability
that this would not occur in nature seemed to be confirmed by the fourth experiment, where air escaped through the middle of the 'down-wind' slope only at one point, and then only after several days of air injection; because of the build-up of internal pressure, however, air might escape through the upper avalanche slopes of larger dunes if the internal bedding geometry so dictates. This would seem to be implied by a comparison of Figures 21 and 22, above the arrow at 85 cm.

3) Although homogenisation of the dune sands was seen to take place as air-filled vugs, or lines of vugs, changed their locations, this process cannot explain the origin of dune laminae whose sharpness of definition has only been weakened rather than destroyed (cf. Glennie and Buller, 1983, Fig. 3 8583'). Weakening of the laminae possibly occurs when there is a rapid rise of water through undercompacted sands to replace large volumes of rapidly escaping air. Individual sand grains may have been lifted slightly by the passing water (possibly less than a grain diameter) so that the laminae look as if they have been dilated slightly.

4) It might be argued that these tank experiments to create soft-sediment deformation structures must be invalid because scaling factors were ignored, especially when the Rotliegend dunes that inspired the experiments were believed to be some 50 m high. To the contrary, any attempt to scale down the grain sizes involved would have prevented the experiments from working because of the vastly different capillary forces associated with finer grained sediments. These forces increase rather than decrease with decreasing grain size. Thus the experimental results are valid only for the grain sizes and grain-size ranges actually used. The observed differential flow of water along sand laminae was just as it would occur in nature, so that the preference for air to escape through a capillary seal composed of coarser rather than finer sands has a direct application to much larger dunes composed of sands of the same grain diameters.
What is remarkable, perhaps, is that the small volume of air that could be trapped in such a little dune was sufficient to break through the overlying capillarity seal of wetted sand. It implies that limited deformation could occur in other natural sand structures of the same approximate size provided the grain-size ranges and sealing capacities were about the same. The only major difference with nature is in the amount of air that could be trapped in larger dunes. In larger dunes, the greater volume and potentially higher pressure of entrapped air would be capable of deforming proportionally greater thicknesses of damp dune sand. The bulk of the trapped air in a large dune, however, would probably escape in one big 'blow' rather than being spread over several days as in the experiments. This leads to the probability that the biggest deformation structures, such as the 20-m high fluid-escape structure depicted by Glennie and Buller (1983, Fig. 11), probably also result from one major escape of air rather than from a long drawn-out dribble; and this probably also applies to many of the smaller sized structures. Repeated escapes will occur in some areas, however, as the air in separate pockets within the dune works its way to the surface.

Most of the experimental deformation resulted from the slow, repeated escape of bubbles of air over a considerable period of time. The most striking deformation, however, the creation of the first structure in any particular part of the 'dune's' surface, bore the closest resemblance to the structures seen in nature.

SOME INFERENCES DERIVED FROM THE EXPERIMENTS

The experiments indicate that soft-sediment deformation is likely to occur whenever clean, dry, and especially relatively coarse unconsolidated sands are rapidly inundated by water. Glennie and Buller (1983) have shown that in the Permian, the Weissliegend deformation must have occurred during the very rapid transgression of the Rotliegend desert basin by the Zechstein Sea. In a desert environ-
ment, flash floods will cause the entrapment of air in sands of both aeolian and fluvial (wadi) origin. The ensuing slow upward escape of the air possibly causing dilation and homogenisation of the sands through which it passes; possible ancient examples of such repeated events can be seen in cores of mixed fluvial and aeolian Rotliegend sequences from the area of the southern North Sea (Glennie, 1972, Fig. 10). A modern example of large-scale deformation of dune sand will possibly be found in those dunes that were submerged beneath the floodwaters of a temporary lake in the Djofra Graben, Libya late in 1963 (see Glennie, 1970; Fig. 49); they are in the correct environment for such an event to have taken place.

De Boer (1979) has demonstrated that convolute deformation can occur in response to the repeated entrapment below, and escape of air through, the tidally-wetted estuarine sands of the Netherlands coast. The creation of soft-sediment deformation by repeated fluctuation of the water table has been proposed by Doe and Dott (1980). Although not tested experimentally, the writer suggests that such fluctuations would induce deformation only if they also involve the entrapment of air, or if especially the upward movement of water is rapid enough to dilate and fluidise the sands through which it passes. In most cases, however, such fluctuations are unlikely to be rapid enough to induce homogenisation, let alone a stronger deformation.

With normal marine transgressions of long duration, or when lakes slowly encroach onto an area of formerly adjacent sand dunes over a period of hundreds, if not thousands, of years, soft-sediment deformation of the type described above is unlikely to occur. This is because different processes come into play. With very slow transgressions, the capillary rise of water is so slow that no air is entrapped. More important, however, is the probability that over such a long time span, storms, or even normal wave action, will result in considerable reworking, and possibly in peneplanation, of the dune sands. Such a situation has been described by Tanner (1970), Kocurek (1981) and by Eschner and Kocurek (1982) for the Late Jurassic Entrada Sandstone of Utah and Colorado.
Finally, apart from their academic interest, these experiments also illustrate the well-understood process of water invasion of a gas-filled reservoir of aeolian sandstone.

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REFERENCES


Fig. 1 Conceptual model of soft-sediment deformation in Weissliegend dune sand by air-escape following rapid inundation by Zechstein Sea.
Fig. 2  Sketch illustrating method of construction of the 'dune' by depositing sand on an avalanche slope
Fig. 3. Rising capillary front of wet sand in artificially-constructed dune. Free water rising at 22 cm/hour. Expt. 1.

Fig. 4. Water flooding ceased; capillary front now almost horizontal and some 7.5 cm above level of free water, and submarine slope now about 30½°. Expt. 1.

Fig. 5. Gentle wave action results in creation of beach. Submarine slope now averages about 28°, and exposed sands slump more with each soaking by waves. Expt. 1.

Fig. 6. As result of 'seismic' activity, middle marker lamina is now convex-upward and the sands exposed above water level have slumped. The upper submarine slope is now about 25°. Expt. 1.
6. Entrapment of pocket of air lead sand near Lv 0 hours after flooding began. Level of free water still 17 cm below top of dune. Expt. 2

Fig. 7. Marked differential capillary effect picked out by serrated pattern of wet sand. Note capillary-induced 'faults'. Nearly 3 hours after flooding began. Expt. 2

Fig. 8. Entrapment of pocket of air-filled sand nearly complete about 6 hours after flooding began. Level of free water still 17 cm below top of dune. Expt. 2.

Fig. 9. A more realistic 'dune' less than one hour after flooding began. 'Leeward' avalanche slope to left. Expt. 3.

Fig. 10. Pockets of air-filled 'dune' sand already trapped beneath capillary-wetted layer 2½ hours after flooding began. Expt. 3.
Fig. 15.
Lobes of avalanche sand up to 10 cm long with relics of circular 'volcanoes' at upper ends. 'Leeward' slope. Expt. 3.

Fig. 16. Large air-filled vug at foot of 'leeward' slope. This vug remained until end of experiment. Note row of enlarged air-filled pores above highest marker lamina. Expt. 3.

Fig. 17. Two hours after air-injection began. The only natural air-induced deformation is visible as small surface perturbation (arrow) and small vug. Deformation of 'leeward' slope has progressed up from large vug. Note end point of injection needle. Expt. 3.

Fig. 18. Deformation after 9 days of air injection. The large vug of Fig. 17 2/3 way up 'leeward' slope had long been replaced by another that is slightly deeper. Expt. 3.
Fig. 19. Differential movement of capillary water, location of horizons with better sealing capacity, and locally trapped pockets of air 135 minutes after flooding began. Expt. 4.

Fig. 20. Location of two linear vugs updip from the nozzle of the air-injection needle (level with 85 cm on scale). The first surface deformation occurred 20 minutes later in line with the right-hand vug. Note that the coarser sands fine updip. Expt. 4.

Fig. 21. The process of homogenisation of sands by air escape near the surface, and by repeated relocation of air-filled vugs in the core of the dune can be imagined from this figure. As with the Weissspiegel of the North Sea area, a vertical well through the crest of the dune would pass in and out of deformed, homogenised and undeformed dune sand. 18 hours after air injection began. Expt. 4.

Fig. 22. A fan of deformed and homogenised dune sand spreading out from just 'upwind' of the end of the air-injection needle (arrow). A similar fan developed against the back side of the tank indicating good internal permeability. In the centre of the dune, however, deformation is largely confined to the top 25%, which suggests that there, the surface seal is better than against the glass sides of the tank. Note the patch of almost undeformed sand. Compare with Fig. 1. After six days of air injection, and six days after water drained from tank. Expt. 4.
A wetted lamina of fine-grained sand is held together by the surface tension of the water. Air can be trapped below this lamina as the capillary pressure in these pore throats are relatively high, normally increasing with decreasing grain diameter. Air will escape only if the internal air pressure is greater than the capillary pressure bonding these fine grains.

Capillary Pressure $P_c = \frac{2 \sigma \cos \phi}{r_t}$

where

$\sigma$ = surface tension, which for water/air interface = 72 dynes/cm at laboratory conditions

$r_t$ = diameter of pore throat = $0.154 \times R_2$ for perfect spheres

and in our case, we assume that the sand grains are completely water wet, so that $\phi = 0$ and $\cos \phi = 1$.

Thus for sand of 0.1 mm (100 $\mu$) diameter,

$$P_c = \frac{2 \times 72 \times 10}{0.154 \times 0.05} \text{ dynes/cm}^2 = \frac{1440}{0.0077} \text{ pascals} = 1.8457 \text{ atmospheres}$$

To overcome this capillary pressure, the buoyancy of a column of air is needed that is at least 18.5 m high assuming that the lamina remains rigid. Since capillary pressure is inversely proportional to grain diameter, the following tabulation shows the pressures needed to break the capillary seal for laminae of differing uniform grain diameter.

<table>
<thead>
<tr>
<th>Grain diameter ($\mu$)</th>
<th>Air pressure (atmospheres)</th>
</tr>
</thead>
<tbody>
<tr>
<td>25</td>
<td>7.4</td>
</tr>
<tr>
<td>50</td>
<td>3.7</td>
</tr>
<tr>
<td>100</td>
<td>1.85</td>
</tr>
<tr>
<td>200</td>
<td>0.925</td>
</tr>
</tbody>
</table>

Obviously the rigidity of the system will be upset if the pressure due to the weight of the sand grains themselves is less than the air pressure, in which case the sand grains will be lifted out of the way permitting the air to escape.
Fig. 24 - Idealised diagram of alternating dry and water-wet laminae of different uniform grainsize. The barrier formed by the one fine-grained water-wet lamina could, in practice, be strengthened by the mechanical rigidity derived from alternations of dry and water-wet laminae of differing grainsize. In nature, each water-wet lamina is probably at least 2-3 grains thick.