RECENT CARBONATE SEDIMENTATION NEAR ABU DHABI,
TRUCIAL COAST, PERSIAN GULF

by

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Abstract

The area studied comprises a series of lagoons, lying along the southern margin of the Trucial Coast embayment, which is a broad, shelf sea. The lagoons are separated from the shelf by a complex barrier of islands, coral reefs and shoal banks. The entire area is one of carbonate sedimentation. Air and sea temperatures are high and rainfall low; high net evaporation produces saline marine conditions (42-65‰ S.). Shelf sediments are dominantly a skeletal (mainly molluscan) facies; lagoon barrier sediments comprise coral reef and oolite facies; lagoon sediments are largely aragonite muds, variously pelleted; stromatolitic algal sediments are abundant in innermost lagoon areas Organism communities have been studied and new ecological data derived concerning communities in saline environments. Observational studies, together with chemical and mineralogical analyses, of fauna, flora and sediments, have delimited the organic contribution to the sediments. Published synthetic data on the co-precipitation of strontium with aragonite has enabled a direct chemical origin to be ascribed to at least 80-90% of the lagoonal sediments. Sea water analyses show a relative loss of calcium to occur from open shelf to innermost lagoon areas, most rapid loss taking place in the turbulent barrier area.

The lagoons are being infilled by a prograding series of sediments, lying just above normal high water and in places more than 10 miles (16 kms.) wide. Evaporation

losses from the upper sediment surface are balanced by inward movement of lagoon waters. Ground waters, at an average temperature of $34^{o}C$, lie at depths of 1-4 feet. As the brines move inward they become concentrated and gypsum first precipitates, raising the m_{Mg}/m_{Ca} ratio to values in excess of 11. Dolomitisation of the original lagoonal sediments then begins and very locally magnesite is developed. Further precipitation of calcium sulphate is mainly as anhydrite - the first recorded natural occurrence of Recent anhydrite. The final brines have a m_{Mg}/m_{Ca} ratio of less than 2 and contain negligible amounts of sulphate ion. Celestite is fairly abundant and results partly from direct precipitation from concentrated brines but mainly as a by-product of the diagenetic reactions.

The sedimentary and diagenetic processes described are thought to have operated in many past back-reef and lagoonal environments.

CONTENTS

CHAPTER I

INTRODUCTION AND GENERAL DESCRIPTION OF THE REGION

Introduction:

The overall research project to study the Recent sediments of the Persian Gulf commenced in September 1961, when G. Evans and the author first visited the region (Figure 2) in H.M.S. Dalrymple. Sediment samples were collected using Dietz-Lafonde grabs and gravity and piston corers. Eight stations were occupied in the axial areas and sediment samples obtained. Later in the season, eleven more deep water samples were collected by the author on a return trip from Kuwait to Abu Dhabi, and at the end of the season, in February 1962, the author collected a further thirteen shelf samples from the Great Pearl Bank, in the vicinity of Jazirat Sir Abu Nu'air. During part of the latter period Evans was engaged in collecting shelf samples in the small vessel Ghazal, on loan to the project from the Iraq Petroleum Company.

The centre for the nearshore studies was the island of Halat al Bahrani (Figure 3), where the Royal Navy had established a camp, from which to survey the sea approaches to the town of Abu Dhabi and the Oil Company

Fig. 1 Motorised launch used for nearshore and lagoon studies. Length 18 ft. (6 ms.); draught 21 ins. (50 cms.); stern mounted derrick with hand operated winch used in conjunction with Dietz-Lafonde grabs or small gravity corers; echo-sounder amidships, capable of recording to within 1 ft. (30 cms.).

jetty at Ras al Dhabayah. The author worked for two months from this camp, studying the area near Halat al Bahrani. Most of the open sea work was completed during a three week period with G. Evans. The sampling was carried out from an 18 ft. (6 ms.) motorised launch of shallow draught, fitted with a small derrick and echo-sounder (Figure 1).

When opportunities arose, the author accompanied naval survey parties to other parts of the coast. In this way, reconnaissance and sampling trips were made to Ras Ghanadha, Ras al Dhabayah and Jazirat Abu al Abayadh. Land reconnaissance was carried out on the Dhabayah Peninsula and along the coastal plain and mainland hills between Tarif and Abu Dhabi causeway. A week was also spent on the island of Jazirat Sir Abu Nu'air; a general examination was made of this salt plug, particular attention was directed to the Recent calcarenitic and reef limestones which form a fringing skirt to the island.

The division of labour within the project resulted in Evans studying the deep water and shelf sediments collected during the first field season. The author concentrated on and extended his studies in the nearshore area south-west of Abu Dhabi.

The second field season saw an expanded research team visiting and working in the region. On this occasion the author and his wife spent five and a half months in the field, being joined for the initial two to three weeks by D. J. Shearman. Field sampling studies were extended over the entire near-shore and lagoon areas between Abu Dhabi island and Ras al Kahf. An extensive study was also made of the islands, and coastal plain and hills of the mainland. Further reconnaissance trips were made beyond the confines of the Abu Dhabi area; to the west beyond Mirfa, inland into the desert dune sand areas and north-east to the Sheikdomsof Dubai, Sharja, Ajman and Fujara, and also to Muscat and Oman.

The field studies comprised sediment and water sampling together with collection of any other specimens and data which were considered to be critical in the evolution of the depositional and early diagenetic developments of the area. Sample stations were fixed by dead reckoning and by boat to shore triangulation using sextant or compass. Stations were plotted in the field on tracings from aerial photographs and on maps drawn from Naval charts and the interpretation of aerial photographs. Observations of the bottom were made using face masks or a glass-bottomed inspection box. As the waters are clear and shallow, submarine features are shown extraordinarily well on the aerial photographs. These features were directly examined during subsequent traversing of the area in the launch.

The field observations and sampling have enabled the nearshore island and lagoon complex to be further subdivided into a series of physiographic units which are paralleled also by differences in organism communities and sedimentary facies. For example, the oolite deltas are well marked sedimentary and physiographic features and are traversed by one or more tidal channels. Offshore, below 18-20 ft. (6 ms.), the shelf sediments are largely of skeletal and detrital origin. Coral reefs occur in both offshore, interdelta positions and behind the oolite deltas. The occurrence of coral reefs is outstanding in view of the high temperatures and salinities suffered by the region, in fact the known temperature and salinity limits of reef coral growth have been extended by this study (Kinsman, 1964). Behind the inner reefs and oolite deltas are wide areas of pellet sands and aragonite sands, which extend to the mainland coast. In some inner lagoon areas tidal swamps and creeks and extensive spreads of intertidal algal mats occur.

The inter-relationships of the organism communities and sedimentary facies were determined, and of outstanding interest is the complexity of the sedimentary surface, at any one moment of time, within this nearshore region of islands and shallow lagoons. The sediment distribution is largely controlled by turbulence of the waters. This is related in part to depth but also to water circulation patterns and to the degree of shelter of an area from the prevailing north-westerly winds. The depth control can be related to the tidal effects within the lagoons and in this context the diminution of tidal range with distance back from the open sea has been considered in some detail.

Laboratory studies included examination of loose sediments, thin section analysis of impregnated sediments and rocks, and X-ray and chemical determinations. These have confirmed and expanded the picture gained from the field work. Different sediment types were established morphologically and these were found to show varying carbonate mineralogies and different chemical compositions. The sediments were analysed for magnesium and strontium and these parameters used as additional evidence regarding the sediment origin. Laboratory data on the co-precipitation of strontium in aragonite have enabled a direct chemical origin to be ascribed to the major portion of the lagoonal sediments. Sea water analyses of samples from the open shelf to the innermost parts of the lagoons show a progressive loss of calcium to occur which can be equated partly with the production of organic carbonate materials but mainly with the development of large amounts of chemically precipitated carbonates which take the form of oolitic encrustation of grains and aragonite needle muds.

The aragonitic and calcitic post-Miocene limestones have enabled early diagenetic studies to be made, studies of early cementation and recrystallisation, of loss of the less stable carbonate components and development of early replacement textures.

The most interesting diagenetic developments are those found in the unconsolidated sediments of the coastal plain and similar areas on some of the islands.

Here, the original sediments were usually pellet sands and aragonite muds, but as a result of precipitation from concentrated sea-water in the pore-spaces and of reaction between the interstitial brines and the early sediments, a series of interesting and important developments were found, some of which are tabulated below:-

1. Gypsum : developed mainly by precipitation from concentrated interstitial brines, but also by reaction between brines and the early sediments. Late stage gypsum is also developed after anhydrite.

2. Anhydrite : the first recorded occurrence of Recent anhydrite: thus a critical field study can be made, for the first time, of the conditions controlling its formation. Morphologically the anhydrite is seen to be a direct precipitate from solution, there being no evidence of the pseudomorphing of gypsum. Chemical studies of the interstitial brines indicate, however, that the large amounts of anhydrite which are developed can only reasonably be accounted for by solution of early gypsum and reprecipitation of the calcium and sulphate ions as anhydrite.

3. Dolomitisation : the largely aragonitic sediments are being fairly rapidly dolomitised. The process is a replacement phenomenon, almost certainly one of solution and reprecipitation. The process is best defined as "early diagenetic, pre-lithification dolomitisation".

4. Celestite : developed in appreciable

quantities, following replacement of the relatively high strontium-bearing aragonitic sediments by a relatively low strontium-bearing mineral assemblage of gypsum, anhydrite and dolomite. Some celestite is also precipitated from brines.

5. Magnesite : developed locally in the high magnesium environment which follows the major gypsum formation yet precedes the massive removal of magnesium by dolomitisation.

6. Huntite : ($Mg_3Ca(CO_3)_4$) : an uncommon carbonate mineral; this evaporitic development is unique among its recorded occurrences.

The mineralogical developments can be closely followed by analysis of the pore fluids. For example, development of gypsum is paralleled by a loss of CALCIUM ~~clacium~~ and sulphate ions and dolomitisation is paralleled by loss of magnesium ions from the brines. The combination of this dual approach has enabled the evolution of the diagenetic changes to be unravelled in some detail.

Glossary:

A few Arabic words are used throughout this thesis and their meaning is given below:-

Abu - father; possessor of.

Al or El - the; the definite article

Halat - dry or drying sand bank; a sandy islet

Jabal or Jebel - hill or mountain

Jazirat or Jezirat - island or peninsula

Khor - arm of the sea, inlet or channel

Ras - headland, cape or point

Sabkha - salt flat; low sandy or muddy areas subject to occasional flooding. (Sebkha has been used recently by Illing (1964).)

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Regional Geography:

The Persian Gulf is a shallow sea, nowhere deeper than about 300 feet (92 ms.), connecting via the Strait of Hormuz with the Gulf of Oman and Indian Ocean. The Gulf is asymetric in cross section, the deep water axis lying fairly close to the Persian shore and sloping gradually but slightly irregularly to the deeper waters of the Gulf of Oman, there being no sill or threshold at the Strait of Hormuz (Fig. 2).

The north-eastern or Persian shore is steep and generally mountainous; inland lie the Zagros Mountains. A series of islands lies along this shore, some of them having a tectonic origin, several being intrusive diapiric structures.

At the head or north-western end of the Persian Gulf are the extensive deltas of the Tigris, Euphrates and Karun rivers. These are the only rivers to flow into the Gulf at all seasons of the year.

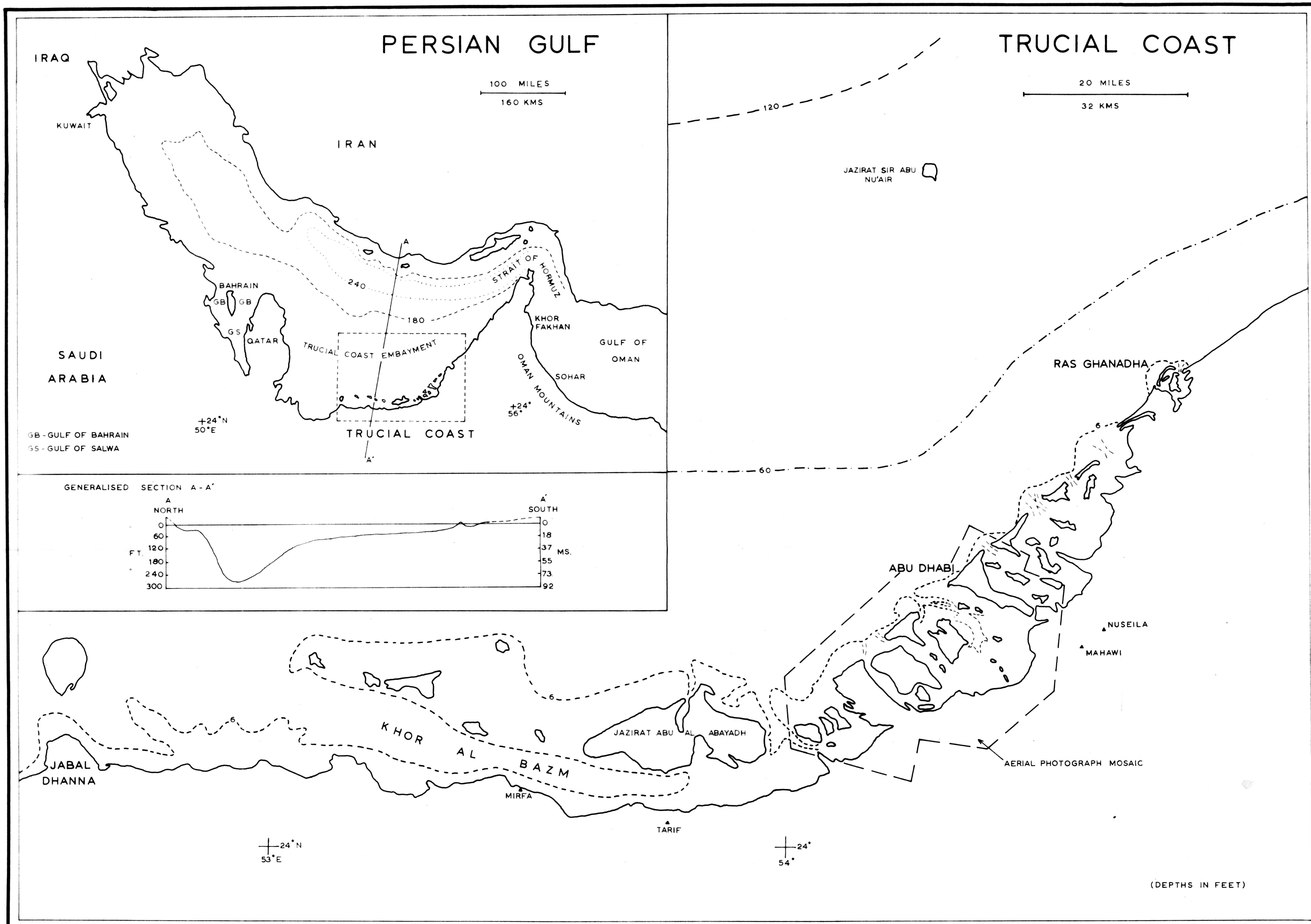
The south-western shore is generally low and inland extend the great deserts of Arabia. Offshore, the water is everywhere shallow and shoals and islands abound. The Trucial Coast embayment, lying to the east of the Qatar Peninsula, has water depths of less than 120 feet (37 ms.) and an abundance of small islands and shoals, some of which owe their origin to diapiric intrusion. The eastern shore of the embayment is flanked by the Oman Mountains and near-shore water depths increase rapidly as the Strait of Hormuz is approached.

Fig. 2

Persian Gulf: showing broad bathymetry and surrounding areas.

Profile A-A'; indicates the asymmetry of the Persian Gulf and broad extent of the shallow Trucial Coast embayment, with lagoons developed along the southern shore.

Trucial Coast: to show the nature of the island and lagoon complex of the southern shore of the embayment. The western part is dominated by Khor al Bazm, a long lagoon running parallel the coast; the eastern part comprises smaller more restricted lagoons. Position of aerial photograph mosaic marks area of detailed study.



PERSIAN GULF
100 MILES
160 KMS
IRAQ
KUWAIT
IRAN
BAHRAIN
GB
GB
GS
QATAR
SAUDI ARABIA
240
180
A
A'
STRAIT OF HORMUZ
KHOR FAKHAN
GULF OF OMAN
SOHAR
OMAN MOUNTAINS
TRUCIAL COAST EMBAYMENT
TRUCIAL COAST
+24°N 50°E
+24° 56°
GB - GULF OF BAHRAIN
GS - GULF OF SALWA
GENERALISED SECTION A - A'
A NORTH
A' SOUTH
FT.
MS.
0
60
120
180
240
300
0
18
37
55
73
92
TRUCIAL COAST
20 MILES
32 KMS
120
60
6
JAZIRAT SIR ABU NU'AIR
RAS GHANADHA
ABU DHABI
NUSEILA
MAHAWI
AERIAL PHOTOGRAPH MOSAIC
JAZIRAT ABU AL ABAYADH
KHOR AL BAZM
JABAL DHANNA
MIRFA
TARIF
24°N 53°E
24° 54°
(DEPTHS IN FEET)

Geological History:

The Persian Gulf region has been the site of extensive sedimentation since Palaeozoic times, great thicknesses of sediments having been laid down in the axial regions. Of particular interest, in the basin margin or shelf areas, is the recurrent development of the calcareous and evaporitic facies, indicating that similar depositional environments were developed many times. For example, on the Arabian shore, anhydrite occurs in sediments and rocks of Devonian, Permian, Jurassic, Cretaceous, Eocene, Oligocene, Miocene and Recent age.

The Persian and Arabian shores differ markedly in their tectonic development. Inland of the Persian shore lie low mountains of gently folded sedimentary rocks, which pass into the complex fold and thrust belt of the Zagros Mountains. The Arabian shore is much less disturbed, folding being gentle and thrusting absent. The Oman Mountains are an exception, however, several lines of evidence pointing to them having been extremely mobile in the recent past.

Diapiric structures are widely developed, being recorded on land in the Persian Mountains, forming islands along the Persian shore and in the Trucial Coast embayment, and as isolated hills along the Trucial Coast.

The effect of the Pleistocene lowering of sea-level is of particular interest. Some geologists have expressed the opinion that the floor of the Persian

Gulf would have been largely exposed when sea-level fell to its maximum extent, this opinion being based on the present-day bathymetry. An alternative suggestion is that fairly rapid infilling has been taking place, depths would therefore have been relatively greater than at present, and the floor not exposed. Until longer cores are available from the axial areas this problem cannot be solved.

Along the Trucial Coast there is abundant evidence of fairly recent fluctuations in sea-level. Beaches and marine erosion surfaces are found in various elevated positions in the Persian and Oman Mountains, but these regions are known to have been tectonically active. During the same period the Trucial Coast would seem to have been relatively stable.

Brief Review of Previous Work:

The first study of the Recent sediments of the Persian Gulf was made by Emery (1956). Temperature, salinity and current measurements were made and their vertical and horizontal distributions determined. Water temperatures in excess of 32°C were recorded for the entire region and water salinity was found to be generally greater than 38‰ . Salinity and current measurements in depth showed more saline, denser Persian Gulf water to sink and flow out through the Strait of Hormz beneath the less saline, incoming Indian Ocean waters. Some sediment samples were collected and a sediment distribution map compiled with the additional aid of about 3,600 bottom notations derived from navigational charts. The sediments were found to be dominantly calcareous, with appreciable amounts of terrigenous detritus occurring only round the Tigris-Euphrates and Karun deltas at the head of the Gulf and along the Persian shore. The finest sediments were found to occupy the deeper water areas against the Persian shore, coral and more sandy sediments occurring in the shallower waters near the Arabian shore.

Houbolt (1958) has made a detailed study of the off-shore sediments to the north and east of the Qatar Peninsula. The shallow water sediments were found to be essentially skeletal calcarenites, dominantly molluscan in origin, which passed laterally into calcilutites towards the deep water axis. The energy balance of waves and currents was shown to be responsible for the distribution of the calcilutite grades of the sediments.

Two coastal lagoons on the Saudi Arabian mainland, due west of Bahrain, have been examined by Bramkamp and Powers (1955). They reported the deposition of calcium carbonate and gypsum-rich sediments from highly concentrated brines entrapped in the shallow lagoons.

Sugden (1963 a & b) has made a study of the sediments of the Gulfs of Bahrain and Salwa and parts of the Trucial Coast embayment, finding a widespread development of aragonite muds, oolites and pseudo-oolites. He considered in detail the problems of formation of aragonitic oolites. A regional survey was also made of the variation in salinity, with particular reference to the remarkably high values occuring in the Gulf of Salwa and along the Trucial Coast, where salinities as much as 60% above that of normal ocean water occur. Sugden considered the high salinities to have had little effect on the diversity and abundance of the marine fauna. The distribution of the high salinity waters was also considered with respect to evaporite deposition.

Wells and Illing (Wells 1962), working in the coastal lagoons and intertidal areas around the Qatar Peninsula, have recorded the occurrence of dlomite and gypsum within the Recent carbonate sediments. More recently (1963) they have described the precipitation of calcium carbonate in the Trucial Coast embayment. The precipitation takes the form of isolated patches of milky water, or "whitings" and is thought to be triggered off by the sporadic, abnormally rapid development of "blooms" or high populations of diatoms in the surface waters. The authors consider this process could well be important in the large-scale formation of fine-grained carbonate sediments.

Description of Trucial Coast Embayment - Trucial Coast:

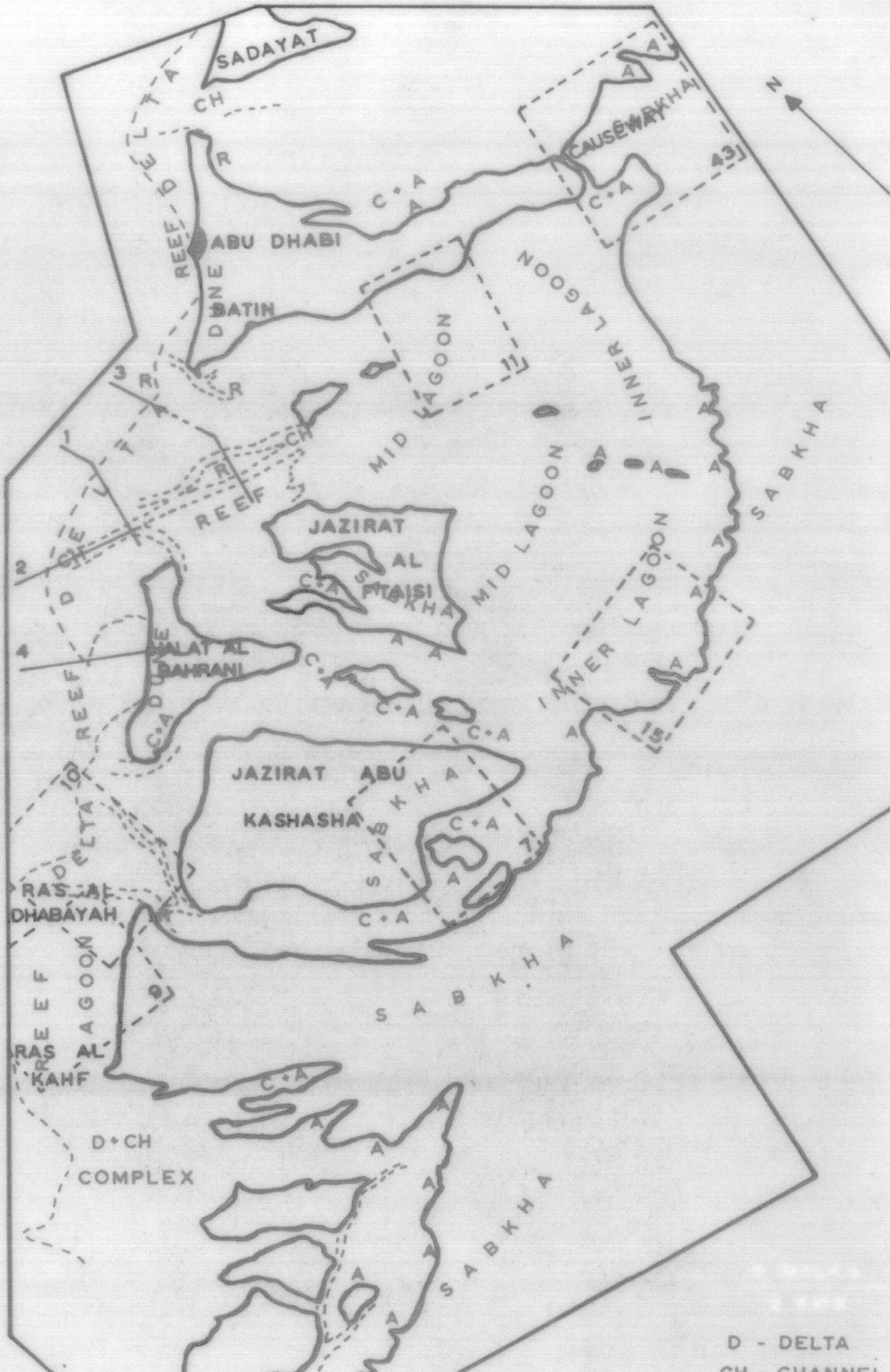
General :

The Trucial Coast embayment is a broad shelf sea with depths rarely exceeding 120 feet (37 ms.). The western and southern shores are of low relief; islands and shoals abound and marine waters everywhere are shallow. Eastwards the land relief rises, offshore islands are absent and nearshore water depths comparatively greater (Fig. 2).

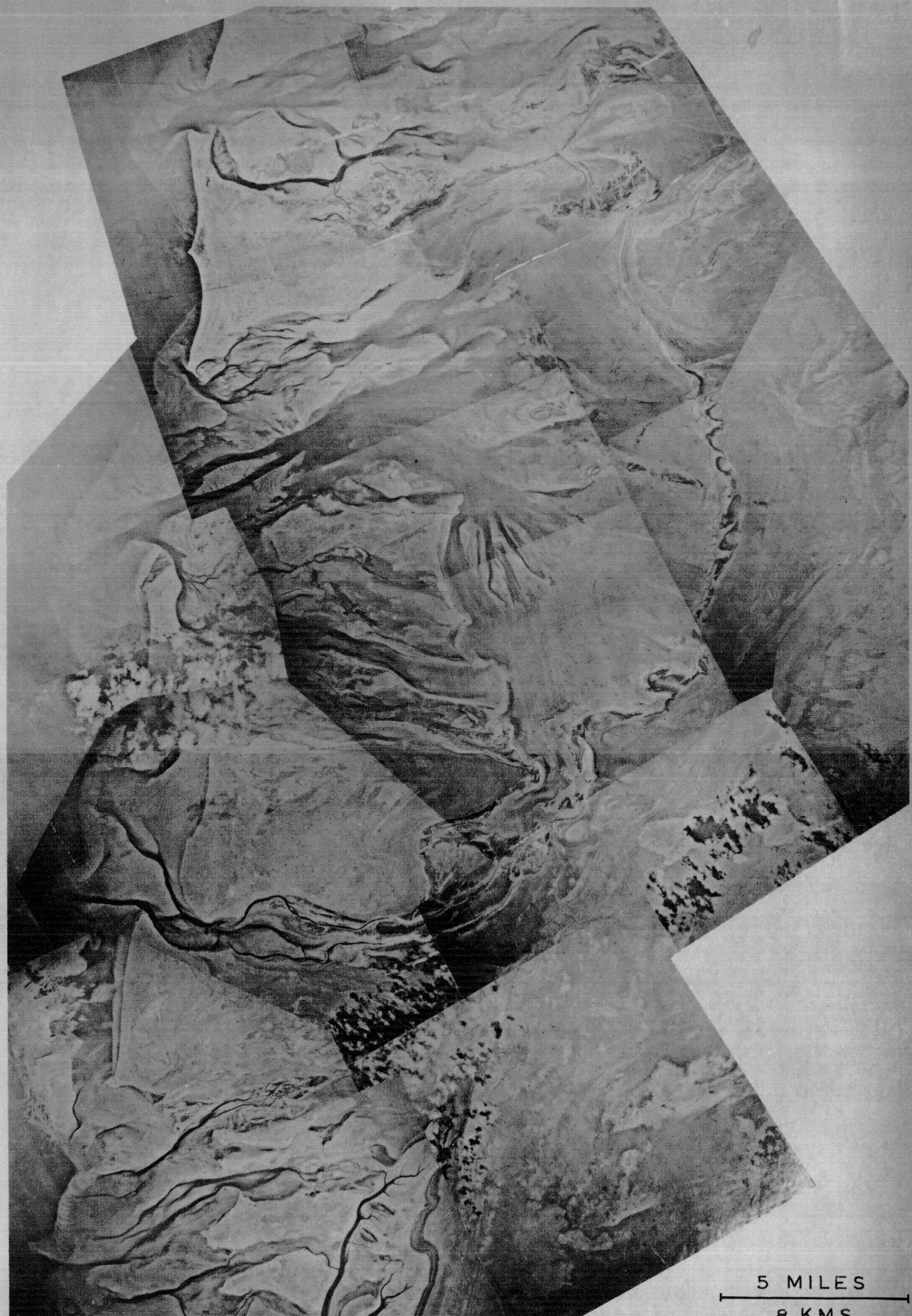
In the north-western parts of the embayment Houbolt (1958) has described a series of offshore terraces which he has related to post-glacial rises in sea-level. Detailed study of available Admiralty soundings would cast doubt on these being well marked features; if indeed, they are developed at all. The general bottom topography would seem to be rather irregular, local surface relief in depths of 120 feet (37 ms.) being as great as 40 feet (12 ms.). A gradual, if irregular, shallowing occurs southwards towards the Trucial Coast shore and at distances of 20-25 miles (32-40 kms.) depths rarely exceed 60 feet (18 ms.).

The form of the southern shore is shown in Fig. 2. The nearshore island and lagoon complex may be conveniently considered in two parts. In the west the islands are joined by sandy shoals and are separated from the mainland by a lagoon about 60 miles (100 kms.) long, almost closed at its eastern end and running parallel to the coast, called Khor al Bazm. Apart from a few small tidal channels which cut across this offshore barrier, the

Fig. 3 Aerial photograph mosaic of Abu Dhabi-Ras al Kahf area. The overlay shows the major physiographic and sedimentary features. Also shown are positions of individual aerial photos comprising figures 7, 9, 10, 11, 15 and 43; positions of profile lines of Fig. 8 are also indicated.



SADAYAT
CH
R
C · A
A
CAUSEWAY
SABKHA
43
C · A
N
ABU DHABI
BATIN
REEF DELTA
DUNE
INNER LAGOON
MID LAGOON
11
3
R
R
CH
1
R
REEF
DELTA
MID LAGOON
A
A
A
JAZIRAT AL FTAISI
C · A
SABKHA
SABKHA
2
4
HALAT AL BAHRANI
C · A
DUNE
REEF
C · A
C · A
C · A
INNER LAGOON
15
10
JAZIRAT ABU KASHASHA
SABKHA
C · A
DELTA
RAS AL DHABAYAH
C · A
A
7
REEF
LAGOON
9
SABKHA
RAS AL KAHF
SABKHA
C · A
D + CH COMPLEX
SABKHA
CH
D - DELTA
CH - CHANNEL
R - REEF
A - ALGAL
C - CREEK


5 MILES
8 KMS

dominant tidal flow is parallel to the mainland coast. Water depths in Khor al Bazm range from about 50 feet (15 ms.) at its western end to less than 6 feet (1-2 ms.) behind the large island of Abu al Abayadh.

East of Jazirat Abu al Abayadh and extending north east to Ras Ghanadha, a distance of 70 miles (110-120 kms), the offshore islands are separated from each other by large, generally deep, tidal channels which connect with the lagoons behind. In this eastern part, tidal movement of waters is typically at right angles to the coast and in the lagoons circulation is more restricted than in Khor al Bazm. This is reflected in the salinities of waters from the two areas, values in Khor al Bazm reaching 51‰ (Fig. 28/a: after Sugden 1963 a & b) whereas in the lagoons south of Abu Dhabi island salinities are in excess of 65‰ .

The islands are composed largely of uncemented Recent carbonate sediments. Limestones of Recent and possibly Quaternary age are exposed at the surface and underlie the sediments in many places and on the mainland often cap or are banked against Miocene rocks. This complex of islands and lagoons is bordered in some localities on its landward margin by broad, low beach ridges and isolated hills of Miocene and younger rocks; in other localities, extensive areas of carbonate muds form the inner coastline. The former development of beach ridges and sandy strand-line sediments, is to be found in general along the mainland coasts of Khor al Bazm, whereas the development of a muddy inner coastline is typically found in the eastern area.

Inland of the inner coast, whatever its character, there is usually a wide, flat, coastal plain called the 'Sabkha'. The sabkha lies just above the level of highest high water and may be up to 15 miles (24 kms.) wide. It is bounded on its inland margins by low hills, often covered with a thin veneer of blown sand. In places, a few flat-topped, steep-sided, isolated hillocks resembling buttes, stand up from the sabkha surface, commonly surrounded by a fringe of windblown sand. To the west, at Mirfa, the hills of Miocene rocks reach the coast and there the sabkha occurs as isolated patches occupying the low ground. The Miocene rocks also approach the coast to the north-east near Ras Ghanadha.

The entire region is essentially one of carbonate sediment accretion but locally erosion is dominant, for example along parts of the mainland coast of Khor al Bazm and on the open sea coast of Ras Ghanadha.

Abu Dhabi Area

1) General

This study is primarily concerned with part of the eastern area in the vicinity of Abu Dhabi, shown in Fig. 3. The area is bounded to the north-east by Abu Dhabi island and to the south-west by Ras al Kahf. The 30 ft. (9 ms.) bottom contour was taken as the seaward boundary and landwards the study was carried across the coastal plain to the hills of Miocene strata.

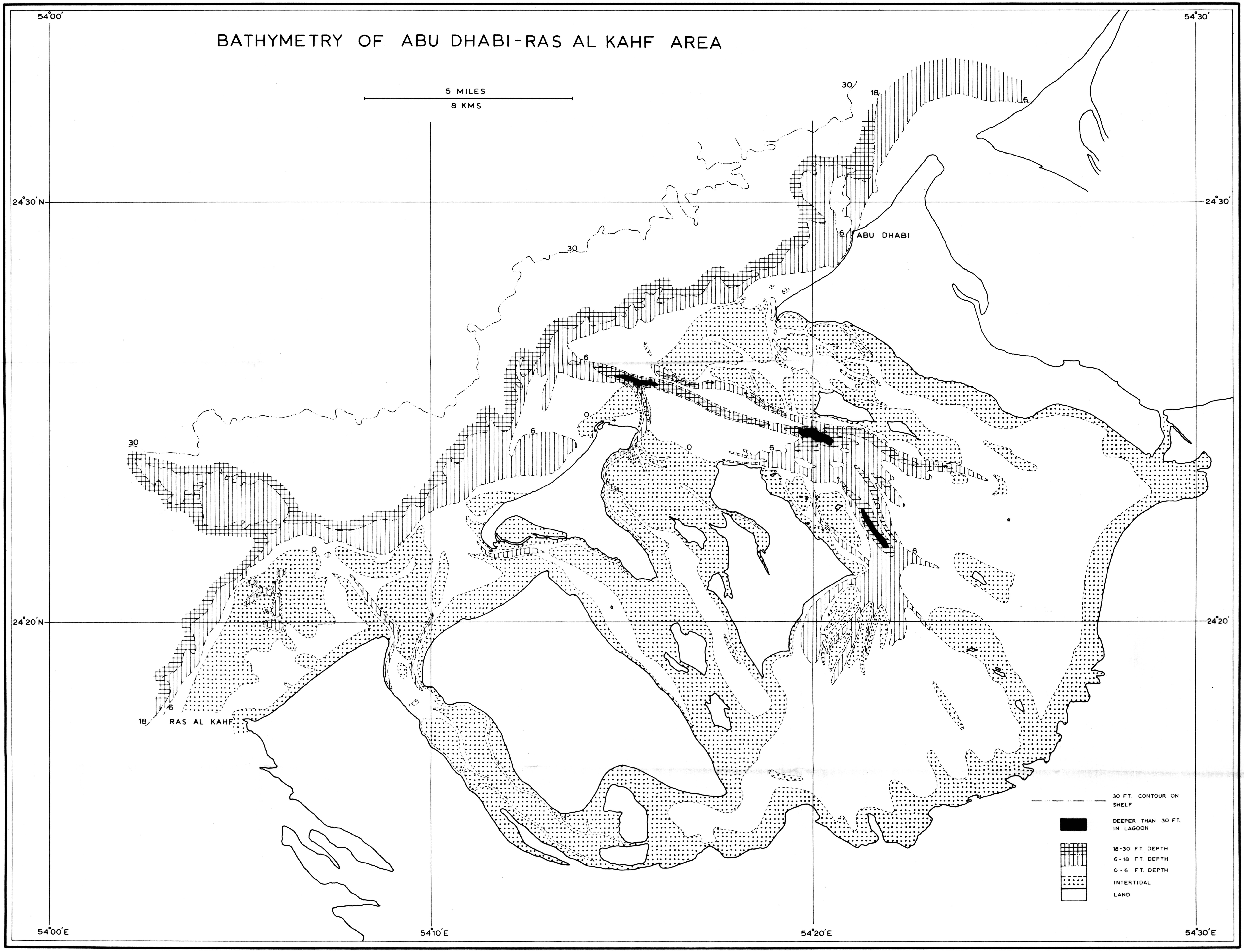
An aerial photograph mosaic comprises Fig. 3 and the overlay indicates the salient features to be noted. The original photographs, at a scale of 1/80,000, were supplied by the Royal Air Force. The seaward parts of the mosaic were keyed to the recent Admiralty survey of the region (1961-62) but the landward parts have no ground control and are therefore of little value for future comparison of accurate shoreline positions or other evidences of accretion or erosion. Subsequently, larger scale aerial photographs (1/30,000) taken by Hunting Survey Ltd. for the Abu Dhabi Petroleum Company, became available for reference purposes and some of these are included as figures.

The detailed bathymetry is indicated in Fig. 4; the seaward data, modified in some details, are from recent Admiralty surveys; the landward data are derived from fieldwork carried out during the present study. All bathymetric data have been corrected for tidal differences and depths are related to low water spring tides. In the seaward areas this coincides with Admiralty chart datum. Mean sea level is four feet (1.2 ms.) above chart datum, thus in the seaward areas an indicated depth of 6 feet (1.8 ms.) will average 10 feet (3 ms.) of water. These relationships do not strictly hold for the innermost parts of the lagoon. Here, the level of low water spring tides departs less widely from mean sea level; it is for this reason that depths have been expressed in relation to low water spring tides, rather than to an arbitrary datum to which, in these areas, the sea would never fall. The indicated depths have been corrected by applying the probable tidal correction pertinent to the area. The tidal correction for any inner locality was derived from a knowledge of the predicted tidal phase at the Abu Dhabi tide guage, together with the approximate time lag involved. Observational deductions in the local area were also important, for example, the distribution of marine fauna and flora indicates closely the limits of the intertidal zone.

Figure 3 shows the open coast to trend south-west-north-east and to comprise the seaward facing coasts of the Dhabayah peninsula and Bahrani, Abu Dhabi and Sadayat islands. Tidal channels connect with extensive back lagoons, and each channel is typically associated with a tidal delta. Between the deltas and fronting the islands are coral reefs of variable development, for example, off Abu Dhabi island the reef is massive and very well developed, off Halat al Bahrani the reef is patchy and less well developed and towards Ras Ghanadha reefs are sometimes completely absent in the interdelta positions. Coral reefs are also present behind the deltas and further back in the lagoons are a diverse series of shoal banks, islands and tidal creeks and swamps. The innermost areas are extremely shallow and Fig. 4 shows how extensive the intertidal areas are in these inner parts.

Fig. 4 Detailed bathymetry of Abu Dhabi-Ras al Kahf area. Note in particular the extensive intertidal areas; also the form of the six and eighteen ft. (2 and 6 ms.) contours with respect to the tidal deltas.

BATHYMETRY OF ABU DHABI-RAS AL KAHF AREA



The main Tertiary outcrop comprises a line of low hills at the inner edge of the coastal plain; this marks the position of at least two earlier shorelines, developed when sea-level was lower than at present. To seaward, the Tertiary rocks were planed off well below present sea-level. The occasional present day outliers of Tertiary rocks would have formed low offshore islands. When sea level subsequently fell a widespread deposit ofaeolian sands was laid down on this platform and was banked against the hills of Tertiary rocks . This deposit extends seawards of the modern coastal islands and occurs at depths in excess of 25 ft. (7.6 ms.) below sea-level. The aeolian grains were largely calcareous and cementation followed fairly rapidly. In accord with local terminology it is called the Miliolite Limestone. A later rise in sea-level to about 8-10 ft. (2.4-3 ms.) above that of the present, resulted in the planation of most of the Miliolite above this level and in the deposition upon this surface of a thin (up to 5 ft : 1.5 ms.) marine limestone. This limestone has not been differentiated by local geologists from the underlying Miliolite aeolian limestone but is here designated the Shell Limestone on account of it commonly bearing coarse skeletal debris. Sea level fell again, the Shell limestone became cemented and both Miliolite and Shell Limestone were widely destroyed by marine erosion acting at about present sea level. A series of thin, often discontinuous limestones were developed somewhat later but whereas the Shell and Miliolite limestones are now almost entirely calcitic, these later limestones are still largely aragonitic. In general, the later aragonitic limestones are developed in the seaward islands and marine areas; some of them are forming today.

This, very briefly, is the origin of the present day disposition of islands and lagoons. The islands are usually cored with erosion residuals of earlier limestones, much modified by subsequent accretion of unconsolidated sediments. The present marine areas are the site of the modern phase of carbonate sedimentation.

2) Land Areas

The islands vary greatly in form and size. Many of them are cored by Miliolite and Shell Limestones and in some areas later limestones are also widely developed. However, much of their combined area results from the accumulation of unconsolidated sediments. The marine sediments are typically deposited to just above the level of highest high water; wind blown material may subsequently modify this initial surface.

The seaward islands and peninsulas have wide seaward faces and taper in toward the mainland coast (Fig. 3). Sadayat, Abu Dhabi and Bahrani islands have fairly steep seaward facing beaches, backed by frontal dune ridges reaching heights of 30 ft. (9 ms.). The steep beach faces are characterised by beach cusps and are fronted by a narrow low tide terrace. At the top of the beaches one or more berms are usually developed, often extensively colonised by large burrowing crabs. The berms are succeeded landwards by a wind stripped terrace which may vary in width from a few feet to 400 yds (up to 400 ms.). Low, hummocky, partly vegetated dunes lead inland from this terrace and merge into the main dune ridge. On Abu Dhabi island, south-west of the town, three parallel ridges are developed. Cropping out between the ridges is a poorly cemented aeolian oolite limestone, probably representing the partially cemented core of an earlier dune. This limestone is better exposed to the north-east of the town, where a dune sand ridge is all but absent and only the eroded core of an earlier dune phase is present.

The Dhabayah peninsula differs slightly as its seaward facing coast is sheltered by an extensive off-shore coral reef. The beach face shelves gently into the back reef lagoon and no cusps are developed. Low, hummocky, partly vegetated dunes lead inland, but no large frontal dune ridge is developed.

Behind the frontal aeolian sediments lie areas of irregular low dunes, patches of Shell and Miliolite limestones, wind stripped surfaces bearing coarse surface lag deposits, or recent marine sediment surfaces. These areas often show evidence of earlier shore-line positions. For example, the old shorelines of the Dhabayah peninsula enable one to follow its development from being an offshore island until it became joined to the mainland

Inner coasts of the islands and peninsulas are gently shelving except where tidal channels run close to the shore. Many coasts have an extensive development of bars and spits of sediment which normally point inwards towards the back of the lagoon, indicating intertidal and shallow water sediment movement in this direction. A wide intertidal zone commonly extends along these coasts. The higher parts of these intertidal areas are usually colonised by small burrowing crabs and these areas have been called the Crab Flats (Fig. 5). Slightly lower in the intertidal zone are areas colonised by gastropods, the so-called Gastropod Flats (Fig. 6). Extensive intertidal algal flats and tidal creeks and swamps adjoin many of the inner lagoon coasts.

The major inner islands are Jazirat al Ftaisi and Jazirat Abu Kashasha. They both owe their present extent to accretion of unconsolidated sediments around cores of Shell and Miliolite limestone. Tidal creeks and swamps surround much of their coasts, this marking the final phases of infilling of the back lagoonal areas. Fig. 7 shows the detail of the near-bridging of the gap between Jazirat Abu Kashasha and the mainland. Algal flat and tidal creek development have all but infilled the remaining marine areas, and the progress of these additions can be seen from the old shoreline positions.

The smaller inner islands have diverse origins. Some are Shell and Miliolite limestone erosion residuals

Fig. 5

A (~~Upper~~ ABOVE) : Wide Crab Flats on the inner coast of Halat al Bahrani, being partially flooded during flood tide. Isolated clumps of *Arthrocnemum glaucum* occur on the Crab Flats but are most common at its inner margins (top right). Note the density of the burrowing pellets of *Scopimera*, produced between two successive high tides.

B (~~Lower~~ OVER) : Detail of *Scopimera* burrowing pellets and stellate pattern of pellets and grooves around the small central burrow.

Fig. 6 Detail of low intertidal Gastropod Flat surface from sheltered coasts of mid lagoon areas; showing trails and abundant Cerithium: surface densely covered with gastropod faecal pellets formed of the fine grades of the sediment.

Fig. 7 Showing infilling of the innermost lagoon areas between Jazirat Abu Kashasha and the mainland - see Fig. 3 for location. Old shoreline positions are indicated (the lineation cutting at a high angle across the latter is wind developed). The development of a sabkha surface is well shown - an almost featureless inner lagoon area passing into a low intertidal creek sequence which gives way to a high intertidal stromatolitic algal mat facies. Old infilled channels can be seen in the sabkha surface. This is the typical sequence of final infilling in the sheltered parts of the inner lagoons. Scale 1/30,000.


OLD SHORELINE
ALGAL FLAT
LAGOON
SABKHA
CREEKS

with greater or lesser amounts of fringing unconsolidated sediments, whereas others comprise creek and swamp areas with occasional beach ridges. The latter islands are extremely low and are almost completely flooded at high water.

The major features of the mainland have already been briefly described, - the coastal plain and hills of Tertiary rocks. These areas will be more fully considered in Chapters V and VI.

The abundant islands in this north-eastern area, together with features of the mainland sabkha, enables earlier sedimentation patterns to be determined. The effect of the islands in decreasing water circulation and turbulence by providing shelter from wind and wave activity is particularly significant. In fact, it can be shown that as the area of islands increased so the back lagoonal sediments became more muddy (Chapter V). The major difference between the north-eastern area of mainly muddy back lagoonal sediments and the Khor al Bazm area where more sandy sediments are accumulating, can be stated to be the difference in water circulation and turbulence.

3) Marine Areas

The marine areas form a series of physiographic units which are described below:

a) Inner Shelf

The inner areas of the open shelf are included in this study. They extend seaward of the 18-20 feet (6 ms.)

bottom contour at the seaward edge of the tidal deltas and offshore coral reefs. Where offshore coral reefs are absent then the shelf extends uninterrupted to the shoreline of the seaward facing islands and peninsulas. This occurs, for example, at Abu Dhabi, where the shelf extends in towards the shore between the offshore reef and the south-western tidal delta. The shelf has a slightly irregular bottom topography but depths otherwise increase gradually seawards. It is probably in large measure a platform eroded into the underlying limestones during a period of lower sea level. It is blanketed irregularly with a variable series of uncemented sediments, oyster reefs and sparse coral growth.

b) Offshore Coral Reefs

The offshore coral reefs are typically developed in interdelta positions, fronting the islands or peninsulas. This development very closely resembles that found in the Bahama region where coral reefs are "best developed on the seaward sides of islands, cays and rocky shoals...." (Newell et al. 1959). The reef off Abu Dhabi is a conspicuous feature and rises fairly abruptly from depths of about 20-25 ft. (6.1-7.4 ms.); the reef top lies just below lowest low water. The seaward face bears a prolific growth of massive corals together with some branching forms; in other parts of the reef the massive forms are less abundant. Behind the reef is a channel where water depths reach 15-20 ft. (4.6-6.1 ms.); the sea floor here is covered with muddy sediments, in contrast to the coarser reef top sediments.

A small, poorly developed reef lies off the front of the tidal delta to the south-west of Abu Dhabi, in the angle between the main delta and the northeastern subsidiary lobe. Occasional small corals are also found extending further to the west, at the front of the tidal delta, for several miles.

The reef off Halat al Bahrani is patchy and poorly developed and consists almost entirely of branching corals. Channels running normal to the reef trend cut across it and break it into several segments. It is a relatively low feature, growing in depths of about 15 ft. (4.6 ms.) and rising only about 6 ft. (1.8 ms.) above this level (Profile No. 4, Fig. 8).

The reef lying off the Dhabayah Peninsula differs yet again; areas dominated by a vigorous growth of branching corals occur as large patches at the seaward margins, but most of these lie at depths of 6-10 ft. (1.8-3.0 ms.). Behind lies an extensive shoal area, sparsely covered with coral heads. Several large depressions, up to 20 ft. (6.1 ms.) deep occur in the reef top, typically surrounded by a rim of living corals and partially infilled with muddy sediments. The reef top is also characterised by large bars and other accumulations of reef derived sand and coarser debris. Behind the reef lies a lagoon, about 6-10 ft. (1.8-3 ms.) deep, in which are abundant microatolls.

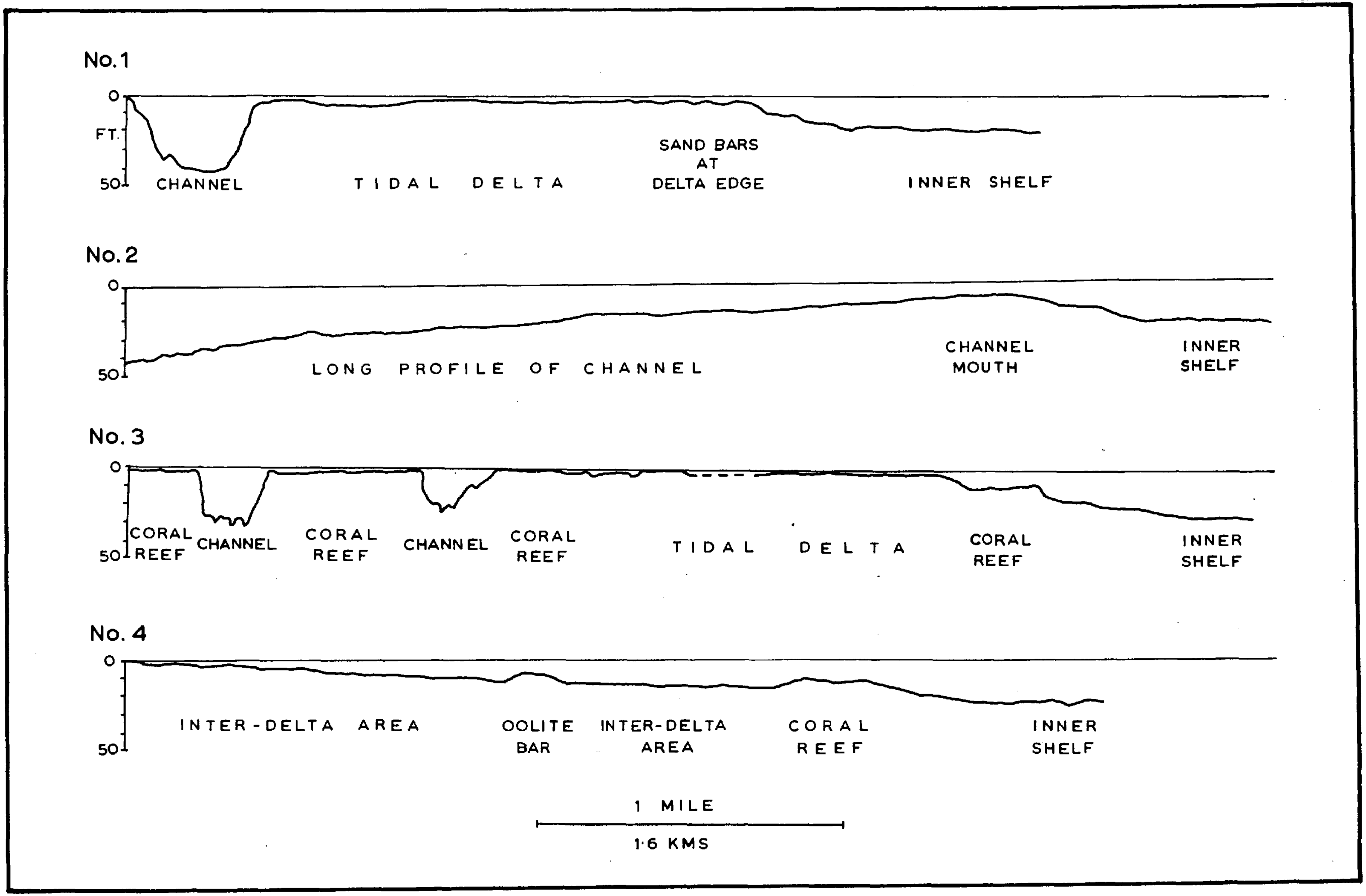
Fig. 8 Profile lines are indicated in Fig. 3.

No. 1 showing maximum development of tidal channel; also tidal delta with large sand bars at seaward edge.

No. 2 Long profile of main tidal channel, showing asymmetric bars across the floor and general shallowing seawards; profile seawards of the channel mouth is identical to that off the edge of the tidal delta.

No. 3 showing inner coral reefs and double tidal channel; also tidal delta with a small offshore coral reef on the delta slope.

No. 4 Interdelta profile with small bar of oolite sand and a low, poorly developed offshore coral reef.



No. 1
0
FT.
50
CHANNEL
TIDAL DELTA
SAND BARS
AT
DELTA EDGE
INNER SHELF
No. 2
0
50
LONG PROFILE OF CHANNEL
CHANNEL
MOUTH
INNER
SHELF
No. 3
0
50
CORAL
REEF
CHANNEL
CORAL
REEF
CHANNEL
CORAL
REEF
TIDAL DELTA
CORAL
REEF
INNER
SHELF
No. 4
0
50
INTER-DELTA AREA
OOLITE
BAR
INTER-DELTA
AREA
CORAL
REEF
INNER
SHELF
1 MILE
1·6 KMS

Each microatoll is characteristically dead in the centre, the live coral forming a fringing rim; the slopes are of sandy materials which pass down into the more muddy, weed covered bottom of the lagoon floor. The reef, reef top depressions, microatolls, lagoon, and a small flood-tide delta can be seen in Fig. 9.

c) Tidal Channels.

Connecting the open sea with the lagoons which lie behind the seaward islands are large tidal channels which are characteristically associated with tidal deltas (Fig. 3). To the north-east of Abu Dhabi a single, wide channel is associated with a simple delta, whereas to the south-west between Abu Dhabi island and Halat al Bahrani and between Halat al Bahrani and Ras al Dhabayah, more than one channel exists and the associated deltas comprise two or more lobes. The channel to the north-east of Abu Dhabi has a maximum depth of 23 ft. (7 ms.) and several shoals partially obstruct the free flow of tidal waters within it. The main channels between Abu Dhabi island and Halat al Bahrani are better defined; they are narrower and deeper, reaching a maximum depth of 44 ft. (13.4 ms.) where the main branches coalesce. Between Jazirat al Ftaisi and Abu Dhabi island, where the constricted channels open into the broad mid lagoon area of deep water, a depth of 45 ft. (13.7 ms.) occurs (Fig. 4). The channels between Halat al Bahrani and Ras al Dhabayah are fairly well defined, are free from shoals and reach depths of 15 ft. (4.6 ms.).

Fig. 9 Showing the offshore coral reef and back-reef lagoon off the Dhabayah peninsula - see Fig. 3 for location. Dark patches at top of figure are depressions in the reef top, up to 20 ft. (6 ms.) deep; the reef top is covered with diffuse coral growth. The seaward edge of the reef top is marked in places by bars of reef sands, similar in form to the bars at the edge of the tidal deltas. The backreef lagoon is densely carpeted with _Halodule_ (dark areas). Abundant micro-atolls can be seen in the lagoon, typically dead in the centre (light coloured) with a peripheral fringe of living coral (dark fringe); the slopes are not vegetated (light coloured) and pass down to the vegetated lagoon floor. A small flood tide delta is also present, developed by flood tide currents flowing into the back-reef lagoon. Scale: 1/30,000.


REEF TOP
REEF FRONT
MICRO-ATOLLS
'HOLES'
BACK-REEF
LAGOON
DELTA
SHORELINE

The channels are thus somewhat variable features but their degree of development closely parallels the size of the associated back lagoons. For example, the Abu Dhabi/ Halat al Bahrani channels connect with a very extensive back lagoon, whereas the lagoon to the northeast of Abu Dhabi island is rather smaller, and that to the south west of Halat al Bahrani smaller again. The associated tidal channels within this sequence become progressively shallower and less well defined.

There are certain features, however, which are common to all channels, in kind though not in degree. When the channels pass between the tips of the islands and are typically most constricted, then the depths are greatest. Tidal currents will reach a maximum in these positions and erosion will be most rapid. The long profiles are all very similar, a gradual shallowing occurring seawards of the deepest parts; profile 2 of Fig. 8 shows this for the main channel between Abu Dhabi island and Halat al Bahrani. In fact, at their mouths, the channels are usually only 2-3 ft. (1 m.) deeper than the adjoining tidal deltas (Fig. 4). In some places the channels lose their identity completely. The mouths of some channels become infilled, as for example, that lying parallel but immediately north-east of the main channel between Abu Dhabi island and Halat al Bahrani.

The channels are typically horizontal slots with widths greatly exceeding depths, depth to width ratios varying from $\frac{1}{40}$ to more than $\frac{1}{80}$. A recent discussion of channels and tidal inlets by Price (1963) indicates that the form of tidal channels from many areas is similar in most aspects to those of the Trucial Coast. Deep holes normally occur in axial and other narrows, and where channels converge, a hole one third to three times as deep as the general channel depth is developed beyond the convergence. Channel depth to width ratios range from $\frac{1}{25}$ to $\frac{1}{75}$. The depth to width ratios and axial deepening of these Trucial Coast channels follow closely his findings, the over-deepening of the channel between Abu Dhabi island and Halat al Bahrani being $1\frac{1}{2}$-2 times the normal channel depth.

Channel walls vary widely in slope and character, some being fairly steep, others much more gentle. Many walls are of loose sediments, others are encrusted with a vigorous growth of coral. The general disposition of uncemented sediments and earlier limestones suggests that the deep channels are incised into the earlier limestones. Price (1963) and Phleger and Ewing (1962) have also found that many tidal channels, associated with similar coastal tidal lagoons, have cut down into Quatarnary deposits.

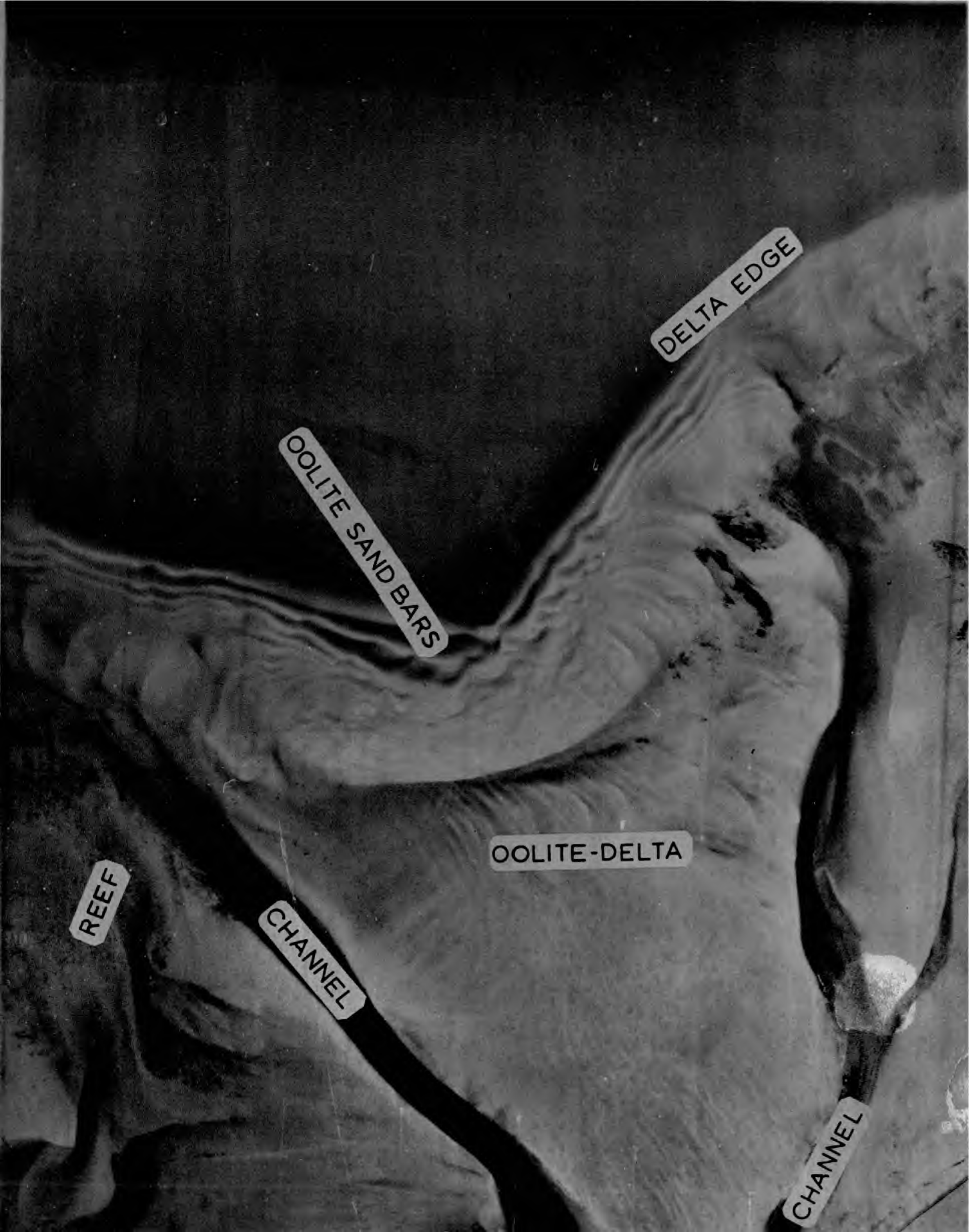
d) Tidal Deltas

The tidal deltas are shoal areas bounded seawards by the 6 ft. (1.8 m.) contour; extensive intertidal areas occur atop the deltas (Fig. 4). The seaward boundary is usually marked by a line of breakers. Seawards of the 6 ft. (1.8 m.) contour the sea bed shelves steeply to 18-20 ft. (6 ms.): Fig. 8, profiles 1, 2 and 3. Below this depth the inner shelf is encountered but the bottom topography still bears some relationship to the deltas; the 24 ft. (7.2 ms.) bottom contour parallels the delta edge. Deeper than this, evidence of the deltas is lacking and the bottom deepens irregularly seawards.

The deltas vary considerably in form, being simple at the mouth of a single tidal channel or complex where more than one channel exists between the islands. They are being deflected partly south-westwards by the prevalent and dominant wave approach direction (north-north-west). For example, the deltas north-east and south-west of Abu Dhabi island are both asymmetric with respect to the islands and tidal channels; the northerly facing edges are being developed at the expense of the westerly edges and the channel mouths are consequently being deflected to the south-west.

At the delta margins large sand bars are developed, with wave lengths of several hundred ft. (200-300 ms.) and amplitudes of 2-4 ft. (0.6-1.2 ms.). The bars take many forms, they may be long, straight, simple, and arranged parallel to one another, or shorter, straight or curved and arranged en échelon; they may also assume the form of petaloid lobes or other configurations. Figure 10 shows the Halat al Bahrani-Ras al Dhabayah tidal delta and the various forms taken by the bars are well developed here. On the two deltas to the north-east, en échelon bars are developed near to and subparallel to the northern margins,

Fig. 10 Ras al Dhabayah - Halat al Bahrani tidal delta - see Fig. 3 for location. Showing tidal channels traversing the delta top and becoming shallower (lighter in colour) near the delta edge. The delta edge is distinct and several forms of sand bars are well developed, including long, straight, parallel bars, smaller, curved, en-échelon bars and petaloid lobes. The dark patches on the delta top are seaweeds growing on a rock floor. Scale 1/30,000.


DELTA EDGE
OOLITE SAND BARS
OOLITE-DELTA
REEF
CHANNEL
CHANNEL

but more irregular bars extend across the delta tops where the latter swing south-west in towards the island coasts. This south-westward movement of the delta sands has resulted in the addition of a large spit, now enclosing a small lagoon, to the northern tip of Halat al Bahrani, since 1958 when the Royal Air Force aerial photographs were taken.

At the northern and south-western ends of Halat al Bahrani the delta top sands are being pushed landwards and between the shoal banks and the island coasts lie shallow lagoons with bottoms bearing a sparse plant growth. Sparse growth of brown seaweeds is also found on the delta top, usually in the troughs between the large sand bars, the floor in these positions occasionally being of limestone.

In addition to the major features, the delta top sediments are extensively ripple-marked; even the delta front between 6-18 ft. (1.8-5.4 ms.) being affected. The various features of the delta top give some indication of the high wave energy afflicted upon these areas.

e) Inner Coral Reefs

Coral reefs of variable development lie behind all the tidal deltas, usually in shoal areas sheltered from excessive wave activity by the islands and channels. The most extensive lies behind the main tidal channels south-west of Abu Dhabi island, fronting Jazirat al Ftaisi. Branching corals predominate and along the channel sides growth is extremely prolific. The reef top lies just below lowest low water, heads of corals being occasionally exposed.

Coral growth is patchy on the reef top and many of the corals are dead and bear luxuriant growths of brown seaweeds.

Small coral areas occur (a) in the channels near Batin, (b) behind the north-eastern tip of Abu Dhabi island, (c) behind Ras al Dhabayah, and (d) in channels along the east and south coasts of Halat al Bahrani.

f) Mid and Inner Areas

The mid and inner areas are loosely definable units and comprise the mid and inner parts of the back lagoons. Depths vary in the mid areas from intertidal to 30 ft. (9.2 ms.). The intertidal areas stretch along island and mainland coasts and also occur as offshore banks. They are generally bare sandy areas, in contrast to the luxuriantly vegetated areas below low water, where the bottom sediments are generally much muddier.

The mid areas comprise the shoal banks and small channels lying behind the oolite deltas and the deeper water lying off the north-east coast of Jazirat al Ftaisi. A typical area is shown in Fig. 11.: shoal banks are light in colour and free from vegetation, deeper areas are darker, where the bottom bears a thick plant growth.

There is no boundary as such between the mid and inner areas; in their inner parts the mid areas are everywhere shoal (less than 6 ft. - 1.8 ms.) and pass insensibly into the inner areas. The inner areas are the very shoal back lagoonal regions where depths gradually decrease until the inner shores are reached.

Fig. 11 Mid lagoon area - see Fig. 3 for location. To show variability in depth and associated development off a plant cover. The dark areas are below low water, are densely carpeted with Halodule and the sediments are always very muddy. The lighter coloured areas, occurring either as shoal banks fringing islands or as isolated shoals are almost devoid of a plant cover and the sediments are normally pellet sands with small amounts of aragonite mud. Scale 1/30,000.


SHORELINE
PELLET SAND SHOAL
DENSE WEED
ARAGONITE MUD

In both mid and inner areas there are few bottom features of interest. The veryshoal banks of the mid areas are occasionally ripple marked but in general turbulence in these parts is very slight.

g) Tidal Creeks and Swamps and Algal Flats

Tidal creeks and swamps are widely developed in some intertidal parts of the mid and inner lagoon. The dominant plant of these areas is a bushy halophyte, Arthrocnemum glaucum (Fig. 5A, 12); the mangrove, (Fig. 13) Avicennia nitida is more local in occurrence. Commonly associated with the tidal creeks and swamps are wide spreads of tough, rubbery mats of blue-green algae, (Figures 7, 14, 15).

Mangrove swamps are developed in some places around Jarirat al Ftaisi and Jazirat Abu Kashasha. They typically lie slightly seaward of the optimum development of Arthrocnemum; the algal mats lie slightly higher again in the intertidal zone. Thus it is common to find a Mangrove/Arthrocnemum association in the low intertidal areas or an Arthrocnemum/Algal mat association in the higher parts of the intertidal zone. It is uncommon to find Mangrove and algal mats together. In the higher intertidal areas algal mats occur alone.

A good example of the creek and algal mat development can be seen in Fig. 7. A few major channels are present together with a multitude of lesser ones which thread their way through the often dense cover of Arthrocnemum

Fig. 12 (~~Upper~~) ABOVE

Creek and swamp area with Arthrocnemum; muddy sediments extensively burrowed by crabs leaving the upper surface extremely irregular.

Fig. 13 (~~Lower~~) OVER

Mangrove (Avicennia) showing entrapment of plant debris around its roots; growing on low intertidal banks of muddy pellet sand.

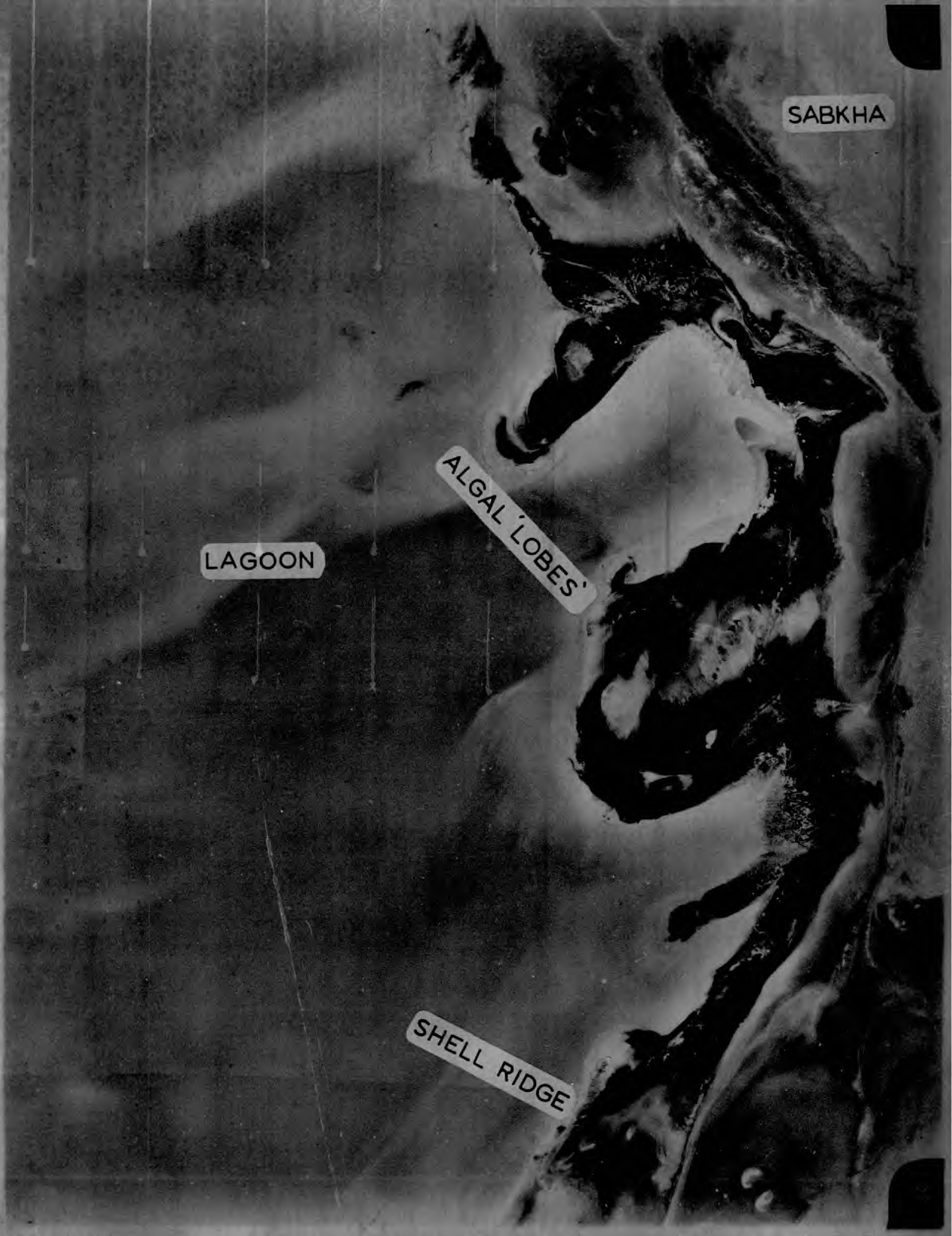
Fig. 14A (Upper)	Detail of upper surface of algal mat showing vermicular structure.
14B (~~Middle~~) Lower	Algal mat broken into polygons with edges upturned.
14C (~~Lower~~) Over	Algal surface of polygons traversed by very shallow (less than 2-3 ins. - 6-10 cms.) channels and pools, often bearing abundant gastropods (Cerithium and Littorina).

Bordering the creeks are the algal mats, some of which are rarely flooded.

Sheltered shoal areas typically bear these associations (Fig. 3), for example, an area exists behind the frontal dune ridge at the south-west end of Halat al Bahrani; other areas exist behind Jazirat Abu Kashasha; extend south-east from the south-eastern tip of Halat al Bahrani, almost to the mainland; occur in various places around Jazirat al Ftaisi; extensive areas occur to the north-east of Abu Dhabi island and behind Sadayat island: much of the inner coast also has this form. Fig. 15 shows part of the inner coast with a development of Arthrocnemum and algal mats forming the landward edges to the inner coast.

Water circulation in these areas is much restricted and the occurrence of tidal creeks and swamps and algal mats marks the ultimate marine environment before final silting up and infilling of any sheltered lagoon area.

Fig. 15 Innermost lagoon coast - see Fig. 3 for location: showing the almost featureless shallow water areas of the innermost lagoon and the mainland shore fringed with wide spreads of algal mats together with some creeks and shell sand bars. Note the manner in which accretion takes place, a bar of pellet and skeletal sand being built, usually on a shoal marine tongue (the lighter coloured areas of the inner lagoon). Behind the bar more muddy sediments accumulate and algal mats develop. Restricted areas between the accreting tongues of sediment develop creeks with <u>Arthrocnemum</u>. Similar developments can be traced inland on the sabkha surface. Scale 1/30,000.


SABKHA
ALGAL 'LOBES'
LAGOON
SHELL RIDGE

CHAPTER II

PHYSICAL AND CHEMICAL CONTROLS OF THE ENVIRONMENT.

GENERAL

The determination of the physical and chemical parameters of any environment of carbonate sedimentation is extremely critical. For example, physical processes may control the form of the coastline and the bottom topography of the nearshore and shoal water areas. The distribution of sediments is greatly affected by physical processes. Physical processes which affect the chemistry of marine waters are obviously significant in the formation of carbonate sediments. Factors affecting the solubility and precipitation of $CaCO_3$ in sea water, such as salinity, pH, temperature and turbulence must all be considered.

WIND

Wind is the source of energy of nearly all coastal changes, and thus the analysis of wind direction and strength is critical in any nearshore area of Recent sedimentation. Waves are the most important nearshore physical phenomenon and are produced by wind acting on the water surface. Air circulation plays an important role in the process of evaporation and is of obvious significance in regions of chemically precipitated sediments.

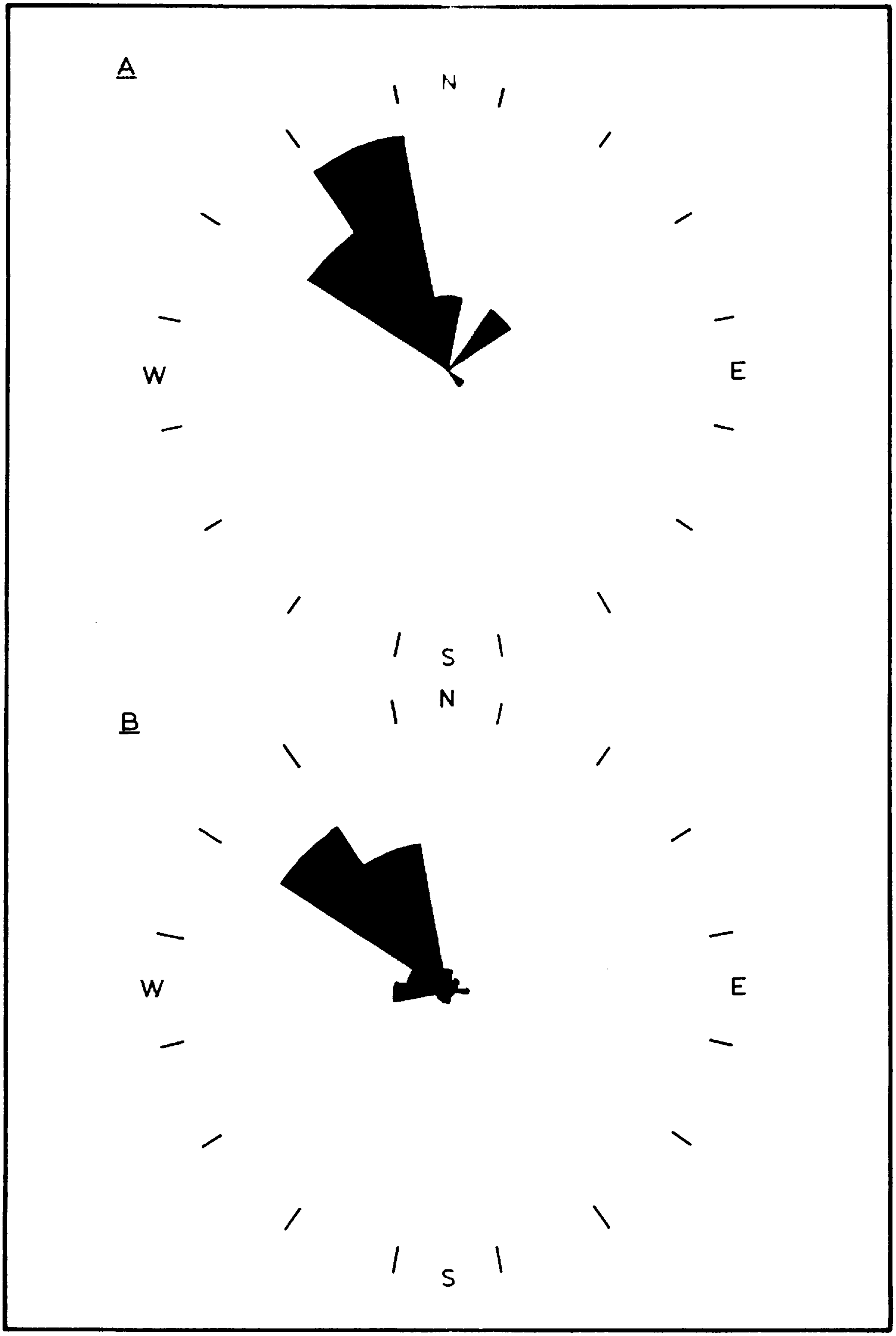
The prevailing wind along the Trucial Coast is the north-north-westerly 'shamal'. Winds from other directions do occur but are less common (Fig. 16). In the Abu Dhabi area the occasional strong winds from the southerly quarter are particularly important as they may bring large volumes of airbourne material from inland, which later come to be deposited in the marine areas.

In addition to the winds of a regional nature there are the diurnally developed onshore and offshore winds. In the absence of any regional air-movement the normal pattern is of gentle offshore winds in the morning, followed later by onshore winds as the inland areas become heated by the sun. The afternoon, onshore winds often become sufficiently strong to move dune sands on land and raise appreciable waves at sea.

Strong and gale force shamal winds blow mainly during the winter months, winds otherwise tend to be light and variable.

Fig. 16 Wind and wave approach directions: 16A for Abu Dhabi area, 16B for Trucial Coast Embayment.

Data are for winds which raise waves of amplitude greater than 3 ft. (1 m.). Data supplied by Meteorological Office.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W
E
S

In shoal areas wind may play a significant role in the movement of marine waters. The wind may back-up or on occasions oppose the normal tidal movements of waters. Along the Trucial Coast the prevalent wind is onshore and thus its effect in the lagoons will be to oppose ebbing waters or back up flooding waters. When strong onshore winds back up high spring tides then vast areas of the sabkha may be flooded, areas which under normal conditions are above the level of high water spring tides.

WAVES

Sea waves are produced by wind blowing across the sea surface. The orientation of the wave front in open water is dependent upon the wind direction, and the wave height or amplitude depends upon the fetch and the wind strength. Some modifications are found to occur in restricted shallow waters.

Meterological Office data for the Trucial Coast (Figure 25 squares 43, 44; note each one degree square is assigned a number which is used when reference is made to it), show that waves of amplitude less than three feet (1 m.) may occur with any orientation; such waves are produced by light and variable winds and have little physical affect on the environment as a whole. Waves of amplitude greater than three feet have a pronounced modal orientation to the north-north-west (Fig. 16A; squares 43 and 44). Data for almost the entire Trucial Coast embayment indicate a mode slightly more to the north-west (Fig. 16B). Wave amplitudes in squares 43 and 44 do not exceed 9.5 feet (3 ms.), although further offshore in squares 52, 53, 54 wave amplitudes of 14 feet (4.3 ms.) are recorded. Except for three records - two north-east and one west - wave amplitudes greater than 6.5 feet (2 ms.) are always associated with north-westerly winds.

The direction of wave approach is critical on the open island coasts and edges of the oolite deltas. The orientation of the coasts and the frontal dune ridges and also the asymmetry of the oolite deltas is a response to the prevailing north-north-westerly wind and wave approach direction.

Except in the tidal channels, wave activity is the major control of water turbulence experienced within these shoal areas. Seawards of the oolite deltas and open coasts the wind has an extensive fetch to the north-west. When the waves approach shallow water they become retarded, wave heights increase, wave lengths decrease and wave velocities decrease. The height to length ratios increase until instability results and finally the waves break and their energy becomes dissipated. Waves commonly break at the edge of the oolite deltas and their margin is marked by a line of breakers during all but the calmest periods. The entire delta tops are subjected to intense wave activity and this is reflected in the character of the underlying sediments.

Inner areas are affected by much less vigorous wave conditions. Waves do not normally cross the tidal channels and thus areas lying inside them are somewhat protected. Further back in the lagoons the islands effectively provide protection by decreasing the fetch of the wind so that only small waves are ever encountered in these parts. It is interesting to note that the small stretch of innermost coast lying south of Abu Dhabi Causeway and which has no island to protect it (Fig. 3), is rather different in character from the coast further south. The differences in sediment type and physical nature between the protected and unprotected inner coasts are sufficiently strong that evidence of this type has been able to be used to reconstruct the palaeogeography of the area at intervals during the present cycle of sedimentation.

Water turbulence is the major physical control determining distribution of sediment grains. Currents would move little sediment if it were not for wave generated water turbulence which raises sediment grains into suspension. However, waves alone would play a negligible role in sediment transport without the lateral translation of particles effected by currents.

TIDES AND CURRENTS

(a) General

The tide is one of the most important ecological factors in lagoons and other areas of shoal water. It is the agent which effects exchange of water and its vertical range determines the extent of algal and mangrove flats and other intertidal areas. Currents in lagoons are almost entirely caused by tidal agencies, except for some wind induced currents.

Tidal waves or 'tides' are merely vertical rises and falls of the surface of the sea and are the resultant of a complex series of partial tides. The periods of the two major components are one lunar day (diurnal tide) and half a lunar day (semidiurnal tide). A diurnal tide is one which exhibits a single high and low water each lunar day; tides of this nature occur commonly a few miles off the eastern coast of the Qatar Peninsula. A semidiurnal tide is one which exhibits two approximately equal high and low waters each lunar day; the tides of the Atlantic coasts are of this type, the diurnal component being small. However, some coasts exhibit mixed tides, with two high and low waters occurring each lunar day but the successive high tides and successive low tides are of quite different heights. The Pacific coast of the United States and most of the Persian Gulf coasts exhibit mixed tides.

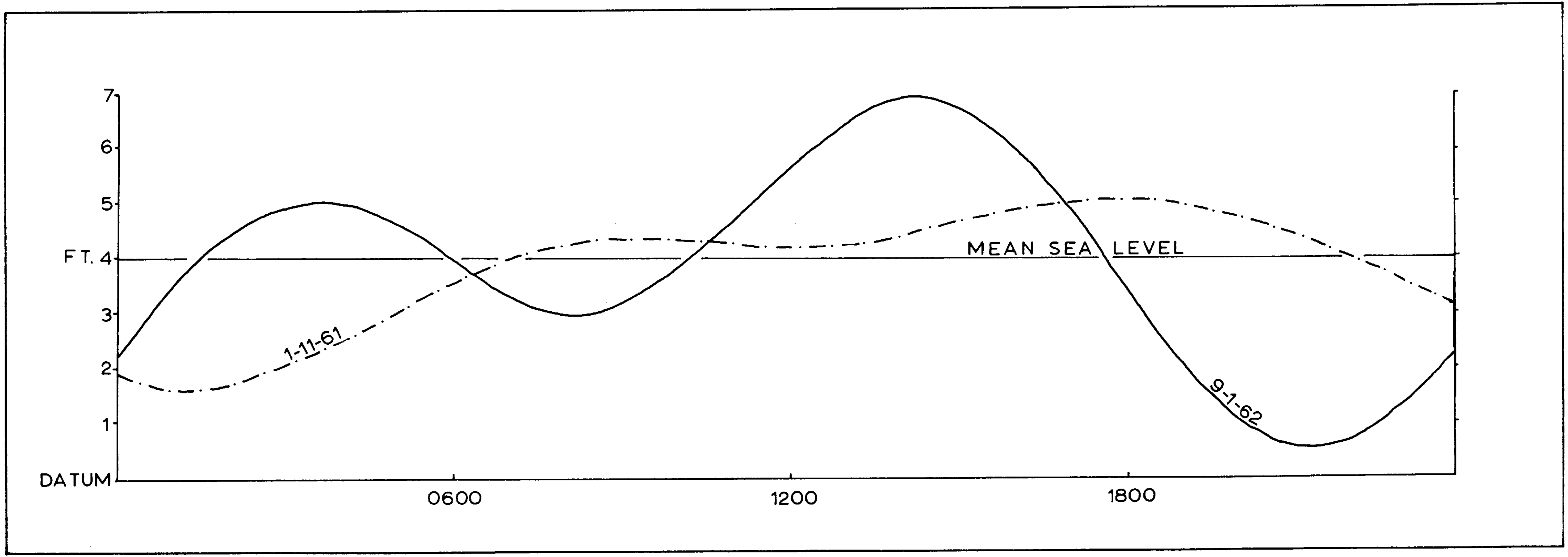
One other tidal complication must be considered - when the sun and moon are in conjunction or opposition higher than normal or spring tides are produced. These occur about every fourteen days and the tidal range is typically about 20% greater than mean tidal range. When the moon is in its quarter phases then lower than normal or neap tides are produced; these are generally about 20% less than mean tidal range. Figure 17 shows the mixed tidal curve for Abu Dhabi and also the difference in tidal range between spring and neap tides.

The mixed tides experienced along the Trucial Coast develop a very strong diurnal inequality at times of tropic tides (when the moon has a high declination north or south); at its most extreme the tide becomes truly diurnal as off the Qatar Peninsula.

Tidal streams or currents represent the horizontal movement of the tidal wave; they generally run parallel to the coast and are a component of the longshore current. They act therefore at right-angles to the main direction of wave action. The tidal streams reverse their direction roughly in phase with the tides and off the Trucial Coast set east-west at about 1 knot (1.8 kms./hour). During periods of strong diurnal inequality (e.g. spring tides), one of the two tidal streams will be considerably stronger than the other, or in its extreme development the streams will turn only once in a lunar day (as off the Qatar Peninsula).

Tidal streams may determine the distribution of finer grained sediments in offshore areas. Great diurnal

Fig. 17 Spring and neap tide curves for Abu Dhabi open coast, data from Abu Dhabi Petroleum Company tide gauge at Abu Dhabi.


7
6
5
FT. 4
3
2
1
DATUM
0600
1200
1800
MEAN SEA LEVEL
1-11-61
9-1-62

inequality in tidal streams will be reflected in an ability to move coarser sedimentary particles during periods of maximum stream velocity. However, it has been noted that in general tidal streams become uncertain inshore and that velocities decrease.

The total longshore current in any place comprises: (a) tidal stream effect (reversible)

(b) Oceanic current effect (irreversible; generally unimportant in nearshore areas).

(c) wind currents

(d) longshore component of oblique wave action.

Thus along the nearshore region of the Trucial Coast the effect of (a) will probably be slight and (b) will be negligible. Components (c) and (d) owe their energy and direction to winds; the prevalent and dominant wind is from the north-north-west and as the open coast near Abu Dhabi trends SW-NE it is not quite an onshore wind. Thus the overall balance of wind and wave produces a slight longshore drift to the south-west. This conclusion is supported by physiographic developments on the oolite deltas.

Amongst the islands and in the lagoons tidal currents will dominate, although at times currents generated directly or indirectly by the wind may become important.

(b) Tidal Data from the Coastal Areas

During both field seasons fairly complete tidal data were available from the tide guage of the Abu

Dhabi Petroleum Company at Abu Dhabi. During the first season tide pole stations were set up at Halat at Bahrani, Ras al Dhabayah and Jazirat Abu al Abayadh by the Admiralty, who were conducting hydrographic surveys in the region. Some information has also been obtained from surveys by Sir William Halcrow and Partners, Development Engineers in Abu Dhabi. The locations of the tidal stations are shown in Figure 18.

The inter-relationships between the tidal curves of the several stations are shown in figures 19 and 20. (Figure 19/A represents average tidal conditions, 19/B spring tide conditions). The Abu Dhabi station is situated on the open coast with no shoal areas to restrict the tidal movement of waters. The Abu Dhabi barge station lies 2½ miles (4 kms.) up channel and the Halat al Bahrani station 4 miles (6½ kms.) up channel, between shoal banks. The Ras al Dhabayah station lies 3 miles (5 kms.) up channel from open water and the Jazirat Abu al Abayadh station lies 4-5 miles (7-8 kms.) up a very wide channel which offers little restriction to the free movement of waters; comparison of the curves for Abu Dhabi and Jazirat Abu al Abayadh (figure 19/B) shows this to be so. Coasts with unrestricted access for tidal waters thus exhibit the greatest tidal fluctuations. in this region reaching a maximum of about seven feet (2 ms.). Stations situated some miles from the open sea reflect their positi
in their
tidal curves (compare curves for Halat al Bahrani and Ras al Dhabayah with those for Abu Dhabi and Jazirat Abu al

Fig. 18 Tidal stations from which data are available. Tide gauge and tide pole data are used; data supplied by the Royal Navy, Abu Dhabi Petroleum Company and Sir William Halcrow and Partners.

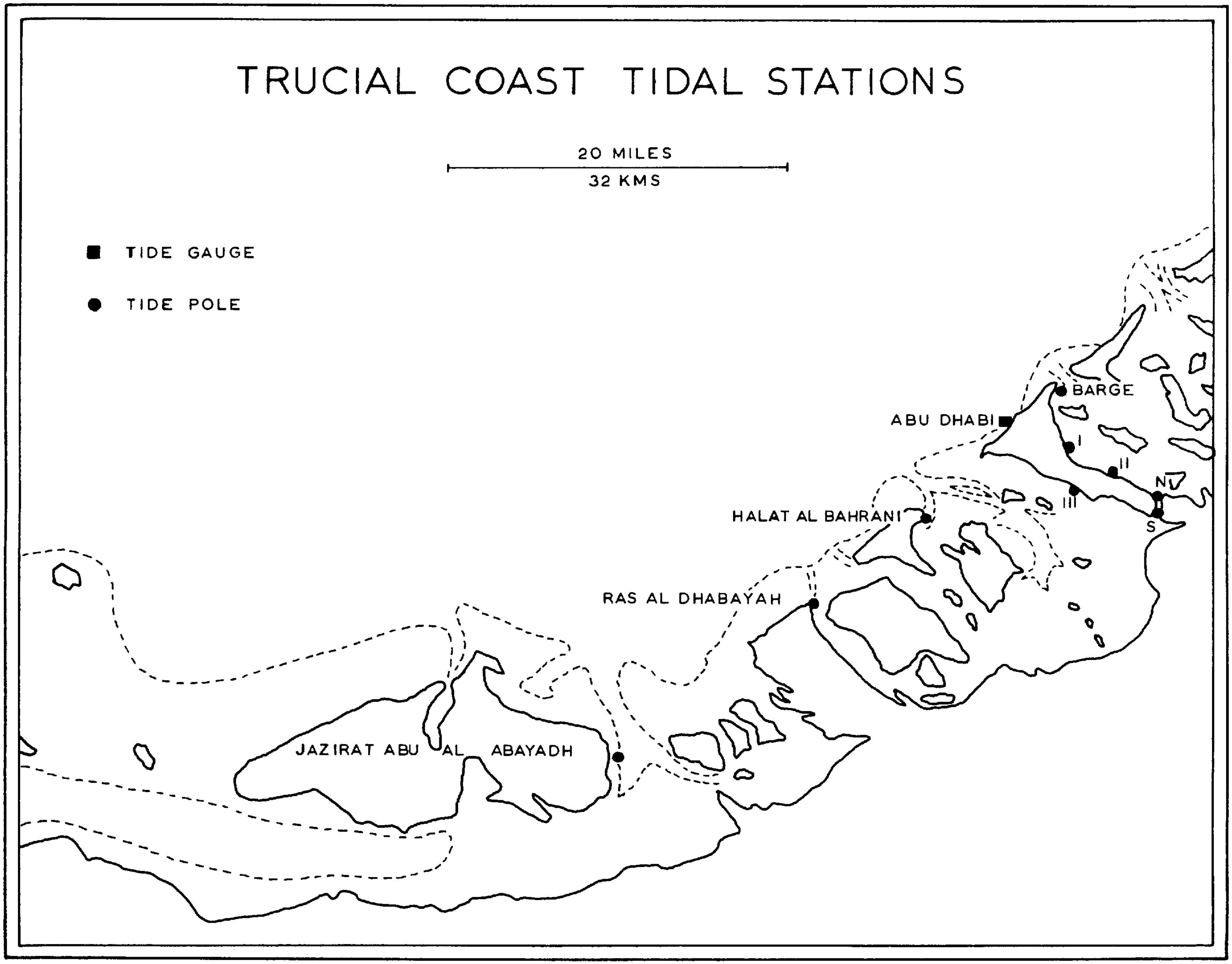

TRUCIAL COAST TIDAL STATIONS
20 MILES
32 KMS
TIDE GAUGE
TIDE POLE
BARGE
ABU DHABI
I
II
III
N
S
HALAT AL BAHRANI
RAS AL DHABAYAH
JAZIRAT ABU AL ABAYADH

Fig. 19 Inter-relationship between the tidal curves for stations at Abu Dhabi, Halat al Bahrani, Ras al Dhabayah and Jazirat Abu al Abayadh.

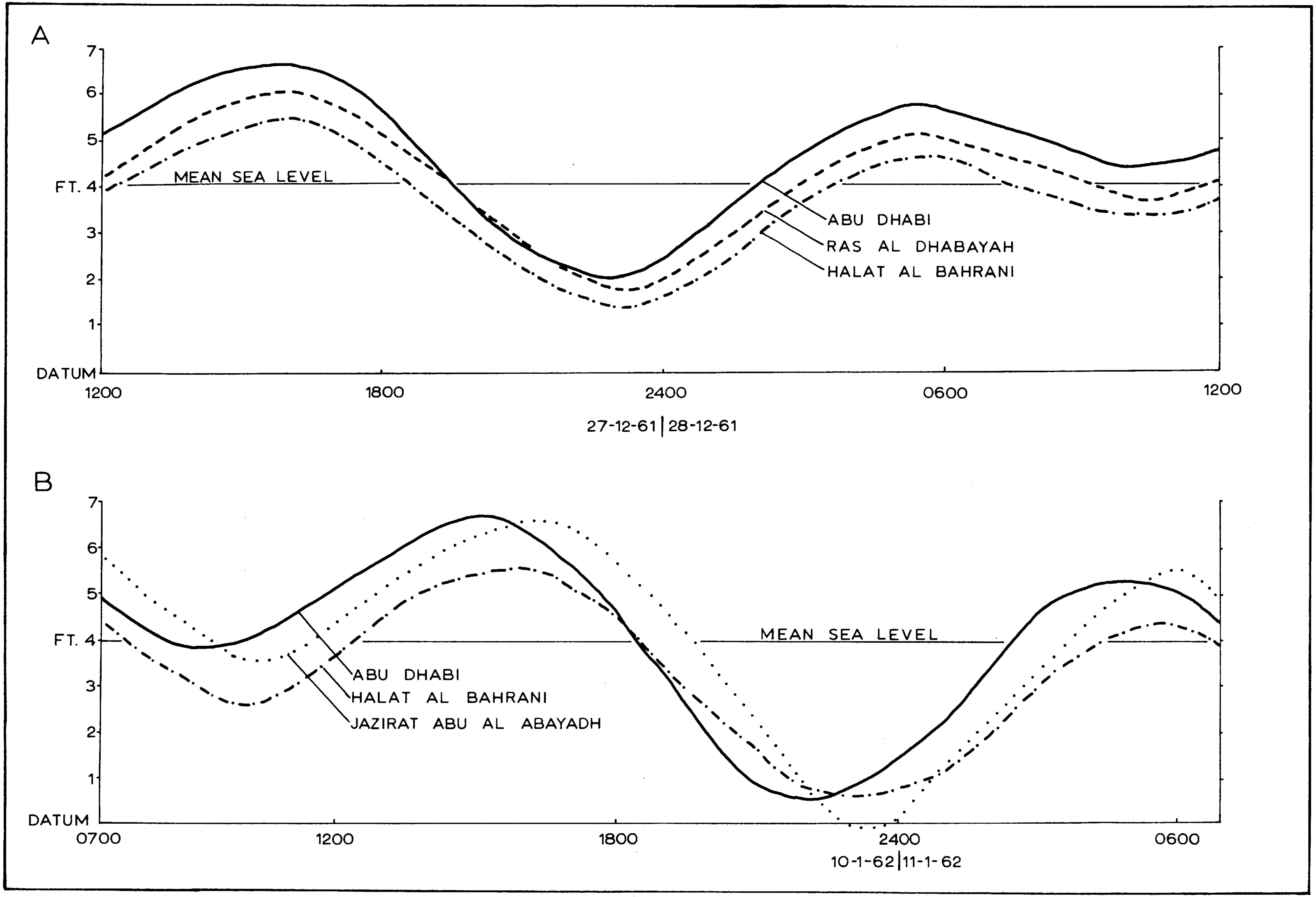
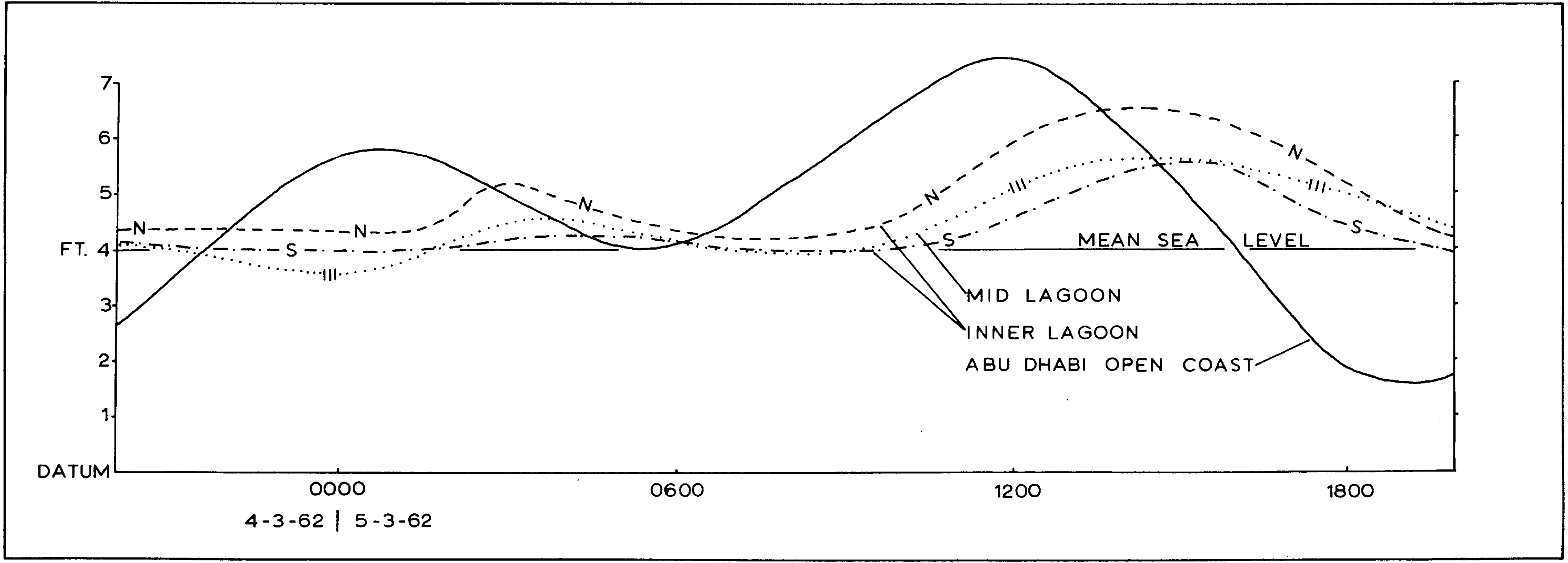

A
7
6
5
FT. 4
3
2
1
DATUM
MEAN SEA LEVEL
ABU DHABI
RAS AL DHABAYAH
HALAT AL BAHRANI
1200
1800
2400
0600
1200
27-12-61 | 28-12-61
B
7
6
5
FT. 4
3
2
1
DATUM
MEAN SEA LEVEL
ABU DHABI
HALAT AL BAHRANI
JAZIRAT ABU AL ABAYADH
0700
1200
1800
2400
0600
10-1-62 | 11-1-62

Fig. 20 Inter-relationship between the tidal curves for stations at Abu Dhabi, mid, and inner lagoon (see Fig. 18 for locations). Stations N. and S. lie north and south respectively of Abu Dhabi causeway.



7
6
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III
III
MEAN SEA LEVEL
MID LAGOON
INNER LAGOON
ABU DHABI OPEN COAST
0000
0600
1200
1800
4-3-62 | 5-3-62

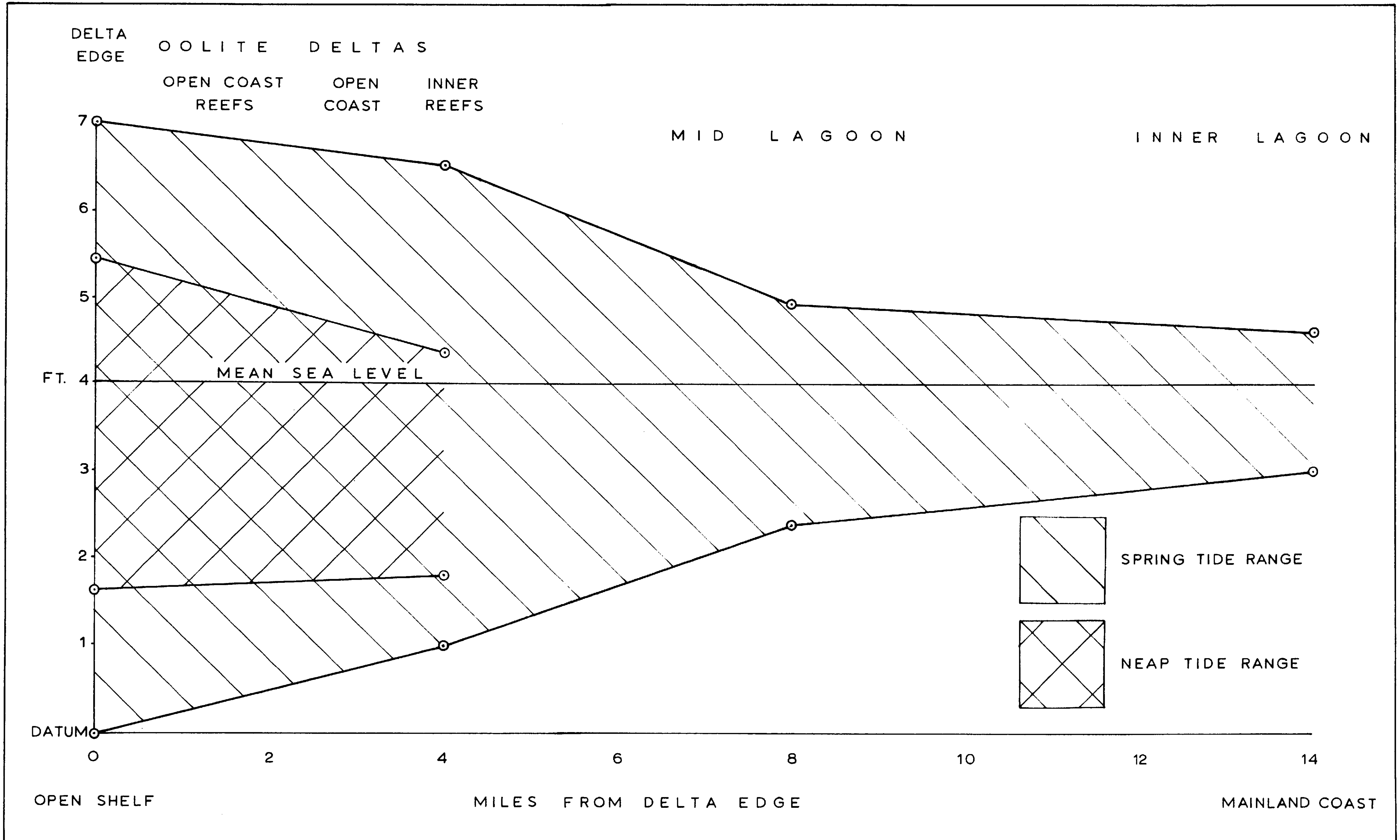
Abayadh).

Simultaneous recordings of tidal fluctuations along the coasts of Abu Dhabi island were made by Sir William Halcrow and Partners in 1961/62. Stations were situated as shown in Figure 18. The barge station has already been described; stations I, II & III were situated along the north and south coasts of the island and stations N and S along the north and south respectively of the causeway joining Abu Dhabi to the mainland. Comparison was also made with the Abu Dhabi tide gauge curve (Figure 20). Station III may be regarded as representative of stations I, II & III; tidal fluctuations are much suppressed and a considerable time lag is involved with regard to attainment of high water (approximately 2½ hours). However, this is not to be unexpected when consideration is taken of the complex routes by which waters reach such positions, flowing through narrow channels and over and around shoal banks. Owing to this complexity of water movement, apparently similar stations within these small embayments may not prove strictly comparable. This is true of station III and stations N and S. The curves for N and S show earlier arrival of flood tide waters north of the causeway than to the south; the absolute high water value for station N is increased by the ponding effect of the causeway. The tidal delay at station N is about 2½ hours, the same as at station II: at station S the delay is about 3½ hours.

Developing the concept of diminution of tidal fluctuation with increase in distance from the open sea or coast it is possible to delimit the vertical extent of the intertidal zone along such a profile (Figure 21). Maximum and minimum values are plotted for Abu Dhabi (open coast), Halat al Bahrani (up-channel), mid-embayment (III) and back embayment (S) stations. Vertical change in level is related mean sea level. Horizontal distance from the coast is indicated using Abu Dhabi open coast as zero; however, the Halat al Bahrani station is actually four miles ($6\frac{1}{2}$ kms.) up channel, the open water lying off the edge of the shoal oolite delta. The Abu Dhabi data are thus moved three miles (5 kms.) seawards and considered as open-sea values. The open coast, oolite deltas and inner coral reef areas suffer maximum tidal fluctuations in excess of 5 feet ($1\frac{1}{2}$ ms.) mid and inner areas suffer a maximum fluctuation of $2\frac{1}{2}$ feet (1 m.). Maximum fluctuations will only be experienced during periods of spring tides; during neap tides fluctuations on the open coast will be less than 4 feet, decreasing gradually inwards. There will be little difference in tidal fluctuation between spring and neap tides in the mid and inner areas.

The consideration of tidal data in this manner leads to the formulation of useful generalizations concerning the tidal effect on such a complex coastline of islands and lagoons. Figure 19 has shown that parallel and near to the open coast the tidal effects are similar and at a maximum. Figures 20 and 21 have shown that tidal

Fig. 21 To illustrate the diminution of the tidal range with distance from the open coast; open shelf and coast tidal range is about 7 ft. (2.1 ms.), mid lagoon range about $2\frac{1}{2}$ ft. (0.8 ms.) and inner lagoon range about $1\frac{1}{2}$-2 ft. (0.5-0.6 ms.). The reference level used is mean sea level. The data also indicate the vertical extent of the intertidal zone with reference to mean sea level.


DELTA EDGE
OOLITE DELTAS
OPEN COAST REEFS
OPEN COAST
INNER REEFS
MID LAGOON
INNER LAGOON
7
6
5
FT. 4
MEAN SEA LEVEL
3
2
1
DATUM
0
2
4
6
8
10
12
14
OPEN SHELF
MILES FROM DELTA EDGE
MAINLAND COAST
SPRING TIDE RANGE
NEAP TIDE RANGE

effects gradually fall off and finally reach a minimum in the inner areas. Although data are only available for a single traverse from the open coast to the inner coast, it is reasonable to assume that similar conditions will extend inwards from the open coast along the entire coastline from Abu Abayadh in the south-west to Ras Ghanadha in the north-east.

The Abu Dhabi-Ras al Dhabayah lagoon is the largest along this eastern stretch of coast. To the north-east, towards Ras Ghanadha, the lagoons progressively diminish in size. As their size decreases so the tidal and current conditions will change as smaller volumes of water are moved during the tidal cycle and distance from the open coast is reduced. Tidal currents will diminish in velocity and tidal range will not show such a marked diminution in the inner lagoon areas.

Spring and neap tide conditions will have widespread and opposed effects on the nearshore and lagoon areas. At extreme high water areas will be flooded which normally remain dry, for example, the higher crab flats and the high intertidal algal flat and sabkha areas. Seaward areas will suffer an incursion of lower salinity, open sea-waters, in fact, every part of the lagoon complex will be affected by the incoming of waters of lower salinity than those which normally prevail. Interstitial ground-waters in algal flat and other intertidal areas and in the sabkha, may suffer dilution and flushing.

During extreme low water areas are exposed which normally are covered by water; exposure may occur during the middle of the day, when exposed floral and faunal elements will suffer extreme dessication. For example, the level of extreme low water will be the effective control of the upper limit of coral growth. Another consequence of extreme low water will be the bringing of highly saline waters temporarily into more seaward areas.

(c) Tidal Currents and Water Circulation

The pattern of water circulation in the lagoons and embayments is complex. Field observations of flood tide currents are shown in Figure 22. The dominant tidal currents are indicated with heavy arrows and reference to the aerial photograph mosaic (Fig. 3) will show that these are the channel positions. On the flood tide water moves in over the edge of the oolite deltas but the greater portion flows along the deep channels. The pattern is reversed during the ebb cycle, although now the channels carry more water and velocities are higher than during the flood cycle. Current velocities across shoal banks are not as high as those in the channels.

Water in the channels, particularly on the ebb tide, exhibits extreme turbulence. Ebbing waters from adjacent shoal banks are much less turbulent and usually flow into the channels almost at right angles to the channel current direction. Ripple marking on the bank top and channel side are usually found to be at right angles.

Fig. 22 Water circulation pattern on the flood tide; heavy arrows indicate main tidal channels, smaller arrows represent flow along minor channels or across shoal banks. The ebb tide pattern is the reverse, except that more water is confined to the main tidal channels.

FLOOD TIDE CURRENTS

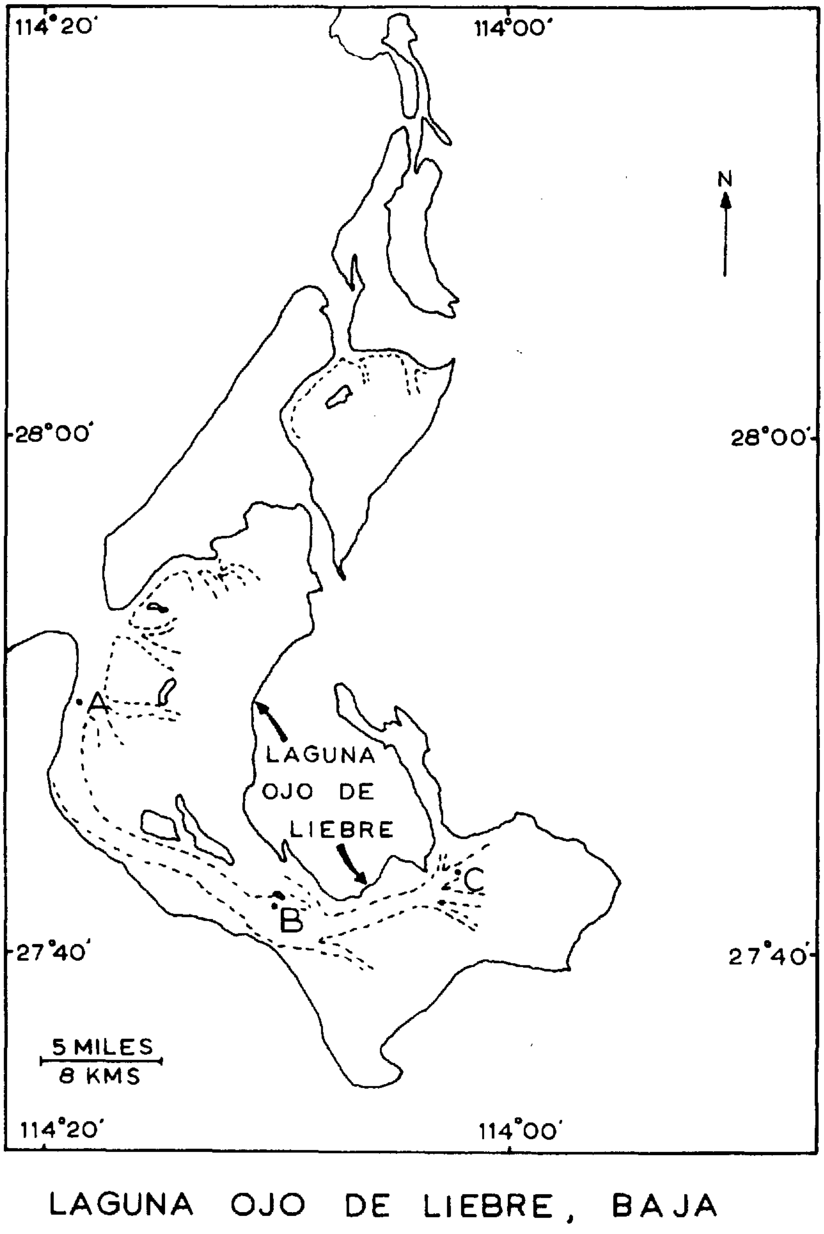
Recently, analyses and theoretical discussions of tides and currents in tidal lagoons and embayments have been made by Phleger and Ewing (1962) and Price (1963). In many ways, data from the former study are similar to those of the Trucial Coast.

Figure 23 (after Phleger and Ewing, 1962) shows the Laguna Ojo de Liebre and three stations at which tidal and current data were obtained. The tidal range is found to vary from three feet (1 m.) to a probable maximum of about nine feet (2½ m.). The tidal delay at B compared with A is 1½-2 hours; the delay at C compared with A is 2½-3½ hours. Average maximum tidal current velocities were found to be: A, 2.5 knots (5 kms./hr.), B, 1.5-2.0 knots (3-4 kms./hr.), C, 0.3-1.5 knots (½-3 kms./hr.).

Current velocities in the channel to the north east of Abu Dhabi island were measured by Sir William Halcrow and Partners and showed values of up to 1.9 knots (3.5 kms./hr.) during the ebb tide. This current velocity was recorded during a period of average tides; during spring tides velocities would be greater. The maximum current velocities were recorded at about half tide or slightly later. This compares closely with the findings of Phleger and Ewing (op. cit. p.171).

The channels to the south-west of Abu Dhabi island are narrower, deeper and better developed than that between Abu Dhabi and Sadayat islands. The area which is flooded or drained during each tidal cycle is rather larger

Fig. 23 Laguna Ojo de Liebre (modified, after Phleger and Ewing 1962). Showing stations A, B and C from which tidal and current data are available



LAGUNA OJO DE LIEBRE, BAJA CALIFORNIA, MEXICO and from observational estimations and comparisons the current velocities are greater.

As a generalisation it may be stated that during neap tide periods water movement is minimal and therefore tidal currents are minimal. During spring tide periods conditions are reversed, great volumes of water are on the move and tidal currents are now at their maximum.

The transport of sedimentary particles is a two-fold process involving water turbulence to lift the particle free from the bottom and then lateral translation of the particle by some form of current. In the shoal areas water turbulence is almost directly related to wave activity. Wave activity does not affect the characteristics of water movement in the tidal channels; water in the channels is characterised by extreme turbulence under almost all wind and tidal conditions. Under quiet wave conditions sediment movement in shoal areas will be minimal, only very fine particles already in suspension will be affected. On shoal banks and on the oolite deltas in particular, great transport of sedimentary particles will occur under storm conditions when turbulence is greatest. Flood tidal currents backed up by mass transport of water due to the onshore directed wave energy will move particles in across the deltas. Under ebb tide conditions, the tidal currents and mass transport currents will be opposed and seaward sediment movement will be restricted. There will therefore be a net movement of sediment in towards the coast and inner lagoons.

In the channels, water flow is always turbulent and potential lateral translation of particles always great. Water movement within the channels is controlled by the tidal conditions, greatest sediment transport occurring during spring tide periods when highest current velocities are experienced. As the ebb current velocities dominate over those of the flood currents the net sediment transport along the channels will be seawards.

Sir William Halcrow and Partners have determined the velocities of the tidal currents to seaward of the open coast of Abu Dhabi and Sadayat islands. It was found that, on the ebb, tidal waters flowed seawards out from the lagoon, almost at right angles to the open coast, for a distance of at least 2½ miles (4 kms.). At this distance from the coast the waters were well clear of the oolite delta, yet still heading seawards, the maximum velocity in this outer area being about 0.7 knots (1½ kms./hr.)

On the flood tide, the tongue of water which had flowed out from the lagoon on the ebb, returned. Unfortunately, no data are available to indicate the amount of mixing of open waters with the lagoon derived waters. With an onshore wind and therefore longshore drift at a minimum, it is possible that little mixing occurs at all.

TEMPERATURE:

Introduction

The Persian Gulf is probably the warmest sea on the surface of the globe and the southern shore abuts onto the great Arabian Desert, a continental area of excessively high temperatures. Thus a study of temperature distribution is in itself of considerable interest. However, if one considers the effect of temperature on floral and faunal development and its effect on the chemistry of carbonate deposition then the importance of such a study assumes even greater stature.

At higher temperatures the overall metabolic rates of plants and animals are found to be substantially increased. Many observations, covering a range of floral and faunal groups, indicate that polar and north temperate, cooler water speies grow more slowly, mature later sexually and live longer than their equatorial, warmer water counterparts (Gunter, 1957).

Temperature plays a major role in determining the productivity of any marine area and in general the crop of plankton in polar and north temperate seas is found to be much heavier than that in equatorial seas. This is perhaps slightly a false picture as although the crop at any one time may be lighter in equatorial waters, the total organic turn-over will be much more rapid.

High temperatures reduce the solubility of $CaCO_3$ in seawater and thus assist in its precipitation.

High rates of organic metabolism together with saturation of the marine waters in $CaCO_3$ result in a massive production of skeletal carbonate materials and therefore in high skeletal carbonate sedimentation rates.

Air and Sea Temperatures

General:

In all oceans the highest values of surface water temperatures, which delimit the so-called thermal equator, lie somewhat north of the geographic Equator. This is probably a reflection of the prevailing circulation in the two hemispheres and also of the presence of the great ice-covered area of Antarctica in the southern hemisphere.

Sea-surface temperatures, on average, exceed air temperatures over the sea by about 0.8°C, a reflection of the surplus radiation received by the oceans. This difference varies seasonally, in winter the air being usually much colder than the sea, whereas in summer air temperatures may be higher than sea-temperatures.

When considering a small sea such as the Persian Gulf due attention must be paid to air temperatures over the adjoining land. In general it is found that air temperatures over water are higher in winter than air temperatures over land, the reverse condition holding in the summer. In shallow water coastal regions sea-water temperatures, air temperatures over the sea and air temperatures over the land will all tend to modify each other.

(1) Air Temperatures over Land

Air temperatures over land along the Trucial Coast are often extremely high. Comprehensive data are lacking for Abu Dhabi but are available for Sharja. Sharja, like Abu Dhabi, is situated on the coast and air temperatures should be broadly comparable, although the few data available from Abu Dhabi do indicate slightly higher values (about 2°C).

Figure 24 shows the average daily, average monthly and absolute maximum and minimum temperatures recorded at Sharja over a period of five years. Of particular interest are the outstandingly high daytime temperatures for the months July-September, temperatures reaching 45°C. (47°C. was recorded at Abu Dhabi in August 1963). The coolest part of the year coincides with the period of greatest air circulation. The great diurnal variation of temperature is also to be noted, values of 11°C. commonly being exceeded.

A knowledge of air temperatures over land is required in order to interpret air and sea temperatures of the nearshore marine areas. These data are also significant when considering the sedimentary and diagenetic processes of the sabkha.

(2) Air Temperatures over the Sea

Data on sea surface temperatures and air temperatures over the sea were obtained from the Meteorological Office for the years 1954-1961. The data were analysed separately for one degree squares (Figure 25). Few vessels

Fig. 24 Air temperatures at Sharja, a coastal station 80 miles (130 kms.) to the north east of Abu Dhabi (Fig. 25). Available data indicate Abu Dhabi maximum values to be about 2°C. higher than those of Sharja. Note the wide diurnal and seasonal range and particularly the high absolute values (reaching 47°C. at Abu Dhabi).

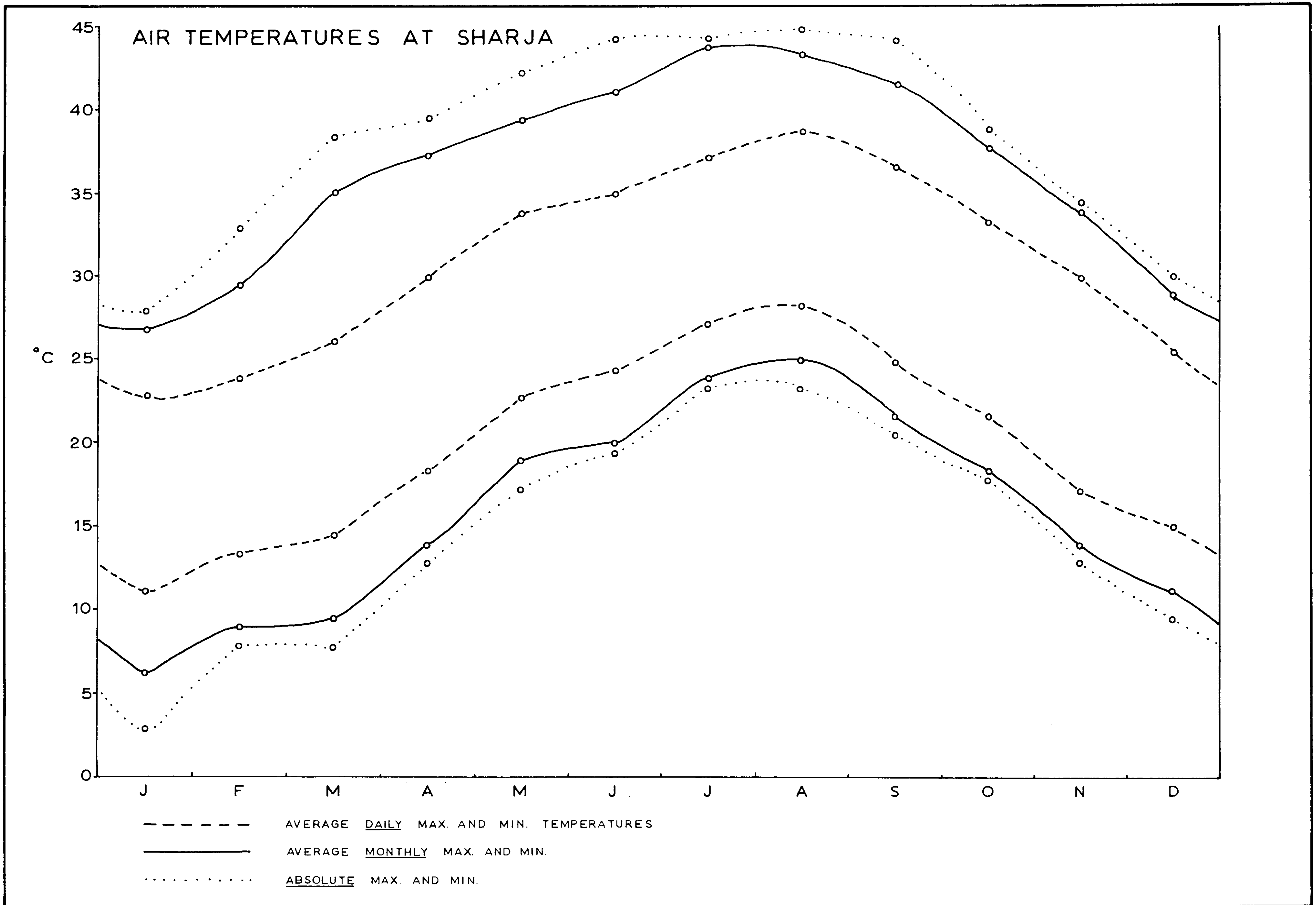
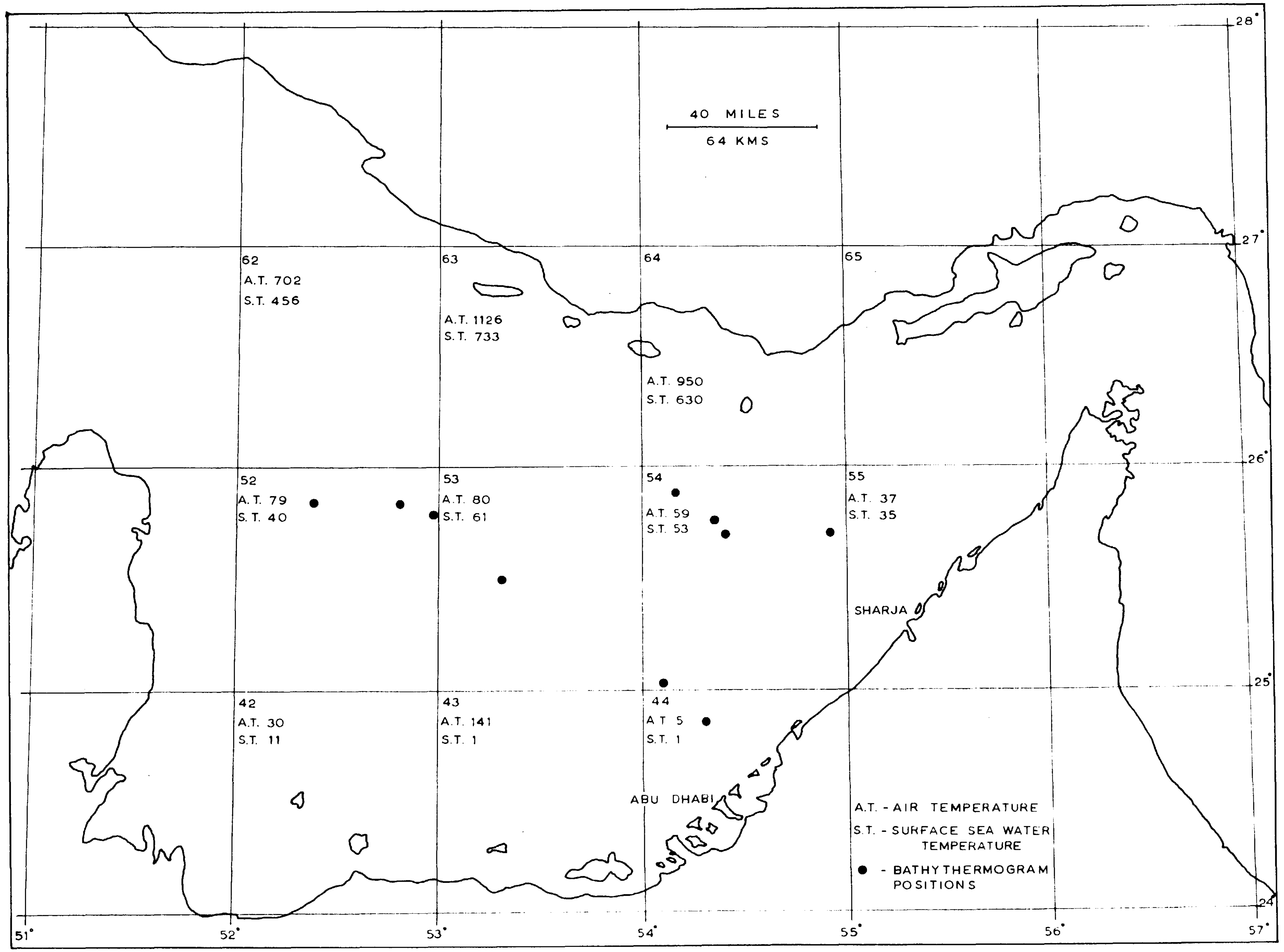

AIR TEMPERATURES AT SHARJA
°C
0
5
10
15
20
25
30
35
40
45
J
F
M
A
M
J
J
A
S
O
N
D
AVERAGE DAILY MAX. AND MIN. TEMPERATURES
AVERAGE MONTHLY MAX. AND MIN.
ABSOLUTE MAX. AND MIN.

Fig. 25 To indicate the 1° squares used in the analysis of air and sea temperatures; each 1° square has a number which is used when reference is made to it, for example, square 62 is 52-53°E and 26-27°N. Each square also bears the number of air and sea temperature readings available; these have been plotted and the data analysed and discussed in the text. The solid dots are positions of bathythermogram data, supplied by the Hydrographic Department of the Admiralty. Data otherwise were supplied by the Meteorological Office.



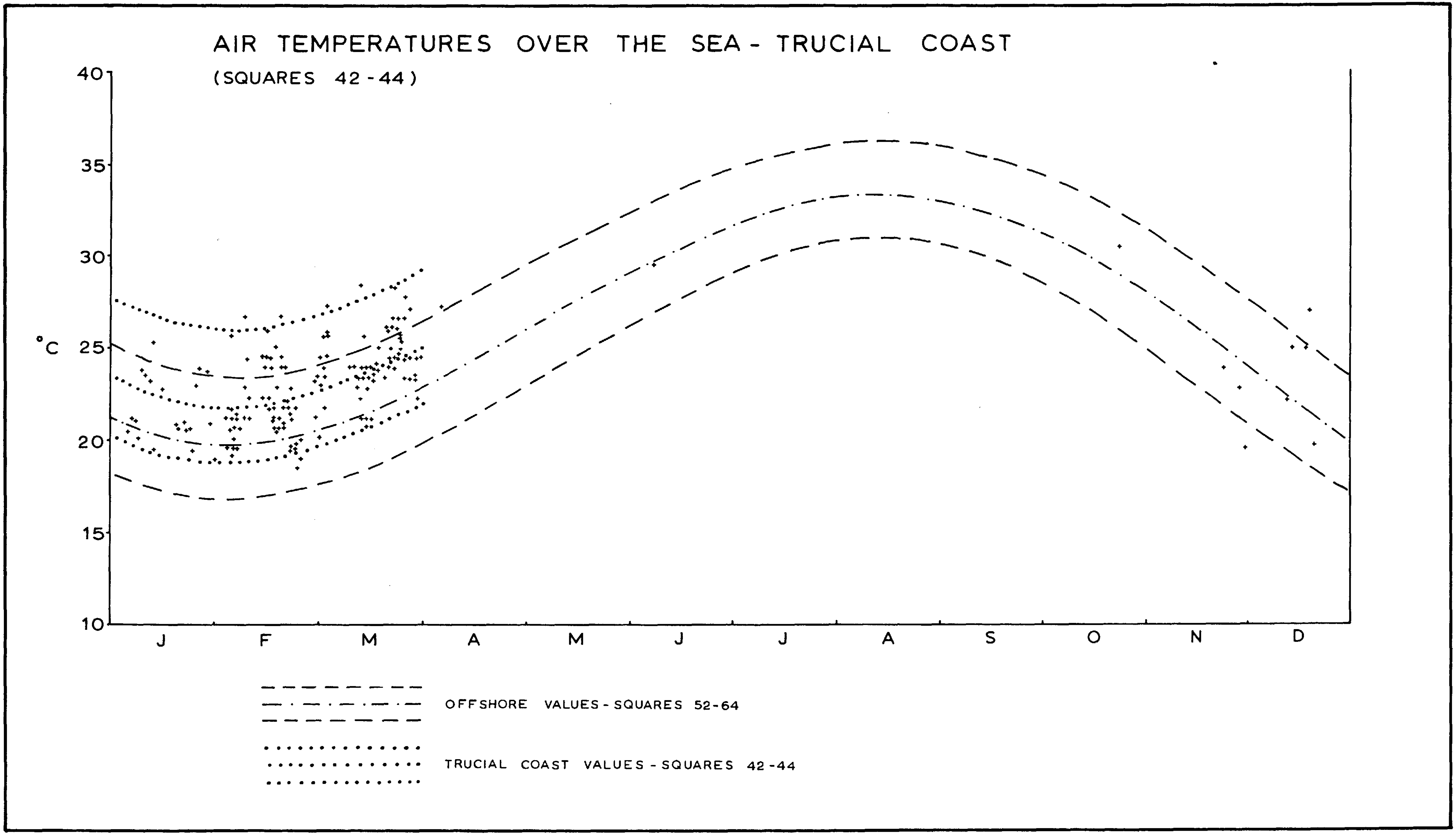
40 MILES
64 KMS
62
A.T. 702
S.T. 456
63
A.T. 1126
S.T. 733
64
A.T. 950
S.T. 630
65
52
A.T. 79
S.T. 40
53
A.T. 80
S.T. 61
54
A.T. 59
S.T. 53
55
A.T. 37
S.T. 35
SHARJA
42
A.T. 30
S.T. 11
43
A.T. 141
S.T. 1
44
A.T. 5
S.T. 1
ABU DHABI
A.T. - AIR TEMPERATURE
S.T. - SURFACE SEA WATER TEMPERATURE
● - BATHYTHERMOGRAM POSITIONS
28°
27°
26°
25°
24°
51°
52°
53°
54°
55°
56°
57°

visit the nearshore Trucial Coast region and data from squares 43-44 are therefore scanty. The squares lying astride the oil tanker routes (62-64) have greatest density of data and soundly established annual temperature variation curves can be derived for them. Temperatures are available for all months of the year and all hours of the day; the resulting curves indicate the combined seasonal and diurnal variations of temperature.

Squares 52-54 have less data but when plotted and compared with those for squares 62-64, the resulting temperature curves are found to be almost identical. Data for squares 42-44, the nearshore Trucial Coast region, are scanty but show some indication of higher temperatures than those recorded further offshore. Data for square 55 have more in common with those of squares 52-54 than those of squares 42-44; this parallels the relationship found between air temperatures on land at Sharja and Abu Dhabi (i.e. slightly increased temperatures near Abu Dhabi and to the west).

Figure 26 shows the open Gulf air temperature curve, the envelope of 6°C. representing seasonal and diurnal variation in temperature (to be compared with 11°C. for the variation in air temperature over the land). The nearshore Trucial Coast data are shown and the probable temperature curves indicated, there being an increase of 2°C. over the offshore values. It may also be seen that the seasonal and diurnal temperature range of the nearshore region is

Fig. 26 Air temperatures over the sea for the Trucial Coast embayment. Nearshore Trucial Coast values are slightly higher.


AIR TEMPERATURES OVER THE SEA - TRUCIAL COAST
(SQUARES 42 - 44)
40
35
30
°C 25
20
15
10
J
F
M
A
M
J
J
A
S
O
N
D
OFFSHORE VALUES - SQUARES 52-64
TRUCIAL COAST VALUES - SQUARES 42-44

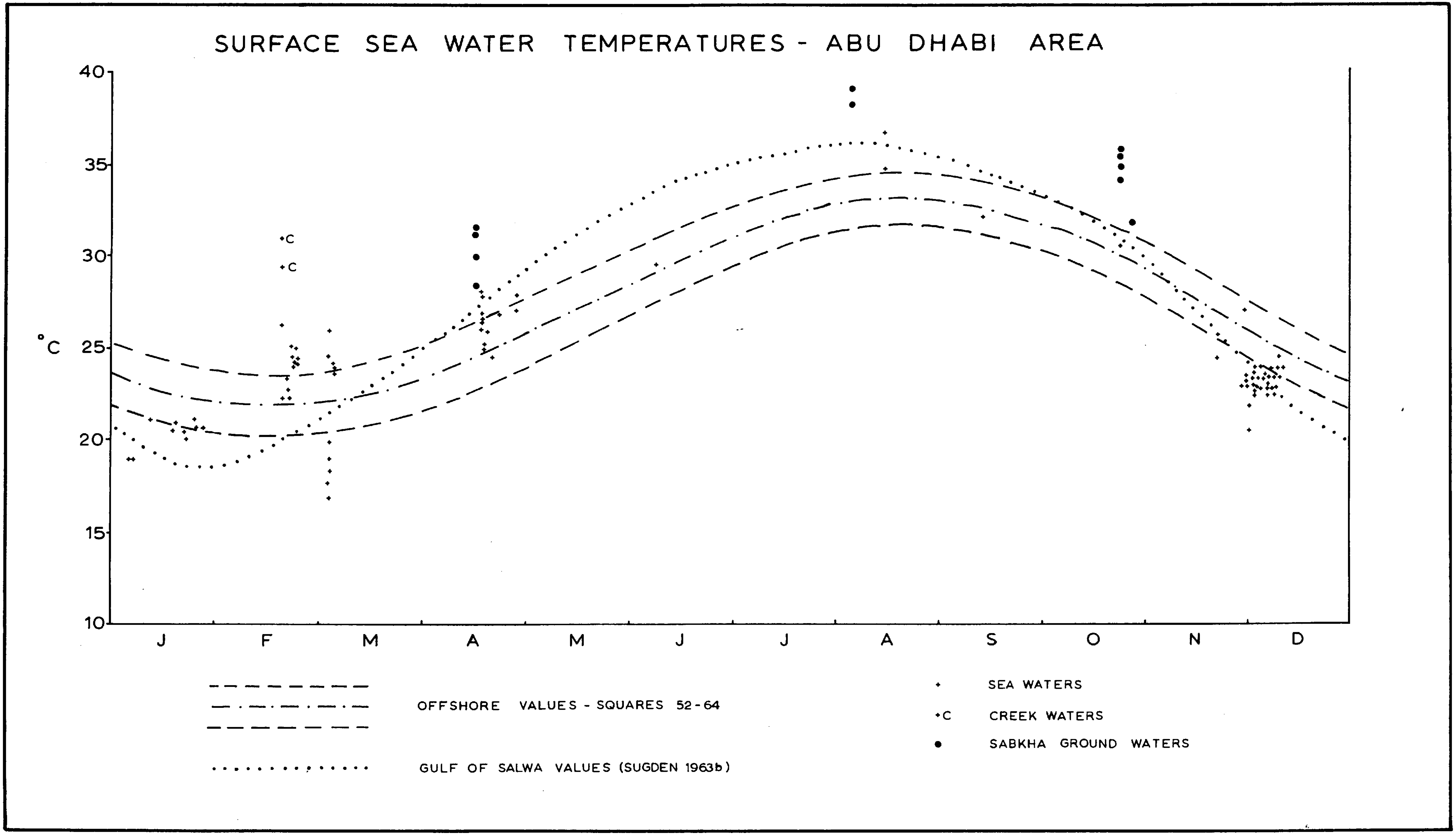
slightly greater than that experienced offshore. Mean air temperatures over the open sea reach a maximum of 33-34°C. in early August and a minimum of 19-20°C. in early February.

(3) Sea Surface Temperatures

Curves of sea surface temperature were derived as explained in the previous paragraph. Figure 27 shows the open Gulf sea surface temperature curve, the envelope of seasonal and diurnal temperature variation being about 4°C. (to be compared with 6°C. for the variation in air temperatures over the open sea). Sea surface temperatures over the open sea reach an average maximum value of about 33°C. in late August and a minimum of about 22°C. in late February.

Recorded on Figure 27 are Meterological office data from the nearshore region (squares 42-44), data from Abu Dhabi recorded during the present study, and the annual temperature variation curve for the Gulf of Salwa (Sugden, 1936b). Many of the Abu Dhabi values are for waters in the restricted environments between and behind the coastal islands; these waters will be greatly affected by air temperatures over the adjacent land. For example, the series of low water temperatures for early March were recorded in the morning when air temperatures over land were low, whereas the two creek-water temperatures, from exceptionally restricted waters, were recorded in the late afternoon in the heat of the day. The nearshore values closely reflect the greater diurnal temperature range experienced by the

Fig. 27 Sea-water temperatures for Abu Dhabi nearshore and lagoon areas. Also included are restricted creek water values and ground water values for the sabkha. The annual temperature variation curve for the Gulf of Salwa (after Sugden 1963b) is also shown.


SURFACE SEA WATER TEMPERATURES - ABU DHABI AREA
40
35
30
25
20
15
10
°C
C
C
J
F
M
A
M
J
J
A
S
O
N
D
OFFSHORE VALUES - SQUARES 52-64
GULF OF SALWA VALUES (SUGDEN 1963b)
SEA WATERS
C CREEK WATERS
SABKHA GROUND WATERS

adjacent land areas.

Also recorded on Figure 27 are temperatures of ground-waters from shallow pits dug in the sabkha; the high values indicate the adsorption of energy by the dark brown sabkha surface and the consequent heating of the subsurface waters at depths of 2-4 feet. The significance of these high values will be discussed more fully elsewhere, (Chapter VI).

(4) Vertical Distribution of Sea-Water Temperature

Few data are available of sub-surface sea-water temperatures for the Trucial Coast embayment. The Hydrographic Department of the Admiralty have kindly supplied some bathythermograms from the region from which a few generalisations may be made. A significant thermocline is only developed in the Gulf during the months July-September, lying typically at about 50 feet (16 ms.) below the surface. Temperatures are usually 32-35°C. above and 27-30°C. below the thermocline. A diurnal thermocline is commonly developed down to about 8 feet (2 ms.), with temperature differences above and below amounting to 2-3°C.

Thus the sea-floor and its associated flora, fauna and sediments lying above 50-60 ft. depth (15-20 ms.) will be affected by a similar seasonal temperature range as is found in open surface waters (22-35°C.); at greater depths the temperatures will be less variable (21-27°C.).

The significant temperature controls operative in the nearshore sedimentary environment may be summarised:-

(a) Range of seasonal and diurnal variations of air temperatures over the land, and particularly the high absolute values; both these aspects affect the nearshore marine areas and the latter are of significance in producing the high temperatures of the early diagenetic environment of the sabkha. During low water intertidal areas will be subjected to the full range of air temperatures experienced over the land.

(b) Range of seasonal and diurnal variation of air temperatures over the sea; will be modified in coastal regions by (a), and both will thus affect the temperatures experienced by the nearshore marine areas.

(c) Sea-surface temperatures in nearshore areas will be modified by (a) and (b). The range of sea surface temperatures will be greatest in very restricted inner lagoonal areas. This range will decrease gradually as water depths and circulation increase, and finally approach the normal open shelf temperature variation values. Sea surface temperatures may be particularly critical in promoting extraction of $CaCO_3$ from surface waters.

(d) Vertical distribution of sea-water temperature; all nearshore sedimentary environments will suffer extremes of temperature changes owing to the shallowness of the water. This may be an important control with respect to temperature sensitive floral and faunal elements. Its effect will also partially control the mineralogy and minor chemistry of the carbonates precipitated.

RAINFALL, EVAPORATION & HUMIDITY

Rainfall

Rainfall is infrequent but when it does occur it is commonly torrential, particularly to the east, and north, in the vicinity of the Oman Mountains. However, the amount of rain generally decreases in a south-westerly direction and thus the Trucial Coast, especially the western portions, are particularly arid.

Rainfall data for Tarif for the years 1958-1964 are given below:

Year	Rainfall (cms.)	Highest Individual Fall
1958	2.87	1.37
1959	6.73	2.44
1960	2.67	1.52
1961	1.70	1.04
1962	0.33	0.30
1963	5.99	2.59
1964 (January only)	5.66	1.98

In May 1963 and January 1964, torrential rains fell along much of the Trucial Coast. Gale force north westerly winds accompanied the rains and the combination of rain plus exceptionally high tides resulted in the flooding of hundreds of square miles of the coastal plain or sabkha.

The effect of heavy rainfall will be to temporarily dilute the lagoon waters. It may also flush capillary waters from the upper levels of the sabkha sediments and dissolve away the surface salt crust.

Evaporation

Evaporation rates in the Persian Gulf are generally high. Privett (1959) states the mean annual evaporation of the whole Gulf to be 144 cms., but that there are indications of rates being higher in the northern half from June to February. Further interpretation of his data has indicated the probable annual evaporation rates of the northern and southern parts of the Gulf to be 165 cms. and 124 cms. respectively. Evaporation rates are high during the winter months and low during the summer months.

In the Atlantic, Indian and other oceans of the world the water salinity remains constant and it is found that the apparent net evaporation is balanced by the precipitation together with run-off and inflow from land areas. But, in the Persian Gulf the amount of precipitation, run-off, and inflow is small. The Gulf if thus a region of net evaporation, losses being made up by active inflow of lower salinity Indian Ocean water at the Strait of Hormuz (Emery, 1956). This regime accounts for the exceptionally high salinity waters which occur in the Gulf.

Sugden (1963 b) has determined the annual variation in salinity at three coastal stations in the Gulf of Salwa. Two of the stations were situated on the open coast (Fig. 28A) and one in a restricted lagoon. The former showed an annual salinity range of about 2‰, being lowest in winter and highest in summer, although random fluctuations caused by wind and tide exceeded the seasonal variation. The latter station, comparable in its position to the restricted lagoonal environments near Abu Dhabi, showed an annual variation in salinity of 7-8‰, the highest values being recorded in the summer. From this indirect evidence it may be assumed that evaporation rates are highest along the Arabian Shore in the summer months.

Humidity

Relative air humidities are generally high, despite the lack of cloud and rain experienced over the region. Early morning values for Sharja are 75-80% during the winter, falling to 60% during the summer; late afternoon values are fairly constant throughout the year at 55-60%.

High humidities and night-time dew provide a fairly constant source of water for the land flora which plays a role in partially stabilising some dune sands. Without this source of water it is doubtful whether the plants could survive.

SALINITY

Studies of the hydrography of the Persian Gulf have been made by Emery (1956) and Sugden (1963). Figure 28 shows the surface salinity distribution determined by Sugden and indicates the extremely high values occurring along the Arabian Shore. To be noted is the gradual fall in salinity offshore and also north-east along the Trucial Coast. At Dubai, coastal salinities have fallen to 40‰ or slightly below. This fall in salinity is marked by the incoming of new faunal elements, forms which the author has found on Masira Island, lying off the Indian Ocean coast of Arabia and also along the Indian Ocean coast of the Oman Mountains at Khor Fakhan and south towards Sohar, (Chapter III).

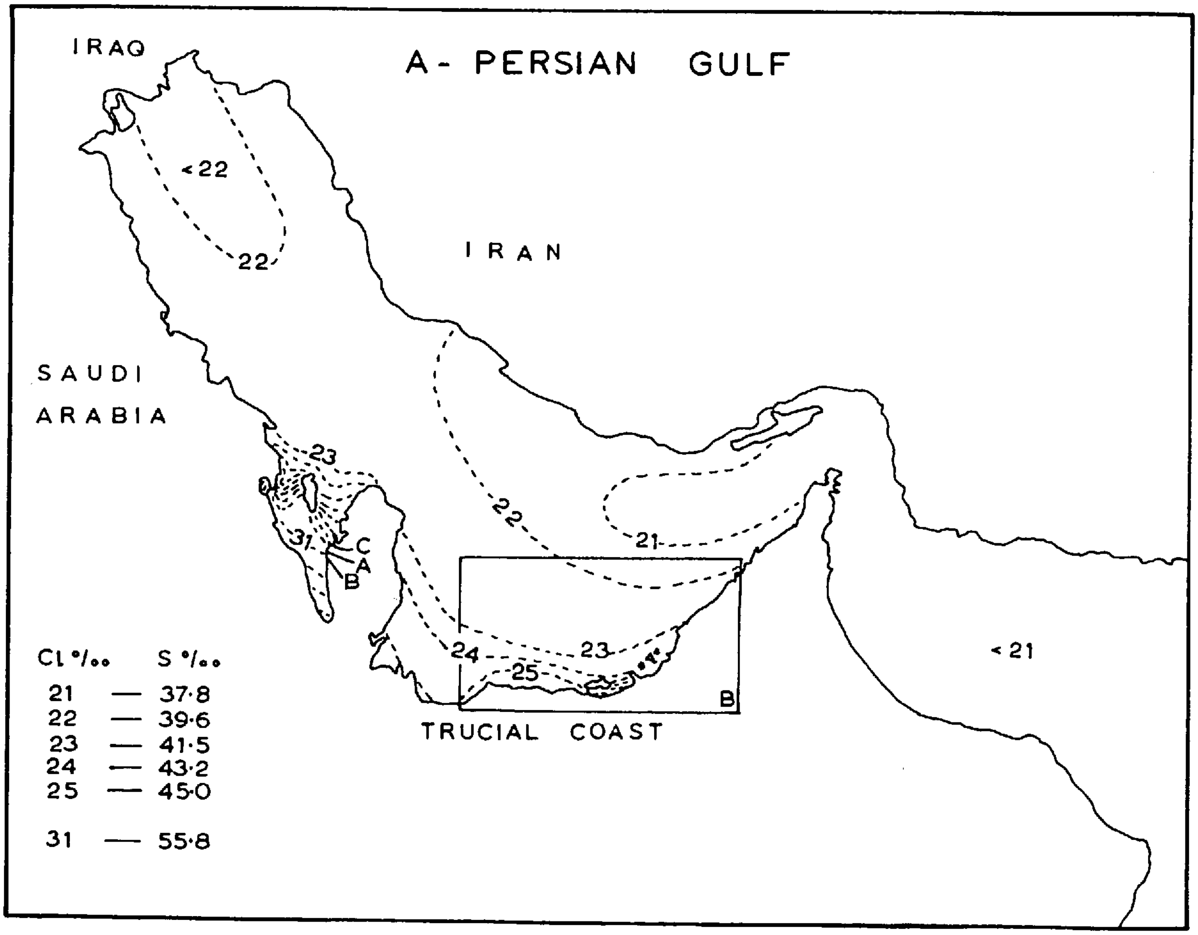
The high salinities of the Persian Gulf are sufficient to distinguish this region of Recent carbonate sedimentation from any others so far studied, for example the Bahama Banks have a salinity range of 36-46‰ , compared with a Persian Gulf range of 37‰ to more than 65‰ .

The region is one of net evaporation and, as Emery has shown, normal salinity Indian Ocean waters (36‰ S), flow into the Gulf at the surface, to balance the loss by evaporation and outflow. Highly saline, dense waters do not accumulate in the deeper parts of the Gulf as there is no sill to pond them back; they thus flow out at low level into the deeper waters of the Gulf of Oman.

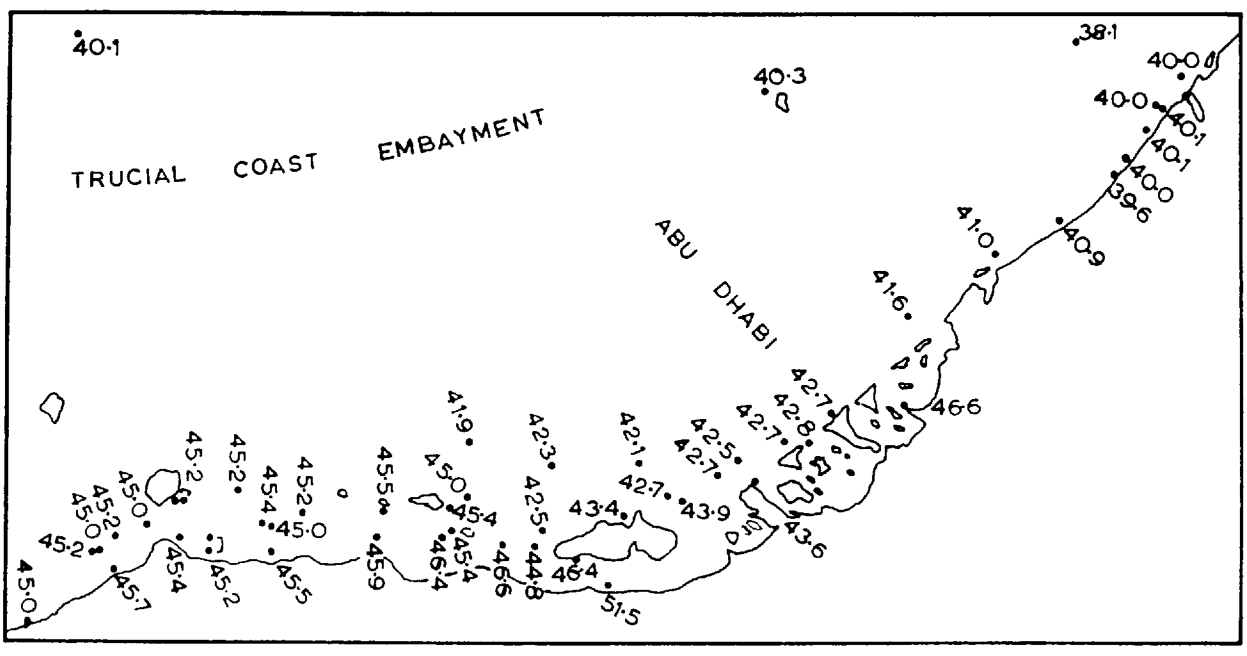
Fig. 28A Mean annual surface salinity in the Persian Gulf (after Sugden 1963b).

Fig. 28B Trucial Coast salinity data from Sugden (1961)

MEAN ANNUAL SURFACE CHLORINITY/SALINITY



B - TRUCIAL COAST



(In this study the concentration of sea-water will be referred to as salinity and expressed in parts per thousand (S.‰). This is the form with which most geologists are probably familiar. Sugden, however, has used chlorinity as an expression of concentration. Salinity and chlorinity are related by an empirical relationship, salinity = chlorinity x 1.8).

A portable National Institute of Oceanography salinity meter was used for field determination of sea and ground water conductivities. The instrument was calibrated to read salinity directly and was periodically checked against a standard sea-water sample. It was necessary to dilute all samples with distilled water as the instrument could be used for direct measurement only up to salinities of 42‰ .

Selected sea and ground waters were brought back to this country in standard sample bottles and their salinity determined using the normal oceanographic chloride titration technique. These samples were also checked in the laboratory on the National Institute of Oceanography thermostat salinity meter. It was then possible to compare laboratory titration and conductivity measurements with conductivities determined in the field.

It was found that with the sea-waters conductivity and titration techniques gave closely comparable results, even though the samples required an initial dilution for conductivity determination. The relationship between chlorinity and conductivity holds true certainly up to salinities of 65‰ . Ground-water values showed poor agreement and with increasing concentration the "chlorinity" was found to bear less and less relationship to the total ion concentration determined by conductivity measurements.

The water sampling stations are shown in Figure 29 which shows field salinity meter data and laboratory salinity meter and titration data. A few additional titration determinations by Sugden and Halcrow are also included.

The sea-water data have been grouped together in Figure 29A to show the range in salinities experienced by the various physiographic and sedimentary sub-environments:-

(a)	Open Sea	42-43‰	
(b)	Open Sea Reefs	42-46‰	
(c)	Oolite Banks	42-45‰	
(d)	Inner Reefs	43-46‰	
(e)	Mid-areas	42-47‰	
(f)	Inner Areas	47-67‰	
(g)	Creek Waters	46-62‰)	areas of very
(h)	Algal Pools	53-89‰)	restricted circulatio

(Ground water values will be considered in a later chapter).

The values are quoted to the nearest whole number as further accuracy is not warranted. The salinity at any place will depend on the state of the tide, on wind

Fig. 29 Map showing salinity distribution in the near-shore and lagoon surface waters near Abu Dhabi. Isohalines are not marked as in any locality the salinity varies seasonally and also depends on the tidal phase.

SURFACE AND GROUND WATER SALINITY

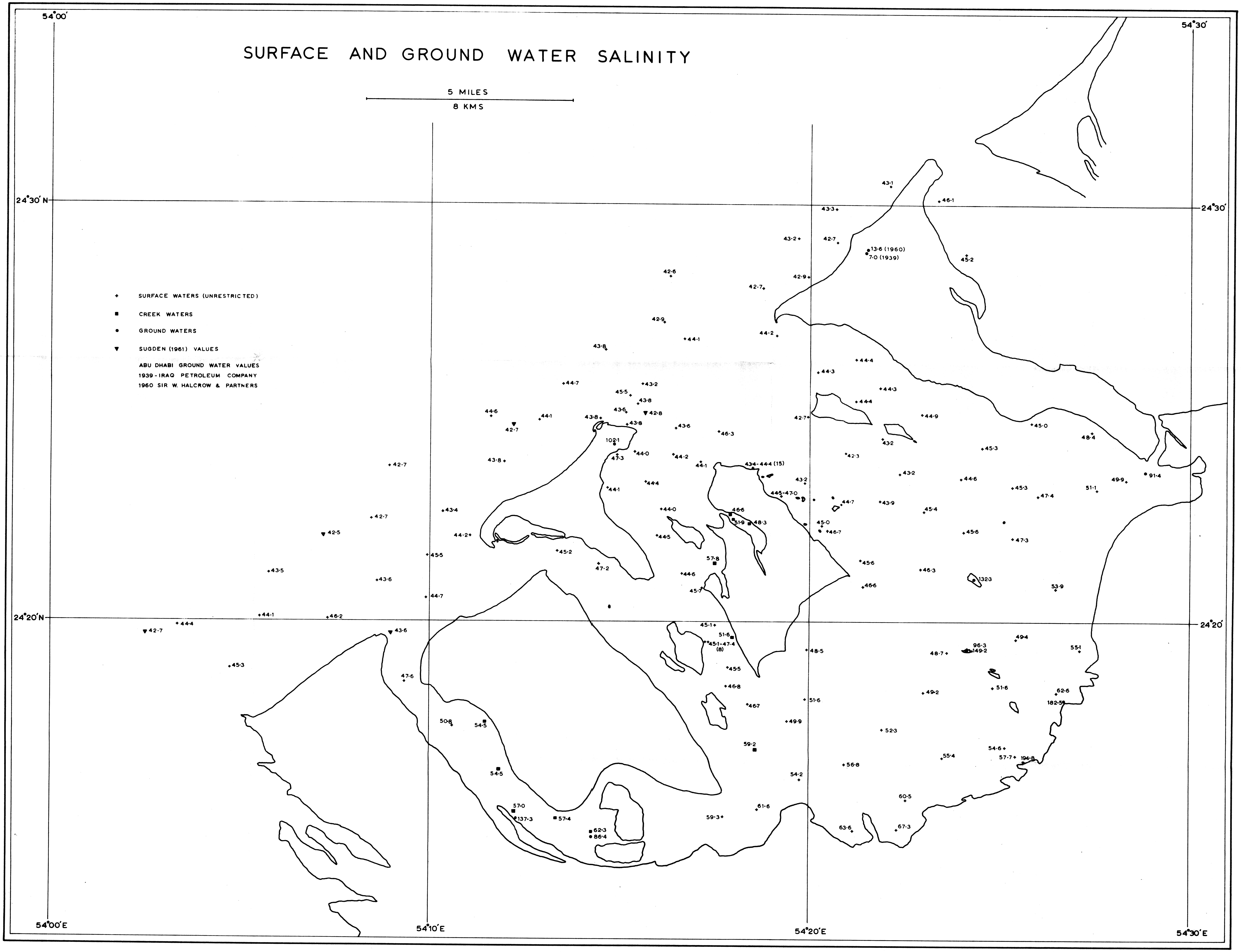
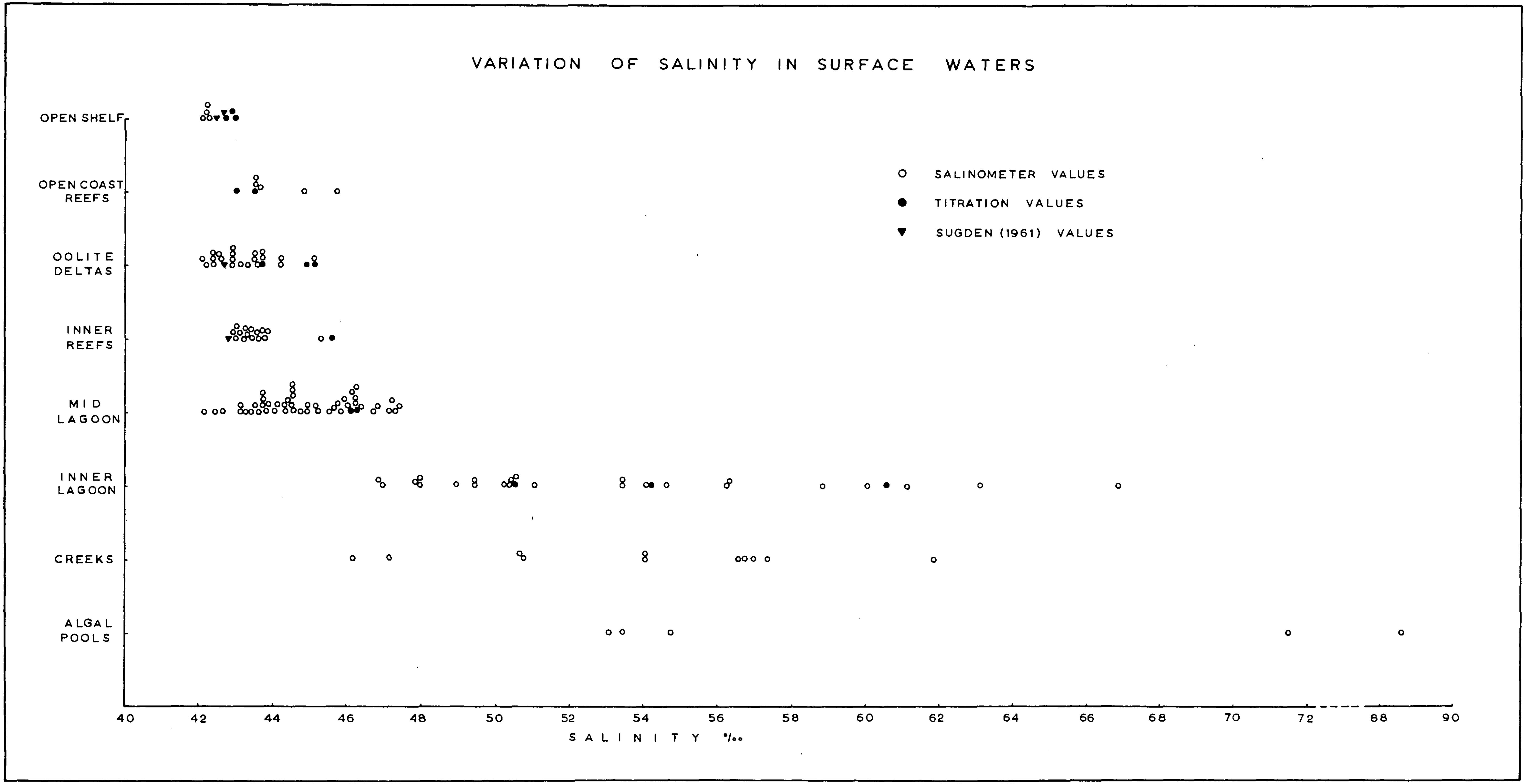


Fig. 29A To show the variation in salinity in the surface waters of the various parts of the near-shore and lagoon areas. Data are from field salinometer measurements and laboratory titration determinations. A few values determined for the area by Sugden (1961) are also included.



VARIATION OF SALINITY IN SURFACE WATERS
SALINOMETER VALUES
TITRATION VALUES
SUGDEN (1961) VALUES
OPEN SHELF
OPEN COAST REEFS
OOLITE DELTAS
INNER REEFS
MID LAGOON
INNER LAGOON
CREEKS
ALGAL POOLS
40 42 44 46 48 50 52 54 56 58 60 62 64 66 68 70 72 88 90
SALINITY ‰

activity, the season, and probably other factors as well. At high tide, values will be lower, and at low tide values will be higher than the average. A station on the north coast of Jazirat al Ftaisi, with fairly free water circulation, showed a tidal variation in salinity of 1‰; a station one mile to the south-east, with a more restricted circulation, showed a tidal variation of 2.5‰. A coastal station off the low island between Jazirat al Ftaisi and Jazirat abu Kashasha showed a tidal variation in salinity of 2.3‰, but of interest here was that at any one time there existed a narrow body of very shallow water swinging along parallel to the shore, which was up to 1.8‰ more saline than water a few yards further offshore. These tidal variations in salinity will reach a maximum in the innermost restricted areas, where the total variation will exceed 10‰ in some places. The narrow zone of very saline water, sometimes exceeding 60-65‰, will move seaward during low water spring tides; during high water spring tides waters of salinity 50-55‰ will temporarily occupy the seaward parts of this innermost zone.

The offshore and seaward areas show a simple relationship, the open sea-waters becoming slightly concentrated in the seaward areas. The effect of the tidal channels in relation to water circulation can be seen from Figure 29. A broad tongue of low salinity water occupies much of the area to the north-east of Jazirat al Ftaisi; this water has evidently reached this position without undergoing the concentration found in the waters crossing the shoal banks. The mid-areas have fairly high salinities, and the effect of the tidal flow of water southward between Abu Dhabi island and the mainland can be easily seen, this inner corner of the lagoon having rather lower salinities than equivalent inner areas to the south. The inner areas show extremely high salinities; the waters probably suffer very little exchange with less saline waters and merely move back and forth under tidal influence, finally becoming interstitial waters in the intertidal and sabkha sub-environments.

Sugden determined the annual variation in sea-water salinity at three stations in the Gulf of Salwa (stations A, B, C, in Figure 28A). Stations A and B, on the open coast and with moderately free water circulation, showed daily variation owing to tidal and wind distribution effects to exceed any annual cycle of salinity variation. Station C, however, in the restricted lagoon, showed a marked seasonal variation, salinities being 51‰ in February and 59‰ in August. The seasonal variation in salinity in the restricted lagoonal areas near Abu Dhabi may well be similar to that recorded at Station C in the Gulf of Salwa..

All salinity measurements were made on surface sea-water samples. However, from the turbulence and effectiveness of the tidal circulation, little vertical variation in salinity would be expected. Phleger and Ewing (1962) determined surface and bottom water salinities in the Laguna

Ojo de Liebre and found the waters to be isohaline at most stations. But at a few stations the bottom waters were more saline, this condition occurring following nocturnal cooling of the highly saline lagoon waters. This effectively increased their density and gave rise to this type of haline overturn, differing from most other types in probably being developed on a diurnal rather than a seasonal cycle as in more open bodies of water.

Offshore in the Trucial Coast embayment, Wells and Illing (1963) have recorded a slight increase in salinity in bottom waters.

pH

Sea water has a typical pH range of 8.1-8.3. Values are usually slightly lower at night and slightly higher during the day.

Daytime pH measurements of unrestricted waters near Abu Dhabi showed a range in values of 8.1-8.4. In more restricted areas, such as algal pools, pH values of up to 9.75 were recorded.

Consideration of pH values in an environment of carbonate deposition is critical because the solubility of $CaCO_3$ is greatly affected by changes in pH.

CHAPTER III

ORGANISM COMMUNITIES AND SEDIMENTS.

GENERAL

The physiographic units described in the earlier chapter are characterised by particular organism communities and sediments. Many organisms typify more than a single physiographic unit but the units can be differentiated from each other on the combined characteristics of their bottom sediments and organism communities.

In this study, nearly 800 surface sediment samples were collected from above high water mark, from the intertidal zone and from the marine areas. Nearly 60 cores, up to one metre in length, were collected, mainly in shallow water, intertidal or land areas. An attempt was also made to determine the distribution of organisms and to assess the part they play in the facies development.

The sampling grid was devised to ensure a reasonable cover over the area; gaps and irregularities in the grid are caused by inaccessibility or rapid local changes in bottom and sediment type. For example, if samples along a traverse are widely and evenly spaced then this reflects a general uniformity of bottom and sediment character. Where local variations were found then these areas were examined and sampled in more detail. Water depth and bottom features were recorded along all traverses. On collection, sediment samples were examined and their superficial features noted. A limited amount of sediment examination with a stereoscopic microscope was carried out in the field and in this way a rough facies map was constructed. Laboratory studies have refined this map but not altered it markedly in any way.

The pH and Eh values of the interstitial waters of the sediments were measured immediately they were collected, using a portable, battery-operated pH meter. This instrument could be read to within 0.1 pH units and to within 10 millivolts Eh units.

Floral and faunal samples were collected at many stations and inter-station distribution of the major forms noted and related where possible to water depth, substrate or other environmental factors. The major elements of the flora and fauna are given below together with ecological data and notes on the part the organisms play in the development of the sedimentary facies.

From the highly saline Gulfs of Bahrain and Salwa, Sugden (1963b) has reported an abundant and diverse fauna of algae, foraminifera, sponges, bryozoa, worms, molluscs, arthropods and fish. Notable for their absence were corals, sharks and turtles. Along the Trucial Coast corals and turtles are common but sharks (except for sand sharks) do not enter the lagoons and are limited to the shelf areas. Dolphins are very common in the Trucial Coast shelf and lagoon areas. There are many similarities between the Trucial Coast and Gulfs of Bahrain and Salwa but also notable differences, which will be considered in more detail when discussing the groups concerned.

FLORA

Land flora is sparse and limited to a few species, but does play a role in the partial stabilisation of some windblown sands. In the tidal creeks and swamps and along some sheltered mid and inner lagoon coasts, *Arthrocnemum glaucum* and *Avicennia marina* (Mangrove) are abundant (Figs. 12 and 13). The former is very widespread and effectively stabilises sediment in the upper part of the intertidal zone throughout much of the inner lagoon.

In the marine areas, the dominant higher plant is *Halodule (Diplanthera) uninervis*, which forms dense mats, often completely obscuring the bottom sediments. Below low water mark the grass-like leaves of *Halodule* grow up to 8 ins. (20 cms.) in length, but in the lower parts of the intertidal zone the leaves are shorter and often only 1-2 ins. (4-5 cms.) long; the roots form a tangled mat in the uppermost 4-6 ins. (10-15 cms.) of the sediment. It is particularly common in the mid lagoon and plays an important role in the entrapment of fine sediment. The entrapment of particles is greatly aided by the growth of small filamentous green algae on the exposed surfaces of its leaves. Along the Trucial Coast, *Halodule* plays the equivalent role to that of *Thalassia* in the Bahamas. Other common marine plants include *Halophila stipulacea* and *Halophila ovalis*; the former is common in shelf and mid lagoon areas whereas the latter occurs mainly in the mid lagoon.

In addition to sediment entrapment, the dense plant cover may assist in the precipitation of $CaCO_3$ by locally raising the pH of the waters during daytime photosynthesis. Below the sediment surface plant debris is not abundant, although a few stems and roots are to be found infilled with mud. Beales (1963) has suggested that this type of trace fossil may be the only indication of the former existence of a plant cover in many areas of carbonate mud accumulation.

Algae are locally abundant. On rock surfaces on the oolite deltas and on dead coral heads on the inner reef, large brown seaweeds grow in profusion. Brown seaweeds are common offshore and are drifted onto open coasts, where large accumulations are to be found near high water mark. On an accreting coastline these produce thin peats within the sediment sequence. In the mid and inner lagoon, brown seaweeds are uncommon. Smaller red and green algae occur throughout the area but do not appear to be significant members of the total community. Two small patches of Avrainvillea c.f. A. amadelpha, were found growing in shallow water off the south-west coast of Jazirat al Ftaisi. Some species of Avrainvillea are known to produce minute aragonite needles in the Bahamas (Cloud, 1961).

The great spreads of tough algal mat which occur in the mid and upper parts of the intertidal zone throughout much of the mid and inner lagoon are perhaps the most important algae from a geological standpoint.

Apart from the calcareous algae of the reefs, they are certainly the forms most likely to leave behind in the sediments direct evidence of their former existence. The algal mats comprise an interwoven network of filaments, dominantly of blue-green algae. The upper surface is a mass of fine, upstanding filaments which entrap and effectively stabilise fine sediment particles (Fig. 33A).

Calcareous algae are largely confined to the coral reefs. Here, a small branching species of Lithophyllum is by far the most abundant and is the source of at least 30% of the reef sediments. Archaeolithothamnion is common locally. An encrusting species of Lithophyllum covers much of the coarse debris of the reefs and channel floors. A delicate branching alga, Jania, occurs throughout the area but is most common in the mid and inner lagoon.

A notable absentee from the calcareous alga group is the aragonitic alga Halimeda, which plays such a significant role in the Bahamas and many other areas of carbonate sedimentation. Except for the very restricted occurrence of Avrainvillea, other algae such as Rhipocephalus, Penicillus, Udotea, and related forms, which may be the source of aragonite needles in Bahaman and other carbonate areas (Lowenstam and Epstein 1957) are also absent.

FAUNA

The fauna has been identified by the staff of the British Museum (Natural History). The lists are incomplete as some identifications have still to be made. The relative abundance and area of occurrence of the forms is indicated as follows:-

Abundance: v.c. - very common: c. - common: u - uncommon: r. - rare.

Location: S - Shelf: O - Oolite delta: C - Channel: R - Reef: M - Mid lagoon: I - Inner lagoon: T - Tidal creeks; algal mats: Z - intertidal zone: L - Land: Ro - Rock surfaces

Corals

v.c.	Acropora sp. cf. A. pharaonis	R, M.
v.c.	Porites sp. cf. P. lutea	R, M.
c.	Platygyra lamellina	R.
c.	Cyphastrea microphthalma	R.
c.	Stylophora pistillata	R.
u.	Favia favus	R.
r.	Coscinaria monile	R.
r.	Siderastrea liliacea	R.
r.	Psammocora (Stephanaria) planipora	R.
r.	Turbinaria sp.	R.
r.	Plesiastrea sp. nov.	R.

The occurrence of reef corals in this area is of considerable interest, as the high temperature and salinity conditions experienced would previously have been considered inimical to their well being. Reef or hermatypic corals are distinguished by the presence of symbiotic, unicellular, dinoflagellate algae or zooxanthellae in their endodermal tissues. 'Deep-sea' or ahermatypic corals lack zooxanthellae and can withstand a much wider range of environmental conditions than can reef corals. Reef corals are limited in their habitats largely because of the mutual interdependence of the coral polyps and their algal symbionts.

Limitations on reef coral growth include factors such as, water circulation, illumination, temperature and salinity:

(a) Water circulation: fairly vigorous water circulation is required to ensure the necessary supply of oxygen and nutrients.

(b) Illumination: reef corals generally live in depths of less than 160 ft. (50 metres) and most vigorous growth takes place in depths of less than 65 ft. (20 metres). The depth control is a reflection of the depth to which light can penetrate in sufficient amount to maintain the metabolic processes of the zooxanthellae. The upper limit of coral growth is normally close to low water mark, as death ensues by desiccation if exposure is too prolonged.

(c) Temperature: water temperature is an important control and reef corals flourish best in the range 25-29°C. (Wells, 1957). They can withstand limited exposure to temperatures as low as 16-17°C. and in the West Indies the maximum endurable temperature for survival and continued growth has been found to be about 36°C. Because of its high rate of metabolism, *Acropora* is thought to be the most sensitive and can endure a maximum temperature of only 32°C. (Wells, personal communication).

(d) Salinity: average tropical ocean water has a salinity of 35‰ and reef corals flourish best within the range 34-36‰ (Wells, 1957). However, they can tolerate dilution to 27‰ or concentration to 40‰ . In the West Indies it is found that *Acropora* can withstand salinities of 40‰ for only a few hours, although *Porites* can survive salinities of up to 48‰ . Above 48‰ all forms die or are damaged (Wells - personal communication).

Along the Trucial Coast, reef corals grow between low water mark and about 30-50 ft. (10-15 metres) depth. However, sea water temperatures and salinities have been found to exceed those recorded in other regions of reef coral growth (Chapter II). Surface and shallow sea water temperatures range seasonally from a minimum of 16°C. to a maximum of over 40°C.; deeper than 12-15 ft. (4-5 metres) the seasonal temperature range is about 20-36°C. Average surface water temperatures from May to October are in excess of 30°C. In late July, August and early September air temperatures over the adjoining islands reach at least 47°C. and during this period shallow water temperatures in excess of 35°C. are commonly encountered. Previous authors (in Wells, 1957) have found that a large diurnal range in water temperature has drastically affected the distribution of corals on reef flats and other very shoal areas. Along the Trucial Coast corals on the reef tops and in very shallow depths endure diurnal temperature ranges in excess of 10°C and a total seasonal range of over 20°C. The reefs of the Trucial Coast are composed dominantly of <u>Acropora</u>, even in very shoal areas which suffer maximum temperature changes; yet <u>Acropora</u> is a form found to be extremely temperature sensitive in the West Indies.

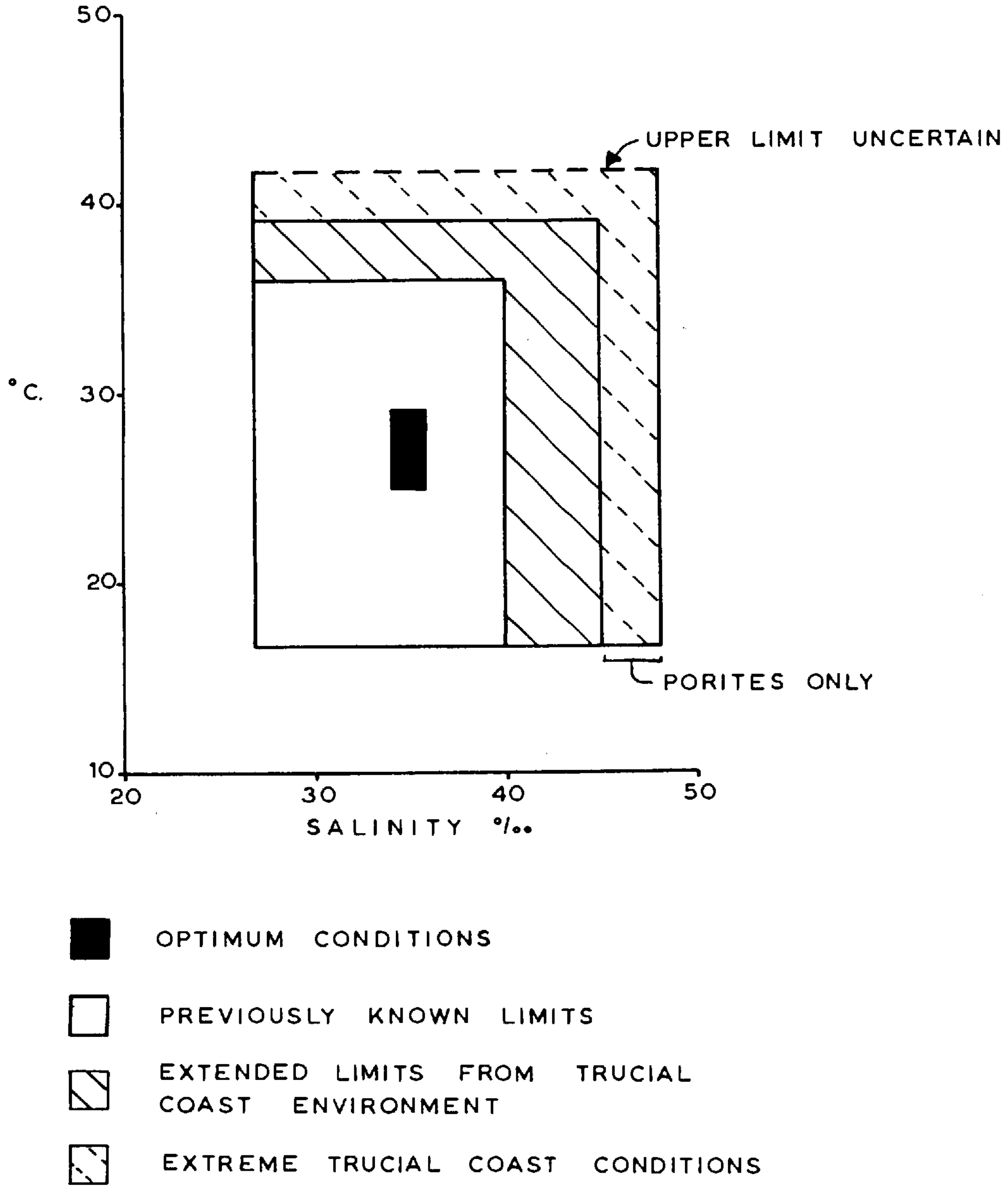
Sea water salinities along the Arabian shore of the Persian Gulf are everywhere high and near Abu Dhabi the open coast salinities are always in excess of 42‰ (Chapter II). Salinities in the reef areas range from 42-45‰ and in the seaward parts of the lagoons large <u>Porites</u> colonies, up to 10 ft. (3 metres) in diameter, occur where salinities reach 48‰. As in the West Indies, it is the massive coral <u>Porites</u> which is capable of withstanding the highest salinities, although above 48‰ even this form cannot survive. Other genera apparently cannot tolerate salinities much in excess of 45‰.

The optimum and previously determined temperature and salinity limits of reef coral growth are indicated in Fig. 30 , also shown are the extensions of the limits determined from the Trucial Coast reef coral environment.

With the exception of the new species of <u>Plesiastrea</u>, all the forms found along the Trucial Coast are

Fig. 30 Temperature and Salinity tolerances of Reef Corals; indicating the previously known limits of reef coral growth together with the extended data from the Trucial Coast environment.

TEMPERATURE AND SALINITY TOLERANCES OF REEF CORALS



known to occur in the Red Sea and Indian Ocean, and most of these species have been previously recorded in the northern Persian Gulf. The genus Psammocora is uncommon but widespread in the faunas of the Red Sea and Indo-Pacific regions but this is the first recorded occurrence from the Persian Gulf. The presence of Siderastrea is also of interest, as it is extremely rare in the living Indo-Pacific fauna, being typically an Atlantic genus.

The total number of genera of reef corals now recorded from the Persian Gulf is fifteen and eleven of these have been found in the Abu Dhabi area. The full Indo-Pacific fauna numbers over 80 genera. The attenuation in the number of genera between the Indo-Pacific and Persian Gulf regions may be explained either as a result of geographical isolation or because of the extreme temperatures and salinities experienced in the latter region or as a combination of these factors. Even within the Persian Gulf the number of genera would appear to be lower along the Trucial Coast than in the northern parts of the Gulf, where water salinities are less than 39-40‰ and summer water temperatures probably do not exceed 35°C. It is therefore probable that the further reduction of genera along the Trucial Coast is a result of the higher water temperatures and salinities of this area.

The distribution of reef corals in areas affected by low temperatures has been remarked upon by several authors (in Wells 1957). Below 18°C the number of genera and species falls and reef development as a whole is less strong than in areas of higher temperatures. In the Abu Dhabi area reef structures are fairly well developed, though probably inferior to those formed under less extreme conditions of temperature and salinity. Despite the reduction in the number of genera, reef structures formed of severl coral genera are still developed where salinities do not exceed 45‰. Above 45‰ reefs are no longer formed and only lone colonies of Porites, often of large size, are to be found.

Sugden (1963b) has considered the absence of corals from the Gulfs of Bahrain and Salwa to be attributable to the low winter water temperatures (18.5°C), but corals grow quite profusely in many regions where temperatures fall to 17-18°C and temperatures of less than 16-17°C are required to completely eliminate reef corals. The present author considers that the most critical control on reef coral distribution along the southern parts of the Persian Gulf is water salinity. Sugden's salinity data for the Gulfs of Bahrain and Salwa (1963b) show values near the northern end of Bahrain Island to be about 45‰ and to increase rapidly to the south. Along the Trucial Coast reef corals flourish until salinities reach 45‰, although single colonies of Porites occur where water salinities are as high as 48‰. The water salinity of the Gulfs of Bahrain and Salwa thus falls at the extreme limit tolerated by reef corals and reasonably accounts for their absence from this area.

In the geological record the presence of coral reefs is usually taken as good evidence that the environment was one in which sea water salinities were close to normal (35‰). However, in the Abu Dhabi area coral reefs are growing under conditions of elevated temperature and salinity. The only indication of extreme conditions is the reduction in the number of genera. Those genera which are present show "no noticeable differences or effects - they might all have come from waters of usual salinity and temperature" (Wells - personal communication). Lying close behind the reefs are the lagoonal carbonate sediments together with early diagenetic dolomite and evaporite deposits. The close juxtaposition of coral reefs, apparently indicating normal salinity waters, and of evaporites and lagoonal 'back-reef' dolomites, normally considered indicative of elevated salinities, is of fairly frequent occurrence in the geological record. But studies on the Trucial Coast have shown that the back reef dolomite and evaporite deposits are a diagenetic development and that excessively high salinities in this environment are not attained by free standing bodies of lagoon waters but by pore fluids within the sediments. Thus the concept of back-reef lagoon with waters of salinity greater than 100‰ is in many instances almost certainly incorrect. However, in order to produce high salinities in the interstitial fluids within the sediments, an area must be one of high net evaporation. High net evaporation will tend to produce higher than normal salinity waters in shallow shelf and lagoon environments (as occurs along the Trucial Coast) but if offshore depths are fairly great and water circulation unrestricted then salinities at the seaward edge of lagoon barriers may rise little higher than those of normal ocean water. Thus, in any area, if fairly deep water conditions can be proved to have existed seaward of the coral reefs then the corals probably lived in waters with salinities close to normal, even though shallow back lagoonal waters might have been fairly saline. But if evidence is found of coral reefs growing at the inner margins of a broad shallow shelf, with saline lagoons behind, then it is quite possible that the corals themselves were living under fairly saline conditions. The latter type of occurrence is exemplified by the present-day Trucial Coast environment. (Most of the above discussion on the relationships of reef coral growth with high temperatures and salinities has been published in a short paper (Kinsman 1964).

Echinoderms

Asteroids

v.c.	Astropecten phragmorus	S, R, M.
u.	Astropecten indicus	O.
v.c.	Linckia multifora	R.

v.c.	Asterina burtoni iranica	R, M.

Ophiuroids

c.	Ophiothrix savienyi	R.
u.	Ophiothrix comata	R, M.
c.	Ophiothrix sp. aff. exigua	C.
u.	Ophionereis dubia	R, C.
u.	Amphioplus hastatus	S.

Echinoids

v.c.	Echinometra mathaei	R.
u.	Echinodiscus bisperforatus	O.
u.	Echinodiscus auritus	O, M.

The asteroids and ophiuroids contribute little to the sediment. The reef echinoid, Echinometra, is often very abundant and contributes appreciable amounts of skeletal materials to the reef sediments. Its activities in the comminution of reef debris are probably quite significant. The burrowing echinoids Echinodiscus bisperforatus and Echinodiscus auritus are not common but their spines are to be found in the oolite sand and seaward pellet sand areas. A fairly diverse echinoid fauna, including E. auritus occurs offshore and spines from these forms form a conspicuous element of the shelf sands.

Molluscs

Lamellibranchs

c.	Arca (Arca) divaricata	M.

c.	Arca (Arca) donaciformis	R.
c.	Arca (Barbatica) lacerata	S, R, O.
c.	Arca sp.	O, M.
u.	Arca uropigmelania	S.
c.	Ostrea cuculata	R, S.
u.	Pinna atropurpurea	M.
u.	Malleus recula	R.
u.	Codakia (Jagonia) divergens	M.
c.	Spondylus exilis	S.
c.	Meretrix meretrix	S.
c.	Petunculus striatularis	S.
v.c.	Circe (Circe) scripta	M.
c.	Circe (Circenita) arabicum	M. R.
u.	Circe (Parmulophoca) corrugata	S.
v.c.	Asaphis (Asaphis) deflorata	S.
c.	Lithophaga sp.	R.
c.	Isognomon legumen	R.
c.	Isognomon sp.	R.
c.	Pinctada radiata	M, S.
c.	Brachydontes (Hormomya) variabilis	O.
c.	Glycimeris arabica	S.
c.	Trachycardium lacunosum	S.
c.	Pitaria (Amiantis) erycina	S.
c.	Glycimeris cf. G. Taylori	S.
c.	Chlamys ruschenbergii	S.

Gastropods

c.	Monilea (Priotrochus) obscura	M, R.
c.	Murex anguliferus	S, O, M.
u.	Murex scolopax	S.
v.c.	Cerithium petrosum	M.
v.c.	Cerithium rugosum	M. I, T.
v.c.	Cerithium scabridum	S, M, I, T.
u.	Cerithium caeruleum	Ro.
v.c.	Cerithidea (Cerithideopsilla) cingulatus	M, I.
c.	Bullaria ampulla	S, M.
u.	Euchelus asper	R.
c.	Drupa margariticola	R, Ro, M.
c.	Nassarius pullus	M.
u.	Nassarius idyllia	Ro.
c.	Nassarius stigmarius	S, M.
r.	Turitella fascialis	M.
u.	Phasianella nivosa	M.
u.	Melampus sp.	M.
u.	Mitrella misera	M.
v.c.	Mitrella blanda	S, Z.
c.	Littorina sp.	Ro, T.
c.	Ancilla cf. A. eburnea	M.
u.	Ancilla cinnamomea	M.
u.	Thais tissoti	Ro.
u.	Thais pseudohippocastaneum	Ro.
u.	Planaxis sulcatus	Ro.

c.	Turbo coronatus	M.
r.	Oncidium peronii	M, T.
c.	Siphonaria sp.	Ro.
c.	Cypraea (Erosaria) turdus	S.
c.	Trochus (Infundibulops) erythaeus	S.
c.	Strombus decorus	S.
c.	Gafrarum pectinatum	R.

Molluscs are abundant and are an important source of sediment. Along exposed coasts, the beach sediments are commonly of molluscan debris. In shelf areas lamellibranchs would seem to dominate but in the lagoons gastropods are the most abundant group. Some species occur over a wide range of environments, for example Mitrella blanda and species of Cerithium are found in shelf and nearly all lagoonal environments. Some species are associated only with coral reefs, and others are found only along rocky coasts. The latter group are very restricted as rocky coasts are found in only a few localities. The commoner shallow water and intertidal gastropods (species of Cerithium, Mitrella, Nassarius) occur along all except the most exposed open sea coasts. In the innermost lagoon and tidal creek areas, Cerithium and M. blanda are extremely common, particularly the former. In these positions they occur almost to the exclusion of other forms. The distribution of Littorina is particularly interesting; it lives along rocky coasts in the upper part of the intertidal zone and also on the surface of the high intertidal algal mats. In the latter position it commonly occurs alone, other forms living nearer mean water level.

Mid lagoon coasts which have extensive crab flats developed in the upper parts of the intertidal zone, commonly have gastropod flats in the mid intertidal zone. The gastropod flats are flooded by every tide, their surface is littered with live and dead gastropod shells and is covered with trails and faecal pellets. The faecal pellets comprise fine carbonate grains and have a characteristic shape and surface ornament. They are very different from the ovoid pellets of the pellet sands.

Sugden (1963b) has discussed the effect of salinity on the diversity of the molluscan faunas from several parts of the Persian Gulf. He concluded that although salinity must have some effects on the fauna, any direct relationship between salinity and diversity of the fauna was largely masked by other environmental factors.

Brief studies have been made by the present author of molluscan faunas from Dubai, where salinities are about 40‰ and from two localities on the Indian Ocean Coast of Arabia (Khor Fakhan and Masira Island) where salinities are about 36‰. It has been found that many species are not represented in the Abu Dhabi area which are abundant in the areas of lower salinity. Even along the Trucial Coast there is a decrease in diversity of the fauna as higher salinities are attained in the innermost lagoon areas. An ultimate Cerithium/Littorina assemblage is found in the most saline areas. This assemblage is found in algal pools where salinities exceed 100‰ . Littorina is found on the almost dried out, salt encrusted, upper surface of many algal mat areas.

Crustacea

1.	Amphipods	
u.	Amphithoe sp.	R.
2.	Decapods (Crabs)	
u.	Ogyrides sp.	S, O.
u.	Anchistus custos	C, O.
u.	Pagurus sp.	O.
c.	Pachycheles sp.	R.
c.	Petrolisthes lamarckii	R.
c.	Petrolisthes sp.	O.
u.	Gonodactylus demani demani	R.
v.c.	Ocypoda aegyptica	L.
v.c.	Scopimera sp.	Z.
c.	Macrophthalmus sp. nr. depressus	T.
v.c.	Cleistostoma nr. dotilleforme	T.
v.c.	Metopograpsus messor	T.
c.	Thalamita sp. nr. poissoni	R, M.

c.	Thalamita sp.	C, O.
v.c.	Neptunus (Neptunus) sanguinolentus	M.
u.	Actaea savigyni	J, M,O.
c.	Chlorodius niger	R, M.
r.	Tetralia glaberrima	R.
u.	Leptodius nr. hydrophilus	R.
u.	Achaeus sp.	C, O.

Many small crustaceans have been ignored in this study. The particular aim regarding the crustaceans, was to determine the burrowing species, as from a sediment contributory point of view this group is unimportant.

The vast number of burrowing crabs in many areas is particulatly worthy of comment and results in the sediments being constantly turned over, to a depth of about 2 ft. (0.6 m.). The large burrowing crab <u>Ocypoda</u> lives at or just above high water mark and makes large burrows up to 4" (10 cms.) in diameter and 2-3 ft. (up to 1 m.) deep in the berm and back beach sediments. The smaller burrowing crab <u>Scopimera</u> is the inhabitant of most sandy beaches except those of the open coast where conditions are probably too extreme. It lives between mid and high water and between each successive tidal cycle completely covers this zone with its burrowing pellets (Fig. 5). Flooding of the surface by the next high water destroys the burrowing pellets of both <u>Scopimera</u> and <u>Ocypoda</u> as they are of an extremely loose texture. The result of this intensive burrowing is the complete destruction of bedding and other sedimentary structures. Both these forms move down the intertidal zone during the period of neap tides and back up the zone during spring tides. <u>Scopimera</u> only exists where the sediments are sandy, more than about 10% of mud being sufficient to cause its absence. <u>Ocypoda</u> lives around all the coasts except those of the innermost lagoon.

In mid and inner parts of the lggoons the swimming crab <u>Neptunus</u> is abundant and disturbs the upper 1-2 ins. (4-5 cms.) of sediment in making its shallow, day-time burrow.

In the creeks and swamps crabs are again abundant, particularly <u>Cleistostoma</u>, <u>Metopograpsus</u> and <u>Macrophthalmus</u>. Each of these forms burrows to a depth of about 2 ft. ($\frac{1}{2}$ m.) in the muddy creek sediments and each burrows in a slightly different manner. The most outstanding is <u>Cleistostoma</u>, whose burrowing mounds are shown in Figure 12. A central burrow 0.5 ins. (1-2 cms.) across is surrounded by a circular mound 2-3 ins. (5-6 cms.) high and 5-10 ins. (15-25 cms.) in diameter; the mound is formed of a superimposed mass of radiating 'fingers' of muddy sediment. Other species make complex, ramifying networks of burrows; the burrows themselves often are lined with a thin layer (0.1 in, 2-3 mms.) of very fine grained carbonate mud and on occasions cementation of the wall occurs and preservation of the burrow is thus assured.

The effect of the crabs is to disrupt the sediment, to impress a bioturbated structure upon it and leave the upper surface far from plane. The characteristic feature of the creeks is the unevenness of the upper sediment surface.

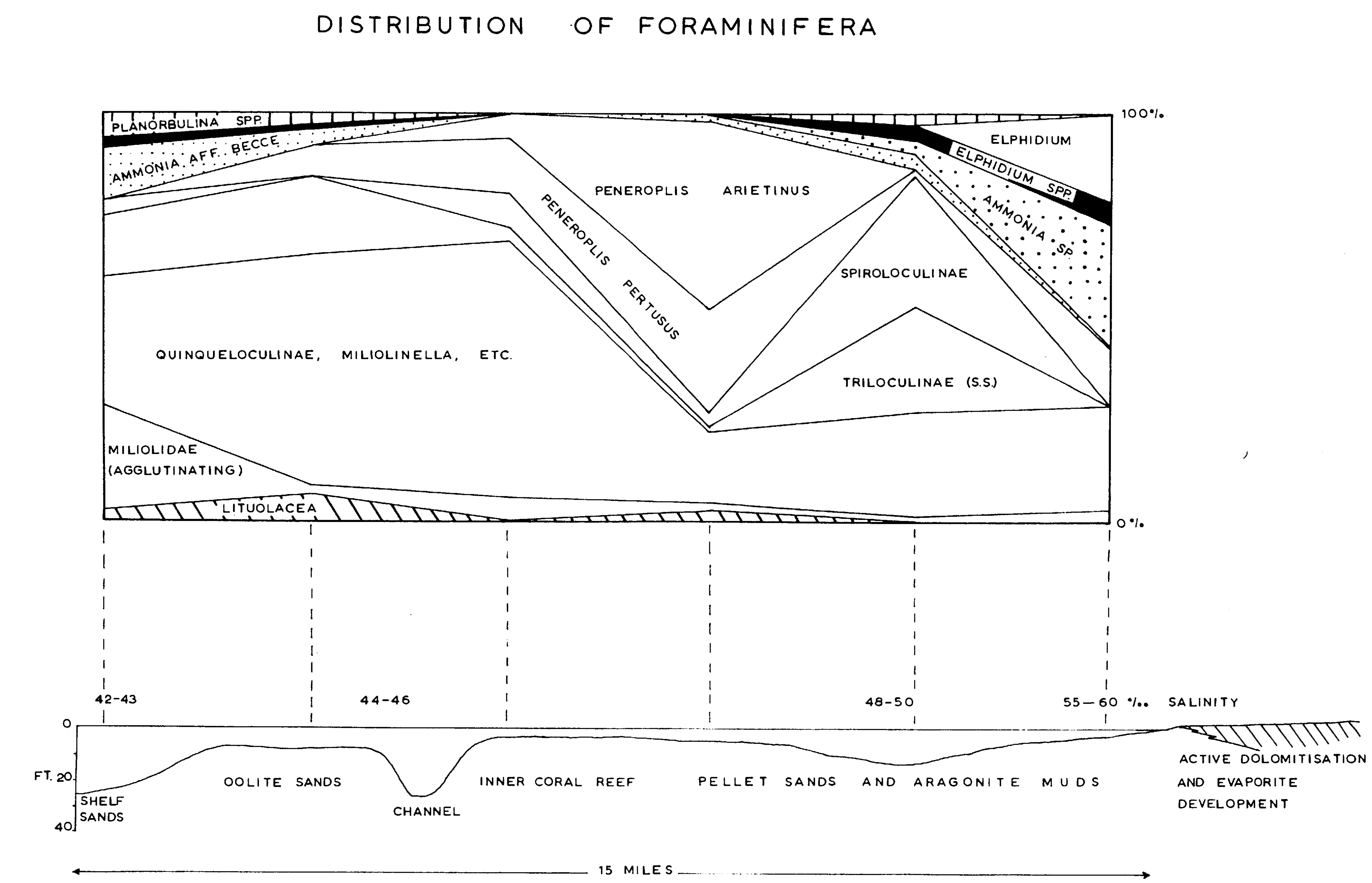
It is significant that apart from a few species of molluscs and worms the crabs are by far the most abundant organisms of the inner lagoon and creek areas. This is so, in spite of the high temperatures and salinities experienced. The physical effect of their activities upon the sediments is immense yet they contribute little skeletal material directly to the sediments. It is possible therefore that the part they have played in the development of these sediments would not be directly recognised in lithified equivalents.

Foraminifera

The foraminifera have been briefly examined from representative samples of the various sediment types by Mr. D. J. Carter. The data are presented in Figure 31. The discussion below was also contributed mainly by him.

In the shelf sands foraminifera are common but small types predominate; in the oolite sands foraminifera of all types are uncommon, but the reef and lagoon sediments show moderately high percentages. In all samples, the foraminifera do not constitute more than 10% of the total sediment.

Fig. 3.1 Distribution of Foraminifera in representative samples : expressed as cumulative percentage graphs of species in the 60-30 fraction. The diagrammatic section shows the inter-relationships between the samples, and the water salinities of the stations.



DISTRIBUTION OF FORAMINIFERA
PLANORBULINA SPP.
AMMONIA AFF. BECCE
PENEROPLIS ARIETINUS
PENEROPLIS PERTUSUS
ELPHIDIUM
ELPHIDIUM SPP.
AMMONIA SP.
SPIROLOCULINAE
QUINQUELOCULINAE, MILIOLINELLA, ETC.
TRILOCULINAE (S.S.)
MILIOLIDAE (AGGLUTINATING)
LITUOLACEA
100%
0%
42-43
44-46
48-50
55—60 ‰ SALINITY
0
FT. 20
40
SHELF SANDS
OOLITE SANDS
CHANNEL
INNER CORAL REEF
PELLET SANDS AND ARAGONITE MUDS
ACTIVE DOLOMITISATION AND EVAPORITE DEVELOPMENT
15 MILES

The inner lagoon fauna is obviously very different from that of the shelf. However, similar assemblages to that of the inner lagoon have been found by earlier workers to typify brackish water lagoonal environments. This assemblage is, therefore, not a brackish water indicator but is merely composed of those forms capable of withstanding the most extreme environmental conditions.

Some lagoonal sequences, associated closely with red beds and thin evaporites have in the past been interpreted as brackish water because of their foraminiferal assemblages. But the close association of 'brackish water' foraminifera and the red beds and evaporites has not proved easy to understand. Using the Trucial Coast data, however, this assemblage may be interpreted as indicating increased salinities, and the occurrence of evaporites is consistent with this.

Other Organisms

Many groups of organisms have not been considered, mainly because they do not contribute obviously to the evolution of the sedimentary facies. Sponges have been found throughout the area but are most common attached to coarse reef and channel debris; they also occur in the shallow mid and inner parts of the lagoons. None of the species collected has been specifically identified but all belong to the

Haploscleridae. Soft bodied forms such as Coelenterates are very abundant. Beche-de-mer (Holothurians) are common locally. The organism responsible for the pelleting of the lagoonal muds is not known. Cloud (1961) considered an annelid worm to be the principal pellet maker in the Bahamas.

These and many other organisms were noted, but apart from the unknown pellet-maker the combined effects of the faunal elements not considered is very slight.

SEDIMENTS

The surface sediment samples have been divided into facies following field and laboratory examination. The parameters of size, sorting and rounding have been crudely determined using comparison tubes or direct measurement of grains. The ratios of components in the samples have been estimated visually.

The following surface sediment facies have been recognised and their areal distribution is shown in Figure 32.

Shelf facies

Reef facies

Oolite facies

Channel facies

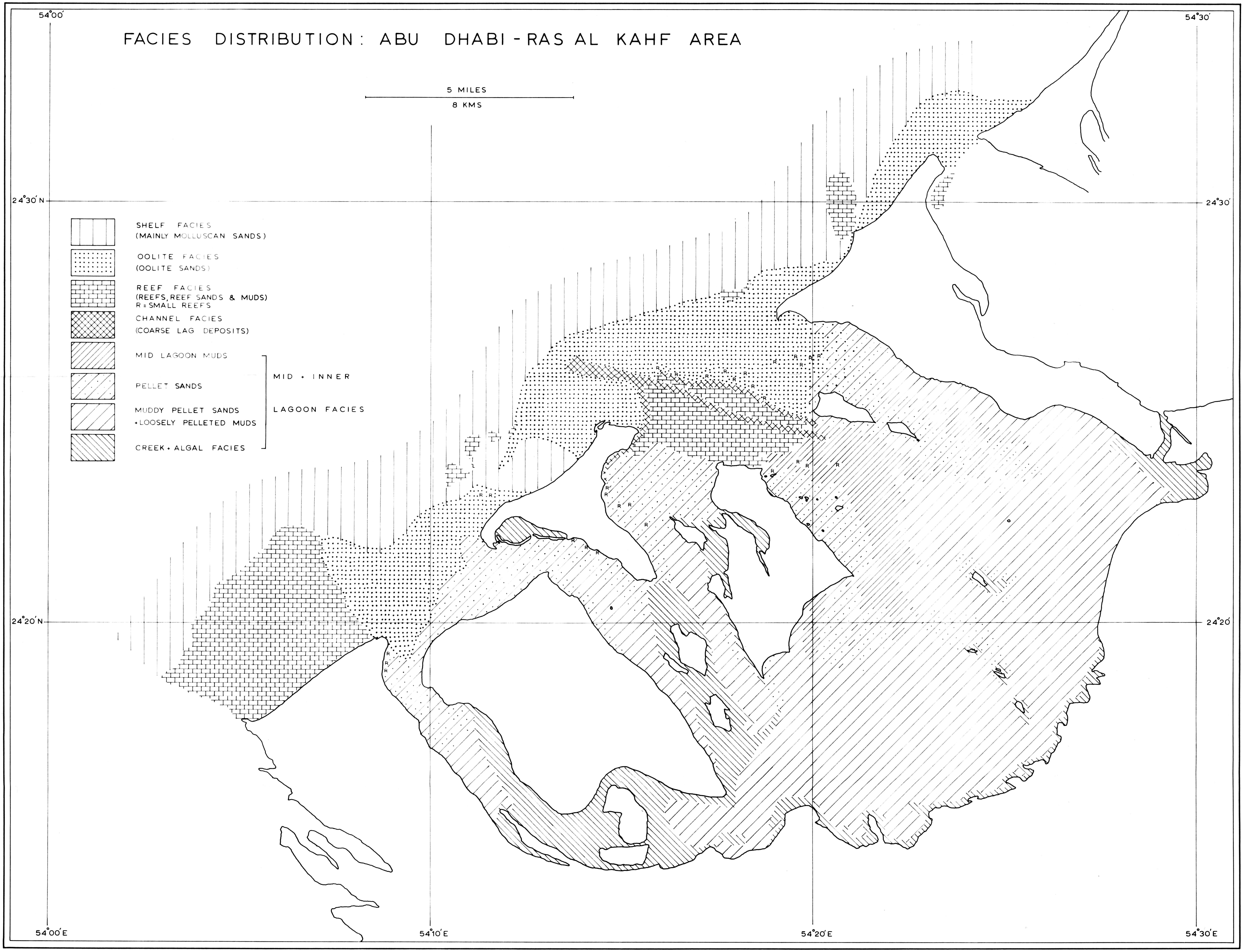
Mid and Inner Lagoon facies

Creek and Algal facies

Shelf facies

The sediments of the shelf facies, although no doubt subjected to wave action during storm conditions, are not as continually affected by wave activity as are the oolite sands of the tidal deltas or the reef top sands of the open coast reefs. Offshore the shelf facies is very variable and sediments range from coarse skeletal sands and gravels to fine muds. The inner shelf sediments are less variable and range from coarse sands to very fine sands with small additions of mud. Local elevations and depressions of the

Fig. 32 Distribution of Sedimentary Facies in Abu Dhabi - Ras al Kahf area.



FACIES DISTRIBUTION: ABU DHABI - RAS AL KAHF AREA
5 MILES
8 KMS
54°00'
54°30'
24°30'N
24°30'
24°20'N
24°20'
54°00'E
54°10'E
54°20'E
54°30'E
SHELF FACIES
(MAINLY MOLLUSCAN SANDS)
OOLITE FACIES
(OOLITE SANDS)
REEF FACIES
(REEFS, REEF SANDS & MUDS)
R = SMALL REEFS
CHANNEL FACIES
(COARSE LAG DEPOSITS)
MID LAGOON MUDS
PELLET SANDS
MUDDY PELLET SANDS
+ LOOSELY PELLETED MUDS
CREEK + ALGAL FACIES
MID + INNER
LAGOON FACIES

sea floor control the distribution of sediments, features of only 2-3 feet (less than 1 metre) in amplitude being critical in this respect.

The sediments are generally medium to dark grey in colour and are usually somewhat foetid, interstitial waters showing a pH of about 7.7 and Eh of about -100 millivolts. The medium and coarse grades are almost entirely molluscan, with some echinoderm, coral and calcareous algal grains. Most coarse skeletal grains are bored and have a corroded appearance. Other shelly materials include foraminifera, ostracods, polyzoa and echinoid spines. Many of the skeletal grains are thin walled and unbroken; many lamellibranchs valves are still articulated, or if separated show little sign of abrasion. Broken shells nearly always have angular edges and evidence of abrasion is everywhere slight. In the fine and very fine grades are appreciable quantities of non-carbonate grains, ranging from 18-34% (average 28%) of the total sample. Coarser sands on local highs of the sea floor contain less than 6% non-carbonate grains; in these sediments the fines have been winnowed away, including most of the non-carbonate grains, which have a medium diameter close to 0.13 ms. Many of the very fine sand and silt sized non-carbonate grains are angular or sub-angular, although occasional fine or medium sand sized, well rounded, typical aeolianquartz grains do occur.

Oolites are rare except immediately below the delta edge. A few ovoid pellets of unknown origin occur in some samples. A few composite grains occur, some of these are limestone grains derived from underlying limestones on the shelf, others are of more recent origin, consisting of grains of the finer grades of the sediment cemented together with granular aragonite; yet others are the cemented infillings of skeletal grains, released after breakdown of the shell. Inside many large skeletal fragments, even single lamellibranch valves, similar cementation of fine sediment grains with granular aragonite is found. Many of the finer grains also bear an irregular and incomplete coating of granular aragonite. Aragonite needles occur inside some skeletal grains, particularly gastropods and foraminifera. The restricted micro-environment inside the skeletal grains would seem to be one in which $CaCO_3$ can be fairly readily precipitated.

The sediments of the interdelta areas are shelf sands with greater or lesser admixture of reef or oolite grains.

Reef Facies

The structure and major forms of the reefs have been described in Chapter I. The reef sediments are those derived almost entirely from the reef fauna, but in places, mixtures with other facies are found.

The more active reefs are dominantly composed of in-place corals with interstitial spaces infilled with calcareous algae. Debris derived from the organic elements of the reef supplies sand and mud which complete the infillig. The less vigorous reefs have more broken and overturned coral clumps but are otherwise similar. Where a reef top bears a fair amount of living coral than the sediments are very ill-sorted. But when much of the reef top is almost completely free from coral heads, then fairly well sorted and abraided sands occur - the latter condition is found across much of the reef top off the Dhabayah peninsula.

The reef sediments comprise coral and calcareous algal grains with minor amounts of echinoid, molluscan and foraminiferal grains. The sediments are generally angular and ill-sorted, fragments ranging in size from entire coral heads to silt and clay sized particles. Interstitial pH and Eh values in the sandy sediments average pH 7.9, Eh +50 Millivolts; in more muddy reef sediments the values average, pH 7.5, Eh -200 to -300 millivolts. The finest grains accumulate in hollows in the reef top or in quiet back-reef hollows and lagoons. The backreef lagoon off the Dhabayah peninsula is floored with reef derived silty and sandy muds, which in the nearshore intertidal zone are being pelleted. The pellets morphologically resemble those of the mid and inner lagoon facies but are derived from reef muds, which differ markedly in mineralogy from the muds of the mid and inner lagoon.

The open coast reefs show a gradational relationship with the shelf facies. Mixed reef and oolite sands occur around the shallow channels and banks near Batin and on the delta top to the north of the main tidal channel between Abu Dhabi island and Halat al Bahrani. Local admixture of reef sediments is found in some parts of the mid lagoon where small areas of coral occur. The interdelta and channel deposits both contain appreciable amounts of reef derived grains.

Oolite Facies

The oolite facies is confined to the delta tops except for some drifted or windblown accumulations. The sediments are largely cream to buff coloured medium and fine grained oolite sands, with oolite generally greater than 60%. At the delta edge medium sized oolites predominate and all grains are well polished. Except on the island beaches, the oolites and skeletal grains farther back on the deltas are not polished and are commonly bored, possibly by boring algae. The shoal banks off the north and south west coasts of Halat al Bahrani are of fine oolite sands. The interstitial pH and Eh values averaged 7.9 and +40 millivolts respectively.

Skeletal grains are subordinate and are mainly of rounded, thick shelled material. Little thin shelled debris occurs except for the occasional very abraded foraminifera and small lamellibranch valve. Those skeletal grains which do occur are all heavily abraded. Limestone grains are not common except where the cover of oolite sand over the underlying limestone is very thin. A few composite grains occur, usually comprising two or three small oolite or skeletal grains joined by a granular aragonite cement.

The non-carbonate content of the oolite sands ranges from 2-15% (average 7%) but is highest in the finest oolite sands. Non-carbonate grains form cores to some oolites, although skeletal grains and pellets form the major core materials. Some of the small non-carbonate grains show partial jackets of aragonite, representing the first stages of oolitisation.

The oolite facies is mixed in places with reef sediments and passes back into the mid lagoon facies with the progressive incoming of pellet sands. Oolite grains are drifted back into these areas or are blown there from the oolite dune ridges. The oolite grains are drifted from the deltas onto the beaches and are then blown inland. In places, large dune ridges of oolite sands are found, for example on Abu Dhabi island and Halat al Bahrani. On Halat al Bahrani the oolites are blown further inland and form oolite-rich sediments off the south-eastern corner of the island. A similar dispersion of oolite grains takes place all along the coast.

Channel facies

The channel sediments may comprise components from several facies, and are often of an extremely heterogeneous character. The form of the channel and the current flow within it largely define the type of sediments to be found. In the main channels between Halat al Bahrani and Abu Dhabi island the surface is littered with a coarse lag of coral and shell debris, thickly encrusted with calcareous algae, bryozoa and sponges and badly bored and abraded. Finer skeletal and oolite sands occupy interstitial positions. The finer materials are carried seawards by the ebb currents and deposited just in from the channel mouth. At the channel mouth itself oolite sands predominate.

The main tidal channels to the north east of Abu Dhabi island and south-west of Halat al Bahrani are less well defined features and do not show such diversity in their bottom sediments. Their sediments are skeletal and oolite sands with only occasional larger fragments. In particular they lack the high percentage of reef derived debris which occurs in the Halat al Bahrani/Abu Dhabi island channels.

Mid and Inner Lagoon Facies

The mid and inner lagoon sediments are mainly aragonite muds and pellet sands, with subordinate amounts of skeletal materials. The latter are largely gastropods and foraminifera. Sediments below low water mark are always muddy and in sheltered areas the intertidal sediments are also muddy. Shoal banks and coasts exposed to the north-west winds have pellet and skeletal sands with little or no mud. In Figure 11 the shoals are of clean pellet sands whereas the weed covered deeper areas have more muddy sediments. The deep area to the north-east of Jazirat al Ftaisi is floored with silty and sandy muds, mixed in two local areas with reef sediments. Most of the muds are loosely pelleted but as the water depth decreases the percentage of hard pellets increases until on the very shoal banks the pellets are quite hard and aragonite mud may be absent. There is an almost complete gradation between loose aragonite mud and hard pellets, the intermediate stages of loosely agglutinated mud and soft pellets being common below low water mark. The intermediate stages also occur in the intertidal zone in sheltered areas. Agglutinated mud may represent partially degraded pellets or primary agglutination of aragonite mud. Much of the aragonite mud is probably pelleted again and again. Pellets above low water mark in exposed areas have most chance of survival as deposition of aragonite takes place, mainly in peripheral positions and firmly cements the pelleted mud together.

As in the other facies the interstitial pH and Eh was found to bear an almost direct relationship to grain size. In sandy sediments values of pH and Eh averaged 7.9 and +50 millivolts; in muddy sediments values averaged 7.5 and -100 millivolts respectively.

The muds consist of aragonite needles less than

1-2μ long, together with more shapeless, granular material, the whole usually loosely agglutinated with an organic film.

Some beaches and islands have local accumulations of coarse shell sands and gravels, often almost entirely of gastropods. The exposed inner lagoon coast, lying opposite the inner end of the main tidal channel between Jazirat al Ftaisi and Abu Dhabi island, and exposed to the shamal winds, has coarser shelly and pellet sands forming the beach and beach ridge sediments.

Areas in the lee of the larger islands receive windblown additions. The oolite-rich mid lagoon sediments to the south-east of Halat al Bahrani have already been mentioned. An unusual addition is that of windblown gypsum-rich sands. For example, along the southern coast of Halat al Bahrani and particularly along the south-eastern coast of Jazirat al Ftaisi, sands with gypsum derived from the island sabkhas is blown onto the beaches and into the marine areas. The dune sands may comprise up to 70% of wind abraded gypsum crystals but although searched for, no gypsum grains were found seaward of the mid parts of the intertidal zone. This is to be expected, as the sea waters of these areas are not sufficienlty concentrated to be saturated with respect to gypsum, which will thus go into solution.

The non-carbonate content of the lagoon sediments varies widely and is directly dependent on the nearness of Miliolite Limestone outcrops. Sediments overlying Miliolite

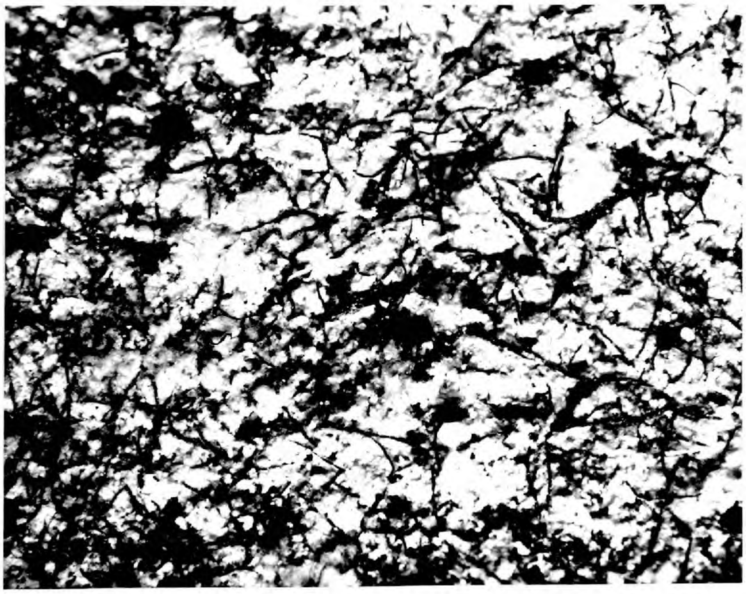
Limestone platforms derive non-carbonate grains directly from the underlying source, but in areas downwind of Miliolite islands the non-carbonate grains are mainly wind derived.

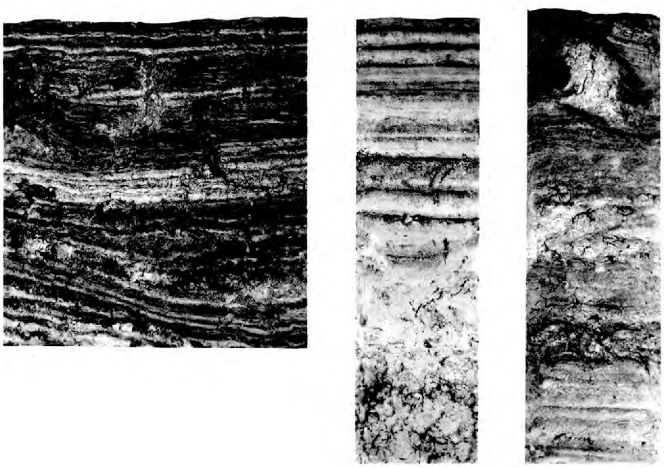
Creek and Algal Facies

The creek and algal sediments occupy the sheltered intertidal zone in much of the mid and inner lagoon. The sediments are generally sandy muds but range from almost pure aragonite needle muds to skeletal sands. Skeletal grain are common on the floors of the creek channels or as beach ridge or strandline accumulations. The latter are composed almost entirely of gastropod shells, most of which are entire and little abraded.

The surface structures of the algal mats can be seen in figures 14A-C. The mats are often broken into polygonal pieces (Fig. 14B), some of which become overturned by tidal waters. Shallow pools and runnels cross the surface (Fig. 14C) and the floors of these are often littered with gastropods. Gastropods also accumulate in the crevices between algal polygons. Minor structures vary greatly, some surfaces being almost plane whilst others support a vermicular growth pattern (Fig. 14A). The algal mats entrap sediment particles onto their upper surface, as can be seen in Figure 33A. The sediments are generally banded, with algal layers alternating with muddy and sandy sediment layers; whole gastropods are common between the algal layers. The algal

Fig. 33 A (Upper)	Upper surface of algal mat showing upstanding 'hairs' and entrapped aragonite mud particles. X 50.
Fig. 33 B (Lower)	Structures in algal banded (stromatolitic) sediments. X $\frac{1}{3}$.





banding may be parallel, sub-parallel or irregular (Figure 33B).

The aragonite muds of the creeks and algal flats are similar to those of the mid and inner lagoon and are mainly composed of aragonite needles 1-2μ in length. The muds are loosely pelleted in many places, but concentrations of hard pellets, such as occur on shoal banks in the mid lagoon, are not found.

Beneath many of the algal mats gypsum is precipitated from interstitial and surface sea-waters, concentrated between tides. The gypsum crystals are often clear but many contain included carbonate particles. The upper 1-2 cms. of the algal sediments is sometimes cemented with aragonite and gypsum, although the sediments below are not affected. As the creeks meander and change their course the earlier sediments may be rederived and many creek sides and floors are littered with broken fragments of loosely cemented limestone.

CHAPTER IV

WATER AND SEDIMENT, CHEMISTRY AND ORIGIN OF THE SEDIMENTS

GENERAL

A knowledge of the mineralogy and chemistry of carbonate sediments may enable the origin of the sediments to be determined and may also indicate the probable course of early diagenetic changes. For example, the sediments originating from a coral reef, even if too fine in grain size to enable the constituent particles to be identified, can still be shown to have a reef origin because of their mineralogy. Work on the solubility of carbonate minerals by Chave and others (1962) indicates that the sequence of post-depositional changes to be expected in carbonate sediments will depend to a large extent on the initial mineralogy.

Carbonate sediments are derived from skeletal sources, together with material contributed by pre-existing limestones and by precipitation from solution. As the skeletal source usually predominantes, the mineralogy, and factors affecting the mineralogy, of the initial skeletons, are rather critical.

The minerals of carbonate sediments are aragonite and calcite. These two paramorphs of $CaCO_3$ have differing crystal structures and minor element chemistry, aragonite having an orthorhombic lattice and being prone to substitution of calcium (Ca) by large ions (greater than 1.0Å), whereas calcite has a rhombohedral lattice and is prone to substitution of Ca by smaller ions (less than 1.0Å). The commonest substitution in marine aragonites is strontium(Sr) which may reach 1-2% $SrCO_3$. Substitution of Sr in calcite is smaller, usually by an order of magnitude. Other substitutions within the calcite lattice may be much greater and the one of greatest concern when considering carbonate minerals derived from sea-water is the substitution of magnesium (Mg) for Ca. This can be quite extensive and may reach 30% $MgCO_3$ in certain calcareous algae. In aragonite the substitution of Mg for Ca is low, being generally less than 1% $MgCO_3$.

Substitution of this kind, where ions of different size replace each other (Ca - 0.99Å : Sr - 1.12Å : Mg - 0.66Å) result in changes in the lattice spacings which, if large enough, may be used to determine the amount of substitution which has taken place. The amount of Sr in aragonite is too small to determine by X-ray diffraction methods, but the amount of Mg in calcites can be determined fairly closely (Chave, 1952 and others).

Skeletal calcites range in composition from almost pure $CaCO_3$ to $(Ca_{70}Mg_{30})CO_3$. This range of compositions has been divided into two parts, low magnesian and high magnesian. It has been found that most Mg-bearing skeletal calcites contain either less than 4-5% $MgCO_3$, or more than 10-11% $MgCO_3$, up to a maximum of about 30% $MgCO_3$. There are very few organisms which produce skeletal calcites in the range 5-10% $MgCO_3$.

Synthetic studies of phase equilibria in the system $CaCO_3$ and $MgCO_3$ by Jamieson (1953), Harker and Tuttle (1955), Graf and Goldsmith (1955), Chave et al (1962) and many others, have shown that only calcite containing less than about 4% $MgCO_3$ is stable under earth surface conditions of temperature and pressure. Aragonite and calcite containing more than 4% $MgCO_3$ in solid solution are metastable phases. Chave (1962) has used the value 8% $MgCO_3$ as the boundary between low and high magnesian calcites, for purely technical reasons, but as there is a general lack of skeletal calcites in the 5-10% $MgCO_3$ range, this is not critical. However, it must be borne in mind that the stable calcites are those with less than 4% $MgCO_3$ and that mineralogical analyses of carbonate sediments should dominantly aim to determine the ratio of stable/unstable minerals.

Aragonite is the major component of shallow water Recent marine carbonate sediments in most parts of the world, and may be of skeletal or non-skeletal origin. Calcite in marine carbonate sediments is almost entirely of skeletal origin, except where pre-existing calcite limestones are a contributory source of sediment. Aragonite is the skeletal carbonate mineral of many groups of organisms (Fig. 34) and along the Trucial Coast the most important are the madreporian corals and molluscs. Low and high magnesian calcites also form the skeletons of many organisms. The amount of low magnesian calcite in shallow water sediments is generally low and along the Trucial Coast organisms with skeletons of low magnesian calcite are not abundant sediment producers. High magnesian calcites form the skeletal structures of several important sediment producing groups, notably the benthoric foraminifera, echinoderms and benthonic algae.

The distribution of aragonite and calcite and of Sr and Mg in the calcareous parts of marine organisms has been extensively studied by Chave (1954a, 1954b, 1962), Lowenstam (1954) and others. Lowenstam found that the principal controlling factors in relation to skeletal mineralogy are generic association of the organism and water temperature. Some species or groups of organisms were found to secrete increasing percentages of aragonite with increasing water temperature. If this may be used as a generalisation , then along the Trucial Coast, where high temperatures prevail, any organism group can be expected to secrete skeletal carbonate with the highest percentage of aragonite possible for that group. In addition, the quantitatively important

Fig 34. Mineralogy of Carbonate Skeletal Materials (after Chave, 1962).

Organism	Aragonite	Aragonite + Low Mg Calcite	Low Mg Calcite	High Mg Calcite	Aragonit High Mg Calcite
Foraminifera:					
Benthonic	R			$\underline{\text{C}}$	
Pelagic			$\underline{\text{C}}$		
Corals:					
Madreporian	$\underline{\text{C}}$				
Bryozoa	C			R	C
Echinoderms				$\underline{\text{C}}$	
Molluscs:					
Gastropods	$\underline{\text{C}}$	$\underline{\text{C}}$			
Lamellibranchs	$\underline{\text{C}}$	$\underline{\text{C}}$	$\underline{\text{C}}$		
Cephalopods	$\underline{\text{C}}$			R	
Arthropods:					
Decapods				C	
Ostracods			$\underline{\text{C}}$	R	
Algae:					
Benthonic	C			$\underline{\text{C}}$	
Pelagic			C		

(C - Common : R - Rare)

Underlined forms are fairly common to abundant along the Trucial Coast).

groups of the corals and molluscs secrete skeletal structures largely or entirely of aragonite. Chemically precipitated carbonate will also be in the form of aragonite. There may thus be expected to be a great preponderance of aragonite over calcite in the Trucial Coast environment of carbonate sedimentation.

The distribution of Mg and Sr in skeletal carbonates has been found to be mainly controlled by the carbonate mineralogy and 'vital' effect of the organism. Thus the major controls on Mg and Sr distribution in skeletal carbonates are those affecting the mineralogy, that is, vital effect and water temperature; but impressed upon the initial mineralogy are found varying amounts of Mg and Sr and the principal controlling factors are again vital affect and water temperature. For example, two similar species of Echinoderm, composed entirely of calcite, may show quite different percentages of Mg, but in general any single species is directly affected by water temperature. Other factors such as salinity, Sr/Ca or Mg/Ca ratios have been invoked by various authors to greater or lesser extent. Forexample, Kulp and others (1952), following Odum (1950), considered the Sr/Ca ratio in a shell to be directly related to the Sr/Ca ratio in solution; the latter ratio was said to be dependent on the salinity and source of the waters. But the Sr/Ca (and Mg/Ca) ratio in sea-waters is fairly constant and does not change appreciably with changing salinity except under the most extreme conditions of Ca withdrawal. Changes in these ratios will therefore be insignificant. The effect of salinity as an independent variable on Sr/Ca ratios in skeletal materials has been considered in an indirect manner by Turekian (1955).

To generalise, it has been found that within a single species or organism group the Mg content increases with rising temperatures. Forms affected by changes in mineralogy due to temperature may run counter to this trend - increasing temperature bringing about an increase in aragonite percentage which will tend to lower Mg values. Conversely, the Sr content will tend to rise as the percentage of aragonite increases. Temperature itself has been found to play a minor role in the Sr distribution in skeletal carbonates, its effect being masked by organic controls.

A new approach to the study of chemically (inorganically) precipitated carbonates has been made possible by recent work of Holland and others (1961, 1962, 1963a, 1963b). They have studied the factors controlling the coprecipitation of metallic ions with precipitated carbonate minerals. The organic control in this case is absent and the mineral phases precipitated should be in equilibrium with their parent solutions. For example, the Sr/Ca ratio of chemically precipitated aragonite has been found to be dependent upon the Sr/Ca ratio of the solution, and the temperature at which precipitation takes place, Preliminary work with NaCl solutions has shown that the relationship is almost independent of this variable, at least within the range of concentrations found in most marine environments. As the Sr/Ca ratio in sea-water is itself almost constant, the Sr/Ca ratio in precipitated marine aragonites will be directly related to temperature.

Laboratory coprecipitation studies are in an early stage and many more data are required, but values to date indicate that at $25^{\circ}C$ an aragonite precipitated from sea-water should contain about 9000 p.p.m. Sr (1.5% $SrCO_3$). At $100^{\circ}C$ the value is about 6000 p.p.m. (1.0% $SrCO_3$). At higher temperatures the amount of coprecipitated Sr is lower but the exact relationship is not yet certain. A probable value for $30-35^{\circ}C$ (the summer temperature of the Trucial Coast waters) is about 7500 p.p.m. Sr (1.2-1.3% $SrCO_3$).

MINERALOGY OF TRUCIAL COAST SEDIMENTS & ORGANISMS

Staining Methods

The mineralogy of calcareous sediments or skeletal materials may be determined by either X-ray diffraction techniques or by staining methods. Loose mounts and thin sections of sediments have been stained using cobalt nitrate solution or Feigl's reagent (Feigl 1958). The opaque precipitate from Feigl's reagent makes its use rather limited, whereas the lilac colouration from cobalt nitrate does not mask the character of the grains. Thin sections were stained by standing them in cold cobalt nitrate solution for 24-48 hours.

Staining tests of this type make use of the relative solubilities of the minerals present. Both cobalt nitrate and Feigl's reagent are supposed to be of use in the differentiation of aragonite from calcite, the aragonite reacting and staining more rapidly than calcite because of its greater solubility. However, high magnesian calcites are more soluble than aragonite (Chave et al, 1962) and thus the stained fraction will comprise aragonite and high magnesian calcite. Nevertheless, this fraction does at least comprise the metastable components. This relationship, obvious when one considers the solubility data, was found quite accidentially when X-ray analysis of a sample of Lithophyllum showed high magnesian calcite (16% $MgCO_3$); yet this calcareous alga had been stained previously and considered to be aragonite. High magnesian calcites from some echinoderms (15% $MgCO_3$) were also found to stain with cobalt nitrate.

X-ray Analysis

A few skeletal materials were examined to determine the amount of $MgCO_3$ in solid solution. Echinodiscus and Echinometra showed 15% $MgCO_3$ and Lithophyllum showed 16% $MgCO_3$.

A quantitative method was devised for the direct determination of three component mixtures of aragonite, low magnesian calcite and high magnesian calcite, within the sediments. Evaluation of the method has been limited but where aragonite forms more than 70% of the sample the ratios can be determined to within $\pm 5\%$; where aragonite is about 50% of the sample then the accuracy is about $\pm 10\%$.

The equipment used was a Philips Diffractometer, in the Department of Geology, University of Durham. Valuable assistance and advice from Mrs. M. Kaye and Dr. H. D. Holland enabled the work to be completed.

For standards, coarsely crystalline aragonite from Sicily, Iceland spar from Reydarfjordur, Iceland and high magnesian calcite from the echinoid Echinodiscus bisperforatus (15% $MgCO_3$) were used. The standard minerals were crushed by hand to less than 200 mesh, taking care not to crush the material more than was absolutely necessary. Mixtures of the minerals were then prepared of known proportions. Each mixture was run three times, being remounted between each run as a check on preferred orientation of grains. Calibration curves were then erected so that the ratios of the minerals in the sediment samples could be determined. As a check, a mixture of 50% aragonite, 25% low magnesian calcite and 25% high magnesian calcite was run under sample conditions and gave values of 47.5% aragonite, 28.6% low magnesian calcite, 23.9% high magnesian calcite. These values were considered sufficiently accurate to warrant use of the technique on the sediment samples.

Representative sediment samples were crushed to less than 200 mesh and run under the same conditions; the aragonite/low Mg calcite/high Mg calcite percentages are shown in Figures 35/A and 35/B.

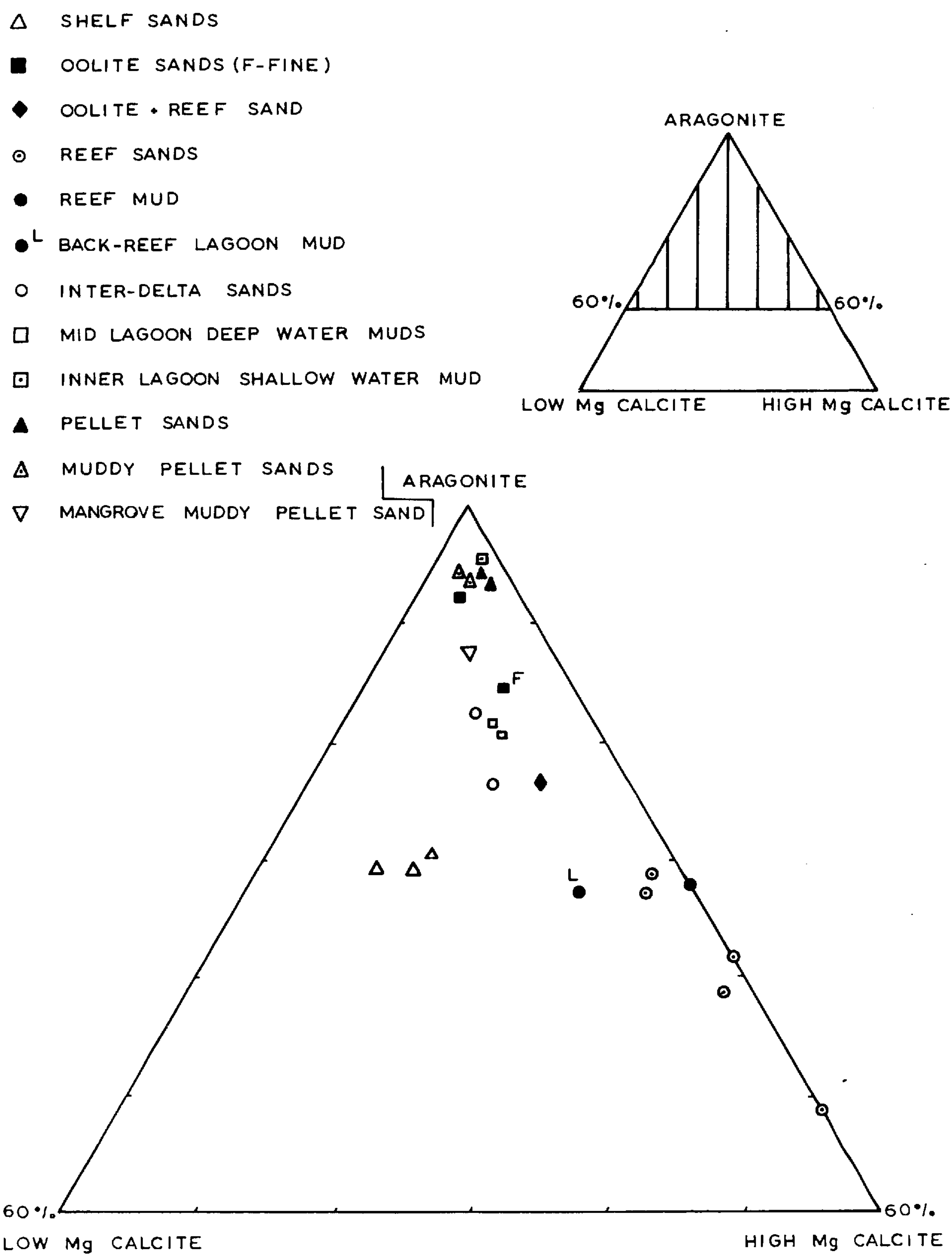
The sediment types differ markedly in general appearance and Fig. 35 shows them to also differ greatly in their mineralogy. The shelf sands show high values of low Mg calcite (about 20%) and this is derived in part from skeletal sources but mainly from submarine erosion of Miliolite Limestone outcrops on the shelf. The oolite sands show fairly high amounts of aragonite, to be expected from their aragonite coatings. The calcitic components of the oolite sands are derived partly from the shelf sands and partly from reef sources. Most of the calcitic grains form cores to oolites. The fine oolite sand has a higher core to jacket volume ratio than the more normal, medium grained, oolite sand; this is reflected in the former having a much higher amount of total calcite.

Fig. 35A Aragonite/Low Mg Calcite/High Mg Calcite Percentage in representative sediment samples.

Sediment Type	% Aragonite		% Low Mg. Calcite		% High Mg Cal	
Shelf Sands	69.5 70.7 69.4	69.9	19.0 17.7 22.2	19.6	11.5 11.6 8.4	10.5
Oolite Sands (Medium) (Fine)	92.2 85.1	88.7	4.3 4.8	4.6	3.5	6.8
Oolite + Reef sands	77.0		6.6		16.4	
Reef Sands	67.5 58.4 48.8 68.4 61.9	61.0	3.6 2.3 0 2.8 0	1.7	28.9 39.3 51.2 28.8 38.1	37.3
Reef Mud	67.6		0		32.4	
Back-Reef Lagoon Mud	67.4		8.2		24.4	
Interdelta Sands	77.3 82.6	80.0	10.0 7.9	9.0	12.7 9.5	11.1
Deep Water Mid Lagoon Muds	81.2 81.7	81.5	7.1 7.1	7.1	11.2 11.7	11.5
Shallow Water Lagoon Mud	95.6		1.5		2.9	
Pellet Sands	94.2 94.6	94.4	1.5 2.4	2.0	3.0 4.3	3.7
Muddy Pellet Sands	93.9 94.6	94.3	3.0 3.2	3.1	2.2 3.1	2.7
Mangrove Muddy Pellet Sand	87.4		6.2		6.4	

Fig. 35/B Aragonite/Low Mg calcite/High Mg Calcite in representative sediment samples.

SEDIMENT MINERALOGY



The reef sands show relatively low amounts of aragonite and low Mg calcite, but a high percentage of high Mg calcite. This is a natural consequence, following on their derivation from aragonitic corals and high Mg calcite forms such as calcareous algae and echinoids. The reef mud sample came from a hollow in the Dhabayah reef top and its mineralogy is the same as that of the reef sands. The back-reef lagoon mud shows a higher percentage of low Mg calcite, mainly because of the addition of foraminifera and ostracods; it is otherwise similar to the reef mud. These two sandy and silty muds look similar in many ways to some of the lagoonal muds, yet the mineralogies of the two groups are very different. Microscopically the reef and back-reef muds are found to have less aragonite needles than the lagoon muds but the physical differences between the two types are not great. It is, however, obvious that both the reef and back-reef lagoon muds are derived largely from abrasion of reef detritus.

The interdelta sands, composed of grains from shelf, oolite and reef facies, show a mineralogy expected from such a mixture.

The mid and inner lagoon sediments are generally high in aragonite, showing more than 80% aragonite in all instances, values being normally greater than 90%. The two deep water, mid lagoon, sandy muds have fairly high numbers of foraminifera and ostracods, which reasonably account for the moderately high amounts of low Mg calcite and high Mg calcite. The shallow water mud is almost pure aragonite with only traces of calcite. The clean pellet sands and muddy pellet sands have similar mineralogies. In Chapter III the pellet sand-aragonite mud sequence was briefly discussed and the similarity in mineralogy between the various phases can be regarded as good corroborative evidence of the development of all the phases from a parent aragonite mud. The muddy pellet sand from the mangrove area has a moderately high amount of molluscan and foraminiferal grains which account for the calcitic components present.

ORIGIN OF THE NON-CARBONATE GRAINS

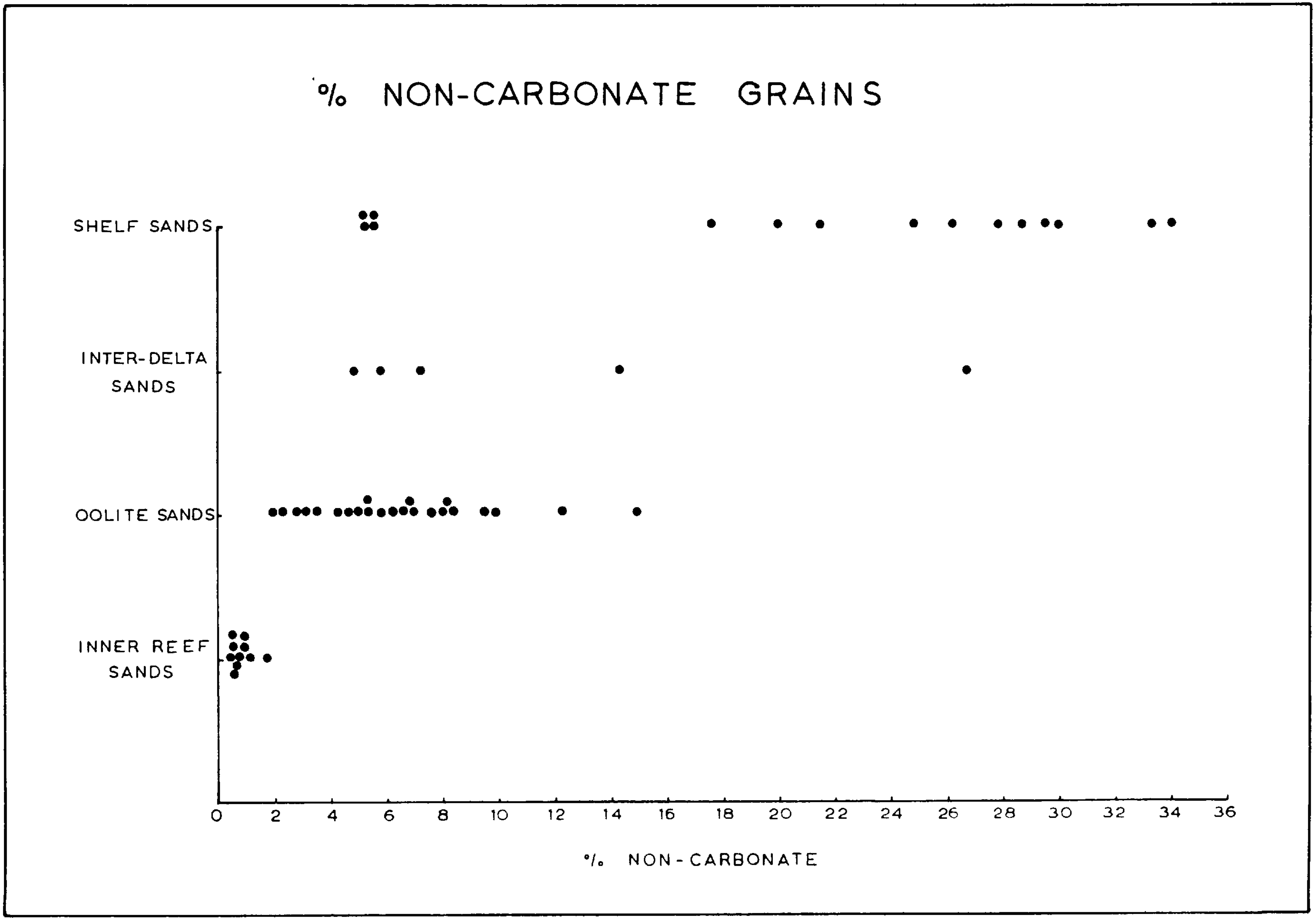
The percentage of non-carbonate grains was determined from about 50 of the seaward sediments, and also from a few limestones. The insoluble residues were examined both as grains and in thin section. The data for the sediment groups are shown in Figure 36 and have been already discussed for each individual facies (Chapter III).

The non-carbonate grains are fine to very fine sands, mainly of quartz and feldspar, with minor amounts of coloured minerals. The majority of the grains are angular or sub-angular but occasional well rounded, typical aeolian grains occur. The non-carbonate grains of the loose sediments, of the most recent limestones and of the Miliolite Limestone are very similar.

There are two possible sources for the non-carbonate grains, either (a) they are of windblown origin, or (b) they are derived from submarine outcrops of Miliolite Limestone.

The addition of non-carbonate grains to the inner coral reef off Jazirat al Ftaisi is only possible from windblown sources. Sediments moving in across the oolite delta cannot cross the tidal channel; this is proved by the complete absence of oolite grains from the inner reef sediments. The Miliolite Limestone outcrops of Jazarat al Ftaisi are fronted by a channel and tidal currents within it run

Fig. 36 Percentage of non-carbonate grains in seaward sediments.


% NON-CARBONATE GRAINS
SHELF SANDS
INTER-DELTA SANDS
OOLITE SANDS
INNER REEF SANDS
0 2 4 6 8 10 12 14 16 18 20 22 24 26 28 30 32 34 36
% NON-CARBONATE

parallel to the coast (Fig. 22). No material derived from the Miliolite Limestone can thus reach the inner reef. The percentage of non-carbonate grains in the inner reef sediments represents therefore a windblown addition and this is obviously small. Local areas near Miliolite Limestone outcrops will gain some non-carbonate grains from them, but as the prevalent wind is from the north-west these grains will tend to be distributed in the mid and inner parts of the lagoon, down-wind from the Miliolite Limestone outcrops.

The high percentage of non-carbonate grains in the shelf sediments represents an accumulation against the foot of the oolite deltas. This concentration has evidently been derived from seaward and could either be part of a thin blanket of windblown grains or be derived directly from Miliolite Limestone on the floor of the shelf. However, in these seaward areas the derivation of sand-sized non-carbonate grains can be reasonably considered only from an adjacent land mass. But offshore winds are infrequent and so this possible source will be small. Miliolite Limestone has been observed on the shelf and it is considered that the greatest part of the non-carbonate accumulation is derived from this source. The Miliolite Limestone is itself a sandy aeolian limestone and the non-carbonate grains within it are identical with those of the surface sediments.

Occasional grains of specularite, probably from the mineralised salt plug of Jazirat Sir Abu Nu'air, occur in some non-carbonate residues. These could either have a windblown origin or have been drifted along the sea bottom in towards the coast, a distance of about 40 miles (65 kms.). Although a number of different rock types occur on Jazirat Sir Abu Nu'air, none of them can reasonably be the source of the non-carbonate grains found in the inner shelf sediments.

The non-carbonate grains forming cores to many oolites and occurring as individual grains atop the oolite deltas, are identical in character to those at the base of the delta slope in the shelf sediments. The shelf sediments thus move up onto the deltas and provide cores for the oolites. A second line of reasoning leading to the same conclusion comes from the low and high Mg calcites of the oolite sands. These minerals form over 30% of the shelf carbonates and are present on the delta tops mainly as cores to oolite grains. Neither the non-carbonate nor the calcitic grains are derived from the limestones directly underlying the oolite sands, as these limestones are aragonitic oolite limestones with a low percentage of non-carbonate grains.

In the mid and inner lagoon sediments the percentage of non-carbonate grains is usually low except downwind of Miliolite Limestone outcrops. In places, Miliolite Limestone immediately underlies the loose surface sediments and supplies non-carbonate grains directly to them.

Six non-carbonate residues were analysed for Sr by an X-ray fluorescence technique and the results are given below. The Sr is probably replacing Ca in feldspar grains. The data effectively show the common character of the non-carbonate grains of the shelf, interdelta and oolite sediments.

Sample	Wt % Non-Carbonate in total sample	Sr. p.p.m.
Shelf sands	26.2	180
	5.3	190
	33.3	200
Interdelta sands	14.3	190
Oolite sands	7.7	180
	6.4	190

SEA WATER ANALYSES

General

Salinity and pH measurements made on sea waters in the field have already been discussed (Chapter II). In addition to these general measurements 14 surface sea water samples were collected for chemical analysis; pH and temperature at the time of collection were noted. The samples were filtered as rapidaly as possible, to restrict evaporation losses to a minimum, into standard water sample bottles and flown to the U.K. Two bottles of each sample were taken, making 220 mls. available for the analyses. Two or three individual samples were collected from each sedimentary sub-environment; the sample locations are indicated in Figure 37.

The waters were analysed for the major constituents, using standard oceanographic techniques. Analyses were made for calcium (Ca), Magnesium (Mg), Strontium (Sr), Potassium (K), Sodium (Na), Chlorine (Cl), Sulphate (SO_4), and titration alkalinity. The data are given in Figure 38. The sediment studies have shown that only carbonate minerals are being formed in the marine environments and therefore Ca, Mg, Sr and alkalinity will be affected, but Na, K, Cl and SO_4 should remain constant.

Alkalinity and chlorinity (gms. Cl/Kgm sea water) analyses were carried out at the Marine Biology Station, Menai Bridge, Anglesey; the remainder of the analyses were completed at the National Institute of Oceanography.

Fig. 37 Locations of sea water samples. Samples are differentiated into the various sedimentary sub-environments from which they were taken.

LOCATIONS OF ANALYSED SEA WATER SAMPLES

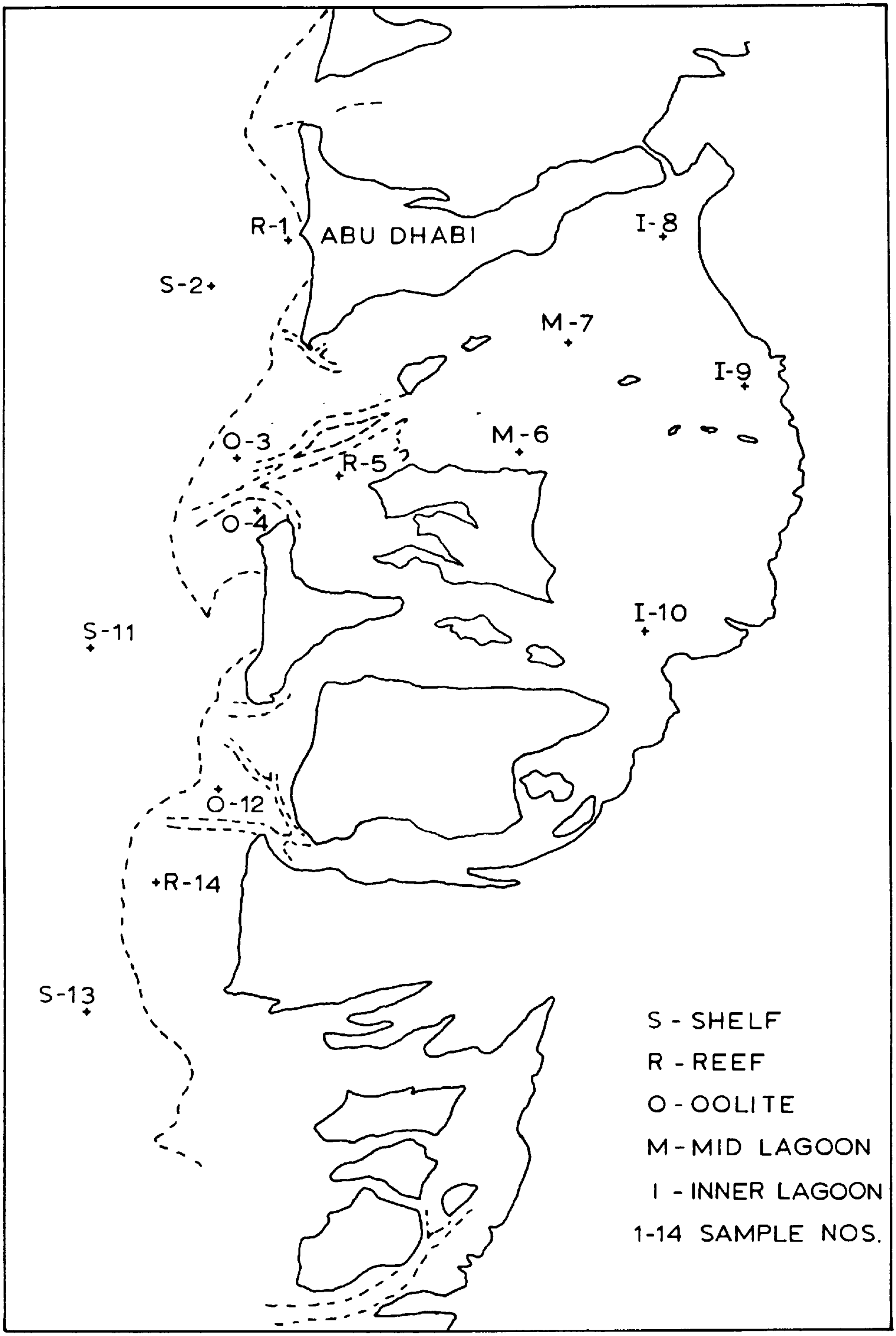


Fig. 38 Field data and laboratory analytical data of sea water samples.

SEA WATER TYPE	OPENWATER			REEF (Open Coast)	(Open Coast)	(Inner)	OOLITE DELTA			MID LAGOON		INNER LAGOON			AVERAGE[5] SEA WATER
Lab. Sample No	3	11	13	1	14	5	3	4	12	6	7	8	9	10	
Field T°C	26.2	27.0	27.0	26.2	28.0	26.4	27.9	26.8	25.5	28.0	25.2	24.9	26.0	24.5	--
Field pH[1]	8.30	8.30	8.30	8.30	8.32	--	8.25	8.25	8.32	--	--	--	--	--	8.1
Dilution[2] Factor	1.215	1.260	1.256	1.223	1.243	1.299	1.243	1.283	1.283	1.313	1.325	1.442	1.733	1.555	--
Density at 22°C	1.0290	1.0304	1.0300	1.0296	1.0296	1.0310	1.0296	1.0304	1.0306	1.0316	1.0316	1.0354	1.0428	1.0381	1.0242
Salinity ‰[3]	42.70	44.14	44.05	42.97	43.57	45.61	43.68	45.14	45.03	46.08	46.37	50.59	60.62	54.38	35.00
Chlorinity[3] gms/kgm (‰)	23.64	24.43	24.39	23.79	24.12	25.25	24.18	24.99	24.93	25.51	25.67	28.01	33.57	30.11	19.37
SO_4 gms/kgm	3.315	3.374	3.383	3.314	3.363	3.512	3.363	3.483	3.459	3.552	3.517	3.889	4.656	4.206	2.712
Ca gms/kgm	0.501	0.517	0.516	0.499	0.508	0.528	0.503	0.521	0.523	0.536	0.530	0.581	0.686	0.620	0.412
Mg gms/kgm	1.615	1.662	1.665	1.620	1.641	1.720	1.651	1.697	1.683	1.739	1.754	1.909	2.287	2.050	1.290
Sr gms/kgm	0.0090	0.0096	0.0096	0.0091	0.0092	0.0099	0.0090	0.0095	0.0095	0.0097	0.0098	0.0107	0.0125	0.0115	0.0077
K gms/kgm	0.493	0.506	0.505	0.493	0.499	0.520	0.497	0.514	0.514	0.528	0.527	0.580	0.698	0.622	0.399
Na gms/kgm	13.07	13.51	13.49	13.25	13.36	13.92	13.27	13.74	13.65	14.05	14.08	15.23	18.54	16.66	10.76
Cl mqs/kgm	666.8	689.2	687.9	671.0	680.4	712.1	681.9	704.7	703.0	719.5	724.1	790.0	946.9	849.3	546.3
Alk mqs/kgm[4]	2.48	2.54	2.51	2.31	2.44	2.26	2.52	2.36	2.46	2.31	2.37	2.35	2.16	2.33	2.43
SO_4 mqs/kgm	69.01	70.23	70.42	68.98	70.01	73.11	69.99	72.49	72.01	73.94	73.21	80.95	96.92	87.54	56.45
Ca mqs/kgm	24.99	25.78	25.75	24.92	25.35	26.33	25.10	26.02	26.11	26.74	26.46	28.97	34.21	30.93	20.58
Mg mqs/kgm	132.8	136.7	136.9	133.2	135.0	141.4	135.8	139.6	138.4	143.0	144.2	157.0	188.0	168.6	106.1
Sr mqs/kgm	0.205	0.218	0.218	0.207	0.210	0.255	0.204	0.217	0.217	0.222	0.224	0.244	0.285	0.263	0.176
K mqs/kgm	12.61	12.94	12.91	12.60	12.77	13.29	12.71	13.13	13.15	13.50	13.48	14.83	17.84	15.90	10.20
Na mqs/kgm	568.7	587.5	586.8	576.5	580.1	605.5	577.4	597.8	593.9	611.2	612.4	662.6	806.6	724.5	467.9

Notes: 1. Field pH not recorded for all samples owing to breakage of instrument.

2. Factor by which original samples were diluted to bring all waters to approximately 35‰S.

3. Chlorinity = Total Halides : Salinity = 0.03 + 1.805 x chlorinity

4. Alkalinity = titration alkalinity

5. Average sea-water data from Sverdrup et al (1940) and F. Culkin, (pers. comm.)

Throughout this work concentrations are expressed as weights or molecular equivalents in a particular weight of solution of sea-water (usually one kilogram). Where salinities, and therefore densities, of water vary greatly then obviously comparable volumes will be of different weights. In order to accurately compare samples, irrespective of their concentration, it is necessary to compare weights or molecular equivalents in identical weights of solution. To this end, densities of the water samples were determined.

Knowing their densities, the water samples were diluted with a known weight of distilled water to bring them all to a common concentration of approximately 35‰ S. This then made all titres more regular and reduced many minor difficulties otherwise caused by differing viscosities, densities and concentrations.

Density

The volume of a pipette was determined by weighing accurately pipetted volumes of distilled water, at a known temperature. Standard data on the volume of one gram of distilled water at the laboratory temperature then enabled the pipette volume to be calculated. Knowing the accurate pipette volume, aliquots of the undiluted sea-waters were pipetted into weighed flasks and their densities determined. At least two determinations were made on every sample and accuracy is within 0.0003 gms./cc.

Chlorinity

Chlorine determinations were made using the standard silver nitrate titration method. The values represent total halide (chlorine and bromine + fluorine + iodine) but as chlorine is present in such great excess the other halides are ignored and computed as if they were chlorine. Chlorinity is expressed in grams of Cl/kgm solution (sea water) i.e. parts per thousand (‰). Salinity is derived from chlorinity by the empirical relationship : salinity = 0.03 + 1.805 x chlorinity (Sverdrup et al, 1942).

Chlorinity analyses are accurate to within 0.05‰. Chlorine is capable of fairly accurate determination and as it is present in such large amounts in sea water (constituting more than55% of total ions) it is useful as an extremely sensitive measure of concentration. Chlorides are not precipitated from sea-water until high concentrations are attained and thus in areas of carbonate deposition, chlorine may be used as a reference against which changes in the relative amounts of other ions may be measured.

Total Cations

The total cations (Ca + Mg + Sr + Na + K) were determined using an ion-exchange resin in the hydrogen form. Separate determination was later made for Ca, Mg, Sr and K and the amount of Na found by difference.

Sodium

Sodium is rarely determined directly because of problems involved in the determination of the alkali metals. Sodium was found by difference: as indicated above. Accuracy of the determination is dependent on the other analyses but the coefficient of variation is about 0.01%.

Potassium

Potassium was determined by the method of Sporek (1956), modified in some details by Dr. R. A. Cox of the National Institute of Oceanography. The potassium is determined gravimetrically as the potassium tetraphenylboron salt. The coefficient of variation is 1%.

Calcium, Magnesium and Strontium

Total Ca + Mg + Sr was determined by standard EDTA titration. Calcium was determined separately by EGTA titration. End points in both cases were determined visually. Strontium was determined separately using a Unicam S.P. 900 Flame-spectrophotometer. Magnesium was then found by difference. The coefficients of variation are : Sr -5%; Mg -0.1%, Ca 0.4%.

The ratios of Ca, Mg and Sr to each other and to chlorine are particularly critical and indicate the course of events in the precipitation of carbonate minerals across the lagoon complex from the open shelf sea.

Sulphate

Sulphate was determined gravimetrically as barium sulphate using the method of Bather and Riley (1954). The coefficient of variation is 0.01%.

Alkalinity

The total alkalinity of the waters was determined by potentiometric titration, using essentially the method of Greenberg et al (1932).

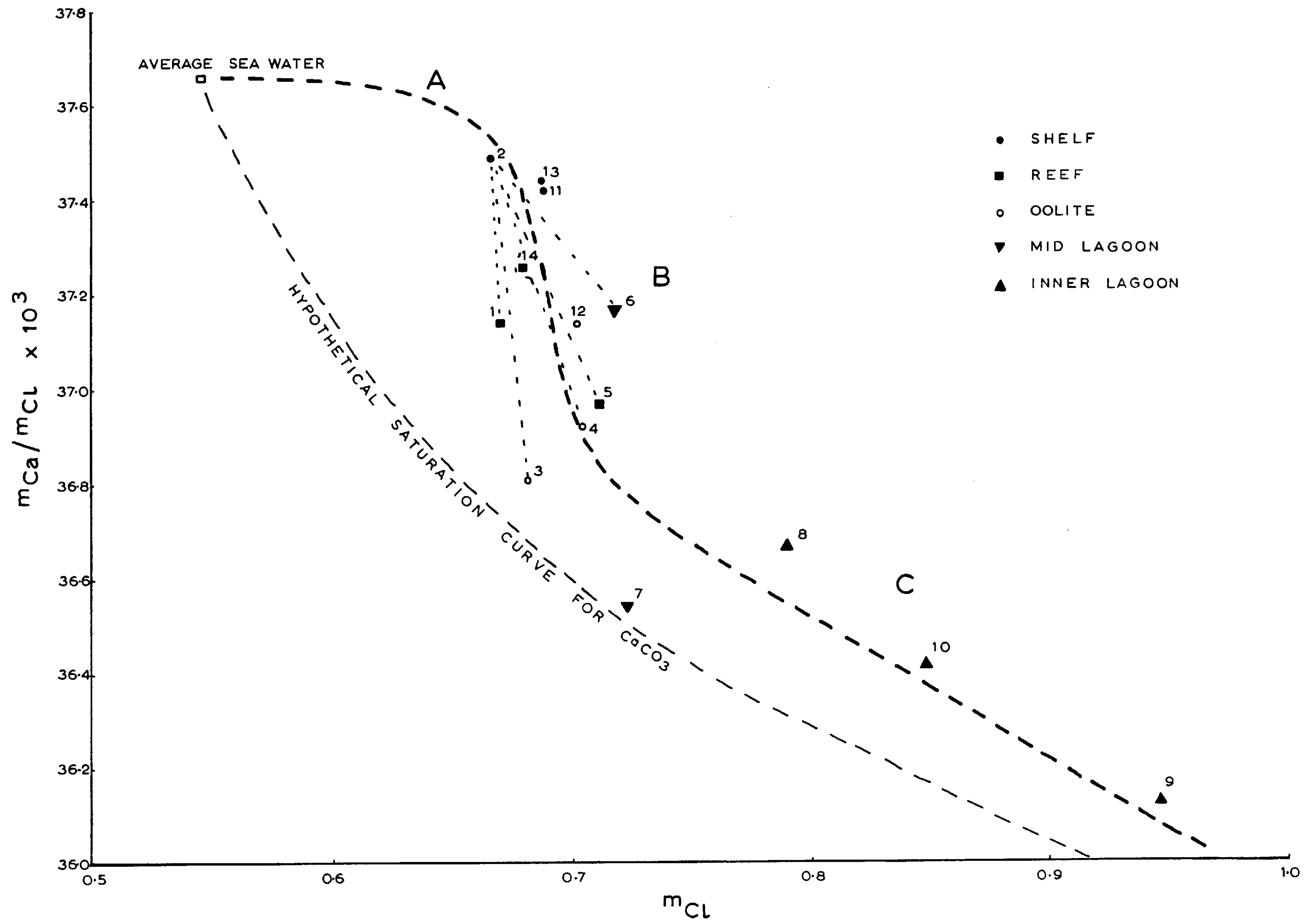
DISCUSSION OF RESULTS

The ratios of potassium, sodium, strontium, magnesium and sulphate to chlorine were found to be constant, within the limits of error of the analyses. Although strontium is being withdrawn from the waters and precipitated with calcium in aragonite the amount lost is masked by errors involved in determining such small amounts of strontium as exist in sea water. Some magnesium is also being precipitated with $CaCO_3$ but again no trend is discernable as the amount lost is very small. However, the calcium/chlorine ratio shows a marked decrease between the open shelf waters and the lagoon waters (Fig. 39). The calcium to chlorine ratio for average sea-water is also shown. The average sea-water value applies to Indian Ocean water entering the Persian Gulf at the Strait of Hormuz, to balance loss by evaporation and outflow at depth (Chapter I). This water becomes concentrated within the Gulf and loses some calcium by the time it arrives off the Trucial Coast (Phase A, Fig. 39). Whitings have been observed offshore by Wells and Illimg (1964) and the author and thus the decrease in calcium is probably caused by both organic extraction and chemical precipitation in the shelf environment.

The waters from each of the physiographic units of the inner shelf and lagoon areas show appreciable variation in both concentration and calcium to chlorine ratios (Phases B and C, Fig. 39). This is to be expected if one considers the complex tidal circulation pattern shown in Figure 22. The three open shelf waters are grouped fairly close together and show the highest calcium to chlorine ratios. Waters from the open coast reefs show little concentration but some loss of calcium. The oolite and inner reef waters show further concentration and appreciable loss of calcium. The two mid lagoon samples differ widely but reference to Figure 29 shows that a tongue of relatively low salinity water reaches far into the mid lagoon; this water flows in along the tidal channels and probably loses little calcium en route to the mid lagoon. Mixing of lower salinity, calcium rich waters with more saline calcium poor waters would explain the fairly high calcium to chlorine ratio of sample No. 6. The low

Fig. 39 To show the relative loss of Calcium between open shelf waters and inner lagoon waters. Note particularly the form of the curve, a feature discussed in detail in the text.

RELATIVE CALCIUM LOSS ACROSS BARRIER AND LAGOONAL ENVIRONMENTS



calcium to chlorine ratio of the other mid lagoon sample (No.7) is difficult to explain; it obviously represents a water which has undergone extensive calcium loss without excessive concentration. The inner lagoon waters show progressive decrease in the Ca/Cl ratio as concentrations increase.

The data repay further examination and of particular interest is the form of the curve (Figure 39). Samples 1, 2, 3, 4, 5 and 6 were collected in that order over a partial flood tide cycle and show progressive concentration with rapid extraction of calcium. The open coast reef water (sample No. 1) shows little concentration compared with the shelf water (No. 2) but calcium has been removed, presumably by the organic elements of the reef. Comparison of sample 2 with samples 3 and 4 shows rapid loss of calcium to take place across the oolite deltas. This loss must represent a chemical precipitation of $CaCO_3$ as there is insufficient density of fauna in these areas to effect equivalent organic withdrawal of calcium. The inner reef and mid lagoon samples (Nos. 5 and 6) probably have a complex history, having previously crossed the oolite deltas or come in along the tidal channels. Samples 12, 13 and 14 were collected in a similar manner and show a similar story of calcium loss as the waters pass over the reefs and oolite deltas. However, here again the water circulation is complex; this is indicated by sample No. 14 (open coral reef) being less saline than No. 13 (open shelf).

In general the relative loss of calcium on the open shelf can be seen to proceed at a fairly slow rate and it is not until the waters are within the high energy belt of the coastal coral reefs and oolite deltas that there is a very rapid extraction of calcium from the waters. Further back in the lagoon calcium loss again proceeds at a relatively slow rate.

Across the shelf the sea-water concentration rises gradually, calcium loss is limited and progressive super-saturation results with respect to $CaCO_3$. In the turbulent inner shelf zone of coral reefs and oolite deltas nucleation takes place, $CaCO_3$ is precipitated and the level of super-saturation rapidly falls. Further slow precipitation occurs as sea-water concentrations gradually rise in the inner lagoon.

An attempt can be made to quantify this relationship. Taking a parent sea-water of salinity 35‰ and calcium 20.58 Meqs./Kgm, over phase A (Fig. 39), which comprises the open shelf environment, the loss of calcium amounts to only 0.04% of the total available calcium, even though the change in concentration is appreciable (x 1.24). Phase B, with little further concentration, shows a loss of 1.96% of the initially available calcium. Phase C, where concentration of the water proceeds to very high salinities, a further loss of 1.90% occurs. The total loss of calcium amounts to 3.9% of the initial amount available.

In most marine environments of Recent carbonate sedimentation sea-water salinities are not as high as those of the Trucial Coast and many areas are not as physically restricted as are the Trucial Coast lagoons. Nevertheless, this general picture of calcium withdrawal from the waters and calcium carbonate precipitation also probably holds true in areas of much lower salinity. For example, Cloud's recent Bahaman data (Cloud 1961: Fig. 34) are capable of interpretation in this way, with most rapid withdrawal of calcium taking place at the bank margins. In the offshore, deep water or open shelf environment calcium is extracted from a relatively large body of water by mainly organic agencies. When the shelf or deep waters move into shallower areas and cross reefs and shoal banks, the warming and turbulence of the waters leads to rapid calcium loss. Further gradual loss of calcium accompanies concentration of the waters in lagoons, as along the Trucial Coast, or in the interior parts of extensive shoal areas such as the Bahama Banks.

In the Bahamas, Cloud (1961) found that both Ca/Cl and alkalinity/Cl ratios decreased across the banks from the open waters. In this study, titration alkalinity determinations showed a decrease in the alkalinity/Cl ratio from the open waters to the inner lagoon but individual stations showed poor agreement between Ca/Cl and alkalinity/Cl ratios. However, the Ca/Cl ratios are sufficient to prove the precipitation of $CaCO_3$ from the Trucial Coast waters.

The Mg/Ca ratio in sea-waters is one of the factors affecting the organic coprecipitation of Mg with $CaCO_3$. In average sea-water the Mg/Ca ratio is 5.15 (ratio of equivalents). In the Trucial Coast inner shelf waters a value of 5.31 was found and in the inner lagoon this increased to 5.50. Calcitic organisms common to areas of varying Mg/Ca ratio should show varying Mg/Ca ratios in their skeletal calcites. Those which live where the Mg/Ca ratio is highest should show highest Mg/Ca ratios in their skeletal calcites. Increase in water temperature also tends to increase the amount of Mg in solid solution within the skeletal calcites. Thus, as the lagoonal areas have higher Mg/Ca ratios and higher annual water temperatures than are found on the open shelf areas, organisms in the former area should show higher amounts of Mg. Along the Trucial Coast calcitic organisms are not abundant in the lagoon sediments. As this source of Mg will therefore be small, it cannot be called upon to partially explain the rapidity of the dolomitisation of the lagoonal sediments (Chapter VI).

The Sr/Ca ratio in the sea waters has been found to be constant. A loss of 4% of the available Ca has been shown to occur and as the Sr/Ca ratio is constant there must also have been a 4% loss of the available Sr. This is an indication that much of the precipitation of $CaCO_3$ takes the form of high strontium-aragonite, as the

Sr/Ca ratio in such aragonite is almost the same as that of sea-water. If appreciable quantities of low strontium-aragonite or calcite were precipitated then the Sr/Ca ratio would rise with increasing loss of Ca.

The Sr/Ca ratio of the sea-waters was found to average 0.84×10^{-2} (ratio of equivalents), but two check analyses by H. D. Holland at Princeton on splits from samples 3 and 4 showed a Sr/Ca ratio of 0.95×10^{-2}. The Princeton value is about 12% higher. Check analyses for Cl, Ca and Mg were also made and agreement between the two laboratories was very close. Further check Sr determinations are now being made by Dr. F. Culkin at the National Institute of Oceanography, Dr. H. D. Holland at Princeton University and Dr. K. Turekian at Yale University.

MAGNESIUM AND STRONTIUM IN CARBONATE SEDIMENTS AND SKELETONS.

Magnesium

A few samples were run spectrographically for Mg, using a semi-quantitative method. The percentage of Mg in the samples was found to be dominantly related to the mineralogy. For example, calcitic organisms usually show high values and reef sediments with appreciable quantities of high Mg calcites are higher in Mg than sediments which are largely aragonitic. The values determined are given below:

Sample	% $MgCO_3$	Sample	% $MgCO_3$
Murex	0.3	Echinodiscus	10.5*
Glycimeris	0.1	Echinometra	10.5*
Glycimeris	1.1	Reef Sands (6)	av. 6.7
Meretrix	0.5	Shelf Sands (4)	av. 3.4
Cuttlebone	0.5	Interdelta Sands (3)	av. 2.8
Lithophyllum	16.0*	Oolite Sands (8)	av. 2.0

* High-Mg calcite.

Lagoonal sediments were not analysed as they normally comprise more than 90% aragonite and the percentage of $MgCO_3$ will be probably less than 0.5%.

Strontium

The percentage of strontium was determined in a few organisms and in about 100 sediment samples, using a standard X-ray fluorescence method. Standards were made using a base of 90% $CaCO_3$ and 10% $MgCO_3$, to which known amounts of $SrCO_3$ were added. Four standards were made containing 1198, 2993, 5973 and 12,000 p.p.m. strontium respectively; these covered the range of strontium expected in the samples. The values are accurate to within ±10%.

The percentage of strontium (expressed as $SrCO_3$) of a few organisms is shown below:

Sample	% $SrCO_3$	Sample	% $SrCO_3$
Murex	0.32	Favia	1.32
Glycimeris	0.35	Plesiastrea	1.32
Glycimeris	0.34	Cyphastrea	1.27
Meretrix	0.23	Porites	1.29
Cerithium	0.35	Platygyra	1.35
Mixed Lagoonal Gastropods	0.32	Lithophyllum	0.47*
Cuttlebone	0.61	Echinometra	0.38*
Acropora	1.26	Echinodiscus	0.41*

*Calcite

The mollusca, even when aragonitic, exert a dominant vital control over the amount of $SrCO_3$ precipitated in their skeletal materials, and values are generally low. The corals, however, exert no vital organic control and show high values of $SrCO_3$ in their aragonitic skeletons;

in fact, the values found are similar to those expected from chemically precipitated aragonite under the temperature conditions prevailing in this area.

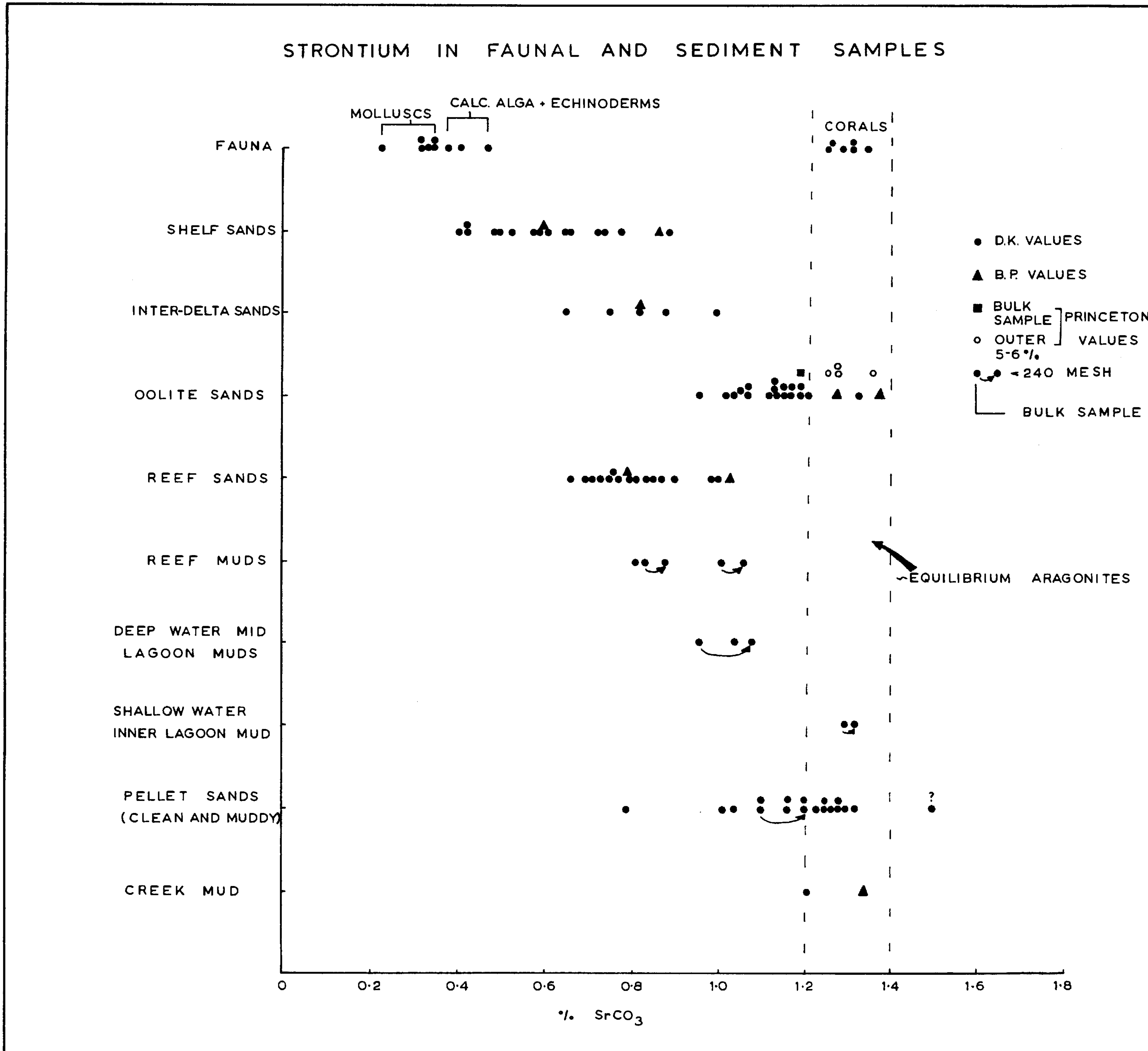
The percentage of strontium, expressed as $SrCO_3$, in the total carbonate fraction of the sediments, is shown in Figure 40. The non-carbonate content of the shelf and interdelta sediments may be as high as 30% and even in the oolite sands it may reach 15%. Thus it was necessary to check the assumption that the strontium of a sediment sample was associated with the calcium of the carbonate minerals and not in any way with the non-carbonate fraction. Six non-carbonate residues were tun for strontium and the values were found to be low enough (less than 200 p.p.m.), to be ignored.

BP Research Laboratories kindly made some check analyses which compared closely with those of the author.

The shelf sediments average 0.6% $SrCO_3$ in the carbonate fraction. These sediments have fair amounts of molluscan debris and of the carbonate fraction less than 70% is aragonite. The presence of low strontium molluscan aragonite and a relatively low percentage of aragonite readily accounts for the low $SrCO_3$ values.

The oolite sands show 1.0-1.4% $SrCO_3$ and average 1.1%. The values are for the carbonate of the bulk sediments and include the aragonite of the oolite jackets and the aragonite and calcite of core grains. H. D. Holland of Princeton University has run one total sample which showed

Fig. 40 Strontium values (as % $SrCO_3$) of sediments and organisms.



STRONTIUM IN FAUNAL AND SEDIMENT SAMPLES
MOLLUSCS
CALC. ALGA + ECHINODERMS
CORALS
FAUNA
SHELF SANDS
INTER-DELTA SANDS
OOLITE SANDS
REEF SANDS
REEF MUDS
DEEP WATER MID LAGOON MUDS
SHALLOW WATER INNER LAGOON MUD
PELLET SANDS (CLEAN AND MUDDY)
CREEK MUD
D.K. VALUES
B.P. VALUES
BULK SAMPLE
OUTER 5-6 %
PRINCETON VALUES
<240 MESH
BULK SAMPLE
~EQUILIBRIUM ARAGONITES
?
0
0·2
0·4
0·6
0·8
1·0
1·2
1·4
1·6
1·8
% SrCO3

a value of 1.19% $SrCO_3$. Further analyses were made of the outer aragonitic oolite jackets using a stripping technique and these gave values averaging 1.28% $SrCO_3$. The latter value falls in the range of values expected from chemically precipitated aragonite under the temperature conditions of this area.

The reef sediments are seen macroscopically and microscopically to be a mixture of coral and calcareous algal fragments. Echinoderm and calcareous algal calcitic carbonates containing about 0.4% $SrCO_3$ are mixed with aragonite coral debris containing 1.3% $SrCO_3$. The resulting reef sediments show $SrCO_3$ values of 0.7-1.0%. The reef muds show similar values to the reef sands.

The interdelta sands, being mixtures of shelf, reef, and oolite facies, show the expected intermediate $SrCO_3$ values.

The mid and inner lagoon sediments are almost entirely aragonitic and typically show a high percentage of $SrCO_3$. The two deep water mid lagoon muds contain about 20% calcite, but the aragonite fraction must be fairly high in $SrCO_3$ to give a bulk sediment value of about 1.0% $SrCO_3$. The inner lagoon, shallow water aragonite mud, the clean and muddy pellet sands and the creek mud all show values of 1.0-1.3% $Sr\ CO_3$. The shallow water inner lagoon mud was passed through a 240 mesh sieve and run again, when it showed a slightly higher value. A few small molluscan and foraminiferal fragments had been left behind on the sieve and the mud was then an almost pure aragonite needle mud. The $SrCO_3$ value of this needle mud is almost the same as that of the oolite jackets and is in the expected range for chemically precipitated aragonite. From their general features and mineralogy it has already been concluded that the clean and muddy pellet sands of the mid and inner lagoon were all derived from an initial aragonite mud. The strontium values corroborate this fairly well. The lower pellet sand values are caused by the presence of admixed skeletal grains, mainly of molluscan origin.

In addition to the shallow water inner lagoon mud, two reef muds, one deep water, mid lagoon mud and one muddy pellet sand were each passed through a 240 mesh sieve and the concentrated fine grades re-run for strontium. All showed an increase in $SrCO_3$ over the bulk sediment value. In the case of the deep water mid lagoon mud and muddy pellet sand, skeletal grains were removed, increasing the percentage of high strontium-bearing aragonite needles which are concentrated in the finer fraction. The reef muds also have a concentration of high strontium bearing aragonite in the finest fractions. In the lagoonal sediments the high strontium aragonite needles are considered to be of chemical origin, whereas in the reef muds they could be of chemical origin or derived from abrasion of coral debris, or both.

ORIGIN OF THE CARBONATE SEDIMENTS

The origin of the recognisable organic contributions to the sediments is obvious, and the source of the non-carbonate grains in this area has already been discussed. The oolite sands and aragonite muds have a rather less obvious origin.

Urey et al (1951) and Epstein et al (1953) showed that under equilibrium conditions the O^{18}/O^{16} ratio of marine carbonates depends on the temperature of precipitation and on the O^{18}/O^{16} ratio of the water from which the carbonates were precipitated. Many organically precipitated carbonates were found in which this relationship was complicated by a vital control exerted by the organism itself. Such carbonates are thus disequilibrium phases with respect to sea water. However, Lowenstam and Epstein (1957) have found that the O^{18}/O^{16} ratios of aragonitic oolites from the Bahamas shows them to be in equilibrium with the O^{18}/O^{16} ratios of the waters at the temperature at which they were formed. The strontium co-precipitation data of Holland (op. cit.) indicate that oolitic aragonite is also an equilibrium carbonate. Both these parameters strongly support a direct chemical origin for the aragonite jackets of the oolites, rather than an origin from organic precipitation.

The origin of aragonite needle muds on the Great Bahama Bank has been comprehensively discussed and studied by Cloud (1961). The aragonite needles have been considered to be either mainly of algal origin or direct chemical precipitates. The aragonite needle muds of some areas can be shown to be entirely of algal origin (Lowenstam 1957). In the Bahamas, Cloud considered the majority of the needles to have been chemically precipitated, the algal contribution being relatively small.

Along the Trucial Coast there is a complete absence of Halimeda, the chief source of aragonite needles in the Bahamas. Other algae which have been shown to be the source of aragonite needles are very rare. Another possible source of aragonite needles is from the abrasion of aragonitic skeletal debris, such as corals and molluscs. However, along the Trucial Coast the coral source is very limited and fine sediments which include abrasion derived reef materials are confined to hollows on and close behind the reefs and to small back reef lagoons. The molluscs are an extremely abundant group in the lagoons , are largely aragonitic and break down fairly readily into individual needles. But the aragonite of the molluscs is fairly low in $SrCO_3$ (less than 0.4%), whereas the lagoonal aragonite muds average 1.2-1.3% $SrCO_3$. Thus in the lagoons, the presence of high strontium aragonite needle muds can only be accounted for by chemical precipitation. Additions from other sources have been found to be slight. At least 80-90% of the mid and inner lagoon sediments must be considered to be chemically precipitated in order to account for the $SrCO_3$ values of the bulk sediments.

The precipitated aragonite need not necessarily have entirely originated within the lagoons. On the shelf, the relative loss of Ca is small but as large volumes of water are involved the amount of $CaCO_3$ precipitated may be quite considerable. Wells and Illing (1964) and the author have noted the occurrence of whitings in these areas. The aragonite precipitated from the whitings may accumulate on the shelf but may also be gradually drifted to the innermost shelf and lagoon areas. Evidence has been given showing movement of non-carbonate and calcitic grains in across the shelf towards the lagoons. Likewise, precipitated aragonite needles may well be moved and finally accumulate in the lagoons.

The loss of Ca over the oolite deltas and offshore coral reefs is relatively great but water depths are small and relatively small volumes of water are involved. The $CaCO_3$ is partly precipitated organically and partly chemically (inorganically). The chemically precipitated aragonite is represented in part or entirely by the development of oolitic encrustations. However, it is more than likely that myriads of individual aragonite needles are also produced, which are then drifted into the lagoons and accumulate in sheltered areas.

Further loss of Ca in the lagoons has been shown to occur. This may take the form of both organic or chemically precipitated carbonate. Chemically precipitated aragonite in the mid and inner lagoon will also accumulate in these areas.

Thus the origin of the aragonite needles may be complex but they are almost entirely chemically precipitated. Although precipitation takes place from waters of greatly differing salinities (38-65‰) the Sr/Ca ratio of the waters will be constant and thus the $SrCO_3$ content of the precipitated aragonites will be dependent on the temperatures of precipitation. As the mean annual water temperatures of the various areas differ by less than 5°C, the $SrCO_3$ content of the various precipitated aragonites will be closely similar. From the data already presented, the amount of organically produced aragonite needles at least along this part of the Trucial Coast must be considered trivial.

C H A P T E R V

PLEISTOCENE AND RECENT EVOLUTION OF THE TRUCIAL COAST.

Before considering the diagenetic developments of the sabkha, the evolution of the present day distribution of islands and marine areas and of the sediments will be discussed. This may be effected by considering the present day sediments and factors affecting their development and distribution, together with data derived from cores and pits dug in the loose sediments. A study of land areas, combined with the analysis of aerial photographs has enabled old shoreline positions to be determined, and together with the other data, a synthesis may be made, and past, present and future developments discussed.

Intimately associated with the uncemented Recent sediments are a series of limestones. The Miliolite Limestone is the earliest of these and the present cycle of sedimentation can be best considered as starting from the erosion of the platform on which the Miliolite Limestone was laid down. In the following discussion evidence is given of relative changes in sea-level but no attempt is made to correlate these with known Pleistocene sea-level fluctuations. The area is tectoncially stable and unwarped, but epeirogenic movements may have ocurred.

The Miocene rocks of the mainland form the basement on which the later sediments were deposited. A platform of marine erosion was cut into this series of marls, dolomitic limestones and gypsum beds and except for a few outlying areas, the Miocene rocks were eroded away to the position of the present day line of hills at the inner edge of the coastal plain or mainland sabkha. The outlying areas, now forming buttes and mesas which project from the sabkha surface, would have been islands at this time. The marine platform was eroded to at least 25 ft. (8 ms.) below present sea level, as off Ras al Dhabayah, Miliolite Limestone occurs at this depth.

Relative sea level then fell and the terrigenous and calcareous sands of the Miliolite Limestone accumulated on the old marine platform. Many of the Miocene rocks are sandy dolomites and the non-carbonate grains are very similar to those of the Miliolite Limestone, being fine to very fine in size, angular to sub-angular, and mainly of quartz and feldspar. It is likely therefore that these grains were rederived by the wind from the marine erosion products of the Miocene rocks and blown southward to form the aeolian accumulations of Miliolite Limestone times. Abundant carbonate grains also accumulated mainly of molluscan origin together with some reef grains (coral, calcareous algal and echinoid). No oolites have been found in the Miliolite Limestone, although some fine-textured, ellipsoidal grains may be pellets of some kind.

The Miliolite Limestone must have a marine equivalent in depth and/or to seaward, but this is nowhere exposed. The only facies exposed is aeolian, and this extends from at least 25 ft. (8 ms.) below sea level, off Ras al Dhabayah, to more than 100 ft. (32 ms.) above sea level where the sands were banked against the hills of Miocene rocks. In most places, the Miliolite Limestone outcrops rarely extend more than 10 ft. (3 ms) above present sea-level as a later plane of marine erosion cuts across it at this height. The available evidence indicates some features of Miliolite Limestone times:-

1. There were no oolite forming areas.
2. Skeletal grains of mollusc and reef origin were abundant, together with fine, non-carbonate grains.
3. Lagoons similar to the present day were not developed. Inland of such sheltered lagoons aeolian accumulations would have been insignificant - this is found in the present environment.
4. The presence of root or rhizome moulds, similar to those of some plants growing on

the modern dunes give evidence of a partial plant cover.

5. Winds were onshore and north-westerly. This is indicated by the constant orientation of the backslope bedding of the Miliolite Limestone dunes. Barchan dunes were often developed, in which case, dips on the cross-bedding may range from south-west to east, with south and south-east predominating.

The data indicate that offshore the environment was probably a simple, shallow shelf, with onshore winds producing a high energy coastline similar in some ways to parts of the present day open coast, where molluscan debris is abundant and where the higher parts of beaches and berms are stripped of their sediments, which are then blown inland and accumulate as dunes.

The Miliolite sands thickly blanketed the area , but as this was an aeolian accumulation the uppr sediment surface was probably far from plane, as evidenced by the great banking of sands against the old shoreline hills at the inner edge of the mainland sabkha. The Miliolite sands became cemented, possibly by a kunkar limestone in the early phases, but cementation was fairly far advanced by the time the oncoming marine phase was initiated.

The Miliolite Limestone today is a calcitic limestone with only occasional aragonitic skeletal grains remaining. The latter often have a fibrous, rotten appearance and readily disaggregate. A thin cement of small, equant calcite crystals surrounds the original grains. The majority of the latter have been dissolved and the rock has a high secondary porosity. Occasional calcitic skeletal fragments still remain.

Relative sea-level then rose and the Miliolite Limestone was largely planed off at about 10 ft. (3 ms.) above present sea-level. This marine platform is developed into the inner edge of the coastal plain but in the Abu Dhabi area has not been found cutting into the Miocene rocks. The outlying Miocene buttes and mesas have generally had any surrounding Miliolite Limestone stripped away and at Mirfa, to the west (Fig. 2) the succeeding Shell Limestone rests directly on Miocene rocks. As with the period of marine planation which eroded away much of the Miocene rocks, but left odd areas remaining to seaward, so with the Miliolite Limestone. Small islands of Miliolite Limestone remained permanently above sea level at Ras al Kahf, Jazirat abu Kashasha, Jazirat al Ftaisi, on the easterly of the two small islands between Jazirat al Ftaisi and Abu Dhabi island, on the south-eastern end of Abu Dhabi island and in places on the present mainland sabkha (for example, the prominent upstanding rock called Nuseila, lying 2-3 miles (3-4 kms.) north of Mahawi). Some of the areas of Miocene rocks which were islands in the pre-Miliolite Limestone stage became islands again during Shell Limestone times.

A retreating sea left behind a coarse skeletal deposit, the Shell Limestone, containing in its lower parts fragments of Miliolite Limestone or where it overlay Miocene rocks, as at Mirfa, containing Miocene fragments. Molluscs in place of growth have been found on the unconformity, together with mollusc borings and many other features proving it to be a plane of marine erosion and deposition. The Shell Limestone sediments in places are false-bedded, with the cross laminae dipping offshore.

The surface evidently closely resembled that shown in Fig. 40 which shows a recent Miliolite Limestone erosion surface on a beach, with fragments of Miliolite Limestone and a concentration of single, abraded lamellibranch valves showing a strong preferred orientation. The dune-bedded Miliolite Limestone with overlying marine Shell Limestone is shown in Fig. 41.

The marine Shell Limestone was laid down over the entire area, although subsequent erosion has largely removed it. Today it is found overlying Miliolite Limestone in isolated outcrops at the inner edge of the coastal plain and in some seaward areas where outcrops of Miliolite Limestone remain which are more than 10 ft. (3 m.) above present sea-level. The Shell Limestone is almost entirely calcitic, all early aragonite having been dissolved. The diagenetic history of the Shell Limestone is similar to that of the Miliolite Limestone and it is now cemented with calcite and

Fig. 40 Modern beach cut into Miliolite Limestone. Cross-bedded Miliolite seen running from top to bottom; surface littered with Miliolite fragments together with single lamellibranch values showing strongly preferred stable orientation. A surface very similar to that developed during Shell Limestone times.

has considerable secondary porosity.

The Shell Limestone is never more than five ft. (1.5 ms.) thick in its marine facies and ranges from a coarse shell gravel to a coarse shell sand. It varies greatly from outcrop to outcrop and no consistent differences are apparent, either parallel or at right angles to the old shoreline. Mollusc and reef grains are the dominant components, although in places oolites and rounded fragments of oolite limestone are fairly abundant. The Shell Limestone was evidently an open coast deposit. The thickness of five feet may be an indication of the tidal range. The eroded debris from the Miliolite Limestone was transported either seawards or landwards, as it does not form a conspicuous part of the Shell Limestone. Possibly the broken down Miliolite Limestone grains were blown inland and added to the desert sands, a relationship paralleled by the derivation of the non-carbonate grains of the Miliolite Limestone from the earlier Miocene rocks.

Where Miliolite Limestone islands were low and had gentle slopes, the normal marine Shell Limestone is found to pass laterally into an aeolian facies, which may be up to 15 ft. (5 ms.) thick. But the constituent particles are all derived from the skeletal grains of the marine facies and the deposit is very different from the underlying Miliolite Limestone. It contains very few non-carbonate grains, often contains oolites, and is generally a medium sized sand whereas the Miliolite Limestone grains are fine to very fine in size,

Fig. 41 Marine Shell Limestone overlying dune bedded Miliolite Limestone.

oolites do not occur and non-carbonate grains may constitute 30-40% of the rock. One common feature, however, is the direction of the aeolian cross-bedding, indicating that in Shell Limestone times the prevalent wind was still north-westerly.

Miliolite islands with steeper slopes show no accumulation of a dune facies equivalent to the Shell Limestone. The Miliolite Limestone outcrops of Nuseila and the ridge to the west of Abu Dhabi causeway were probably islands of this type.

Again, there seems to have been no lagoon development and the environment was one in which coarse skeletal deposits could accumulate - perhaps again a simple open coast with a shallow offshore shelf. Oolite formation may have been an intertidal beach phenomenon or may have occurred on offshore shoal banks. The former suggestion is unlikely as no accumulations of oolitically coated grains have been found: the oolites, at most, constitute 10-20% of the sediment. The presence of well rounded oolite limestone grains indicates that some oolite limestone was being formed and rederived, possibly similar to some of the present day, thin, aragonitic oolite limestones. Except for the absence of oolites, the modern beach sands and gravels of the offshore island of Jazirat Sir Abu Nu'air have much in common with the Shell Limestone sediments, which adds weight to the suggestion that the Shell Limestone is an open coast beach deposit.

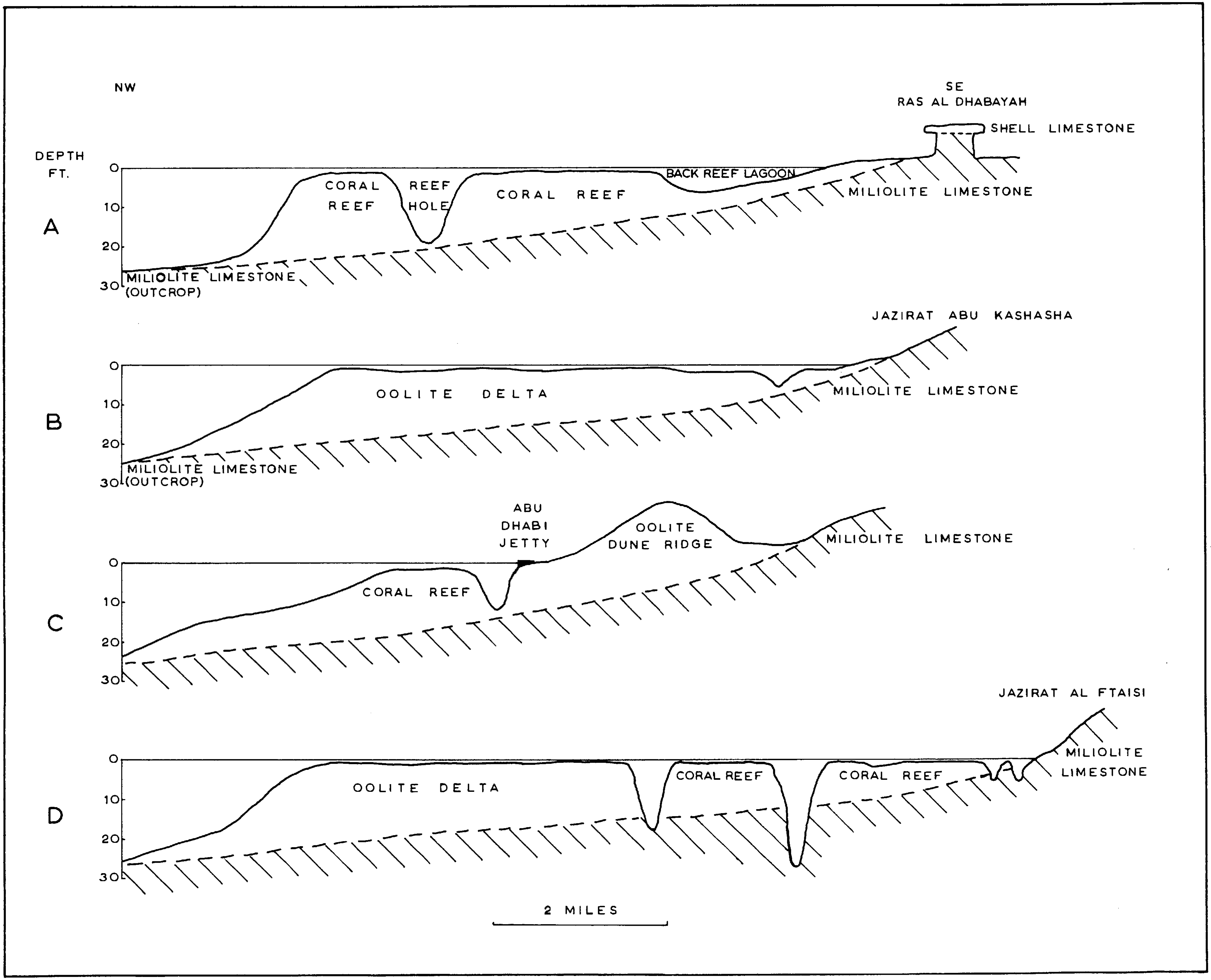
The Miliolite and Shell Limestones have been examined along 250 miles (400 Kms.) of the Trucial Coast from west of Mirfa to north-east of Sharja and it has been found that the developments so far described, affected the entire area. Succeeding changes are more complex and resulted finally in the formation of the very different near-shore features of the area today. As only the Abu Dhabi area has been extensively studied the following discussion is confined mainly to this area - to the development of the barrier islands and lagoons.

Relative sea level certainly fell again after deposition of the Shell Limestone. Extensive marine erosion took place, resulting in removal of nearly all of the Shell Limestone and much of the underlying Miliolite Limestone. The marine erosion again planed off much of the sediments back to the inner edge of the present coastal plain but in general the Miliolite Limestone was not breached and except for outlying buttes and mesas (islands at the time) the Miocene rocks were not eroded. The present day distribution of Miliolite and Shell Limestones results largely from this period of erosion. Later erosion has modified this in some ways. The nature of the area at this period is now in large masked by overlying sediments. Some indications may be gained by examining profiles across the area (Fig. 42). Miliolite Limestone occurs on land at Ras al Dhabayah and Jazirat abu Kashasha and at 25 ft. below sea-level seaward of the oolite delta and offshore coral reef. But it has not been found in the back lagoon or on the reef top or oolite delta. Profiles A and B indicate that the offshore reefs and oolite deltas were both developed on a Miliolite Limestone foundation which sloped gently seawards. Although Miliolite Limestone has not been directly observed on the shelf to the north-east, the offshore profiles (C and D, Fig. 42: and others not included) are all very similar. This relationship of reefs, oolite deltas and dune ridges, all developed on a gently seaward sloping foundation of Miliolite Limestone probably holds true between Ras al Kahf and Ras Ghanadha.

If the area can be envisaged without the present day accumulations of loose sediments and without the coral reefs and oolite deltas, then it can be seen that the erosion of the present day lagoons may be equated with this period. Water circulation would have been fairly open and all islands small in size.

The origin of lagoon barriers in areas of carbonate sedimentation is likely to be more complicated than the origin of barriers of clastic sediments. Barriers of the latter type have been discussed by Phleger and Ewing (1962) who considered the following essential for their formation: (a) an abundant supply of sand, (b) a gently sloping foundation, and (c) wave action on an exposed coast. Factors (b) and (c) have been shown to hold for the Trucial Coast but little is known of factor (a). Derived carbonate

Fig. 42 Diagrammatic Profiles across the oolite deltas and coral reefs between Abu Dhabi and Ras al Kahf.


NW
SE
RAS AL DHABAYAH
SHELL LIMESTONE
DEPTH FT.
0
10
20
30
A
CORAL REEF
REEF HOLE
CORAL REEF
BACK REEF LAGOON
MILIOLITE LIMESTONE
MILIOLITE LIMESTONE (OUTCROP)
JAZIRAT ABU KASHASHA
B
OOLITE DELTA
MILIOLITE LIMESTONE
MILIOLITE LIMESTONE (OUTCROP)
C
ABU DHABI JETTY
OOLITE DUNE RIDGE
MILIOLITE LIMESTONE
CORAL REEF
JAZIRAT AL FTAISI
D
OOLITE DELTA
CORAL REEF
CORAL REEF
MILIOLITE LIMESTONE
2 MILES

sands would fall into this category but active reefs and areas of oolite development could also give rise to barriers. The present day Trucial Coast barriers comprise islands, which are in part beach and in part dune accumulations, coral reefs and oolite deltas. The dune ridges are entirely of oolite sand and thus post-date oolite formation. The offshore reefs show a maximum of 20-25 feet (7-8 ms.) of vertical growth during the present sedimentary cycle. The oolite deltas could be developed on earlier coral reefs but this is unlikely as the present day, offshore coral reefs and oolite deltas are sharply defined entities. Thus in their lower parts the oolite deltas could be composed of skeletal sands, reef debris (unlikely) or oolite sands.

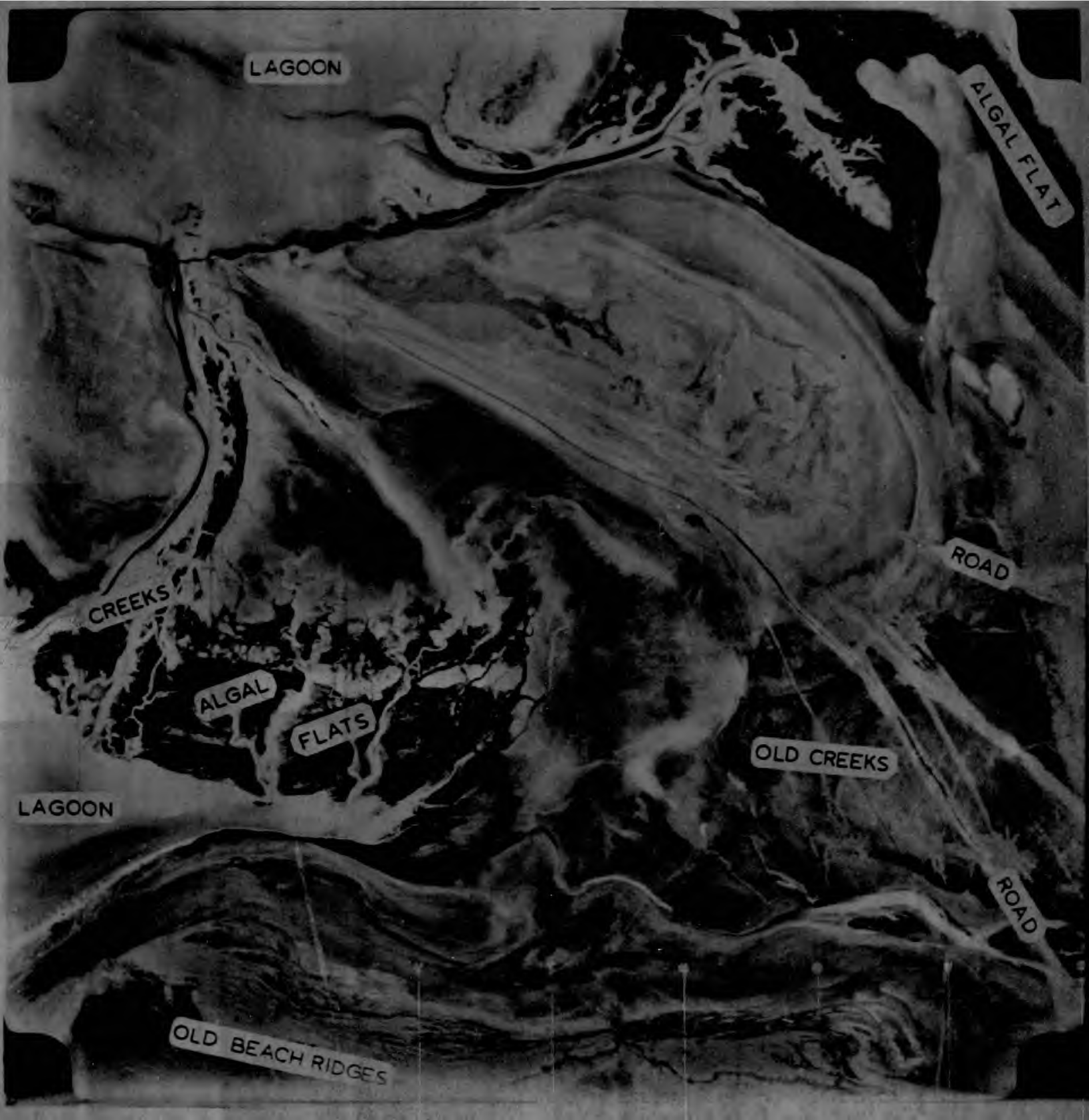
As the barrier developed, tidal movement of waters in and out of the lagoons became confined to certain areas and tidal channels were eroded. It was not until this length of coast became divided into individual lagoons by deposition of loose sediments in the lee of the larger islands that the tidal channels became deeply entrenched, as we see them today. The oolite deltas are intimately related to the tidal channels and as the latter are of fairly late development the deltas themselves may represent merely small lobes, added to an earlier, more linear barrier.

Before deposition of the later sediments began, the area was fairly open and wind and wave activity considerable. In the lee of islands, small sheltered areas existed and these, together with deeper water areas below active wave base would have been the sites of accumulation of the finer, muddier sediments. As sedimentation continued, the islands increased in size and extensive shoal and intertidal areas came into being, which were sheltered from wind and wave activity. The early phase of infilling should therefore show a general lack of the finer-grained, muddier sediments. At the inner edge of the mainland sabkha and against many of the early islands, the initial sediments are found to be skeletal sands or even gravels. Similar intertidal sands are found today along the southern shores of Khor al Bazm and along some of the inner lagoon coast south of Abu Dhabi causeway (Fig. 43). The sands are dominantly skeletal (molluscan) and in places gastropod gravels, resembling modern beach ridge accumulations, are to be found.

As accretion continued, the islands grew in size and the area of the lagoons diminished. The fetch to any inner shoreline became reduced and intertidal sediments became more muddy. In sheltered intertidal areas nearly pure aragonite muds accumulated, particularly in the developing creek and swamp environments. The later mid and inner lagoon marine sediments were muds, variously pelleted, with minor skeletal additions. The earlier lagoonal marine sediments probably had a higher skeletal fraction, the mud fraction becoming more important as the lagoon became more enclosed.

The development of the main tidal channels

Fig. 43 Sabkha development near Abu Dhabi causeway; see Fig. 3 for location. Modern coast backed by beach ridges in lower left; old beach ridges extend inland showing former coastline positions. Coarse beach ridge sands also accumulated around the early Miliolite island close to the road. Constricted areas have algal flat and creek facies development, with more muddy sediments. Old creeks can be seen extending back into the sabkha. The complex final infilling of inner lagoon areas is readily apparent. Scale 1/40,000.



LAGOON
ALGAL FLAT
ROAD
CREEKS
ALGAL
FLATS
OLD CREEKS
LAGOON
ROAD
OLD BEACH RIDGES

between Halat al Bahrani and Abu Dhabi island reached its maximum with the linking of the Dhabayah peninsula to the mainland. Tidal flow of waters in and out of the lagoon became concentrated in the channels and they gouged deeper into the underlying Miliolite Limestone. However, as the lagoon became progressively infilled the tidal exchange of waters decreased and in the channels erosion probably gave way to limited deposition. Future developments will all lead to further reduction in lagoon area and depth with concomittant infilling of the tidal channels. The channels are at present floored with coarse shell and coral lag deposits: later additions can be expected to be finer, and oolite sand, at present washed into the channels but largely transported seaward by ebb currents, will accumulate, in addition to shelly debris. Lateral coral growth will tend to decrease channel widths along much of their length.

Meanwhile, the mid and inner lagoon will become progressively infilled with aragonite muds and pellet-sands. In later phases pellet sands will be absent and molluscan contributions will decrease. The final lagoon phase will be a creek and swamp series of muddy sediments. To seaward of the lagoon barrier, the oolite deltas will become degraded features and the tidal channels will become broad, shallow and ill-defined.

These later developments can be reasonably inferred from study of the narrowing series of lagoons to the north-east. Towards Ras Ghanadha the lagoons progressively diminish in size, aragonite muds and pellet sands fill the lagoons, and tidal channels and oolite deltas become poorly differentiated. The final phase can be seen at Ras Ghanadha where creeks and swamps, with aragonite muds, lie immediately behind a dune ridge of oolite and skeletal sand, which parallels the open coast. The tidal channel has become degraded and is now a broad creek channel and the oolite delta has become poorly marked. In fact, the coast here has become one of active erosion. The dune ridge overlies creek muds, the latter being exposed as a small cliff at mid-water on the open coast beach. Aerial photographs show the dune ridge to truncate large creek channels.

Thus if relative sea level remains constant, the lagoons between Abu Dhabi and Ras Ghanadha can be expected to go through the stages of infilling suggested above. Once infilled, if relative sea level remains constant then erosion from seaward will start destroying the lagoonal sediments. Relative elevation of the area would result in erosion of the lagoonal sediments, unless elevation was very rapid when they could become covered with a continental series of sediments. Depression of the area could result in a similar cycle of lagoonal sedimentation.

The cycle of lagoonal sedimantation can be seen to show certain general features which are constant but in detail great variation is found. The earliest sediments have been shown to be largely skeletal in nature and later in the cycle, as the lagoons became progressively restricted, chemically precipitated aragonite muds became dominant. The final phase of infilling will almost everywhere be a creek and swamp phase of aragonite muds. Stromatolitic algal sediments will be most common during mid and later phases but the associated carbonate sediments may range from coarse skeletal sands to aragonite muds. Meanwhile, in the barrier area, coral reefs have actively grown throughout the sequence and development of oolite deltas has resulted in great spreads of oolite sands in barrier and immediately adjacent lagoon areas.

The lagoonal sediments in the Trucial Coast area are associated with diagenetic evaporite deposits, but under less extreme conditions of net evaporation, this sequence could well be developed without the evaporite association. For a lagoonal cycle of this nature the following requirements are necessary:-

(a) a gently sloping foundation on which a barrier and lagoon complex can be built.

(b) an onshore wind, to develop the lagoon barrier.

(c) climatic conditions under which fiarly rapid production of skeletal and chemically precipitated carbonate materials can take place.

Factor (b) need not be important if the barrier results from reef development. Factor (c) covers temperature and salinity conditions, for example, low net evaporation would probably be paralleled by a lack of an evaporite phase. Higher rainfall than occurs along the Trucial Coast could result in addition of land derived sediment to the lagoonal carbonate sediments. A similar lagoonal cycle to that of the Trucial Coast could probably develop even if sea-water salinities did not exceed 40-45‰. In such a cycle, diagenetic evaporites would probably be absent and coral reef growth in the barrier areas might be the dominent control of barrier development.

CHAPTER VI: EARLY DIAGENESIS OF LAGOONAL SEDIMENTS - THE SABKHA ENVIRONMENT

General: The present cycle of sedimentation has probably been in operation during the past 3-4000 years, since the post-glacial rise in sea-level brought the sea near to its present level. Prolonged accretion during this period has resulted in the development of extensive areas of uncemented sediments, which typically lie just above normal high water. These areas are usually salt encrusted and constitute the sabkhas Sabkhas extend seaward from the mainland outcrops of Tertiary rocks and form the greater part of many of the offshore islands

Marine sedimentation in any locality proceeds to the level of highest high water. In the most sheltered inner lagoon areas the normal tidal range is less than 2 feet (0.7 m.) and high water less than 1 foot (0.3 m.) above mean sea-level. On more exposed coasts the normal tidal range may be more than 4 feet (1.3 m.) and high water more than 2-3 feet (0.7-1 m.) above mean sea level. Tidal ranges are increased during periods of strong, onshore winds, particularly along exposed coasts. Thus sabkha surfaces derived from sedimentation along exposed coasts will be two feet or more higher than those derived from sedimentation along the most sheltered coasts. In addition, the sediments of the sheltered coasts are much finer grained than those deposited along more open coasts. A leveled profile across the sabkha near Abu Dhabi causeway showed a surface relief of less than two feet (0.7 m.), the height varying from 3-5 feet (1-1.6 m.) above mean sea level, except where it rose gently at its inner margins against the Tertiary rocks The lowest parts of the profile coincide with old creek and swamp facies, which formed an area of muddy sabkha; areas of sandy sabkha lay slightly higher

The intertidal zone of the sheltered coasts is gently shelving and passes without break into the sabkha surface. Along more exposed shores the intertidal zone is more steeply shelving and is commonly backed by one or more beach ridges, often composed of gastropod shells.

The sabkha is only occasionally flooded as it lies above the level of normal high water However, when strong onshore winds coincide with spring tides then many hundreds of square miles of sabkha are flooded. The infrequent rains are often accompanied by gale force, onshore winds and the resulting flood waters covering the sabkha are then in part rain water and in part sea water. Such conditions occurred in May 1963 and in January 1964. After flooding, great sheets of water lie on the sabkha surface for up to 1-2 weeks. Most of the water drains away seaward but some percolates down into the sabkha sediments and some is lost by evaporation. The wind may drive great sheets of water across the sabkha surface during flooding. Near-surface erosion then takes place as the protective salt crust becomes dissolved.

The water table within the sabkha sediments lies slightly above mean sea level and lies generally at depths of 1-4 feet (0.3-1.2 m.), depending in part on the height of the sabkha surface. It was found from the leveled profile that the water table rises slightly in the innermost sabkha areas. A very narrow zone adjacent to the shore shows a fluctuating water table, dependent on the tidal phase operative in the lagoon. Over most of the sabkha, however, the water table is fairly constant. Even after flooding, the level of the water table undergoes no appreciable change, a good indication that downward percolation of flood waters is not very extensive. Above the water table is a continuous capillary zone where all interstitial pore spaces are filled with water. Higher again is a discontinuous capillary zone where only thin films of water cover the sediment particles, and in the highest levels the sediments are often completely dry. The drawing of water into the interstitial pore spaces above the permanent water table proceeds to a higher level in the finer, muddy sabkha sediments than in the more open, sandy sediments.

Wind is an important agent of change, particularly in the areas of sandy sabkha. Wind deflation, however, can proceed only to the level of the water table, or more correctly, to some level above the ground water table determined by the upward capillary movement of pore waters within the sediment. As the ground water table lies slightly above mean sea level in both muddy and sandy sabkha sediments, and as the upward capillary movement of waters, which controls the base level of wind deflation, proceeds to higher levels in the muddy sediments, then the base level of wind deflation is obviously lower in sandy than in muddy sabkha sediments. This, coupled with the higher initial level of the sandy sediment surface readily accounts for the observed wind deflation of the sandy sabkha and relative lack of deflation in the more muddy sabkha areas. In fact, all areas of muddy sabkha have a windblown sand layer, a few inches thick, overlying the muddy inter-tidal sediments. Another factor leading to the deflation of the sandy sediments concerns the surface salt cemented crust, which acts as a protective capping. The salt crust of the higher sandy sabkha sediments is much thinner and more friable than the salt crust found in areas of lower, muddy sediments, and is a much less effective protection against the wind. Other factors may play a part in the determination of the height and character of the sabkha surface but the relative significance of these is not known. For example, post-depositional compaction will be greater in the muddy than in the sandy sediments, and the processes of interstitial precipitation or solution of minerals may physically raise or lower the sabkha surface in some places. It is fairly certain that the precipitation of minerals from the interstitial brines causes a vertical growth of the sabkha surface of 1-3 feet (0.3-1 m.).

Evidence of wind deflation is readily seen in areas of beach ridge formation. The seaward ridge may be 4-5 feet (1.2-1.6 m.) above the general sediment level but successive ridges become lower inland and the site of the original ridges becomes marked by a coarse lag of skeletal debris. However, deflation is severely limited except for the beach ridge accumulations, and it is doubtful whether more than 2 feet (0.7 m.) of the sabkha surface are ever removed by the wind. Evidence of the sheet removal of sand is seen in areas where gypsum crystals and anhydrite nodules, which originally grew within the sediments, now litter the surface, or on occasions stand on small pedestals. Windblown sand forms the uppermost few inches of the sabkha sediments in many areas but in general it does not accumulate until it reaches the hills of Quaternary and Tertiary rocks, at the inner edge of the sabkha plain.

Temperatures of the sabkha are high when compared with most earth surface sedimentary or early diagenetic environments. The algal mats are dark olive green in colour and the sabkha is brown, sandy sabkha being light brown and muddy sabkha typically darker brown. The dark surfaces will tend to absorb radiation. Seasonal air temperatures range from 16-47°C., averaging 28°C., and it is more than likely that the dark algal and sabkha surfaces may reach temperatures of 60°C., or even higher during the summer months. The upper surfaces suffer greatest temperature ranges, the seasonal variation probably being over 40°C., and diurnal variation over 20°C. A thin upper zone will suffer almost equivalent temperature ranges but slightly deeper, the diurnal temperature range will become less pronounced. However, seasonal temperature ranges in excess of 10°C. affect the sabkha sediments and ground waters to a depth greater than 4 feet (1.2 m.), as ground water temperatures have been recorded of 39°C. in August and 26°C. in February. The average annual temperature of groundwaters within the sabkha is 34°C., about 5°C. higher than the marine waters of the lagoons (Fig. 27).

The sediments of the sabkha have already been discussed, the general sequence being a series of upper intertidal zone sediments of stromatolitic algal facies, creek and swamp facies or shelly beach and beach ridge facies, overlying lower intertidal skeletal or pellet sands with greater or lesser amounts of aragonite mud. These in turn overlie lagoonal sediments which are usually aragonite muds and pellet sands, but which may be largely skeletal along exposed coasts. The lateral and vertical sequence may vary greatly in detail but the sabkha sediments comprise essentially mixtures of two end members. (a) a sandy sequence, occurring around the inner edge of the eastern mainland sabkha, around former islands, and forming most of the sabkhas to the west of the Dhabayah peninsula, and (b) a muddy sequence, typically a sheltered coast deposit, occurring almost exclusively in the eastern sabkha areas.

The sandy sabkha sediments are largely skeletal and typically contain 2-3000 ppm. Sr. In contrast the muddy sabkha sediments, which are dominantly the product of chemical precipitation, contain higher amounts of Sr (~ 7000 ppm.) Mineralogical analyses of inner lagoon sediments near Abu Dhabi have shown them to comprise more than 90% aragonite (Fig 35) and analyses of 18 samples from 14 algal flat cores in the same area have shown a similar mineralogy. Likewise, the sedimentary and organic structures of the sabkha sediments are those of the marine environments to seaward.

The only organic structures unique to the sabkha are the 'burrows' illustrated in Fig 46/A and Fig. 52, core 1. Figure 46/A shows an anhydrite bed with near vertical 'burrows', 1-2 mms. in diameter, infilled from above with windblown sand. The 'burrows' definitely post-date the formation of the anhydrite. Figure 52, core 1, shows similar 'burrows', infilled from above with windblown sand, developed in a fine-grained carbonate mud (now entirely dolomitised). The 'burrows' are developed in wet sabkha areas, are almost certainly organic in origin and may be made by some form of crustacean.

In studying the diagenetic developments of the sabkha environment an initial premise was made. It was considered that the diagenetic minerals of the sabkha were developed from the initial marine sediments, together with contributions from concentrated sea-waters, and that except for Co_2 the system is essentially closed. Evaporation rates from the upper sabkha surface are not known but compared with the rainfall (~ 5 cms./year) the rates are very high. Thus the sabkha ground waters can only be derived from inward moving lagoon waters. This assumption has been found to hold true for the island sabkhas and for most of the mainland sabkha. At the innermost edge of the mainland sabkha the ground waters are possibly of mixed origin, being partly derived from concentrated sea-waters and partly derived from ground waters within the Tertiary rocks.

Sabkha Mineralogy: The diagenetic minerals of the algal flat and sabkha environments are aragonite (and possibly calcite), gypsum, anhydrite, celestite, halite, dolomite, magnesite and huntite. These minerals occur in varying proportions together with the aragonite, low Mg calcite and high Mg calcite of the original marine sediments. In addition, occur windblown grains of a variety of minerals, the commonest of which are quartz and feldspars. The Tertiary rocks are a possible source of detrital dolomite, gypsum and the rare anhydrite grains. The mineral distribution along a composite profile across the mainland sabkha is shown in Fig. 44.

Aragonite

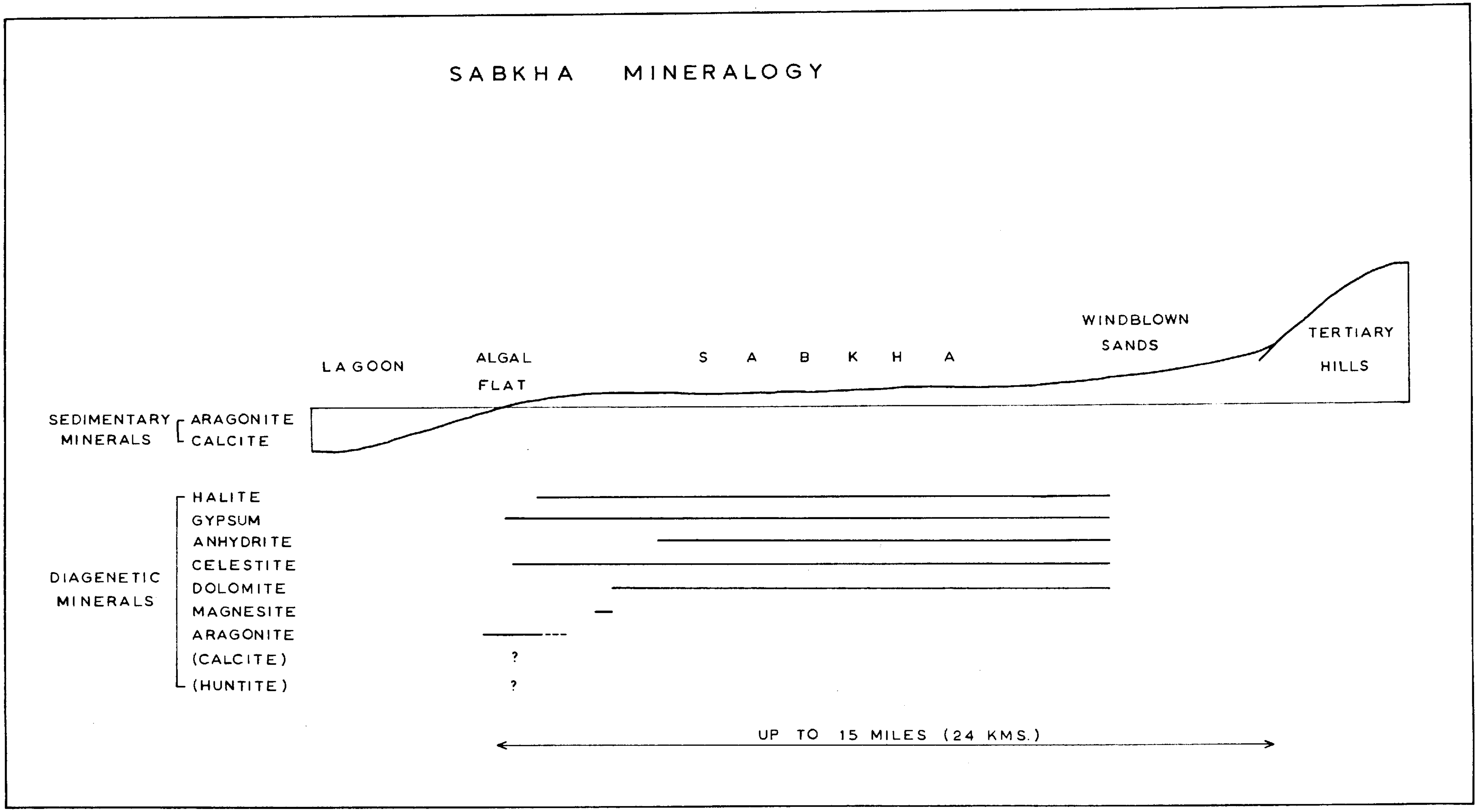
Aragonite is present in the algal flat and sabkha environments mainly as a primary carbonate mineral associated with the earlier phases of marine sedimentation. Aragonite of diagenetic origin is precipitated in interstitial positions within the algal flat sediments. A common development is the aragonitic cementation of the uppermost centimetre or so of the algal flat sediments. Similar limestones have been found within the sabkha sequence several miles from the present shoreline. The aragonite may be present either as needles or as near equant grains, in both instances less than 5μ in size.

As a result of the diagenetic reactions, aragonite diminishes in amount across the sabkha with increasing distance from the shore. In some places the replacement has gone to completion and aragonite is entirely absent. The initial aragonite loss is confined mainly to the aragonite needles, which are typically less than 1μ in length.

Calcite

Calcite of both low and high magnesian varieties is present as a primary marine carbonate mineral of organic origin. The Miliolite and Shell Limestones are both calcitic and are a source of low-Mg calcite, mainly

Fig. 44: Schematic distribution of diagenetic minerals across the mainland sabkha. Island sabkhas only bear the seaward part of the sequence


SABKHA MINERALOGY
LAGOON
ALGAL FLAT
S A B K H A
WINDBLOWN SANDS
TERTIARY HILLS
SEDIMENTARY MINERALS
ARAGONITE
CALCITE
HALITE
GYPSUM
ANHYDRITE
CELESTITE
DOLOMITE
MAGNESITE
ARAGONITE
(CALCITE)
(HUNTITE)
DIAGENETIC MINERALS
?
?
UP TO 15 MILES (24 KMS.)

to areas downwind of their outcrops. On Jazirat al Ftaisi fairly high amounts of low-Mg calcite of this origin are found in the uppermost few centimetres of the sabkha sediments. In the seaward parts of Abu Dhabi island, where low salinity ground waters occur, oolitic sediments are being cemented with low-Mg calcite. Low and high Mg calcite is present in most sabkha samples and it is possible that some may result from in situ precipitation. Mineralogical analyses of a few samples show slightly higher percentages of calcite than would be expected were it all derived from the calcite of the initial marine sediment. There is some evidence of the high Mg calcites being preferentially dissolved, but in general both low and high Mg calcites are present in most sabkha samples.

Gypsum

Gypsum occurs as a diagenetic mineral in the upper parts of most algal flats or other high intertidal areas, and is present throughout the sabkhas of the islands and mainland. In some places the gypsum forms more than 50% of the upper 3-4 feet (1 m.) of the sediment. Gypsum is associated with both sandy and muddy marine sabkha-sediments, and also with windblown sabkha-sediments. Several distinct stages of gypsum formation may be distinguished, and the associations and habits of the different types are fairly distinct.

The earliest gypsum occurs in the upper

algal flats close to mean high water mark, beneath the surface algal mat. The crystals first appear immediately beneath the uppermost algal mat but with increasing distance from mean high water mark the gypsum is found to occur at increasingly greater depths below the surface. The gypsum crystals range up to 1 cm. in diameter and are commonly clear and free from inclusions. Many, however, bear included carbonate particles, often arranged in zonal patterns. The crystals are discoidal in shape, flattened normal or nearly normal to the 'c' axis and are similar to those described by Masson (1955) and others from the Laguna Madre. Many of the gypsum crystals are deeply corroded by subsequent partial solution. Cores in the uppermost parts of the algal flat environment commonly show successive algal mats underlain by clear discoidal gypsum crystals, associated with greater or lesser amounts of carbonate sediments. Some algal flat areas have gypsum crystal 'mushes' extending 2-3 feet (up to 1 m.) below the surface. Similar crystal mushes occur throughout the width of the sabkha, sometimes associated with stromatolitic algal sediments. The crystals of the mushes may be variously disposed with marine or windblown sediment or they may occur almost pure. Commonly, almost pure gypsum layers alternate with gypsum free or gypsum poor layers, at least to depths of 4 feet (1.2 m.).

A later form of discoidal gypsum crystal is represented by the large gypsum sand crystals which occur within the upper sabkha sediments or litter the surface as deflation lag deposits in some places. These crystals have only been found in areas of sandy sediments and are locally common to the west and east of Mirfa and also in areas to the north-east of Abu Dhabi causeway. The gypsum sand crystals grow within the sediments in all orientations and reach at least 25 cms. in diameter. Included within them are a great diversity of carbonate and non-carbonate grains - whole gastropods commonly occur. Traces of the original sedimentary stratification can often be seen running through the sand crystals. The sand crystals are of a similar habit to those of the algal flats, being discoidal and flattened normal or near normal to the 'c' axis.

A third type of gypsum comprises flattened, bladed crystals, up to 10 cms. in length, found so far only growing in muddy sabkha sediments, at depths of 1-4 feet (0.3-1.2 m.). The crystals are complex intergrowth forms and are rarely free from fine-grained included carbonate, commonly arranged in a zonal manner. The crystals grow in all orientations within the sediment and often bear traces of algal laminae or other original sedimentary stratification (Fig. 45). Associated with the bladed gypsum crystals are uncommon rosettes, up to 5-6 cms. in diameter, which are intergrowth forms of the

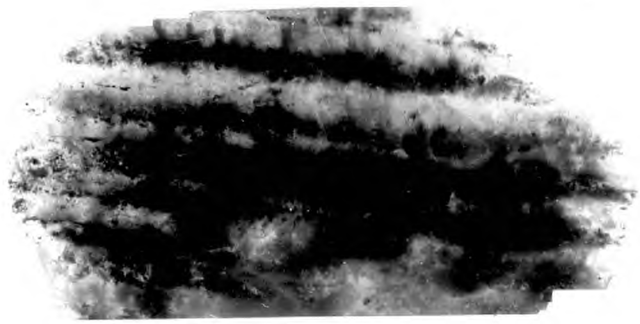


Fig. 45: Bladed gypsum crystal from 1 m. depth, showing algal laminae and original bedding traces. Natural size

basic discoidal crystals, together with occasional single discoidal crystals up to 5 cms. in diameter. Discoidal crystals of this size (> 1.5 cms.) are a sub-surface development, always later than the early phase of discoidal crystal formation, found in the algal flat and seaward sabkha areas.

A fourth type of gypsum comprises euhedral crystals, flattened in the 010 plane and similar to the classically described 'selenite' crystals. This crystal habit has only been found as a development after anhydrite and is crowded with anhydrite inclusions (Fig. 51/B). Such crystals have been found up to 6 cms. in length but have only been found in one locality, near the base of the Dhabayah peninsula.

The first formed gypsum crystals are thus the small discoidal forms of the upper intertidal zone. But similar, small discoidal crystals form throughout the diagenetic sequence in the uppermost parts of the sabkha. An example of an extremely late development is illustrated in Fig. 51/A which shows discoidal gypsum crystals with cores formed by small anhydrite nodules. The bladed gypsum crystals are later than the discoidal crystals of the algal flats but are coëval with later discoidal crystal development in the sabkha. The gypsum sand crystals are of fairly late development and those forms which post-date anhydrite are obviously very late in the diagenetic sequence.

Mechanical concentrations of gypsum crystals are common on the sabkha surface. Wind deflation removes the finer sediment particles, the coarser gypsum crystals becoming concentrated as a surface lag deposit. On the offshore islands, dune sands, downwind of sabkha areas, are commonly found to be rich in wind abraded gypsum crystals. In fact, some dune sands comprise more than 70% gypsum crystals.

Anhydrite

The occurrence of anhydrite as a diagenetic mineral within the sabkha environment is of particular interest as this is the first recorded natural occurrence of Recent anhydrite. The mineral was first found in a sabkha core collected by the author in January 1962. Subsequent field studies have shown it to be of widespread occurrence and to be developed on a geologically significant scale.

Anhydrite has so far been found to occur only in the sediments of the mainland sabkha. The fairly extensive sabkhas of Jazirat al Ftaisi and Jazirat Abu Kashasha have failed to reveal its presence. As with the gypsum, the initial occurrence is within the uppermost centimetre or so of the sediment. The anhydrite has been found to always have gypsum lying between it and the intertidal zone. The distance inland from the intertidal zone, to where anhydrite is first encountered, cannot be defined in absolute units as this distance may vary from 2-300 yards (2-300 m.) to more than a mile.

The earliest anhydrite, in the uppermost few centimetres of the sabkha sediments, has been found as small nodules 1-2 mms. in diameter, often associated with a gypsum mush. The nodules comprise a felted mass of anhydrite laths, measuring 25x5x3μ. Thin sections of core samples have revealed single crystals and multiple aggregates of anhydrite laths to be developing within the sediments. Figure 50/B shows the early development of an anhydrite nodule, the aggregate being only 400μ across.

Further anhydrite development takes a variety of forms and the mineral has been found to extend to depths exceeding 4 feet (1.2 m.) in mid and inner parts of the mainland sabkha. In general, the anhhdrite occurs within the sediments above the permanent water table but a few occurrences have been found extending below. In places the upper 3-4 foot (1 m.) of the sabkha is at least 50% anhydrite but local variation is great and the detailed sequence in ore pit is often not paralleled in a closely situated pit.

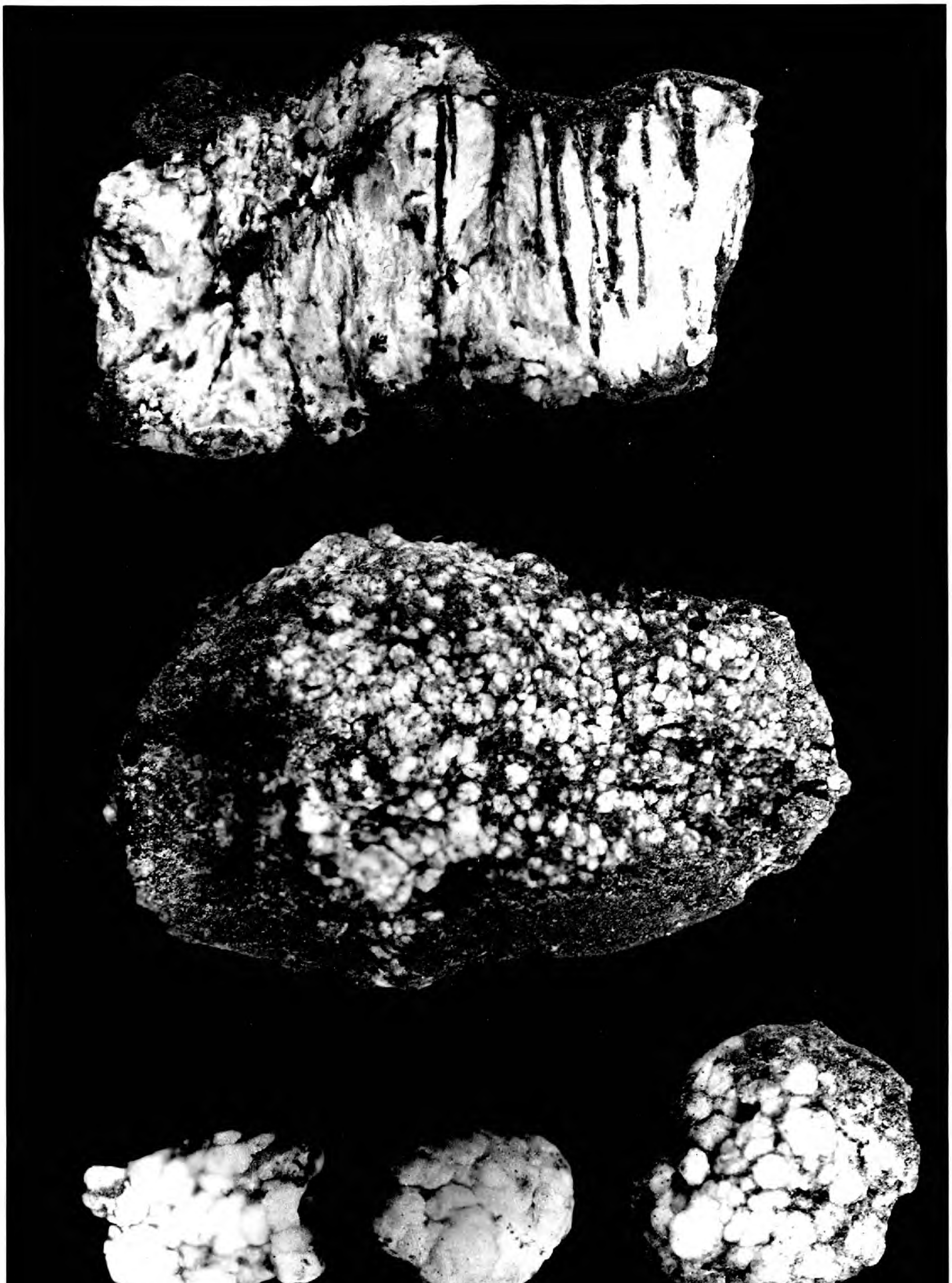
Some of the major forms taken by anhydrite are illustrated in Figures 46, 52, 53. Thin beds and nodules are the major forms. The beds may be up to 15 cms. or so thick, but average about 7-8 cms. They may have single or complex marginal relationships with the sediments

Fig. 46 - A (top) Bed of anhydrite with vertical texture, enclosed above and below by windblown sand. Vertical burrows infilled with sand.

B (middle) Nodular anhydrite; part of a discontinuous layer of anhydrite composed of small nodules up to 5 mms. in diameter. Small nodules surrounded by windblown sand.

C (lower) Nodular anhydrite showing large nodules composed of smaller nodules with thin films of sand between. Individual crystals can be seen glistening on the surface of the central nodule. Nodule on left has been planed flat to compare with the well known 'chicken-mesh' structure exhibited in Fig. 47.

Scale - natural size.



in which they occur, and may be horizontal or variously cross cutting. Within a bed the texture is usually massive, although commonly a vertical texture is developed (Fig. 46/A). The majority of beds are free from foreign grains and appear brilliantly white; on occasions, carbonate and other material occurs mixed with the anhydrite and the bed is then brownish and may have quite diffuse boundaries with the adjacent sediments.

Nodules vary greatly in development. They may range from less than 1 mm. to more than 20 cms. in diameter; shape ranges from spherical to strongly flattened and they may occur as individuals or as complex, intergrown masses of hundreds of nodules. The latter development commonly gives rise to beds of anhydrite nodules, an example is shown in Fig. 46/B. Beds of flattened, plate-like nodules are fairly common but this is a growth form; beds of nodules sometimes show a slight tendency towards flattening but this is uncommon - this flattening may well be mechanically induced. Nodules may be packed tightly together with little or no interstitial material (Fig. 46/C), or they may be separated by greater or lesser amounts of foreign material (Fig. 46/B). The former occurrence is common in many ancient back reef and lagoonal deposits, being termed net or chicken-mesh structure; a Mesozoic example from the Persian Gulf is shown in Fig. 47.

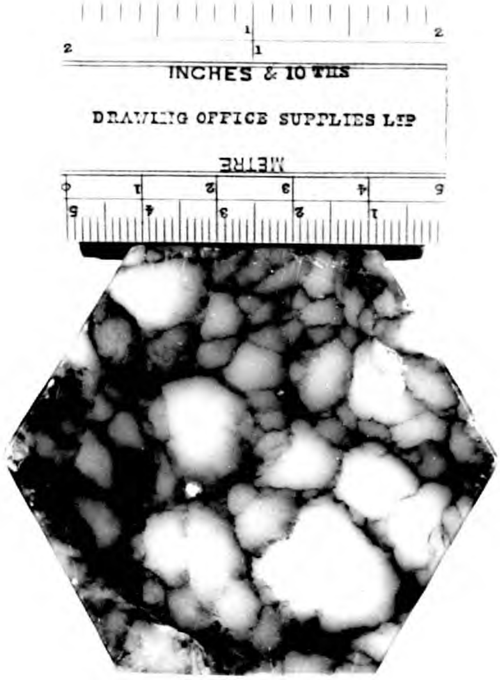
Nodules are usually free from foreign grains but on occasions examples are found replete with carbonate and non-carbonate sabkha sediment grains (Fig. 53/A). During growth the nodules are evidently able to shoulder the enclosing sediments aside.

Optical and X ray examination has shown the anhydrite to be normal and to compare closely with published data for ancient anhydrites. The crystals are lath-like, colourless, transparent, free from inclusions, are elongated parallel to the 'c' axis and flattened parallel to 100. Refractive indices are n_{α} = 1.570, n_{γ} = 1.576; n_{β} was not determined owing to the orientation of the crystals. The laths may be extremely thin (2-3μ) and show a good pseudo-cubic cleavage (Fig. 48). Crystal edges are usually straight but on occasions ragged edges are found, the crystals having undergone some secondary solution. Single laths range from less than 2-3μ in length up to a maximum of 3-4 mms. Normal length to width ratios vary from about 5:1 to 10:1. Thickness is dependent on crystal size but even the largest crystal laths do not exceed 50-60μ in thickness.

Orientated thin sections of nodules and core samples have shown crystal textures in most instances to be random. Fig. 49/A shows random texture in a developing anhydrite nodule. Fig. 49/B, however, shows a fairly strong parallel orientation of the larger anhydrite laths parallel to the bedding, although much of the finer background anhydrite has a fairly random texture; this sample is from

Fig. 47 (top): Polished core section of nodular anhydrite showing chicken mesh structure (compare with Fig. 46/C): of Mesozoic age. Supplied by M. Morton, Abu Dhabi Petroleum Company.

Fig. 48 (lower): Recent anhydrite crystals - a water mount, to show pseudo-cubic cleavage and crystal form. Crystals are typically lath-like and up to 3 mm. in length (normally 50-300, long). Scale x 200.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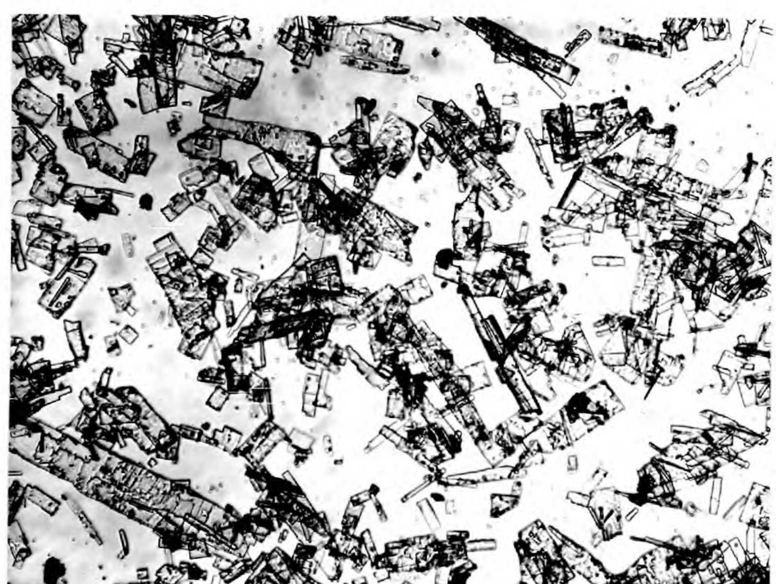


the upper few centimetres of core 2 of Fig. 52. An example of an exceptionally coarse-grained anhydrite nodule is shown in Fig. 50/A where the large laths have a random orientation. Many anhydrite nodules and beds show arcuate, sinuous patches of laths with sub-parallel orientation; an unusual example is shown in Fig. 51/B. Near foreign grains orientated laths may either show a 'wrap-around' relationship or show no disturbance at all.

Sampling over a wide area of the mainland sabkha has shown that there is little overall sequence in the arrangement of anhydrite beds and nodules, one form predominating in one area, yet being absent in the next. Similarly, detailed microscopic study has revealed no constant relationship between the crystal form and size and the major field structures. The smallest crystals have been found forming almost pure beds of anhydrite, and the coarsest crystals have been found in anhydrite nodules, but the wide range of intermediate sizes occurs indiscriminately in both beds and nodules.

An unusual occurrence of anhydrite is illustrated in Figs. 51/A and 51/B. The former shows early anhydrite nodules forming cores to small, discoidal gypsum crystals; the latter shows an S-shaped area of anhydrite laths within a large gypsum crystal. Both forms of gypsum obviously post-date the anhydrite.

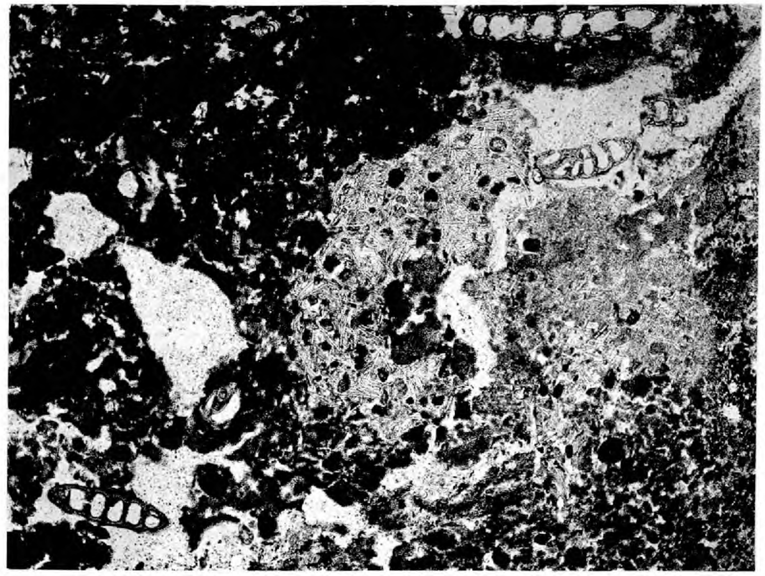
Sabkha cores are illustrated in figures 52

Fig. 49/A (upper): Anhydrite nodule (thin section), showing laths with random orientation growing in carbonate sediment with Peneroplid foraminifera.

Scale x 20.

Fig. 49/B (lower): Thin section of orientated core sample, bedding horizontal; showing near horizontal development of many laths but a general background of randomly orientated anhydrite laths: upper levels of core 2, Fig. 52.

Scale x 20.



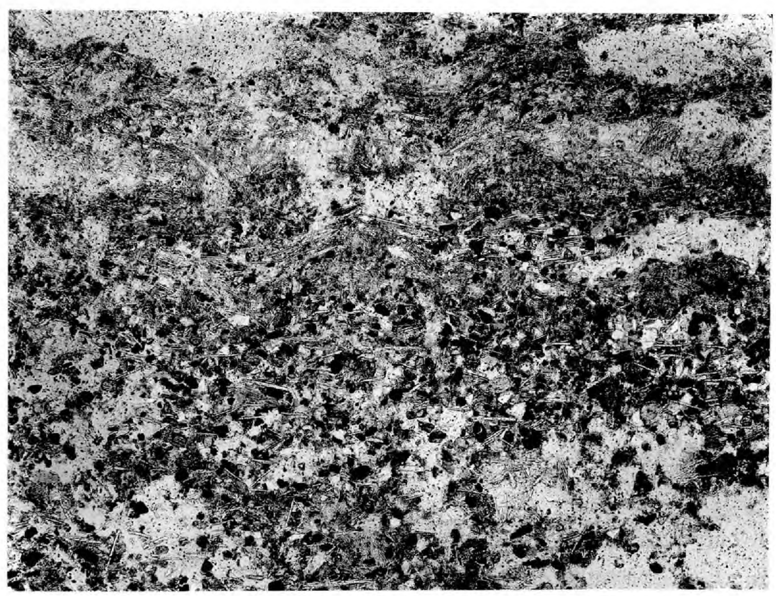
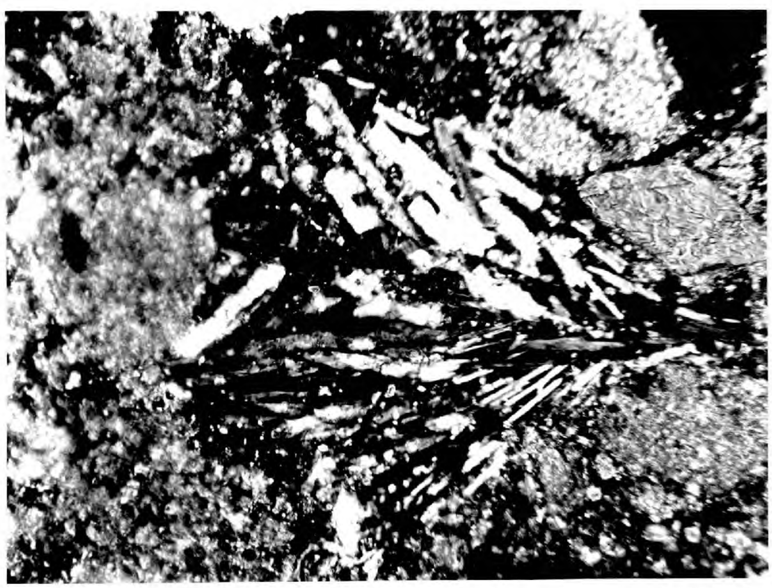


Fig. 50/A (upper): Thin section of exceptionally coarse grained anhydrite, crystals up to $1\frac{1}{2}$-2 mms. in length. To show random orientation within a nodule.

Scale x 25.

Fig. 50/B (lower): Thin section showing incoming of anhydrite as very small sheaths of crystals. Further growth leads to nodule development. This stage is preceded by the development of single crystals and then small clusters of laths appear. Matrix of dolomite and detrital grains (rhomb in east is not dolomite).

Scale x 200 : crossed nicols.



and 53 and show typical forms of anhydrite occurrence, together with associated sedimentary and diagenetic features.

Wind deflation and erosion of the sabkha surface during flooding expose anhydrite at the surface. The nodules are fairly resistant and commonly litter some surfaces of erosion. However, large quantities of anhydrite crystals are distributed across the sabkha surface, particularly by the wind, and are a component of the near-surface detrital layers.

Celestite

Celestite is a fairly abundant mineral throughout the sabkha sediments of the islands and mainland. It first occurs in the uppermost algal flats associated with gypsum, and the intimate association of gypsum and celestite is found throughout the sabkha environment. Where the initial carbonate sediments have been dolomitised celestite is particularly abundant. It occurs only rarely in close association with anhydrite; in general it is found in the sediment surrounding the anhydrite nodules or beds, rather than within them. In places the celestite locally forms up to 2-3% of the sediment, but the normal abundance is less than 1%.

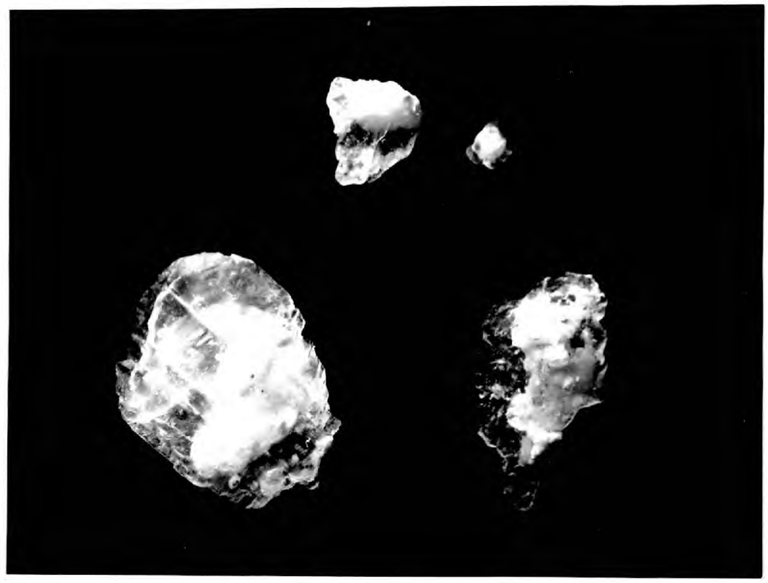
The celestite occurs in a dispersed manner as single and complex aggregates of crystals, varying in length from less than 2-3μ to more than 700μ. The crystals are free from inclusions and usually show well formed crystal

Fig. 51/A (upper): Gypsum crystals with anhydrite nodules as cores. Discoidal gypsum crystals, showing good cleavage, occurring in upper few centimetres of sediments in inner sabkha position.

Scale x 3.5.

Fig. 51/B (lower): Thin section of gypsum crystal, cleavage running NW-SE. Sinuous mass of anhydrite crystals included within the gypsum crystal. This is a very late stage gypsum crystal, formed after anhydrite.

Scale x 60.



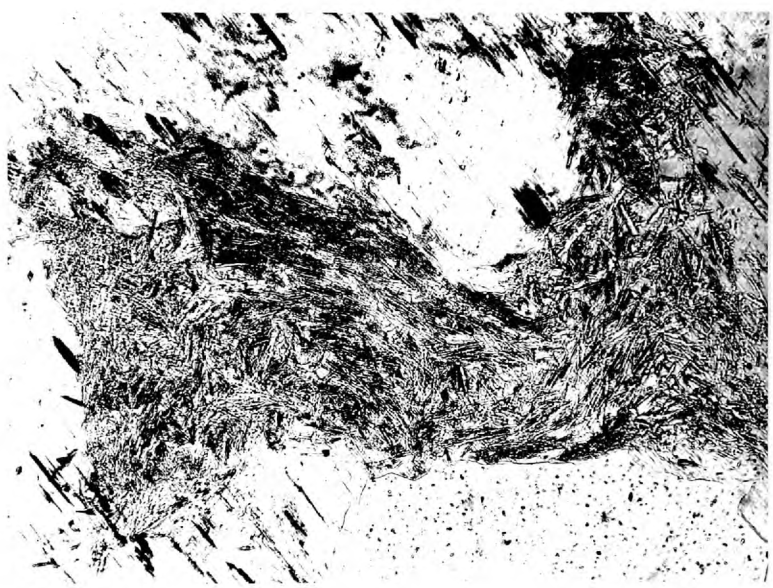


Fig. 52. Sabkha cores, from muddy sabkha sequence, across the mainland sabkha near Abu Dhabi.

(A = Anhydrite; G = Gypsum; D = Dolomite)

Core 1: Diffuse anhydrite, growing between gypsum layers in windblown sand, at top. Top of marine sediments entirely dolomitised, and burrowed by unknown sabkha organism. Marine sediments below are partially dolomitised. Celestite abundant in upper dolomite.

Core 2: Anhydrite as small nodules and thin bands (2 mms.) in upper parts. Algal banded sediments below with gypsum and celestite.

Core 3: Anhydrite nodules at top with pure dolomite below. Dolomite throughout but amount falls off downwards. Good algal banding.

Core 4: Nodular anhydrite at top, some gypsum below. Dolomite with celestite and large bladed gypsum crystals developed below.

Core 5: Gypsum crystal mush at top and bottom with anhydrite nodules between; nodules in a matrix of anhydrite and dolomite. Celestite abundant in lower gypsum.

Core 6: Gypsum crystal mush at top with nodular anhydrite below: anhydrite cored, discoidal gypsum crystals near contact (Fig.51/A). Mixed anhydrite, dolomite and celestite below. Near base are algal bands with large gypsum crystals, some dolomite and some celestite.

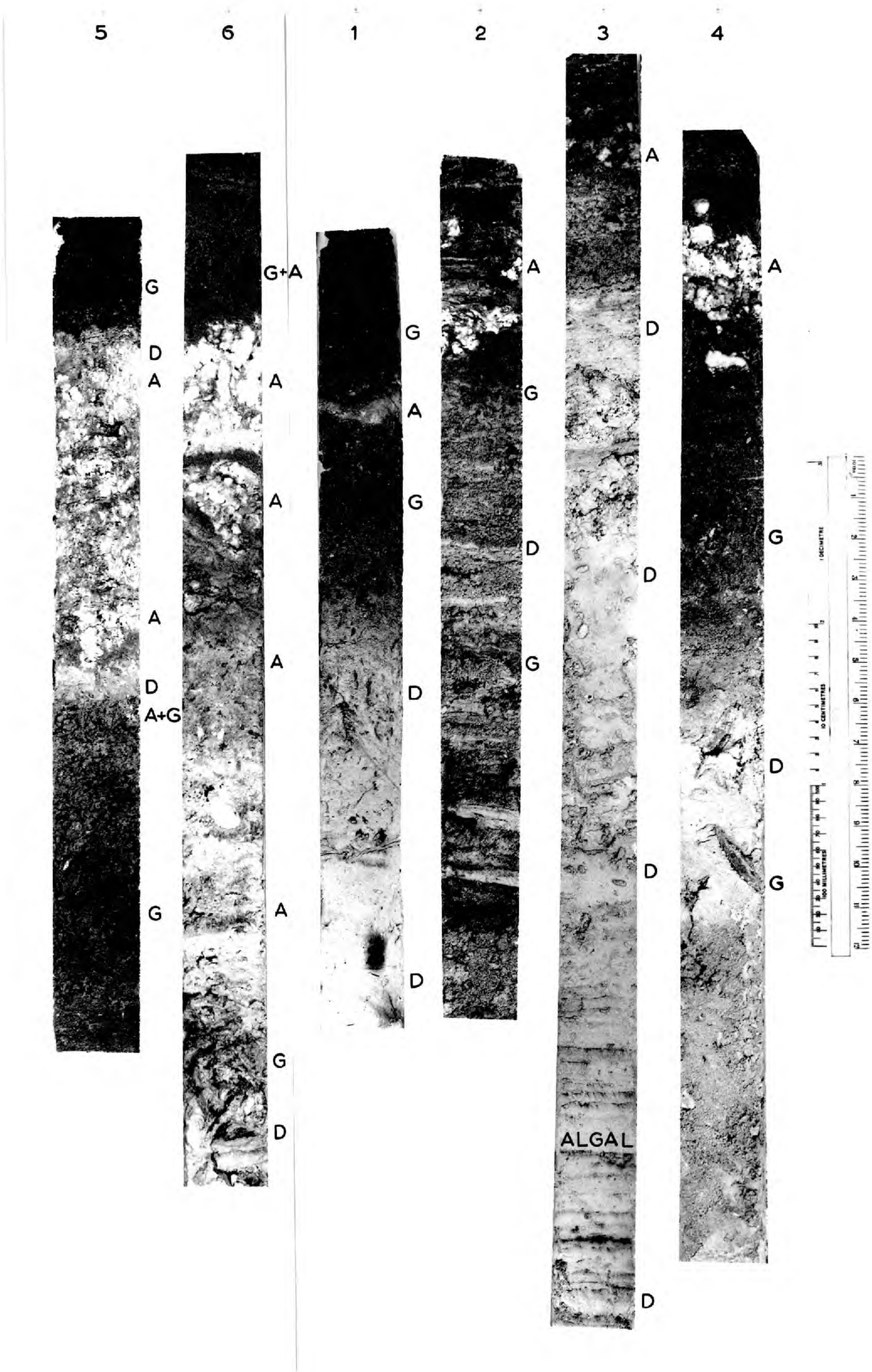

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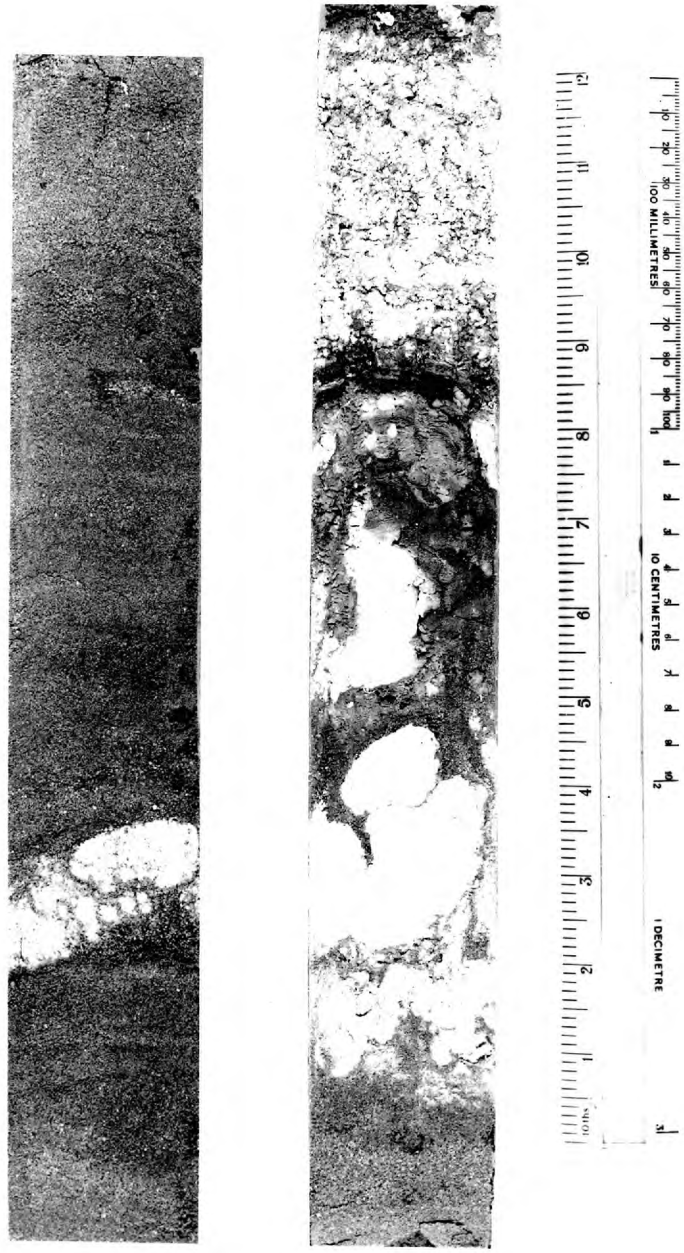
Fig. 53: Sabkha cores from sandy sabkha sequence.

Core 1: Cross cutting band of nodular anhydrite, some nodules free from foreign grains, others with many included grains.

Core 2: Small, diffuse anhydrite nodules in matrix of anhydrite and dolomite at top. Below are larger nodules in similar matrix; some celestite associated with dolomite.

1 2

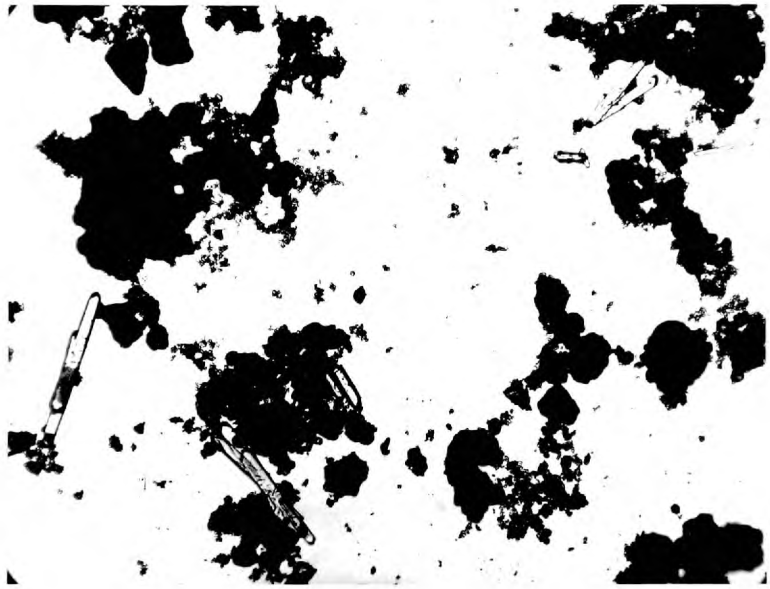


faces. Sometimes partial solution imparts a somewhat ragged look to crystal faces and edges. The crystal habit is the same in all associations, being elongate parallel to the 'b' axis and showing a combination of prism and macrodome; rarely, very small brachydomes are developed. Single crystals and aggregates of crystals with the same habit are shown in Fig. 54. Single crystals may range from a few microns to 700μ in length and may be relatively long and slender or short and stout. Aggregates of varying degrees of complexity are the most common and all stages between single crystals and complex radiating sheaves are to be found. A wide range of these forms is often to be found together in a single sample.

The occurrence and distribution of celestite has been briefly treated by Evans and Shearman (1964). They considered there to be three states of aggregation of the celestite crystals, with particular distributions, (a) large, single, euhedral crystals occurring in the near surface sabkha sediments, (b) simple radiating aggregates of a few individuals, found associated with late stage gypsum sand crystals, and (c) radiating sheaves, which occurred together with early gypsum beneath the upper algal mats. The three forms of celestite were treated as being distinct types with distinct associations within the diagenetic sequence.

Detailed study by the present author has shown that the distribution of habits suggested by Evans and

Fig. 54: Celestite crystals from 8 cms. depth, Jazirat at Ftaisi. Showing slender and stout single crystals, simple and complex aggregates. Large crystal aggregate in lower photo is 300μ in length. Small equant crystals in upper photo are rhombs of dolomite.



Shearman applies in only a very few instances. Combinations of all types commonly occur together, one particular type dominating in one locality or at one level and another type dominating in another locality or level. The single, euhedral crystals have been found associated with all the minerals of the sabkha, at various depths down to 4 feet (1.2 m.), and range from less than 10μ to more than 500μ in length. The radiating sheaves have been found to be the end stage of a multiple development.

No attempt can yet be made to account for the overall character of the celestite crystals in any locality but factors such as porosity, availability of Sr-bearing solutions and ease and rapidity of crystallisation are probably involved.

Halite

Halite occurs as a near surface mineral throughout the sabkha environment, in the uppermost dry or nearly dry zone of the sediments. Surface pools, on evaporation, leave a halite crust up to 7-8 cms. in thickness. Upward capillary movement of pore waters results in the precipitation of halite near the surface. The halite crust firmly cements the upper sediment surface, particularly in areas of muddy sabkha. The crust is typically rucked and contorted and commonly broken into rough polygons. Crystals are rare and the mineral is an ephemeral development, being dissolved whenever the sabkha is flooded.

Dolomite

Dolomite has been found throughout the sabkhas of the islands and mainland, and occurs down to depths in excess of 4 feet (1.2 m.). So far, it has been found associated mainly with areas of originally fine-grained, muddy sediments; occurrences in sandy sabkha sediments are rare. The mineral typically occurs as a white or natural coloured mud composed of small rhombs and equant grains, less than 2-3μ in diameter. After some practice it is possible to identify the dolomite optically but in general X-ray diffraction methods must be used.

Dolomite has not been found in sediments from the marine areas. The mineral first appears in the upper few centimetres of the sediments at the algal flat-sabkha junction. In this position it lies in the capillary zone above the permanent water table. Further inland dolomite is found at increasing depths, and in inner sabkha positions occurs at least 2 feet (0.7 m.) below the water table. The dolomite is obviously a replacement of the original carbonate sediment, and as the percentage of dolomite increases so the percentage of aragonite needles decreases. At some levels the replacement has gone to completion, even small skeletal grains being dolomitised. In general, the amount of dolomite in any position tends to fall off downwards, but it is still present at depths of 4 feet (1.2 m) in inner sabkha areas (the depth of the deepest pits dug during this investigation).

X-ray diffraction studies have shown the dolomite to differ somewhat from standard dolomite. The crystal structure is slightly disordered and an excess of Ca is found in almost all instances, the composition ranging from $Ca_{50}Mg_{50}$ to $Ca_{55}Mg_{45}$. The latter features will be discussed in more detail later.

Magnesite

Magnesite has been identified by X-ray diffraction in samples from the upper 10 cms. of the sabkha sediments near the centre of Jazirat al Ftaisi. The mineral has so far not been found in the mainland sabkha. The magnesite is fine grained ($<2\mu$) and occurs dispersed within the upper, windblown part of the sediments, together with gypsum and celestite and traces of dolomite.

The magnesite has an expanded unit cell and in this feature is comparable with other Recent magnesite occurrences described by Alderman et al. (1961) and Graf et al. (1961). The significance of this feature will be discussed in a later section.

Huntite - $Mg_3Ca(CO_3)_4$

Huntite has been found in the upper 30-40 cms. of the sabkha sediments over a fairly wide area near the base of the Dhabayah peninsula. The huntite occurs as irregular blebs up to 7-8 mms. in diameter or as flattened wisps and layers, up to 2-3 mms. in thickness. It forms about 5-10% of a band, about 20-30 cms. thick, lying 10-15 cms. below a surface gypsum lag and salt crust. The mineral is associated with a sandy mud of very fine wind-blown grains of quartz and carbonates, together with abundant, corroded crystals of anhydrite, celestite and gypsum.

The huntite is white in colour and inside the blebs or layers the only associated mineral is halite. The crystals are less than 1μ in size, a feature common to all occurrences of huntite so far described. The mineral was identified by X-ray diffraction.

This evaporitic occurrence of huntite is unique and will be considered in more detail in a later discussion.

<u>Ground Waters</u>

Determinations of chlorinity/salinity were made on ground-waters in the field and on samples brought back to the laboratory. The locations of all but the mainland sabkha groundwaters is shown in fig. 29. Four samples were taken across the mainland sabkha between Abu Dhabi causeway and the inner sabkha margin, 1-1½ miles (2 kms.) from the Tertiary hills. In taking a sample a pit was dug to just below the water table and the water allowed to settle. Samples were taken, and those for laboratory analysis were filtered and stored in standard glass sample bottles. Water pH and temperature were recorded, as well as depth of the water table below the surface. At all water sample stations a sediment core was taken and a field examination of the local sediment character and mineralogy was made.

As the sabkha is an area of high net evaporation, and as the sabkha ground waters are derived from the sea-waters of the lagoons, the concentration of the ground waters should increase away from the shoreline. In addition, the concentration should normally increase upwards from the ground water table, through the capillary zone, as evaporation takes place from the upper sabkha surface. This inward and upward concentration of waters reflects the normal cycle of pore water movement within the sediments. Flooding of the sediment surface by rain or sea water will create a 'sandwich', with more concentrated pore waters lying between the less concentrated waters of the ground water table and of the surface flood waters. On these occasions, downward percolation of the surface flood waters will dilute the near-surface, most concentrated pore waters and a pore water of mixed origin will result. In the algal flat zone, where flooding of the surface is not too infrequent, the surface algal mat, even though cracked and breached, would seem to act as a shield and to preclude excessive mixing of surface lagoon waters and the pore waters below. There is abundant evidence, both in algal flat and sabkha areas, that surface flood waters do cause flushing and dilution of pore waters in the uppermost few centimetres, but that their effect does not normally extend much deeper. The flushing and dilution of pore fluids may be expected to proceed to greater depths in areas of sandy sabkha than in areas of muddy sabkha sediments. In muddy sabkha areas there is evidence of flushing and dilution in the near-surface, windblown, sandy layer, but not below: the very fine grained, muddy sediments would seem to act as an impermeable barrier to the downward percolation of waters.

The chemical composition of the parent brine is known from the analyses of the lagoon waters. Diagenetic changes involving precipitation from the concentrated brines or reaction between brines and original sediments, should be reflected in the brine chemistry. Thus, at the inner edge of the algal flats, where gypsum has been found to be developed in fair quantity, the interstitial brines should show a relative loss of Ca^{++} and SO_4^{--}. In areas where dolomitisation is taking place, the brines should show a relative loss of Mg^{++}. However, all the diagenetic reactions first take place in the upper levels of the sediment, above the permanent ground water table. Thus to follow their course in detail, samples of pore fluids from the capillary zone are required, in addition to samples from the ground water table. Even so, ground water table samples should indicate the general picture, in a 'broad-brush' manner, and in this reconnaissance survey of the sabkha diagenetic environment, only ground water table samples have been studied. In one particular, these samples will be less likely to give local anomalous results. Samples from upper levels will, on occasion, represent a mixture of surface flood waters and capillary pore waters, whereas ground water table samples should be free from this uncertainty. Only in the innermost mainland sabkha areas, where waters from the Tertiary rocks mix with sabkha derived waters, should the ground waters be anomalous.

One other source of anomaly of ground water concentration and composition derives from the nature of the sabkha sediments. The sabkha sediments in any area are by no means uniform in grain size, and therefore differ in porosity and permeability. In areas of muddy sabkha sediments, occasional old beach accumulations of more sandy sediment occur, and in sandy sabkha areas the sediments of ancient beach ridge accumulations and inter-ridge accumulations differ in grain size. Bars of sandy sediments in muddy sabkha areas will provide paths of easy and rapid passage for pore waters, whereas lateral pore-water movement through the muddy sediments may be expected to be much slower. Similar differences in ease of lateral water movement will exist in all sabkha areas and may complicate the simple relationship of increasing concentration with distance from the shore. For example, an old beach ridge, inclined at an angle to the present shoreline, could transmit ground waters of relatively low concentration inland of areas where, owing to slower lateral passage, the ground waters were more concentrated. After flooding, the process could well work in reverse, concentrated sabkha ground waters moving seawards through these 'channels'. Such a mechanism could well account for anomalously high ground water concentrations in seaward areas, following sabkha flooding.

Ground water samples were collected in April 1963. From rainfall data and local information, it seems that the mainland sabkha plain had not been flooded for 3-4 years. For the sabkha plain to be extensively flooded twice within one year (May 1963 and January 1964) seems to be quite exceptional. The samples collected and analysed in this study, therefore, represent waters which have undergone little mixing with surface flood waters and may represent the peak concentration conditions.

In all localities, no water seepage occurred from levels above the ground water table. Perched water tables might well exist after flooding of the sabkha, particularly at the marine mud/windblown sand junction in areas of muddy sabkha.

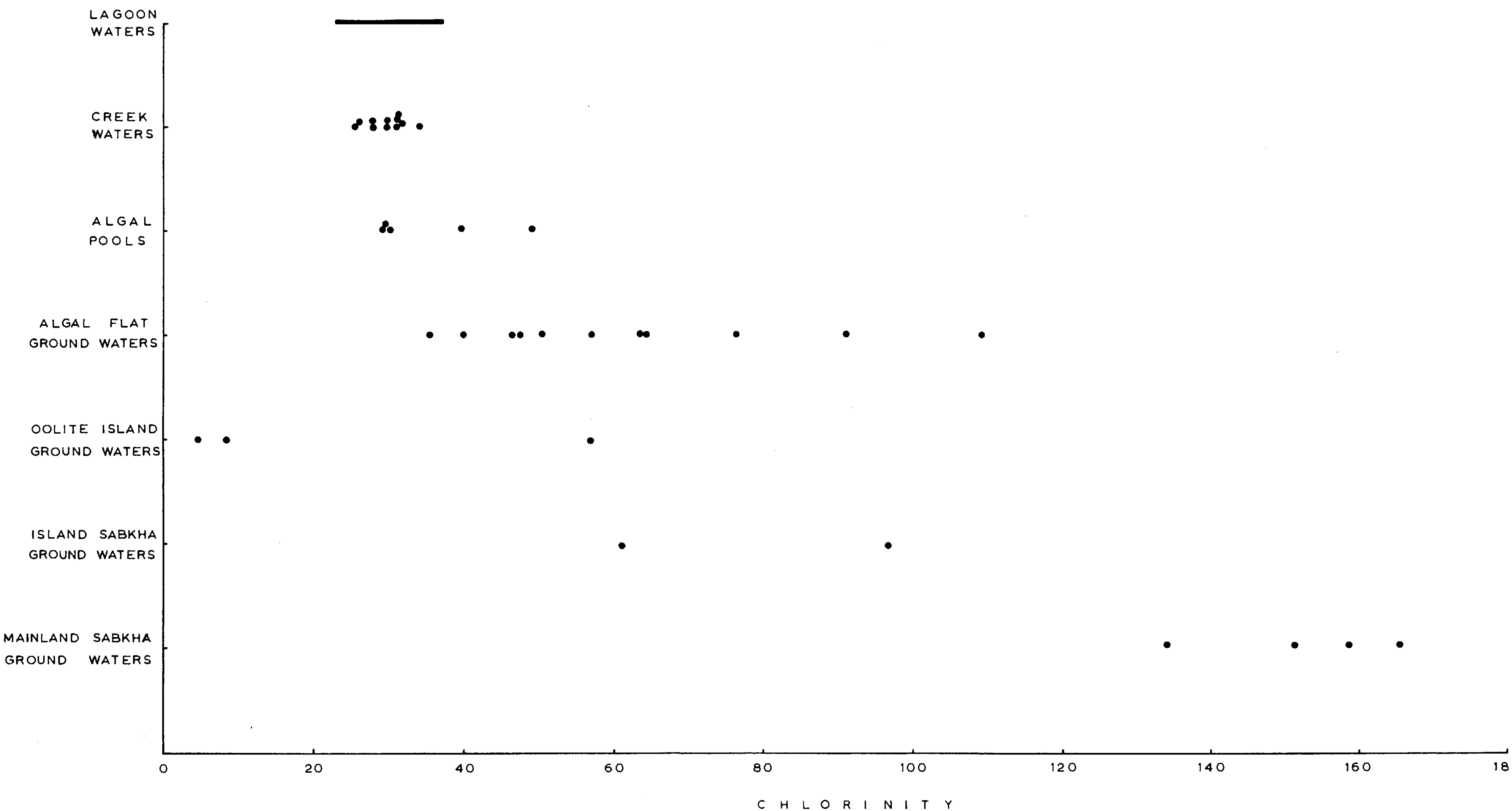
Ground waters consistently showed lower pH values than the sea waters. Algal flat and island sabkha ground waters typically showed values of 7.0-7.5, decreasing with distance from mean water mark. In the mainland sabkha values decreased still farther, ranging from 6.0-6.8, averaging about 6.3.

Salinometer and laboratory titrations gave the chlorinity/salinity distribution indicated in Fig. 55. The surface lagoon, creek and algal pool water values are included for comparison.

In the discussion of ground water concentrations, chlorinity will be used, as this is a measurable parameter, which below values of about 210‰ (when NaCl precipitates) gives a reasonable measure of the concentration. As ions are progressively lost from the brines, so the empirical relationship between chlorinity and salinity becomes more and more unacceptable. In discussion of the analyses the term 'concentration' will be used. A standard sea water (S. S.W.) with salinity 35‰ has a chlorinity of 19.37‰, and this obviously represents a concentration of x1 (pure water would be x0). Similarly a concentration of x5 represents a chlorinity of 96.85‰.

Ground-water chlorinities within algal flat sediments range from 35‰ to more than 110‰. Ground waters from seaward islands in areas of oolite accumulation vary greatly. On Halat al Bahrani a chlorinity of 56‰ was found but wells near the Sheik's Palace at Abu Dhabi showed values of 4.4‰ in 1939 (Iraq Petroleum Company) and 7.6‰ in 1960 (Sir William Halcrow & Partners). The

Fig. 55: Chlorinity/Salinity of surface and ground waters.


LAGOON WATERS
CREEK WATERS
ALGAL POOLS
ALGAL FLAT GROUND WATERS
OOLITE ISLAND GROUND WATERS
ISLAND SABKHA GROUND WATERS
MAINLAND SABKHA GROUND WATERS
0
20
40
60
80
100
120
140
160
180
CHLORINITY

Halat al Bahrani value is what might reasonably be expected but the Abu Dhabi values are near brackish water values. There would seem to be a local accumulation of rain water in the base of the oolite dunes. Local aquifers of this type are widely recorded inland in the desert areas. The difference in the ground water concentration on the two islands is exhibited in the cementation of the Recent oolite limestones in both areas. On Halat al Bahrani the carbonate deposited from the waters of high concentration is aragonite, whereas on Abu Dhabi island calcite is the interstitial cementing carbonate deposited from the low concentration waters. In both places, the oolite grains are still aragonite.

Ground water chlorinities within the sabkha range from 60‰ to more than 166‰. Only the mainland sabkha has so far been found to have water chlorinities greater than 110-120 ‰.

Density determinations, included in Fig. 56, showed only the most concentrated waters to depart widely from previously published data on the density of concentrated sea water. This divergence is reasonably accounted for, as published data are for concentrated sea waters from which various mineral phases were directly precipitated, whereas the sabkha waters take part in complex reactions with the sediments and thus their chemical composition will differ markedly from that of waters of equivalent

Fig. 56: Field data and laboratory analytical data of ground water samples

Ground-water type	Algal Flat			Island Sabkha	Mainland Sabkha			
Lab. Sample No.	19	20	22	21	15	16	17	18
Water Table depth (cms.)	30	15	18	46	81	76	76	61
Field T.^{o}C	27.5	-	26.7	-	28.4	31.4	30.5	31.2
Field pH	7.10	7.45	-	7.55	6.20	6.10	-	-
Dilution Factor	5.181	3.463	3.506	5.507	7.904	9.014	9.499	9.978
Density at 22^{o}C (Gms/cc)	1.1245	1.0865	1.0847	1.1277	1.1834	1.1922	1.2032	1.2113
Concentration (x St.sea-water)	4.695	3.300	3.295	4.971	6.943	7.829	8.214	8.565
Chlorinity Gms/Kgm	90.92	63.92	63.82	96.31	134.47	151.64	159.08	165.90
SO_4 Gms/Kgm	11.683	8.531	7.658	8.567	2.506	1.978	0.436	0.391
Ca Gms/Kgm	0.927	1.123	0.922	1.216	2.482	2.559	8.995	10.433
Mg Gms/Kgm	6.492	4.337	4.066	5.996	5.532	6.808	9.508	15.530
Sr Gms/Kgm	0.0223	0.0204	0.0179	0.0325	0.0451	0.0568	0.2147	0.2445
K Gms/Kgm	1.635	1.177	1.241	1.803	2.736	3.839	3.132	4.152
Na Gms/Kgm	50.28	35.48	35.92	53.30	75.08	81.36	75.11	65.10
Cl Meqs/Kgm	2564.5	1802.9	1780.0	2716.4	3792.8	4277.1	4487.1	4679.4
Alk. Meqs/Kgm	2.55	2.35	1.83	1.28	1.24	2.03	0.57	0.94
SO_4 Meqs/Kgm	243.19	177.59	159.40	178.32	52.18	41.17	9.08	8.14
Ca Meqs/Kgm	48.50	56.02	45.99	60.66	123.84	127.70	448.87	520.60
Mg Meqs/Kgm	533.9	356.7	334.4	493.1	454.9	559.8	781.9	1277.2
Sr Meqs/Kgm	0.508	0.466	0.408	0.741	1.028	1.296	4.899	5.579
K Meqs/Kgm	41.82	30.10	31.73	46.12	69.95	98.19	78.94	106.18
Na Meqs/Kgm	2187.2	1543.2	1562.5	2318.6	3266.1	3538.9	3267.2	2831.7

concentration but which have not been involved in brine/ sediment reactions.

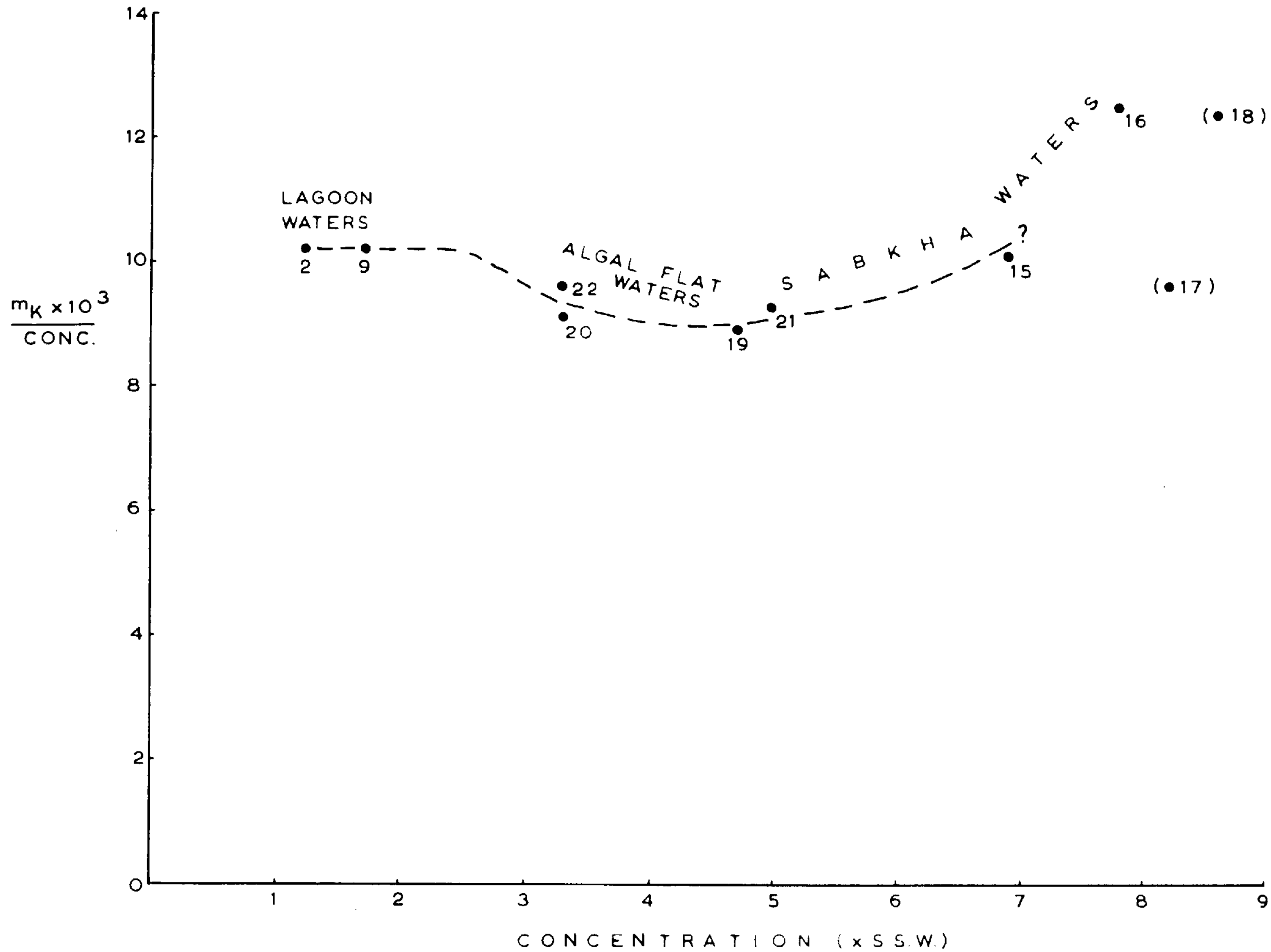
The ground water samples were all diluted to a chlorinity close to 20‰ and were analysed for calcium, magnesium, strontium, potassium, sodium, chlorine, sulphate and titration alkalinity, using the methods described in Chapter IV. The data from the samples form a straightforward sequence, except for the two samples from the inner parts of the mainland sabkha (Nos. 17 and 18).

The chlorine and sodium values give m_{Na}/m_{Cl} ratios close to 0.86 (the ratio in sea water) except for the two innermost mainland sabkha samples, which give ratios of 0.73 (No. 17) and 0.61 (No. 18). No analyses of ground waters from the Tertiary rocks are available from close by, but several analyses near Tarif have shown m_{Na}/m_{Cl} ratios of 0.84-1.02. If the waters from the Tertiary rocks near Abu Dhabi are similar, then admixture of this component cannot be called upon to explain the relatively low amounts of Na. If the low ratios are a reflection of a loss of Na and Cl as halite, then the concentrations of these waters must once have been x11.7 (No. 17: to be compared with a concentration of x8.2 from chlorinity) and x18 (No. 18: to be compared with a concentration of x8.6 from chlorinity). But the present chlorinity is too low for halite precipitation. The only reasonable suggestion would seem to be that halite has been precipitated and the brines diluted after the precipitation. The concentrations used in the discussion of the analyses are those derived from the chlorinity; these are minimum figures, correct values could be much higher.

Potassium analyses (Fig. 57) of ground waters of the algal flats and island sabkhas, show a relative deficiency averaging 10% . No K bearing minerals have been identified and it can only be concluded that the algal mats, through which much of the sabkha ground water must pass, are responsible. Marine algae are reported to concentrate and fix K within their tissues (Rankama and Sahama, 1950). Samples 15 and 17 show values close to normal, yet samples 16 and 18 are about 20% too high. Within the sabkha sediments are abundant algal intercalations. The organic tissues of the algal layers, on rotting, may be expected to release some of the K concentrated within them, and this could well explain the erratic K values. At an early stage in the investigations it was hoped that K would prove useful as a measure of concentration in brines from which halite had precipitated, but obviously it has little value in this respect.

In the Laguna Ojo de Liebre area, a rather similar environment to the sabkha, W.T. Holser (personal communication) has found early diagenetic gypsum altered to fine-grained polyhalite. The interstitial brines

Fig. 57: Potassium analyses of ground water samples.

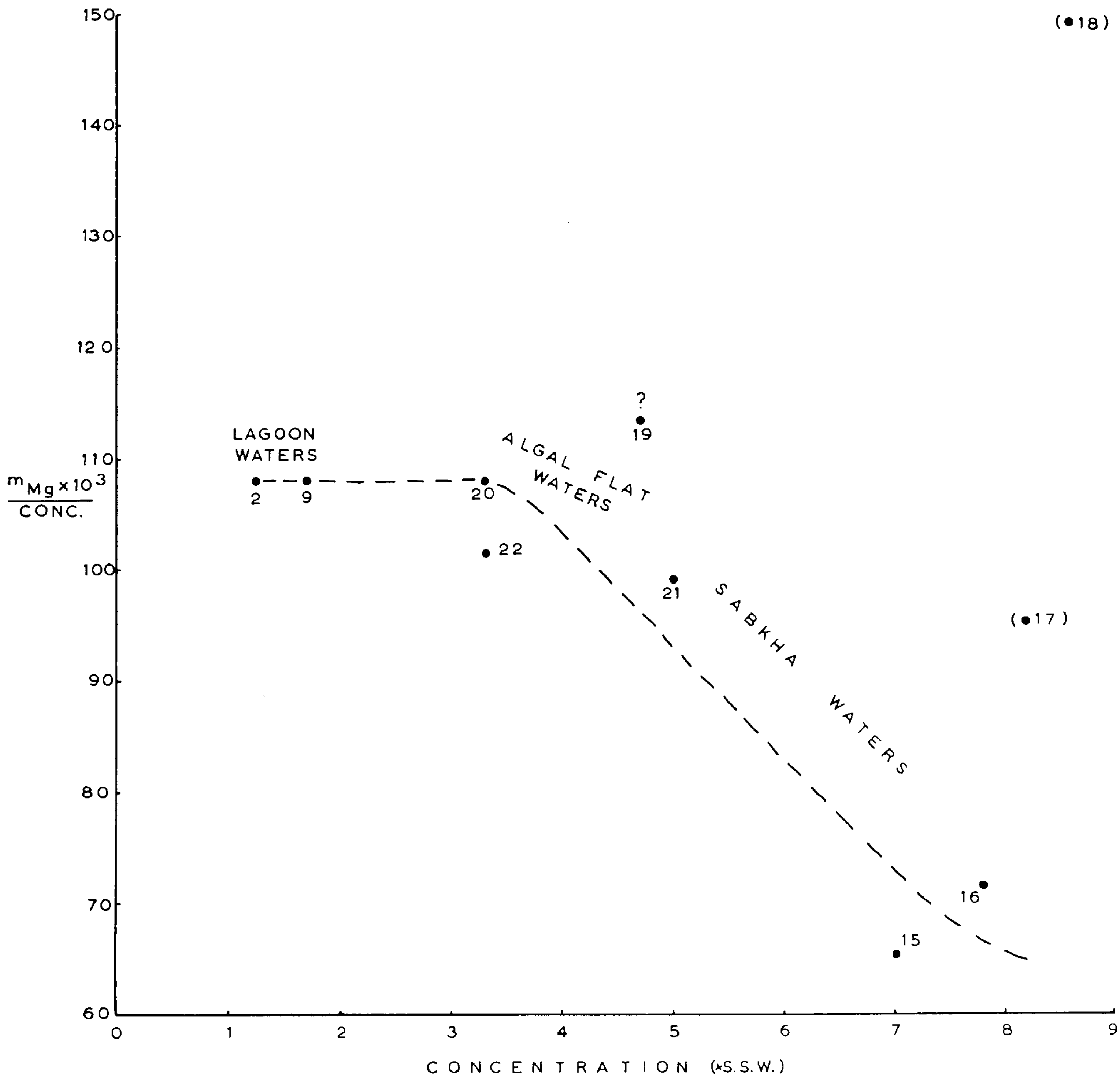


14
12
10
8
6
4
2
0
$\frac{m_K \times 10^3}{CONC.}$
LAGOON WATERS
ALGAL FLAT WATERS
S A B K H A W A T E R S
?
2
9
22
20
19
21
15
16
(18)
(17)
1
2
3
4
5
6
7
8
9
CONCENTRATION (x S.S.W.)

reach concentrations of x60 and are of the potash-magnesia facies. A brine of equivalent concentration from the Trucial Coast sabkha would have a very different chemistry, because of primary sediment differences. In theLaguna Ojo de Liebre area the sediments are sands and clays of terrigenous origin, with only minor amounts of carbonates; sediment/brine reactions are, therefore, rather limited, and under such conditions highly concentrated waters may evolve, still rich in Mg and SO_4 ions. Equivalent Trucial Coast brines would have negligible amounts of SO_4 and low Mg, because of widespread sediment/brine reactions such as dolomitisation. Thus in the Trucial Coast environment, polyhalite is not to be expected.

Magnesium analyses (Fig. 58) indicate clearly the progressive loss of Mg with dolomitisation. Sample 20 shows no loss of Mg and X-ray diffraction indicates no dolomite in associated core samples. Sample 22 has traces of dolomite developed, and the island sabkha sample (No. 21) has both dolomite and magnesite in the associated core. Sample 19 has no dolomite associated with it but is slightly anomalous, being about 4% too high in Mg; the reason for this is not known. Sample 22 is from a high algal flat position and could represent a mixed sample; this would account for the relatively large loss of Mg yet low concentration. The mainland sabkha samples 15 and 16 show a relative loss of 35-40% of the available Mg. Cores

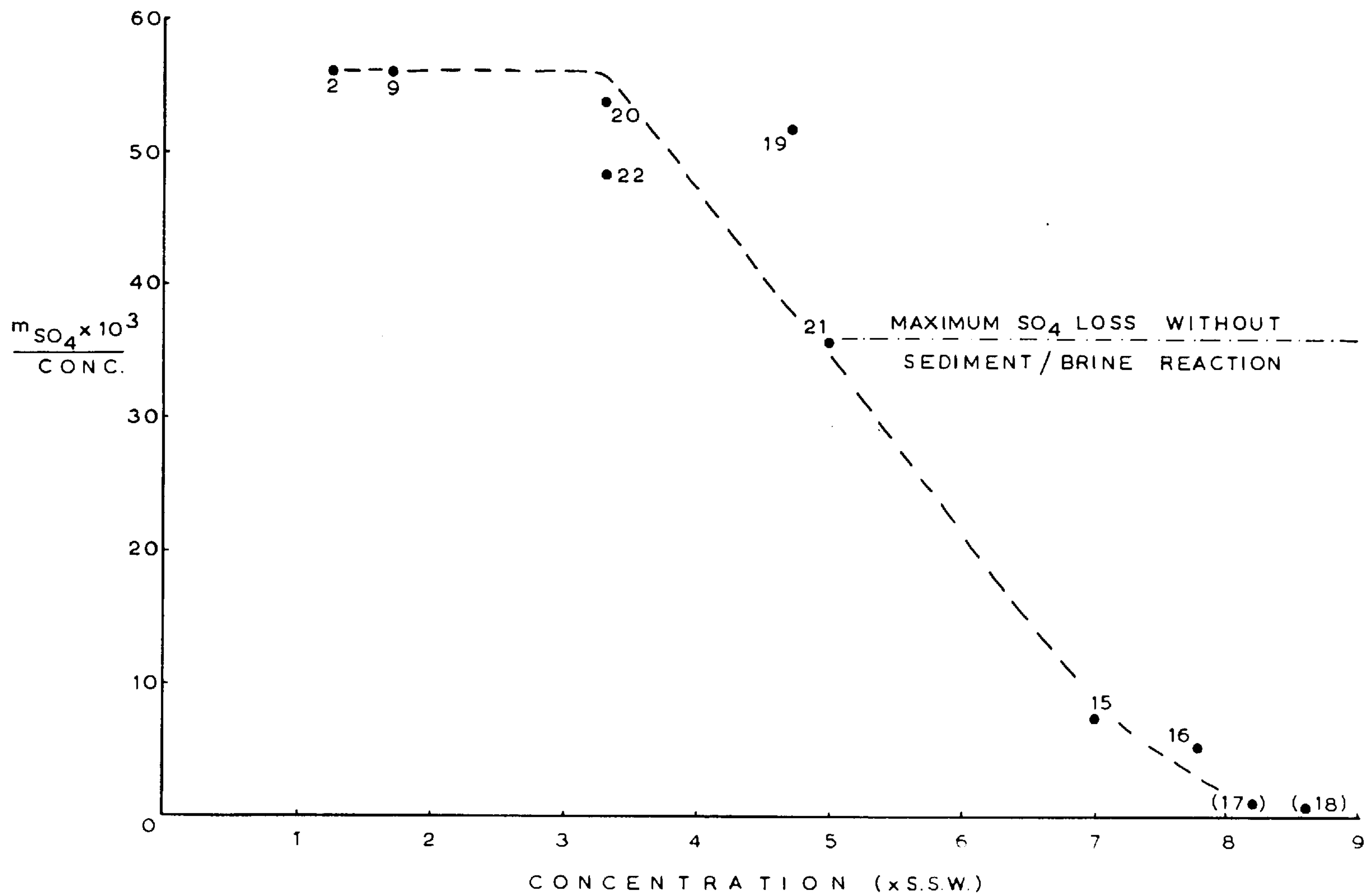
Fig. 58: Magnesium analyses of ground water samples.


150
140
130
120
110
100
90
80
70
60
m Mg x 10 3 / CONC.
0
1
2
3
4
5
6
7
8
9
CONCENTRATION (×S.S.W.)
LAGOON WATERS
ALGAL FLAT WATERS
SABKHA WATERS
2
9
20
22
?
19
21
15
16
(17)
(18)

from these positions show abundant dolomite to be developed, and cores to seaward show dolomitisation to have taken place across the whole width of the sabkha to the upper algal flats. Samples 17 and 18 show relatively high values of Mg but if concentrations based on the m_{Na}/m_{Cl} ratios are used (No. 17, x 11.7 : No. 18, x 18) then values of 66.9 (No. 17) and 71.1 (No. 18) are found. These values would seem more realistic, as otherwise the high Mg values would seem to necessitate postulating the solution of some dolomite, which is unlikely. Addition of waters from the Tertiary rocks is precluded as these have m_{Mg}/conc ratios close to those of samples 15 and 16.

The formation of gypsum, anhydrite and celestite, within the sabkha sediments, is paralleled by a relative loss of SO_4 from the brines (Fig. 59). The algal flat ground waters all show a slight loss of SO_4 which can be equated with the early phases of gypsum precipitation. The island sabkha sample (No. 21) has fair amounts of associated gypsum and shows a loss of about 35% of the available SO_4. This loss corresponds, fortuitously, to the maximum SO_4 loss (as $CaSO_4$, $CaSO_4 2H_2O$) which can take place from a concentrated sea water, without sediment/brine reactions. Further loss of SO_4 is only possible if Ca is released by some reaction (such as dolomitisation) between the carbonate sediments and the interstitial brines. Samples from the mainland sabkha all show

Fig. 59: Sulphate analyses of ground water samples.



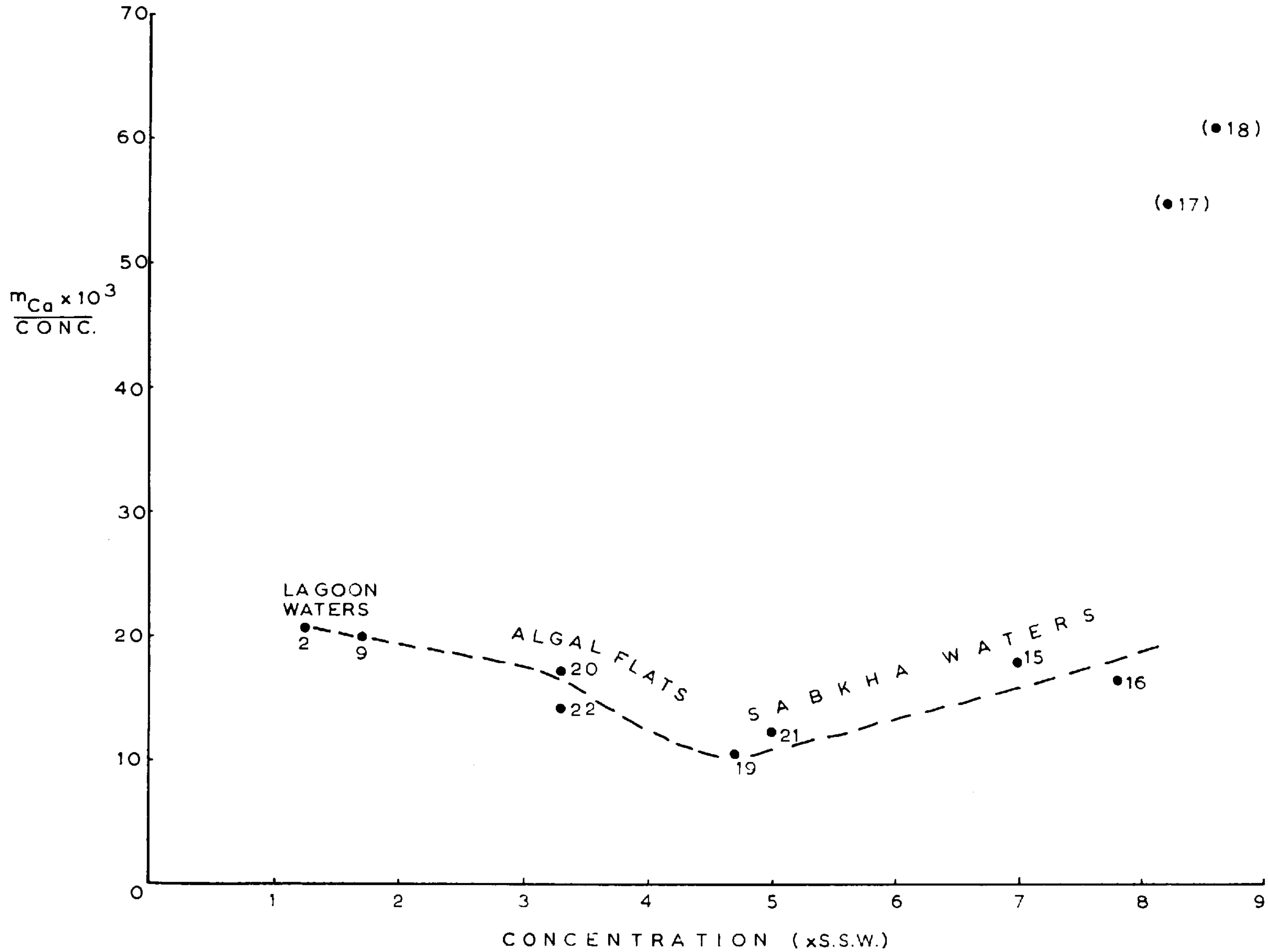
60
50
40
30
20
10
0
1
2
3
4
5
6
7
8
9
m SO4 x 10^3 / CONC.
CONCENTRATION (x S.S.W.)
2
9
20
22
19
21
15
16
(17)
(18)
MAXIMUM SO4 LOSS WITHOUT
SEDIMENT / BRINE REACTION

extremely low SO_4 values and indicate massive production of Ca to be taking place by sediment/brine reactions.

A relative loss of SO_4 from pore fluids within unconsolidated sediments is commonly reported, the loss usually increasing with depth. The loss occurs where there is often no evidence of the production of sulphate minerals. The loss has been attributed to bacterial activity, the SO_4 being reduced to H_2S, which is usually discernible by smell. In only one locality within the sabkha was there any evidence of the foetid smell of H_2S, and there is generally little evidence of the subsurface environment being strongly reducing. It is therefore considered that relative loss of SO_4 from the interstitial brines can be directly equated with the volume of sulphate minerals developed.

Analyses of lagoon waters have shown a progressive loss of Ca to take place across the lagoon complex. This is largely brought about by the precipitation of aragonite, and evidence has been given of further formation of minor amounts of aragonite in the algal flat environment. However, the loss of Ca is relatively small (10-15%) compared with that associated with the formation of gypsum. This is shown in Fig. 60 by a steepening in the curve between samples 20-19. Gypsum and anhydrite are actively precipitated at least to the concentrations of samples 15 and 16, as is shown by the curve of SO_4 loss.

Fig. 60: Calcium analyses of ground water samples.


70
60
50
40
30
20
10
0
$\frac{m_{Ca} \times 10^3}{CONC.}$
1
2
3
4
5
6
7
8
9
CONCENTRATION (xS.S.W.)
LAGOON WATERS
ALGAL FLATS
SABKHA WATERS
2
9
20
22
19
21
15
16
(17)
(18)

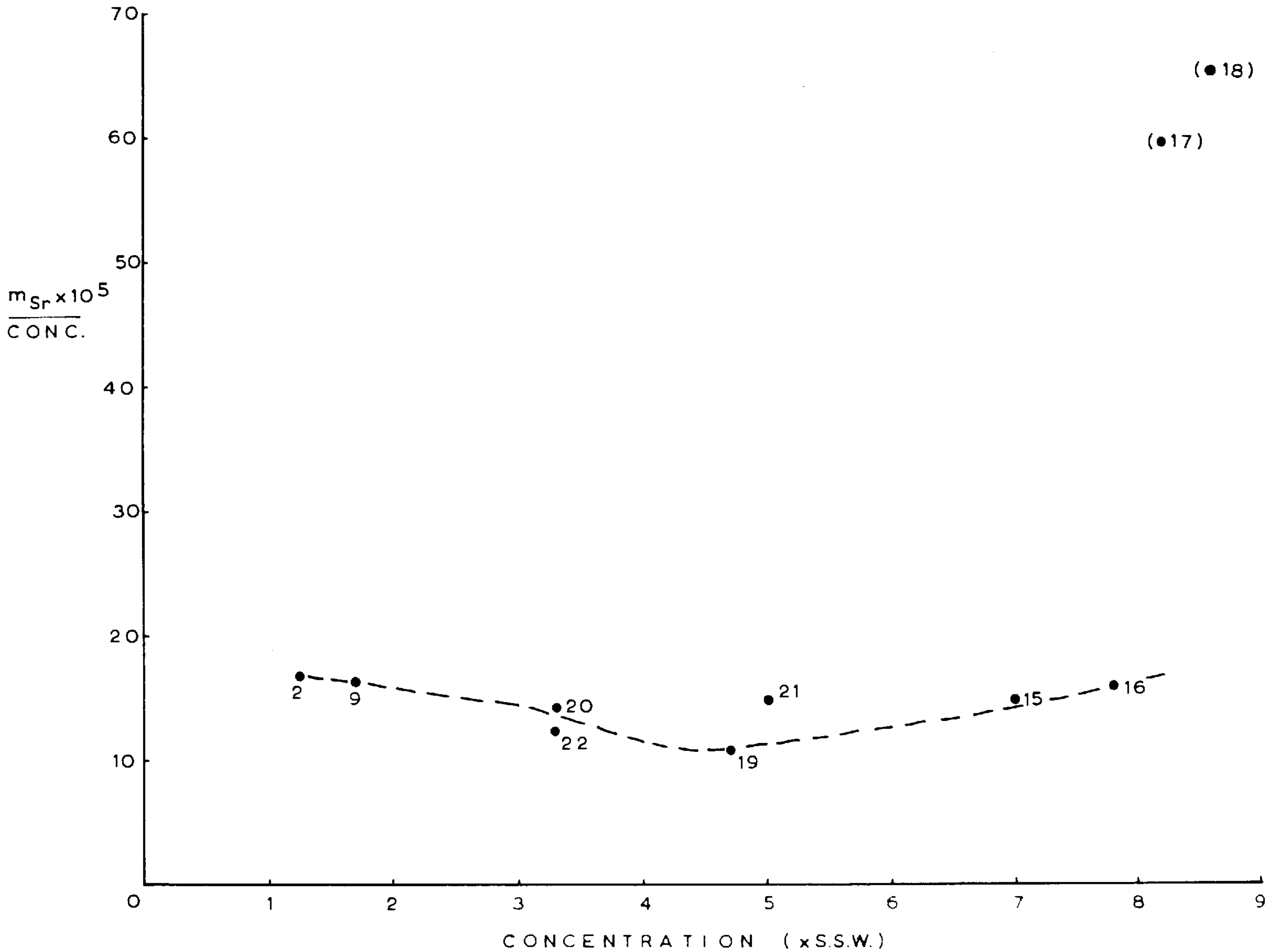
Yet the Ca curve shows a minimum near a concentration of x$4\frac{1}{2}$-5 (50% Ca loss) and then actually increases at greater concentrations. This relative increase in Ca is associated with dolomitisation and represents the excess of Ca which is not precipitated as sulphate minerals. Samples 17 and 18 are anomalous, and if concentrations of x11.7 and x18 are used (derived from m_{Na}/m_{Cl} ratios) the values still show a relative increase. It is possible that this may be equated with the very low values of SO_4 in these waters, the Ca being able to rise to fairly high concentrations owing to the great solubility of $CaCl_2$.

The strontium data (Fig. 61) show a similar pattern to that exhibited by Ca. The initial Sr loss is associated with aragonite and later loss takes place as celestite. The increase after the minimum is related to release of Sr during dolomitisation. Samples 17 and 18 show similar anomalously high values to those for Ca and may also reflect the almost complete lack of SO_4.

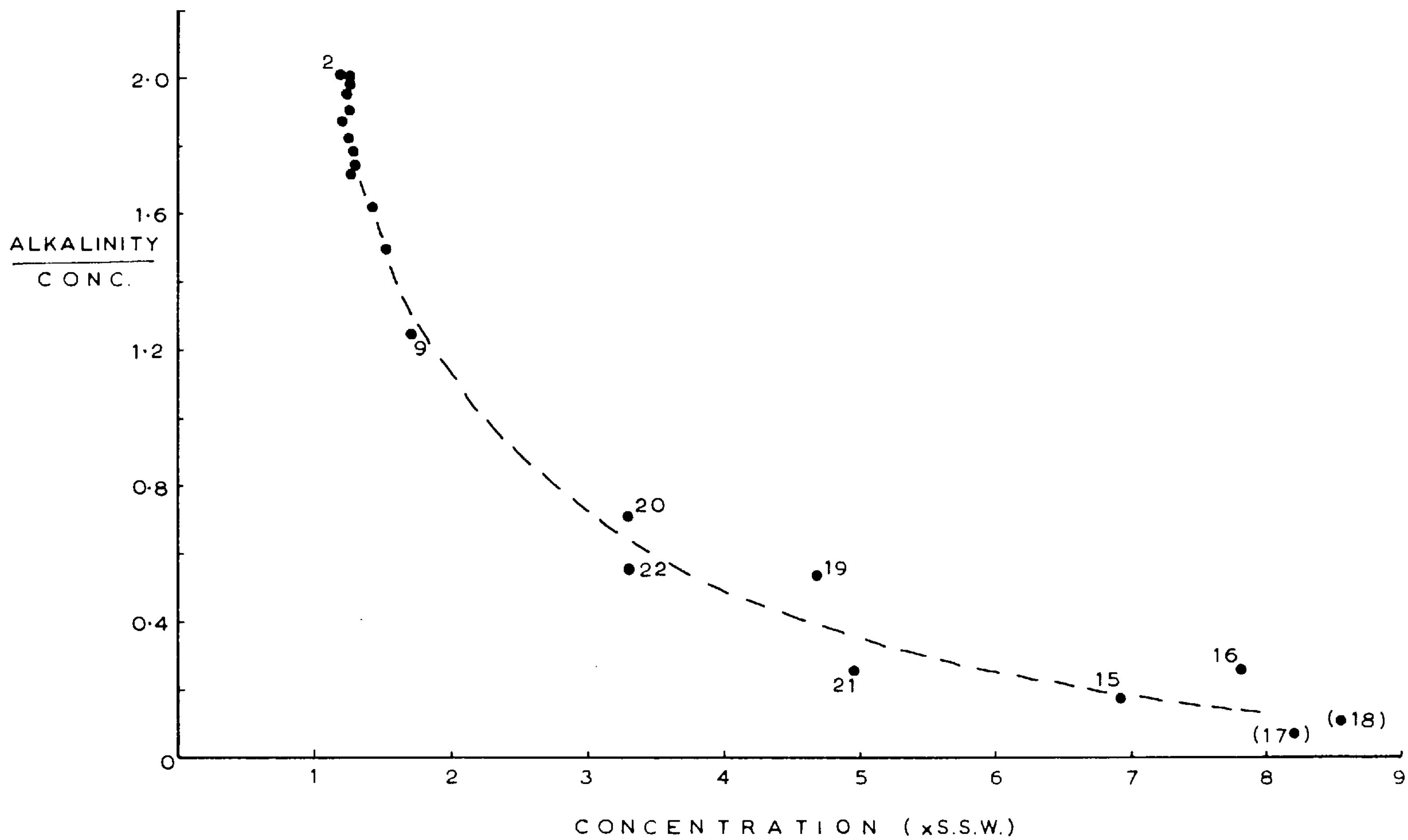
Alkalinity/concentration ratios are plotted in Fig. 62, and show decreasing values with increasing brine concentration. The significant feature here is that within the ground waters alkalinity is low and this obviously will have a limiting effect on the diagenetic development of precipitated carbonates.

A generalised diagram showing the trends in evolution of the algal flat and sabkha brines, with respect

Fig. 61: Strontium analyses of ground water samples.



70
60
50
40
30
20
10
0
1
2
3
4
5
6
7
8
9
$\frac{m_{Sr} \times 10^5}{CONC.}$
CONCENTRATION (xS.S.W.)
2
9
20
22
19
21
15
16
(17)
(18)



(ALKALINITY AS MEQS./KGM.)

Fig. 62: Plot of alkalinity/concentration ratios in sea and ground waters.

to Ca, Sr, Mg and SO_4, is given in figure 63. The curves for Mg and SO_4 show loss of these components to both start together at chlorinities of about 65‰, yet field evidence indicates that gypsum formation definitely preceeds dolomitisation. The gypsum and dolomite are both first developed in positions above the water table and thus water table samples may tend to lag behind some stages of the diagenetic mineral developments.

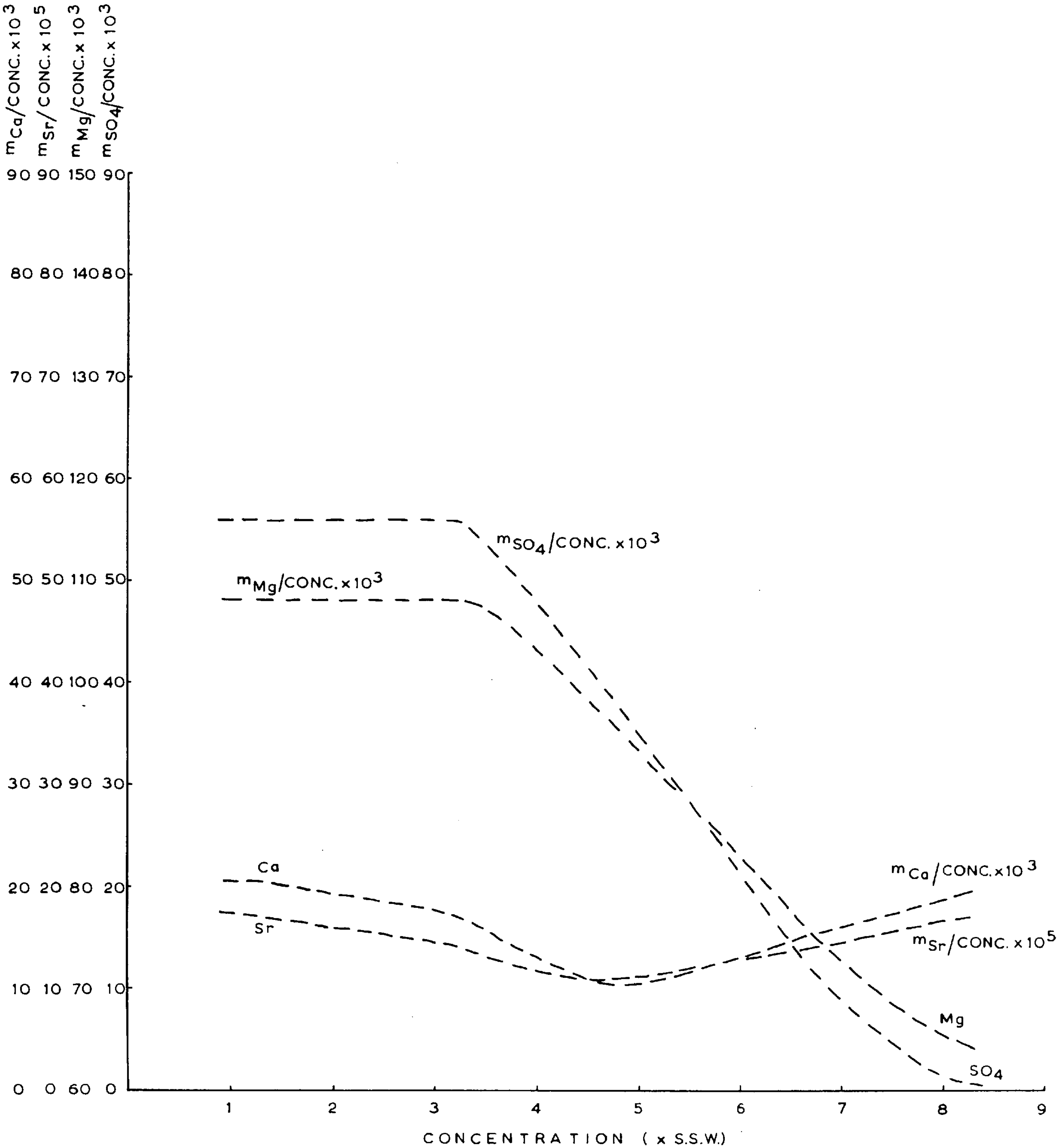
General Discussion

Having now presented what is considered to be the critical and relevant information regarding the sabkha environment, and having described the mineralogies developed and the progressive changes observed within the pore fluids, the ground is prepared for a discussion of the process of dolomitisation, of magnesite and huntite development, of gypsum/anhydrite relationships, and of the strontium cycle.

Dolomitisation

As previously described, it is the fine-grained, aragonite muds which become most rapidly dolomitised, and the process may be followed microscopically by noting the gradual loss of the aragonite needles. The needles are slender and typically less than 1μ in length. In their place are developed dolomite crystals, occasionally as well shaped rhombs, or more commonly as roughly equant grains, in both instances 2-3μ in size.

Fig. 63: Summary diagram of ground water evolution with respect to Calcium, Magnesium, Strontium and Sulphate.

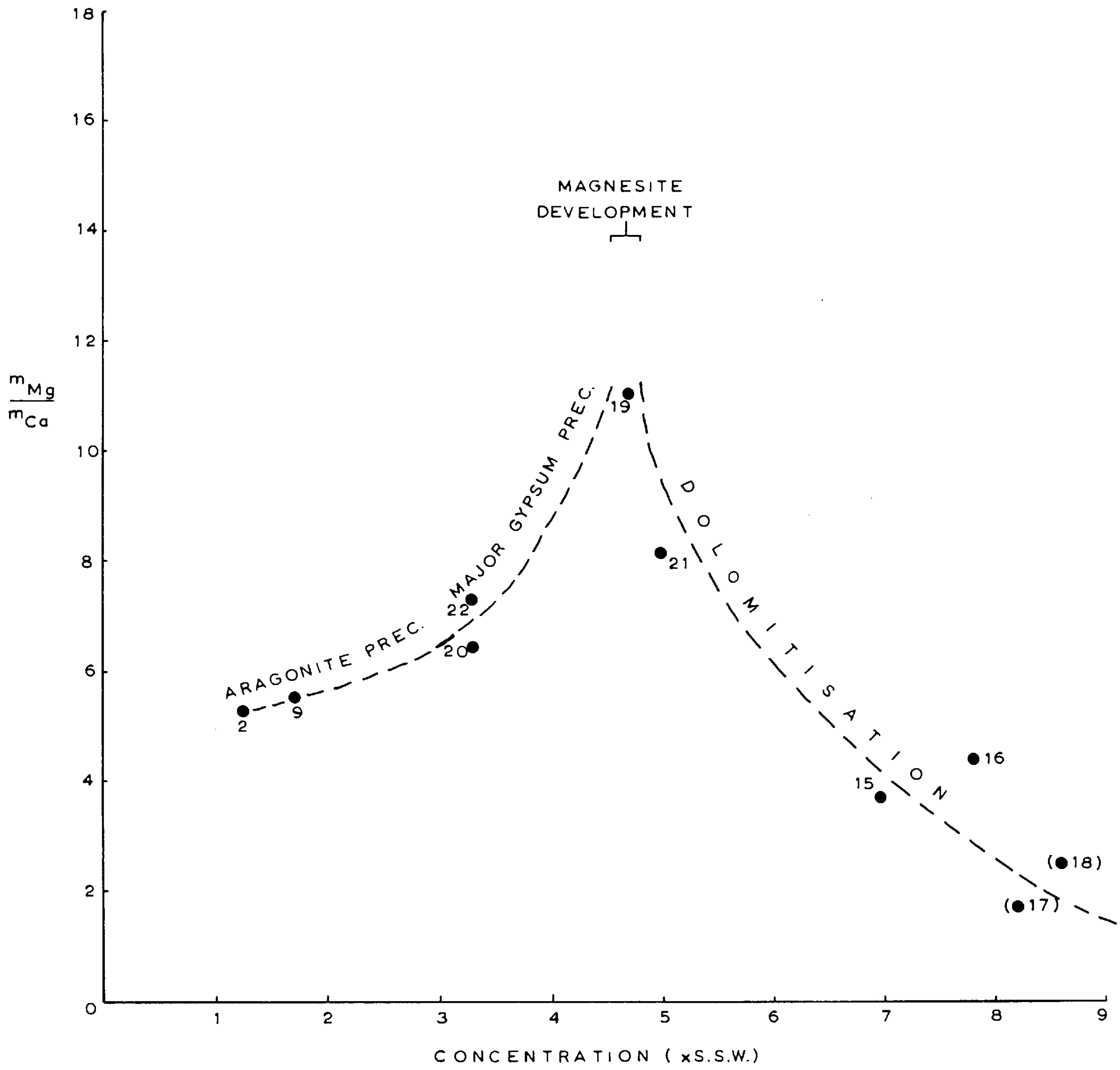


m_{Ca}/CONC. x 10^3
m_{Sr}/CONC. x 10^5
m_{Mg}/CONC. x 10^3
m_{SO_4}/CONC. x 10^3
90 90 150 90
80 80 140 80
70 70 130 70
60 60 120 60
50 50 110 50
40 40 100 40
30 30 90 30
20 20 80 20
10 10 70 10
0 0 60 0
1 2 3 4 5 6 7 8 9
CONCENTRATION (x S.S.W.)
m_{SO_4}/CONC. x 10^3
m_{Mg}/CONC. x 10^3
Ca
Sr
m_{Ca}/CONC. x 10^3
m_{Sr}/CONC. x 10^5
Mg
SO_4

The pore fluids obviously react with the aragonite sediments. The average ground water temperature of $34^{\circ}C$, and the small size of the aragonite needles, will both be factors tending to increase the rate at which the reaction proceeds. From the available evidence, the dolomitisation is a process of solution of original aragonite needles and grains, and precipitation of dolomite as rhombs or equant grains. Degens and Epstein (1964) have examined the carbon and oxygen isotope ratios of several Recent dolomites and have concluded that the growth of the dolomite took place under solid state conditions from a crystalline $CaCO_3$ precursor. This conclusion is thought to be disproved by the present author by textural and other evidence, but cannot, as yet, be dismissed out of hand.

The brines which bring about the reaction are fairly concentrated and dolomite does not appear in the sediments until the ground waters concentration reaches chlorinities of about 65‰ (Fig. 63). The m_{Mg}/m_{Ca} ratio is critical in the dolomitisation process (Fig. 64). In the shelf to lagoon sequence the ratio has been shown (Ch. IV) to rise from 5.3 to 5.5, owing to precipitation of aragonite. Further precipitation of aragonite raises the ratio slightly higher and when gypsum precipitation starts at chlorinities of about 65‰ , the m_{Mg}/m_{Ca} ratio rises steeply to values in excess of 11. In studying a similar diagenetic environment, on the coast of the Qatar

Fig. 64: Plot of m_{Mg}/m_{Ca} ratios in algal and sabkha ground water samples.



MAGNESITE DEVELOPMENT
MAJOR GYPSUM PREC.
ARAGONITE PREC.
DOLOMITISATION
$\frac{m_{Mg}}{m_{Ca}}$
CONCENTRATION (xS.S.W.)
0
2
4
6
8
10
12
14
16
18
1
2
3
4
5
6
7
8
9
2
9
22
20
19
21
15
16
(18)
(17)

Peninsula, Illing and Wells (1964) recorded m_{Mg}/m_{Ca} ratios as high as 17. Close sampling in both areas might reveal even higher ratios. The formation of magnesite would seem to fit into the very local environment where the m_{Mg}/m_{Ca} ratio in the ground waters is at its peak. In the capillary zone above such a position, further gypsum precipitation will occur and it is possible that m_{Mg}/m_{Ca} ratios of 20 or even 30 may be exceeded. It is within the upper part of the capillary zone that the magnesite has been found, and the dolomite also first appears at this level.

Once started the process of dolomitisation would appear to proceed fairly rapidly. For example, sample station 21 (Fig. 64) has fair amounts of gypsum, together with moderately high amounts of dolomite. The ground water from this position is obviously past the peak m_{Mg}/m_{Ca} ratio, but in the algal flat samples 19 and 20, close examination has shown that there is no dolomite developed and thus these samples are situated at or to the left of the peak value. Sample 22 has traces of dolomite developed, yet from other evidences would appear to also be on the left of the peak value. This interpretation cannot be examined in too great detail as the number of samples is insufficient and there are minor anomalies in the Mg values of samples 19 (high) and 22 (low) which cannot be fully explained by the local mineralogies. The problem may in part be one of local migration of ground waters so that the ground waters and diagenetic minerals in any position need not necessarily be intimately related.

The dolomitisation of the original carbonate sediments proceeds to completion in some places or at some levels. The continued loss of Mg and fall in the m_{Mg}/m_{Ca} ratio across the sabkha supports this suggestion. The final m_{Mg}/m_{Ca} ratio falls to less than 4, and if the values of samples 17 and 18 are accepted, it falls to less than 2. This ratio is close to what would be theoretically expected, in fact the equilibrium ratio for dolomite would be 1.

Two dolomitisation reactions may be envisaged:-

1. $2CaCO_3 + Mg^{++} \longrightarrow CaMg(CO_3)_2 + Ca^{++}$
(aragonite) (brine) (dolomite) (brine)

2. $CaCO_3 + Mg^{++} + CO_3^{--} \longrightarrow CaMg(CO_3)_2$
(aragonite) (brine) (brine) (dolomite)

Reaction 1 requires Mg^{++} from the brines to react with the original aragonitic sediments, the reaction products being dolomite and Ca^{++}. Reaction 2 requires by Mg^{++} and CO_3^{--} in the brines to react with the original aragonite; in this case the reaction product is only dolomite.

It has been shown that the loss of SO_4 (Fig. 59) from the brines, with parallel development of $CaSO_4$ minerals, requires massive production of Ca as a by-product of some replacement reaction. Dolomitisation by reaction I would satisfy this requirement. However, the excess Ca for $CaSO_4$ mineral development could also be derived by direct reaction between brines and sediments, and it can be shown (see later discussion of gypsum/anhydrite) that limited reaction of this type certainly does take place. Mass balance calculations involving relative loss of Mg, Ca and SO_4 from the brines show that the dolomitisation is effected chiefly by reaction 1. For example, the only sulphate minerals precipitated are gypsum, anhydrite and celestite, and of this total celestite is not more than 1% and can be ignored. Thus for every equivalent of SO_4 removed from the brines, one equivalent of Ca will be removed (as $CaSO_4$ or $CaSO_4.2H_2O$). Similarly, in the process of dolomitisation ($2CaCO_3 + Mg^{++} \longrightarrow MgCa(CO_3)_2 + Ca^{++}$), for every equivalent of Mg lost from the brines there will be an equivalent gain in Ca. To test this suggestion, samples 20, 21 and 15 will be examined, values of m_{Ca}/conc., m_{Mg}/conc. and m_{SO_4}/conc. being read from figures 60, 58 and 59 respectively:-

	Sample 20	Sample 21	Loss	
$\frac{m_{Ca}}{conc.} \times 10^3$	17	12	5	14
$\frac{m_{Mg}}{conc.} \times 10^3$	108	99	9	
$\frac{m_{SO_4}}{conc.} \times 10^3$	54	36	16	

The relative loss in SO_4 is 16 Meqs.; the Mg loss may be converted to an equivalent Ca gain (16 Meqs.); the total Ca lost is 14 Meqs. The agreement is fairly close. The additional Ca required to balance SO_4 could be derived from a sediment/brine reaction involving direct solution of $CaCO_3$.

Similarly:	Sample 21	Sample 15	Loss or Gain
$\frac{m_{Ca}}{conc.} \times 10^3$	12	18	+6 (Gain)
$\frac{m_{Mg}}{conc.} \times 10^3$	99	66	33 (loss)
$\frac{m_{SO_4}}{conc.} \times 10^3$	36	28	28 (loss)

In this instance the SO_4 loss is 28 Meqs.; the Mg loss is 33 Meqs., equivalent to a gain of 33 Meqs. of Ca, of which 28 Meqs. combine with SO_4, leaving a surplus of 5 Meqs., which is close to the observed gain of 6 Meqs. Ca.

These two examples prove fairly conclusively that the major dolomitisation process is a straightforward replacement of half of the Ca of the original $CaCO_3$ by Mg from the brines (reaction 1).

Studies of aragonite muds by Sir W. Halcrow & Partners, as part of their programme of road building within the Sheikdom, have indicated a typical sample to comprise almost 50% pore space. At a density of 2.9 gms/cc, 1 cc. of aragonite sediment = 1.45 gms., or 0.58 gms. Ca. On dolomitisation, assuming a $Ca_{50}Mg_{50}$ product, 0.174 gms. of Mg are required, and this amount of Mg is contained in 130 ccs. of normal sea water. This volume is reduced to 20-25 ccs. if a brine concentration x5-6 is assumed (the concentration at which dolomitisation is well under way). But as the brine is rarely stripped of more than 40% of the available Mg, this volume must be increased to 50-60 ccs.

As the initial sediments are over 90% aragonite, and amounts of high Mg calcite are never greater than 4-5%, the Mg content of the original sediment will be negligible. To dolomitise 1cc. of such a sediment it has been shown above that it is necessary to pass through it 50-60 ccs. of a brine of concentration x5-6 (equivalent to 250-350 ccs. of normal sea water). To carry out this process some kind of circulation system must be developed. In the sabkha environment, high rates of net evaporation cause a uni-directional flow, the lateral movement of waters proceeding inwards from the lagoons into the sabkha sediments. There is no definite evidence of any kind of a return system or reflux. If the brines of the mid and inner sabkha positions only reach limited concentrations then this might be regarded as evidence of a reflux.

The maximum thickness of the sabkha sediments probably does not exceed 20-30 feet (6-9 ms.) and a more likely average maximum value is 10 feet (3 ms.).

This column of sediment is moderately permeable and thus downward movement of the dense interstitial brines of the upper levels is physically possible. The sabkha sediments are overlying the fairly permeable Miliolite Limestone, which could well provide a path for the seaward movement of the brines. The initial marine sediments will have pore spaces filled with waters of normal lagoon type and not until sedimentation has built up a surface above mean high water will there be any development of sabkha conditions and highly concentrated (high density) brines. The concentrated, dense, interstitial brines of the sabkha are thus underlain in depth by pore fluids of lower concentration and density; the displacement of the latter could well be achieved and a reflux system developed.

On the Quatar Peninsula, Illing and Wells (1964) found the dolomite to almost disappear at depths of 2-4 feet (1 m.), and certainly this is the picture found in the seaward parts of the Trucial Coast sabkha. But further inland, where the diagenetic processes have been operating for a longer period, dolomite is still abundant at depths of 4 feet (1.3 ms.). If the relationship found by Illing and Wells prevailed throughout the sabkha environment then it could be postulated that the dolomitisation and other diagenetic changes are limited to levels near to or above the ground water table, and that they are limited to these levels essentially because, only here are brines of sufficient concentration produced. But the lack of very concentrated brines and the development of dolomite at depths of 2 feet (0.7 m.) and more below the water table would indicate some kind of reflux system to be operating. Although the concentrated, high density brines are, indeed, the product of the near surface levels, once concentrated they may participate in the displacement of less dense pore fluids at deeper levels. Thus, on slender evidence, it is suggested that a reflux may be operative within the sabkha environment and that all the sediments of the present cycle of sedimentation and perhaps even the underlying Miliolite Limestone, may be being dolomitised.

The development or otherwise of a reflux system has great significance with reference to the diagenesis effected. If there is no reflux operative then dolomitisation will be limited to the upper levels, near to and above the ground water table. If a reflux is operative then the complete sequence of lagoonal sediments will become dolomitised. Whether or not a reflux system is operative, the initial dolomitisation takes place in the capillary zone above the water table. Later dolomitisation at depth is dependent on the operation of a reflux system.

Recent or late-Glacial dolomite has now been described from several parts of the world:-

1. Alderman et al. (1957-1963) - South-east of South Australia.
2. Taft (1961) - South Florida.

3. Miller (1961) - Inagua, Bahamas.

4. Graf, Eardley and Shimp (1959, 1961) - Great Salt Lake, Utah.

5. Wells (1962), Illing and Wells (1964) - Qatar Peninsula, Persian Gulf.

6. Shinn and Ginsburg (1964) - Florida and Bahamas.

7. Deffeyes, Lucia and Weyl (1964) Lucia, Weyl and Deffeyes (1964) Bonaire, Netherlands Antilles.

In addition, dolomite of no great geological antiquity has been reported from several coral atoll borings.

The Great Salt Lake occurrence is unusual, being in an inland lake where water compositions differ widely from that of sea water. The dolomite is very fine-grained, has a poorly ordered crystal lattice structure, and has been dated as 11,300 before present. This occurrence and the dolomite of the atoll reef rocks will not be considered further, apart from commenting that the atoll dolomites are all Ca rich and show weakened order reflections (Goldsmith and Graf, 1958).

Of the remaining examples, only the Australian occurrence would seem to be truly primary. In this locality are a series of restricted lagoons and ephemeral lakes, with waters whose composition approaches that of sea water, variously diluted or concentrated. Plant photosynthesis brings about high pH conditions (up to 9.2: rarely up to 10.3) and the dolomite has been found to precipitate from the surface waters when m_{Mg}/m_{Ca} ratios are about 7-8. In addition to the fine-grained dolomite, some aragonite and fairly large amounts of high Mg calcites are also precipitated. The dominant factors controlling the mineral phases precipitated are believed to be pH and m_{Mg}/m_{Ca} ratio. The precipitated dolomite is Ca rich and has a partly disordered lattice, conforming to the protodolomite definition of Graf and Goldsmith (1956). The composition ranges from $Ca_{50}Mg_{50}$ to $Ca_{56}Mg_{44}$. Later diagenesis converts the protodolomite to an ordered, stoichiometric dolomite. Radiometric analyses of the dolomite have given ages of less than 600 years and about 2500 years, proving it to be Recent (Skinner, Skinner and Rubin, 1963).

The dolomite described by Taft (1961) from southern Florida was found most commonly near the carbonate sediment/water interface, as euhedral rhombs, ranging from less than 1μ to more than 60μ in size. The rhombs typically had dark, often rhomb-shaped, inclusions as cores. Little data was available to explain the occurrence, but on textural evidence Taft considered it to be a product of Recent diagenesis. X-ray diffraction showed the dolomite to have normal lattice spacings and to be presumably stoichiometric. Dolomite from this area was analysed radiometrically (Deffeyes and Martin, 1962) and found to show no measurable C^{14} activity, indicating an age in excess of 35,000 years.

It was concluded that the dolomite was of detrital origin, and it was warned that textural evidence may on occasions be very misleading in the assignment of a Recent origin to occurrences of dolomite in unconsolidated sediments. However, Taft (personal communication, June 1964) has indicated that new C^{14} data show some of the dolomite to be definitely active ($<$ 35,000 years), although material with dusky cores is probably detrital.

Miller (1961) has found clear, pink dolomite rhombohedra, up to 140μ, associated with aragonite and calcite in the sludge scraped from evaporating pans on the island of Inagua in the Bahamas. The dolomite occured in sediments which had remained above the water table for 8 years. Miller considered the dolomite to be younger than the enclosing matrix, although there was no evidence whether it was a replacement of $CaCO_3$ or a void precipitate. In the evaporating pans, although high temperatures and concentrations are achieved, no dolomite is precipitated. No data is available regarding composition or degree of ordering of the dolomite. As Deffeyes and Martin (1962) have warned, textural evidence is not necessarily proof of Recent origin, and this occurrence may therefore not be as straight forward as it seems. Certainly the crystal development would seem excessive, over a period of 8 years, when compared with other similar diagenetic dolomite occurrences of Recent age.

Wells (1962) has described dolomite from the carbonate sediments above high water, on the Qatar Peninsula, Persian Gulf. The occurrence has since been described in more detail by Illing and Wells (1964), and can be seen to closely parallel that described in this study. The environment and mode of formation are almost identical. The dolomite occurs as rhombs 1-5μ in size, and two samples gave C^{14} dates of 2,670 and 3,310 years, proving it to be Recent. No data have been given on the relative degree of ordering of the dolomite, but in the earlier paper, an X-ray diffraction powder photograph shows very weak ordering reflections and an expanded lattice, consistent with a composition $Ca_{54}Mg_{46}$.

Shinn and Ginsburg (1964) have described dolomite development from Florida and the Bahamas, in the latter area the process taking place over hundreds of square miles. The primary carbonate sediment sequence is similar to that of the Trucial Coast, with stromatolitic algal sediments, aragonite muds, and abundant evidences of shallow water conditions. The dolomite is a replacement of the carbonate sediments, including skeletal fragments, occurs as crystals less than 3μ in size, and has been dated as Recent. The occurrence is similar to that of the Trucial Coast, except that there is a lack of associated evaporitic minerals. Pore fluid concentration by surface evaporation losses, and dolomitisation (and possibly gypsum precipitation?) probably take place during the dry season, but during the rainy season (May-September: 40-60 inches of rain per year) the more soluble minerals (gypsum?) will be leached away and dolomite alone will remain. During the present cycle of sedimentation (4-5000 years) up to 5 feet (1.6 ms.) of dolomite has been formed. No data is available on the degree of ordering or composition of the dolomite.

Dolomitisation of Recent and Plio-Pleistocene sediments has been reported from the island of Bonaire, Netherlands Antilles by Deffeyes et al. (1964) and Lucia et al. (1964). The process is one of replacement of $CaCO_3$, brought about by concentrated sea water brines in which the m_{Mg}/m_{Ca} ratio has been raised by gypsum precipitation. A reflux of the dense brines has been shown to occur, the brines flowing downward through underlying carbonate sediments and limestones and bringing about the replacement of the original $CaCO_3$ by dolomite. The dolomite in the loose sediments typically occurs as crystals, about 2μ in size, shows at least a partially ordered structure, and compositions range from $Ca_{54}Mg_{46}$ to $Ca_{56}Mg_{44}$. Brines become concentrated in both interstitial positions and in ponded, surface lakes; the brines are shown to have lost $CaCO_3$ and $CaSO_4$, but no loss of Mg can be demonstrated. The ratio m_{Mg}/m_{Ca} rises as high as 30. However, a lack of halite and other late evaporite minerals, and the fact that the brines reach concentrations of only about x7, has been interpreted as indicating that a reflux of the brines must be taking place. Mass balance calculations by the authors have shown that the suggested rate of reflux is capable of carrying out the dolomitisation which has been found. This study is an excellent demonstration of the inferred process of dolomitisation suggested by Newell et al. (1953) for the Permian Reef Complex rocks, and by Adams and Rhodes (1960).

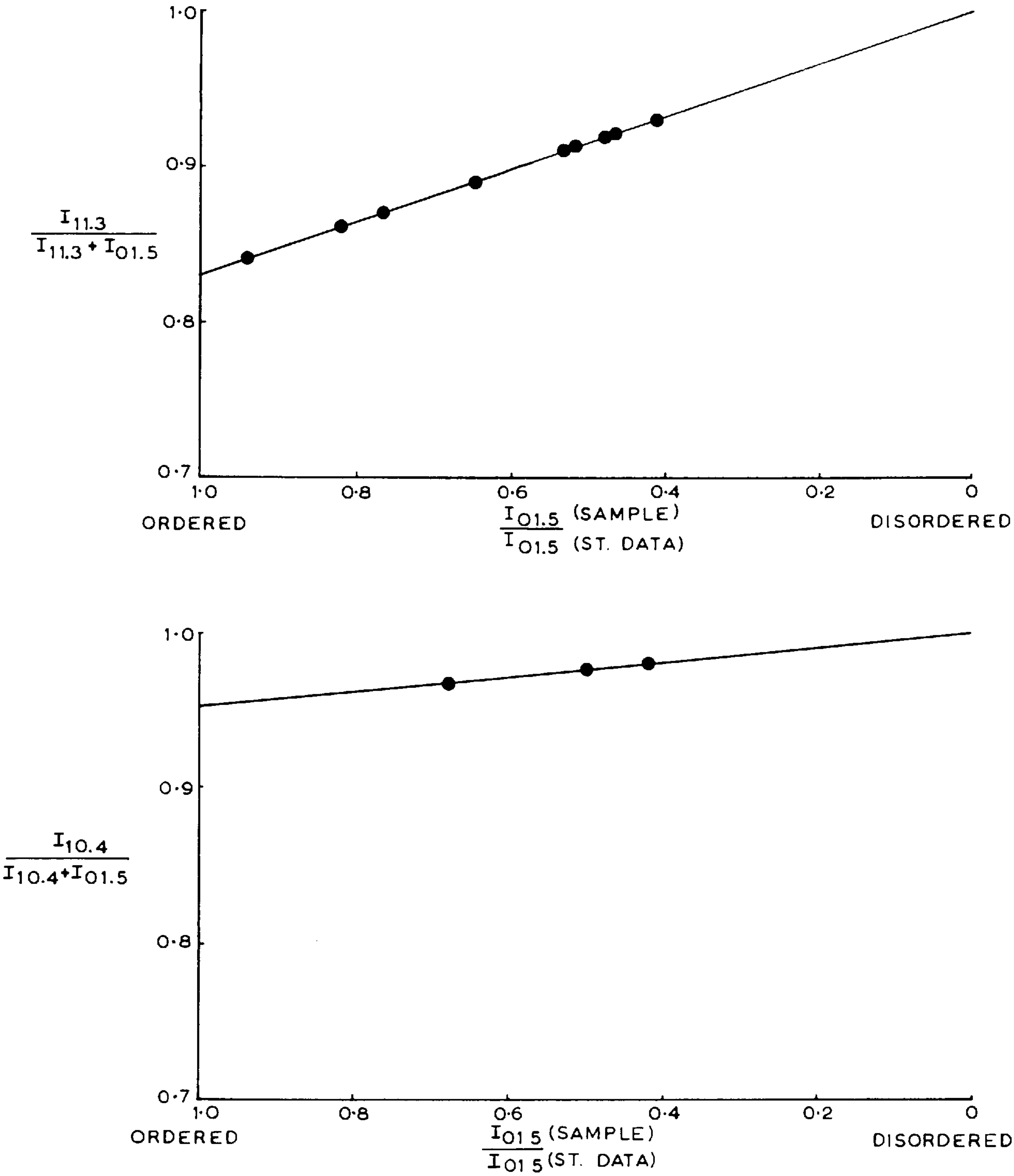
Except for the Australian occurrence, all these studies, including that from the Trucial Coast, indicate that dolomite of early diagenetic origin is a common development and is to be expected in almost any area of carbonate sedimentation, particularly where net evaporation occurs. The net evaporation need not necessarily be an annual (absolute) net evaporation, as a prolonged dry season, with high rates of evaporation, is sufficient to bring about dolomitisation of carbonate sediments, particularly if the sediments are fine grained. Owing to its low solubility, a short rainy season, bringing about surface and interstitial brine dilution, will not necessarily result in destruction of the dolomite. Where data is available the replacement or precipitation process is largely dependent on the m_{Mg}/m_{Ca} ratio of the waters, a value of about 8 having to be exceeded. Factors such as high temperature ($>30^{o}C$) and small grain size of the original carbonate sediment will bring about an increase in the rate of the dolomitisation process, although several authors have shown that even fairly coarse skeletal fragments have been replaced within the past 4-5000 years.

The initial dolomite, in nearly all instances, shows an excess of Ca and an expanded lattice. The lattice ordering of the Ca and Mg ions is often poor, as is indicated by the poor development of the ordering reflections (10.1, 01.5 and 02.1). The Trucial Coast dolomite ranges in composition from $Ca_{50}Mg_{50}$ to $Ca_{55}Mg_{45}$. The degree of lattice ordering is variable but an attempt has been made to quantify this relationship, rather than use terms such as 'poorly ordered' or 'well ordered', which are beginning to fill the literature on this topic.

D.L. Graf (personal communication) has suggested that the placing of carbonate disorder on a numerical scale is complicated by the suspicion that there is both point and layer disorder. However, although later work may prove it to be oversimplified, an empirical parameter may be used, relating the intensity of one of the ordering reflections to the intensity of some other lattice reflection (not an ordering reflection). In Fig. 65, the ordering reflection used is 01.5 and the other structural reflection either 11.3 or 10.4, the choice of the latter being determined entirely on convenience. This method makes the assumption that the degree of ordering is paralleled by an equivalent development in the intensity of the

Fig. 65: Relative order or disorder of Dolomite structure.

DOLOMITE ORDER / DISORDER



STANDARD DOLOMITE DATA FROM GRAF (1961)

ordering reflections. The end points on which the two reference curves are based are derived from standard dolomite data (Graf 1961). The fully disordered structure is displayed by minerals such as calcite, where any cation can occupy any cation position in the structure (obviously it makes little difference if the cations are all the same, as they are in calcite). If a fully disordered dolomite were to be found, then all the cation positions would be occupied by an entirely random distribution of Ca and Mg, and the order reflections (as with calcite) would be absent. This is not found in practice, all dolomites showing some degree of cation ordering. In a fully ordered dolomite, the Mg occupies one plane within the structure and the Ca occupies another; mixed planes do not exist.

The Trucial Coast dolomites show varying degrees of ordering, ranging from 40 to 95%. No correlation has been found between degree of ordering and composition or between either of these factors and position within the sabkha sequence. It is reasonable to assume that at an equivalent level the inner sabkha dolomites are older than those to seaward. Studies of the Australian dolomites (Alderman et al. 1963) have shown that they become ordered with time and develop a stoichiometric composition, but this relationship has not been discerned in the sabkha.

Students of the 'dolomite problem' are aware of the difficulties and problems surrounding the formation of dolomite, and the mineral has yet to be synthesised under natural conditions. An intelligent discussion of some of the possible reasons attaching to the difficulty of dolomite formation has been forwarded by Graf and Goldsmith (1956). They consider the problem of forming a mixed cation crystal and its nature when finally formed. They suggest that the ordering requirement of the structure reduces the rate of nucleation and the rate of growth upon already formed nuclei. The occurrence of Ca-rich, partially ordered initial phases would therefore seem reasonable, and is summarised in the 'protodolomite concept' (protodolomites being the suggested precursors of the fully ordered, stoichiometric dolomite).

Much has been written concerning the conditions under which dolomite may form and data from the natural occurrences gives an idea of the requirements. The precipitation of dolomite from open waters will be dependent upon $a_{Ca^{++}}$, $a_{Mg^{++}}$ and $a_{CO_3^{--}}$. This gains expression in the m_{Mg}/m_{Ca} ratio, and pH of the waters, and the Australian occurrence indicates the precipitation to occur at m_{Mg}/m_{Ca} ratios of 7-8 and pH values of about 9. Replacement dolomite is rather different and the reaction is dependent entirely on $a_{Ca^{++}}$ and $a_{Mg^{++}}$. The reaction $2CaCO_3 + Mg^{++} \rightarrow CaMg(CO_3)_2 + Ca^{++}$, obviously does not involve $a_{CO_3^{--}}$ and thus the pH of the interstitial brines is not critical in this reaction. The controlling factor in the development of replacement dolomite is the m_{Mg}/m_{Ca} ratio, and values of 8-10, indicated from the Persian Gulf studies, must be achieved for the reaction to get under way.

Magnesite Development

The occurrence of the magnesite has already been described. It is associated with detrital, windblown grains, including abundant calcite, together with diagenetic gypsum, celestite and traces of dolomite. Ground waters in the area have an m_{Mg}/m_{Ca} ratio of 8. In the upper levels of the capillary zone, where the magnesite occurs, further gypsum precipitation probably raises the ratio much higher.

X-ray diffraction shows the magnesite to have a significantly expanded lattice, a feature common to other Recent and late-Glacial magnesites described from Australia (Alderman et al. 1961) and the Great Salt Lake (Graf et al. 1961) respectively:-

	S.E. Australia	Great Salt Lake	Trucial Coast	Standard Magnesite
a_o	4.66	4.669	4.657	4.6330 Å
c_o	15.23	15.21	15.108	15.016 Å

The magnesites all show expanded lattices, although the Trucial Coast magnesite is less expanded than the others. The expansion of the Trucial Coast magnesite could be accounted for by a substitution of 5 Mol % $CaCO_3$. The expanded lattice of the Great Salt Lake magnesite has been discussed by Graf et al (1961), who found that the total amount of cations present, other than Mg, failed by an order of magnitude to account for the lattice expansion. They suggested that the expanded lattice results from the intimate association of water with the solid.

In the Great Salt Lake, a thin bed of fine-grained dolomite was found 1 foot below the surface. Shoreward, a fine grained aragonite/magnesite assemblage occurred at the same level as the dolomite. The magnesite and dolomite were considered to be either chemical precipitates or diagenetic alteration products. The parallel was drawn with waters from dolomitic terrains, evaporating in caves, and precipitating hydromagnesite and either calcite or aragonite. It was suggested that a hydromagnesite-aragonite assemblage was formed in one part of the lake whilst dolomite was formed elsewhere, in water of somewhat different composition. The hydromagnesite has since converted to something akin to magnesite. The possibility was also suggested that the hydrated magnesite may have originally precipitated with the aragonite.

The Australian magnesite occurrence (Alderman et al, 1960, 1961, 1963) is regarded by the authors as primary, a mixed dolomite-magnesite precipitate being recorded, when water pH values were about 10 and m_{Mg}/m_{Ca} ratios about 16. The authors suggested that the magnesite may represent the final Mg-rich product of a reaction series:-

Age of Lake	Sediment	Approx. m_{Mg}/m_{Ca} in water at time of maxm. pH	
	Aragonite + Mg calcite	5	
	Mg calcite	6	
increases	Mg calcite + Ca dolomite	8	Reaction Series
	Ordered dolomite	10	
	Dolomite + magnesite	16	
	Aragonite + hydro-magnesite	20	

Following Graf et al (1961), the authors suggest the possibility that the aragonite-hydromagnesite assemblage may eventually develop into the dolomite-magnesite assemblage.

The Trucial Coast magnesite occurrence is not associated with hydromagnesite-bearing sediments and the data available would indicate that it has developed as the equilibrium carbonate when m_{Mg}/m_{Ca} ratios are at their maximum. As sedimentation gradually increases the width of the sabkha, the zone of peak m_{Mg}/m_{Ca} ratios will move seawards. Thus, although so far only found on Jazirat al Ftaisi, the magnesite may be of a fairly widespread occurrence in the upper sediment levels throughout the sabkha. Whether the magnesite has replaced pre-existing $CaCO_3$ or whether it is a void precipitate is not known.

Studies by Garrels et al (1960) have indicated that calcite and magnesite are unstable with respect to dolomite. Thus the magnesite may be a mineral phase developed only in the narrow environment of high m_{Mg}/m_{Ca} ratios and later in the diagenetic sequence its place may be taken by dolomite.

Huntite Development

The main interest attaching to the huntite is that this is the first occurrence of the mineral associated with evaporitic deposits, and as such huntite must now be added to the list of possible minerals associated with the evaporitic facies.

Huntite has only been described previously from seven localities (in Vitaliano and Beck, 1963) and is usually limited to near surface weathering zones, overlying dolomite, magnesite or brucite rocks. One exception to the more common occurrence is that described by Baron et al (1957), who found the mineral as a fine-grained 'mountain-milk' in a cave in the Herault region of France.

The mineral usually occurs as small veins and nodules. The crystals are white and are always less than 1-2μ in size; X-ray diffraction must be resorted to for identification. The unit cell dimensions are similar in all occurrences, including that from the Trucial Coast:-

	Graf & Bradley (1962)	Skinner (1958)	Trucial Coast
a_o	9.498	9.502	9.499 Å
c_o	7.815	7.818	7.819 Å

All the occurrences so far described are associated with Mg rich rocks and the mineral is thought to be precipitated from Mg rich solutions in the near surface weathering zone.

The Trucial Coast huntite is associated with anhydrite, gypsum celestite and traces of dolomite, but little is known of the chemistry of the solution from which it was precipitated. It resembles anhydrite in its formation as aggregates of crystals, which are able to shoulder aside the enclosing sediments. The diagenetic studies of the sabkha have shown that the brines of much of the seaward part of the sabkha are rich in Mg, and that m_{Mg}/m_{Ca} ratios may be quite high. The sediments of this part of the sabkha are sandy and almost unaffected by dolomitisation. Yet abundant anhydrite is developed. The ground waters at one time must certainly have reached concentrations of about x6, to bring about anhydrite precipitation, yet in areas of muddy sabkha, by this stage in concentration, appreciable dolomitisation has occurred and m_{Mg}/m_{Ca} ratios have fallen to less than 5. Thus in areas of sandy sabkha, where dolomitisation is retarded, Mg rich, concentrated brines may be produced and it is suggested that is is within this solution environment that the huntite is precipitated.

The equilibrium field of Huntite has been suggested by Graf and Bradley (1962) to be fairly restricted, but even so, the required conditions are obviously achieved, in at least some parts of the sabkha. Studies by Garrels et al (1960) suggest that huntite is not stable with respect to dolomite and magnesite. Thus during later diagenesis the huntite may be replaced by other, more stable, Mg-bearing carbonate minerals.

Gypsum-Anhydrite Relationships

The observational data have already been given, together with the brine analyses. It has also been shown that the gypsum precipitation starts at concentrations of about x3.3 and that the major gypsum precipitation is completed by the time ground water concentrations reach x6-7. At these concentrations the brines have been shown to contain only about 20% of the initially available SO_4^{--}. Yet appreciable amounts of anhydrite are found within the sabkha sediments. Although gypsum occurs throughout the sabkha, much of it must go into solution and be reprecipitated as anhydrite. No evidence has been found of anhydrite pseudomorphing earlier gypsum yet from mass balance calculations this process of gypsum solution and anhydrite precipitation definitely occurs. As there is no evidence of pseudomorphing, the remobilised Ca^{++} and SO_4^{--} therefore migrate to new sites before anhydrite precipitates. Early anhydrite crystallisation has been observed in the form of isolated laths or aggregates of laths, in both instances occurring as a primary mineral phase. In addition, the forms taken by the anhydrite (beds, nodules) are rather different from the modes of occurrence of the gypsum.

It has been shown that appreciable amounts of Ca are released by dolomitisation and are subsequently precipitated, together with SO_4, as either gypsum or anhydrite. There is also evidence from the brine analyses of direct replacement of $CaCO_3$ by gypsum. In areas of muddy sabkha sediments, where dolomitisation is fairly rapid, the release of Ca as a by-product of this reaction is fairly substantial. Consequently, large amounts of calcium sulphate minerals can be precipitated and the brines nearly stripped of SO_4. However, in areas of coarser, sandy sabkha sediments, where dolomitisation is retarded, the maximum SO_4 loss is determined by the amount of Ca available from the concentrated sea-water brines. Assuming little loss of Ca as $CaCO_3$, this amounts to only about 35% of the SO_4. Thus in these areas, if sediment/brine reactions do not proceed, the total development of calcium sulphate minerals will be only about one third of that developed in muddy sabkha areas. Field studies have not been sufficiently detailed to enable this to be confirmed. Sandy sabkha sediments may show rather more evidence of a direct sediment/brine reaction involving merely solution of the original $CaCO_3$ of the sediments (brine pH of 6-7 will promote this reaction). It is possible, therefore, that the additional Ca required for calcium sulphate precipitation, in sandy sabkha areas, may originate from a different diagenetic reaction than that which produces the excess Ca in muddy sabkha areas.

Comparison of the field and laboratory data with published synthetic data, has shown that the distribution of the gypsum and anhydrite and the conditions under which they occur, can be reasonably accounted for. Data from synthetic studies of gypsum-anhydrite relationships have been summarised by Stewart (1963), and rather more usefully by W.T. Holser (1961: unpublished report). Holser's summary plot of the equilibrium fields is shown in Fig. 66 (not to be reproduced without permission of W.T. Holser). The temperature range of the sub-surface sabkha sediments is superimposed, together with concentration data of samples 15-22. As has been stressed previously the concentrations are those of the ground water table brines, whereas at higher levels in the sediment increased concentrations will be achieved.

The mineralogical and brine data indicate that gypsum is precipitated both within its own stability field and within the lower part of the anhydrite stability field. Once precipitated, the gypsum can exist (as a metastable phase) throughout the anhydrite stability field. From field evidence the anhydrite does not precipitate until ground water concentrations of x6 are reached, which, at the temperature of the sabkha environment, is well within the anhydrite stability field.

The direct precipitation of anhydrite has proved a difficult synthetic problem and published data

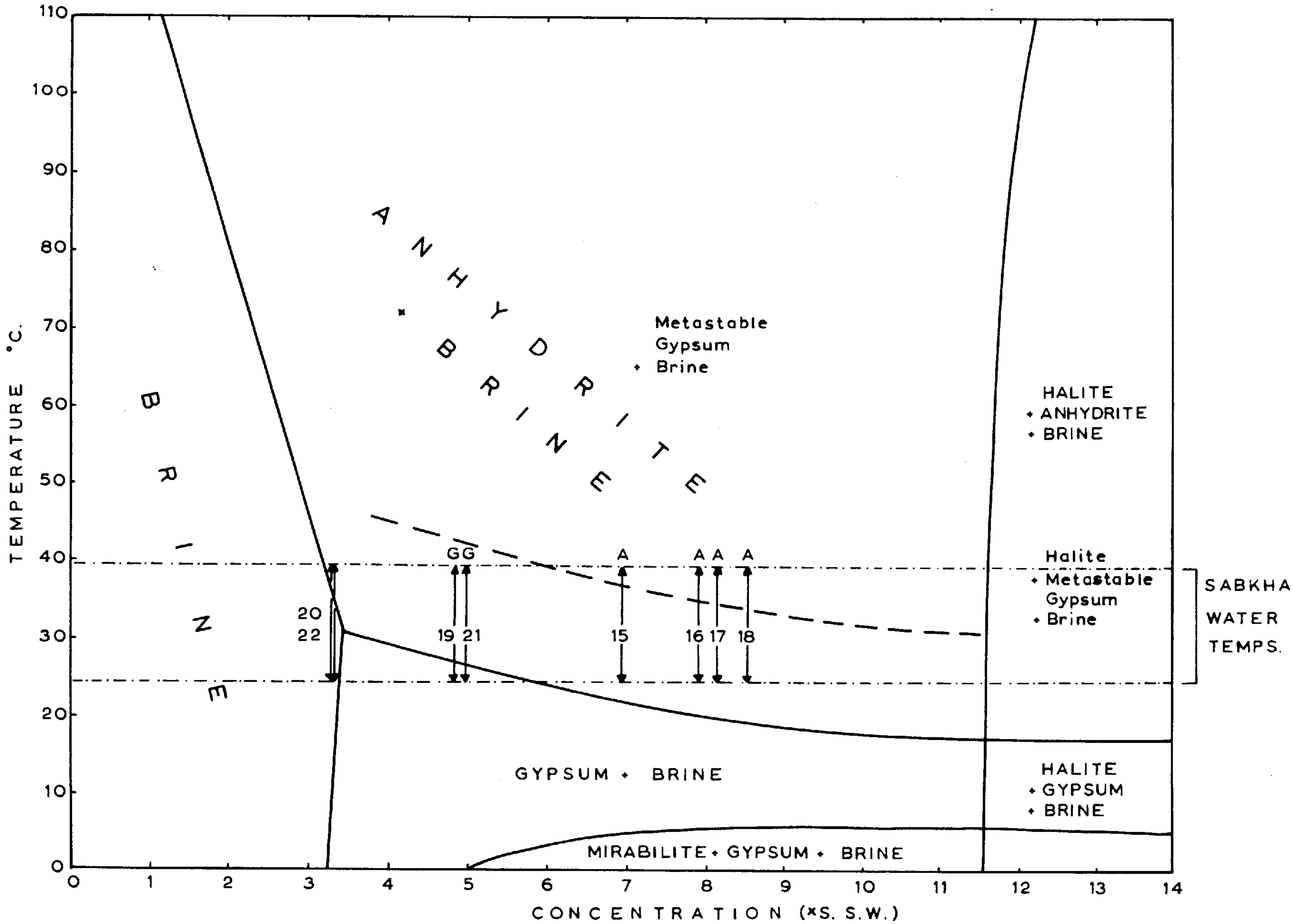
Fig. 66. Equilibrium relationships of calcium sulphate minerals. Stable relationships are shown, together with metastable gypsum relationships (other metastable relationships omitted).

Heavy dashed line, gypsum $\rightarrow$ anhydrite equilibrium boundary determined from Trucial Coast studies (see text).

A = Anhydrite; samples 15-18

G = Gypsum; samples 19, 21

Samples 20, 22 lack gypsum or anhydrite



TEMPERATURE °C.
CONCENTRATION (×S. S.W.)
ANHYDRITE + BRINE
BRINE
Metastable Gypsum + Brine
HALITE + ANHYDRITE + BRINE
Halite + Metastable Gypsum + Brine
SABKHA WATER TEMPS.
GYPSUM + BRINE
HALITE + GYPSUM + BRINE
MIRABILITE + GYPSUM + BRINE
GG
A
A A A
20
22
19 21
15
16 17 18

so far indicate that it can only be precipitated if anhydrite nuclei are added to the solution, within the anhydrite stability field. There would appear to be a kinetic barrier in the direct nucleation of anhydrite, and this is expressed within the sabkha environment by the anhydrite not precipitating until conditions are well inside the anhydrite stability field. In fact, from the sabkha data, it would seem that the gypsum $\rightarrow$ anhydrite boundary should lie as indicated by the dashed line in Fig. 66.

If anhydrite precipitated in accord with the equilibrium data, then in the Trucial Coast areas of high temperatures and fairly rapid achievement of high brine concentrations (owing to high rates of net evaporation), little gypsum would be precipitated. Gypsum precipitation would occur only in the cooler months of the year (November-April), and then only in a very narrow, high intertidal/ sabkha edge zone. In fact, anhydrite would be precipitated in many high algal flat areas.

The gypsum/anhydrite boundary has only been approached directly from the anhydrite $\rightarrow$ gypsum direction (hydration of anhydrite), as the gypsum $\rightarrow$ anhydrite reaction must be preceded by the addition to the solution of anhydrite nuclei. The equilibrium data determined from the hydration of anhydrite are well known. Hydration of anhydrite in the sabkha has been shown in the development of anhydrite cored gypsum crystals.

The combined data indicate the following gypsum/anhydrite equilibrium relationships for the Trucial Coast environment:-

1. Initial precipitation of gypsum occurs as determined from synthetic studies, at a concentration of x3.3.
2. Gypsum precipitation continues into the lower parts of the anhydrite equilibrium field.
3. At concentrations of x6, anhydrite precipitates. The suggested precipitation boundary of the anhydrite field is indicated in Fig. 66.
4. On brine dilution, anhydrite will not be hydrated until concentration and temperature conditions fall below the synthetically determined equilibrium boundary.

In other words, the gypsum → anhydrite (precipitation and solution/precipitation) reaction proceeds at higher concentrations/temperatures than the anhydrite → gypsum (hydration) reaction. It has been suggested that difficulty in nucleation of anhydrite may account for this relationship.

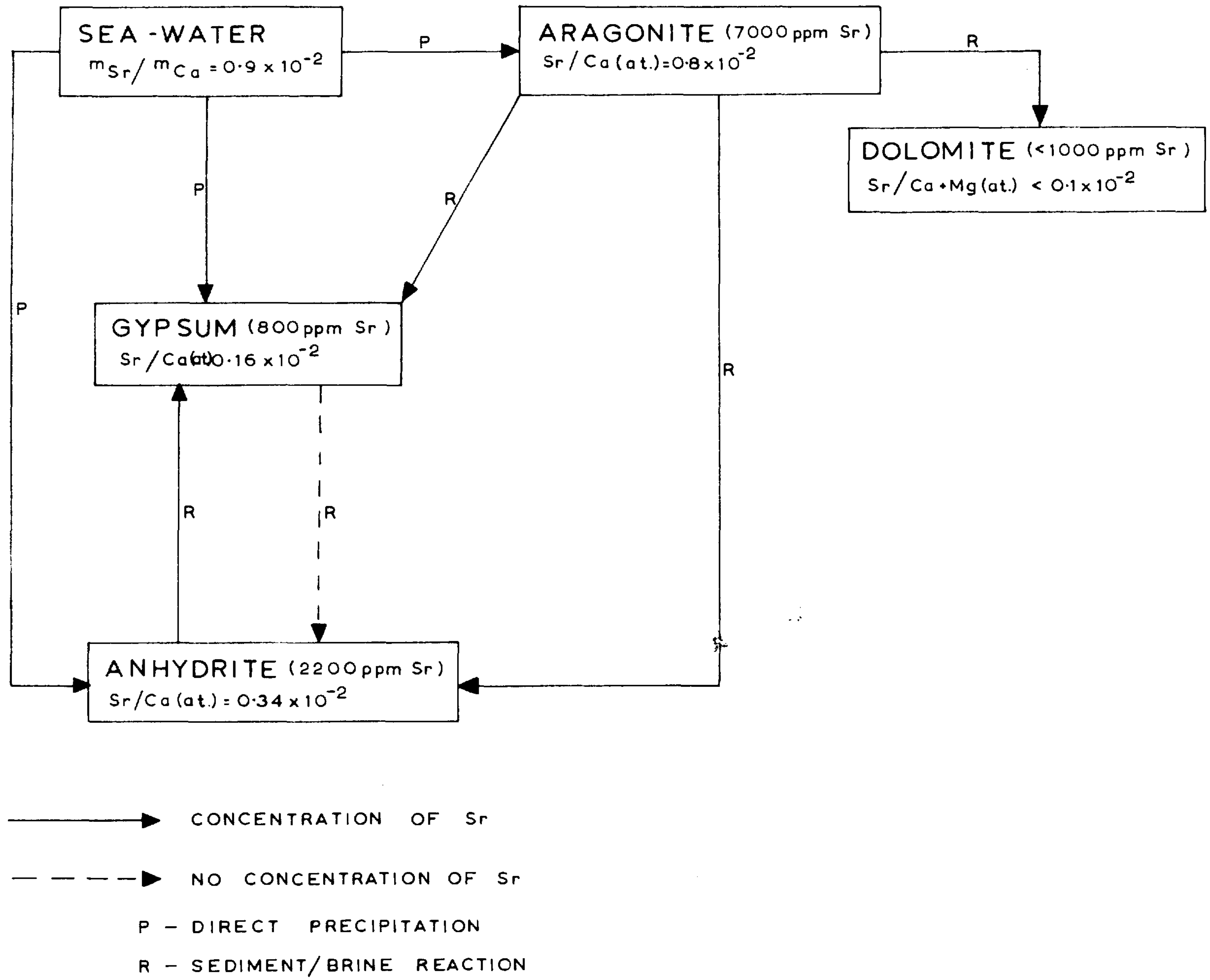
The Strontium Cycle

The strontium cycle within the sedimentary and diagenetic environments of the Trucial Coast is summarised in Fig. 67.

In sea water, the ratio $m_{Sr}/m_{Ca} = 0.9 \times 10^{-2}$, although as discussed in Chapter IV, this value is being checked by further analyses The chemically precipitated

Fig. 67. The Strontium Cycle, within the sedimentary and early diagenetic environments of the Trucial Coast.

THE STRONTIUM CYCLE



aragonites have been shown to contain $\sim$7000 ppm. Sr, and the Sr/Ca (at.) ratio = 0.8 $\times 10^{-2}$, indicating that there is a very slight concentration of Sr in the inner lagoon waters following aragonite precipitation. Molluscan aragonite averages $\sim$ 2000 ppm. Sr, giving a Sr/Ca (at.) ratio of 0.23 $\times 10^{-2}$. Eight gypsum samples, leached free from aragonite and analysed for Sr by X-ray fluorescence averaged 800 ppm. Sr, a Sr/Ca (at.) ratio of 0.16 $\times 10^{-2}$. Ten anhydrite samples averaged 2200 ppm. Sr, giving a Sr/Ca (at.) ratio of 0.34 $\times 10^{-2}$. By studying the diagenetic reactions of the sabkha, and using the above data, it is possible to account for the widespread development of celestite.

Aragonite, gypsum and anhydrite all precipitate from sea-water or concentrated sea-water brines. Taking the m_{Sr}/m_{Ca} ratio in sea-water as 0.9 $\times 10^{-2}$, it can be seen that the distribution coefficient $k_{Sr}^{Aragonite}$ is 0.88; k_{Sr}^{Gypsum} is 0.18; $k_{Sr}^{Anhydrite}$ is 0.38. Thus when aragonite is precipitated there is little concentration of Sr in the remaining sea-waters and no Sr minerals are directly precipitated. When gypsum precipitates from the concentrated brines there is an appreciable concentration of Sr left behind, and celestite is consequently precipitated. The precipitation of anhydrite also brings about a concentration of Sr in the remaining brines, but not as great as that brought about by gypsum precipitation.

If chemically precipitated aragonite is replaced by either gypsum or anhydrite then a concentration of Sr results. Similarly, the replacement of anhydrite by gypsum also effects a Sr concentration. Only the replacement of gypsum by anhydrite does not back up the brines in Sr.

The dolomitisation is also responsible for the production of large amounts of excess Sr. It was not possible to prepare a dolomite sample which was definitely free from celestite and so a mixed sample was prepared and analysed on the electron probe. Sensitivity is unfortunately low and the only certain result is that there is less than 1000 ppm. Sr in the dolomite. Consideration of crystal structure and lattice spacing permit one to guess that the amount of Sr in the dolomite is probably not more than 100-200 ppm. Whatever the absolute value, it can be seen that the production of dolomite from a high Sr aragonite mud, will result in large quantities of excess Sr.

These chemical considerations compare very closely with observational studies. They show why the dolomite is always packed with celestite crystals, why gypsum is usually associated with celestite, and why less celestite is associated with anhydrite than with gypsum. If much of the anhydrite results from reprecipitation of dissolved gypsum, then this accounts for the lack of celestite associated with the anhydrite. In addition, the original carbonate sediment values indicate that much more celestite will be produced during diagenesis of the muddy sabkha sediments (chemically precipitated aragonite muds) than of the sandy sabkha sediments (rich in molluscan, skeletal grains).

Diagenesis of the limestones, resulting in the development of a low-Mg calcitic limestone from an originally aragonitic limestone, is also paralleled by a loss of Sr. Although many of the limestones are rich in molluscan grains and therefore initially low in Sr, some limestones rich in oolites (now calcitic) have been examined which would initially have been high in Sr. In all instances the Sr values of the calcitic limestones were less than 2000 ppm. Aragonitic oolite limestones showed values the same as or very slightly higher than those of the oolite sands. This is reasonable, as the cementing aragonite is of chemical origin and thus high-Sr bearing.

No data are available on the solubility of celestite in sea-water but its occurrence in the diagenetic cycle shows it to precipitate just after the initial gypsum (x3.3). Gypsum is often found without celestite in the high intertidal, algal mat areas, but celestite is never found alone. Assuming a concentration of about x 3.5 for saturation with respect to $SrSO_4$, a solubility of about 0.0074 gms. $SrSO_4$/Kgm. concentrated sea-water is indicated (correcting for Sr coprecipitation with gypsum).

This compares with a value of 0.0114 gms/Kgm pure water at 30° C (Lange, 1961).

Celestite "has been observed to precipitate with the carbonates from the waters" in the south-east of South Australia (Skinner, 1963). Skinner considers the pH of the water to be important in celestite precipitation and that the optimum value is that at which the associated carbonates (dolomite and calcite) precipitate (pH 8.5-9.2). In fact, the water pH should have a negligible effect on the precipitation of celestite; precipitation will occur when saturation with respect to $SrSO_4$ is reached and is thus dependent on a_{Sr}^{++} and $a_{SO_4}^{--}$. Saturation depends therefore on water concentration and on the amount of coprecipitation of Sr with associated precipitating mineral phases.

Celestite has been reported from many limestone and evaporite sequences but if limestones are assumed to have been initially in most part aragonitic, and to have altered to low-Mg calcite during diagenesis, then celestite should be much more common than has been reported. Possibly, isolated celestite crystals have been mistaken for quartz crystals in many instances, the birefringence of the two minerals being identical.

Concluding Remarks

This study has been concerned with the development of a series of carbonate sediments, of shelf facies, barrier facies and lagoonal facies. The barrier of coral reefs, shoal oolite deltas and islands, is a high energy belt, as the dominant wind and wave approach is onshore. The protected lagoon sediments are largely chemical precipitates, although along more exposed inner lagoon coasts molluscan sands and gravels may dominate. One of the features of the lagoonal infilling is the diachronous relationships of all the facies.

The diagenetic studies have shown that a fairly broad knowledge of the original marine sediments is a necessary prerequisite. The chemical and mineralogical studies were directed to this end.

The diagenetic changes within the sabkha have also been shown to be diachronous, the belts moving seaward as the lagoons are infilled. One of the more interesting phenomena is the difference in rate and character of the diagenetic changes which may arise owing to differences in initial sediment type (factors such as grain size, mineralogy and chemistry being important in this respect). For example, dolomitisation proceeds rapidly in muddy sabkha sediments, but is relatively retarded in sandy sabkha sediments. As a result, the ground water brines will differ greatly, the former having lost much

Mg, the latter very little. Partially dependent on dolomitisation is the development of calcium sulphate minerals, there being a greater production of gypsum and anhydrite in areas of rapid dolomitisation.

It is quite probable that the diagenetic differences between the wide Trucial Coast sabkhas and the more restricted sabkhas of the Qatar Peninsula, studied by Illing and Wells (1964), are partially dependent on initial sediment differences, in addition to differences in extent. Illing and Wells report brine concentrations as high as those of the Trucial Coast, yet ultimate m_{Mg}/m_{Ca} ratios are higher and dolomite development more restricted, indicating that the process of dolomitisation is relatively retarded and more restricted. Celestite has not been reported from the Qater area but must occur. Anhydrite has not yet been reported but its absence is reasonably accounted for by retardation of the diagenetic processes. If anhydrite does occur in the Qatar area, the Trucial Coast data indicate that it will be a marginal development.

No other area so far described, (except possibly Laguna Ojo de Liebre: Phlegar and Ewing, 1963) would seem to offer a similar environment to that of the Trucial Coast, the high temperatures and rapid rates of brine concentration bringing about rapid rates of diagenetic reactions and being responsible for anhydrite precipitation.

The sedimentary and diagenetic processes described in this study have almost certainly operated in many ancient barrier and lagoonal environments. The developments along the Trucial Coast are on a scale which is geologically significant, particularly when one considers that they have probably resulted only during the past 4000 years.

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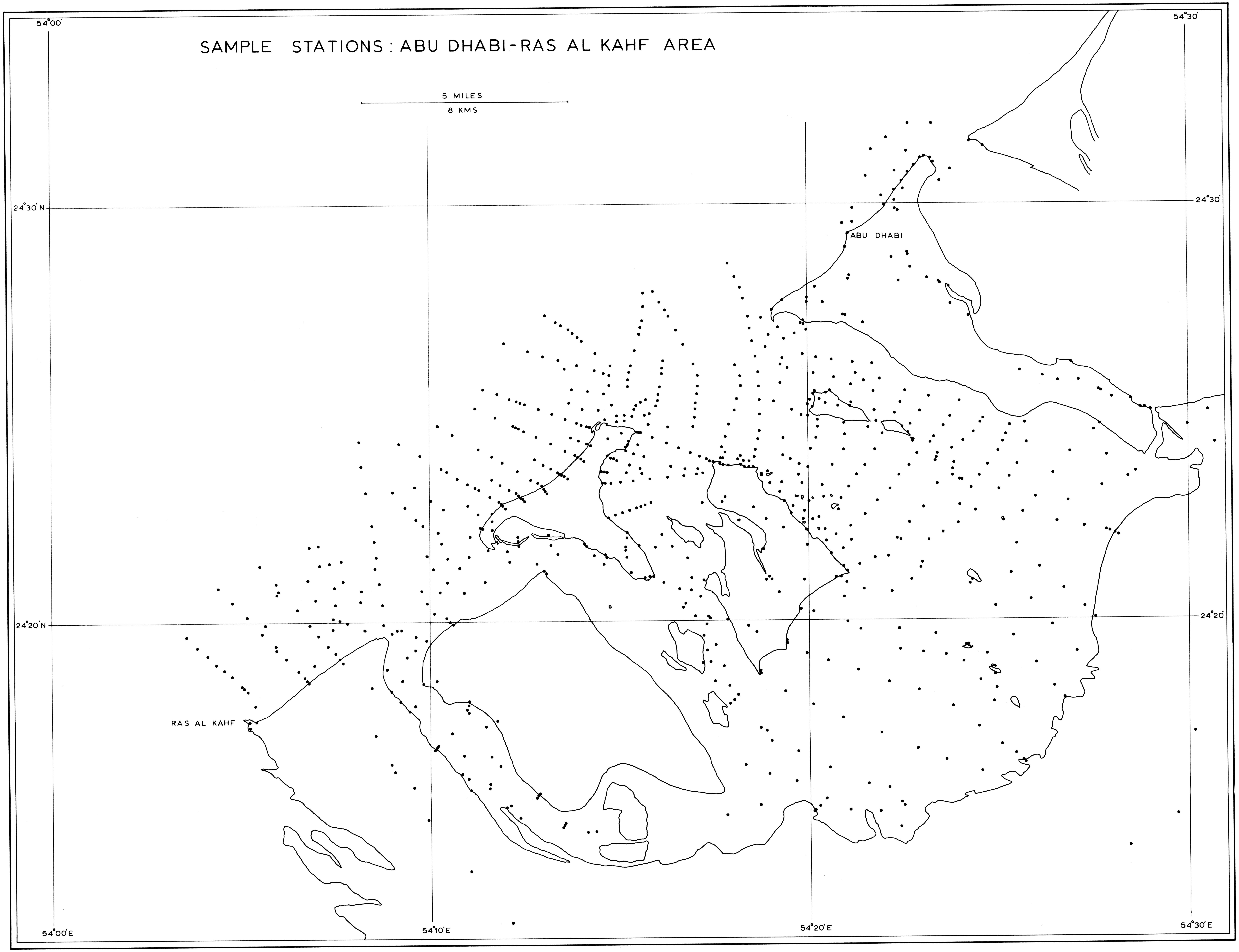
APPENDIX

Map of Abu Dhabi-Ras al Kahf area showing sample locations.


SAMPLE STATIONS : ABU DHABI-RAS AL KAHF AREA
5 MILES
8 KMS
54°00′
54°30′
24°30′N
24°30′
24°20′N
24°20′
ABU DHABI
RAS AL KAHF
54°00′E
54°10′E
54°20′E
54°30′E