THE ORIGIN, NATURE AND DISTRIBUTION OF

GYPSUM CRUSTS IN DESERTS

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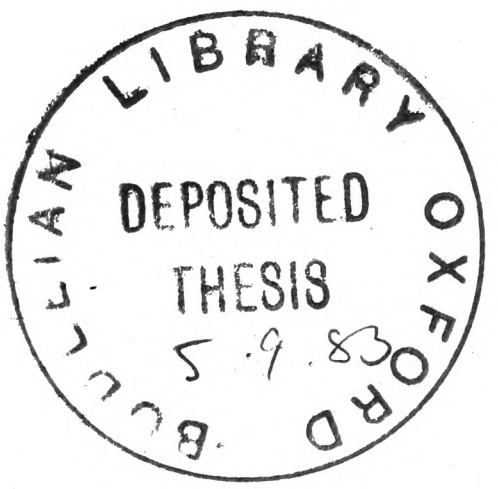

A B S T R A C T

The origin, nature and distribution of gypsum crusts in deserts.

Submitted Michaelmas Term, 1982, by Andrew Watson, Wolfson College, University of Oxford, for the Degree of D. Phil.

All the warm deserts of the world exhibit gypsum crusts in favourable localities, generally areas with a source of gypsum and less than 250 mm of rainfall annually. The features, comprising loose powdery or cemented crystalline accumulations of calcium sulphate dihydrate, are found at the surface or within the uppermost 10 m of regolith. Thicknesses vary from a few millimetres to several metres and purities range from about 15% to nearly 100% gypsum.

A review of the literature on gypsum soils and crusts reveals that the classification of types requires both standardizing and simplifying. A preliminary classification of crusts into three types can be made on the basis of structural and stratigraphic characteristics of examples from southern Tunisia and the Central Namib Desert. The types consist of subaerial aqueous evaporites, two subsurface and three surface forms. Detailed physical and chemical analyses justify a broadly similar genetic classification though one subsurface form is believed to be a primary pedogenic type of which the surface forms represent relics at various stages of solutional deterioration.

Examples of subaerial aqueous evaporites are all products of shallow-water sedimentation in seasonally flooded chotts and sabkhas. The non-pedogenic subsurface type, croûte de nappe, is produced by displacive crystallization at the surface of a fluctuating water table. Gypsum precipitation results from either evaporation or mixing of saline waters causing saturation with respect to gypsum. Surface gypsum crusts are subsurface illuvial accretions exposed by erosion of overlying material. The gypsum is derived from solution of surface aeolian, atmospheric, colluvial or alluvial deposits by meteoric waters which percolate into the upper soil zone replenishing the antecedent soil moisture deficit. Subsurface accumulation results from displacive crystallization at host sediment grain contacts when gypsum saturated soil moisture evaporates.

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INTRODUCTION

The examination of the origin, nature and distribution of any geomorphic feature necessarily involves the combination of studies of process and form in a spatial context. While gypsum crusts have been recorded in all the warm deserts of the Earth, information regarding their structural characteristics and genesis is scarce. For this reason, the approach adopted here is to examine the global and regional distribution patterns of the crusts before describing the physical and chemical nature of examples analysed by other workers and of a wide range of types and forms identified in two field areas. These areas are southern Tunisia, in the vicinity of the Chotts el Rharsa, Djerid and el Fedjedj, and the Central Namib Desert near Walvis Bay. On the basis of these examinations the various hypotheses that have been proposed to account for the formation of gypsum crusts will be evaluated. Finally, a series of models are outlined which describe the processes involved in the genesis of the various types of gypsum crust, particular emphasis being placed on the most common type which is also the most enigmatic. While the models are extensions and syntheses of several models proposed by other workers, they incorporate modifications which replace erroneous concepts employed in the earlier designs. Many of these shortcomings have become apparent only as a result of the present survey.

The aim of this study is to refine our understanding of the processes which result in gypsum crust formation and evolution through an analysis of their geography, stratigraphy, structure and chemistry. The significance of this objective lies in the fact that gypsum crusts are not only extensive features in terms of their distribution but are also of great relevance to the geology, geomorphology and pedology of arid areas.

Gypsum crusts are natural accumulations of calcium sulphate dihydrate at the

surface of the Earth or at a depth of less than 10 m within the regolith. The term "crust" is ambiguous in that they are not necessarily indurated horizons. They may, for example, consist of powdery deposits over 1.0 m thick which mantle a desert landscape. Thicknesses may also vary considerably, some crusts merely a few millimetres thick play an important role in stabilizing sand dunes through superficial cementation. Crusts up to 5.0 m thick have still more pronounced effects on the landscape, preserving relic topography by resisting erosional processes. The gypsum content may be between 15% and nearly 100%. The crystalline structure ranges from minute, almost cryptocrystalline, particles forming an alabastrine mass to large disc-shaped crystals up to 0.20 m or 0.30 m in diameter which twin and interlock to form desert roses.

In part, this physical and chemical diversity is related to the mineralogical and crystallographic characteristics of gypsum. For this reason, a preliminary account of these properties is warranted. In chapter one some of the most important characteristics of gypsum ($CaSO_4.2H_2O$) and the related minerals hemihydrate ($CaSO_4.\frac{1}{2}H_2O$) and anhydrite ($CaSO_4$) are examined. The great variability in their tendencies to dissolve and recrystallize and to dehydrate and hydrate under different chemical conditions must be rationalized on at least a simple level. Having achieved this it is possible to formulate genetic models based on the theoretical effect soil and groundwater chemistry has on gypsum crust form and structure. In chapter two, the observations on the chemical and crystallographic characteristics of the calcium sulphate minerals are employed to explain how they are deposited naturally and how the sediments evolve in different geochemical environments. While some of these genetic and diagenetic processes are not directly involved in the formation of gypsum crusts, they provide insights into the nature of the source materials for many crusts. Furthermore, the techniques of petrographic analysis developed by geologists in their examinations of bed rock gypsum deposits are a valuable asset in this study. Gypsum crusts have long regarded by many sedimentologists

and geomorphologists as spatially insignificant and temporally ephemeral, being, in most cases, in the domain of the pedologist.

This attitude is unjustified. Their geomorphic significance has already been alluded to. They are also far from restricted in their distribution over the globe. While they are limited to areas with generally less than 250 mm of rainfall annually and where potential evaporation exceeds actual precipitation throughout the year on a monthly basis, it should be pointed out that about 22% of the land surface of the Earth lies within these bounds. The scarcity of reports of gypsum crusts of other than littoral or lacustrine origin in the geological record may reflect misinterpretation of the origin of gypsiferous strata rather than their absence. The spatial distribution of gypsum crusts is closely linked to sources of gypsum hence many climatically suitable zones are devoid of them owing to the lack of source material. In chapter three the areal extent of crusts and their regional distributional characteristics are described. It is apparent that in any single area exhibiting crusts there may be several sources or merely a solitary source of gypsum. This trait is well exemplified when the two field areas are compared. In chapter four, the areas are described in detail and the climatic, geologic, geomorphic, pedologic and botanic differences outlines. In southern Tunisia, gypsum is present in bedrock, in groundwater, in coastal and continental sabkhas and in colluvial, fluvial and aeolian sediments. In contrast, none of these potential sources appears important in the Central Namib Desert where the gypsum forming the crusts may be authigenic. The diversity of environments represented by the two field areas has proved to be of great value in the understanding of the processes involved in crust genesis. Despite the environmental contrasts, the gypsum crusts in the two areas may be subdivided according to the same system and all the sub-groups are represented in both field regions.

The subdivision of gypsum crusts adopted here identifies three main types according to a classification based on their geomorphic and stratigraphic occurrence and large scale structural characteristics. Within this framework a detailed examination of the field associations between crust forms and other sediments is undertaken in chapter five. Micropetrographic analyses enable refining of the structural classification and some significant observations on the genetic differences among the types and forms. Additional information is provided by chemical analyses of the gypsum accumulations and stratigraphically or geomorphologically related deposits. On the basis of these detailed field and laboratory investigations it is possible to extend the preliminary physical classification to incorporate some genetic parameters. Furthermore, it can be shown that certain forms, even belonging to different types, have microstructural and chemical similarities indicative of genetic relationships among them. Hence much of the essential information required to facilitate a greater understanding of gypsum crust formation has been provided. The next step is to evaluate previous models in the light of these new data and, if necessary, modify them or construct new ones.

In chapters six and seven, the applicability of the existing models is assessed. They are subdivided into five broad groups, namely those incorporating subaerial aqueous evaporation; per ascensum, or upward movement of gypsum and subsequent crust accretion; lateral movement of source material; authigenic gypsum accumulation; and per descensum or net downward translocation and illuviation. The advantages and disadvantages of each of the models are evaluated in terms of their applicability to the various types and forms of crust which the foregoing analyses identified. In some cases, such as that of laminated evaporite sequences, the mode of formation is relatively straightforward. In others, none of the existing models appears to be able to explain crust genesis adequately. While they often help describe the mechanisms of primary formation, mobilization and transport of gypsum, most of the models

fail to provide detailed simulations of the processes of accretion. The genesis of the most common type of crust in both field areas is least well explained by the existing models. In chapter seven a model is formulated which fulfils the stipulations of the field evidence and the laboratory analyses. Of great significance is the fact that the models of genesis for the various types of crust are valid in the two field areas even though environmental conditions and the source of gypsum are markedly different.

Finally, in chapter eight, the implications of these genetic models are evaluated. They extend beyond the realm of gypsum crust evolution in that the models are relevant to other types of desert crusts and many aspects of desert geomorphology. Of particular significance are the implications the pedogenic model has on the extent and rate of aeolian sedimentation and erosion in the areas under consideration and, indeed, any areas where surface gypsum crusts are encountered. Furthermore, the study provides evidence of major subsurface displacive gypsum crystal growth and crust accretion. This in turn is of great importance in the understanding of the processes of salt weathering. The need to solve problems associated with aeolian processes and salt weathering is becoming increasingly urgent as desert regions experience economic development and the demands that it places on construction. Besides having to tackle these problems, engineers have to contend with the difficulties associated with volumetric fluctuations related to hydration and dehydration of calcium sulphate minerals. While previously these problems were encountered only in areas of anhydrite or gypsum bedrock, in many desert areas they have become almost ubiquitous owing to the presence of gypsum crusts. Similarly, the desire to extend agricultural production into marginal areas has prompted the exploitation of land previously regarded as unfavourable owing to the presence of soil crusts.

The study of the origin, the nature and distribution of gypsum crusts

presented here is largely limited to an examination of examples from two contrasting field areas. Though the models of crust genesis that are proposed are improvements on previous examples many questions remain unanswered. Two important areas open to future research into the origin of gypsum crusts are laboratory and mathematical simulation of crust genesis and accretion. In addition, studies of the isotopic characteristics of the crust forming minerals offer great potential for the interpretation of the origins of gypsum crusts.

CHAPTER 1

CALCIUM SULPHATE CHEMISTRY

a. Introduction.

In order to understand the processes involved in the formation of natural gypsum deposits it is essential that the complexities of the chemistry of calcium sulphate are acknowledged. This applies not only to the interpretation of the origin and post-depositional alteration of thick evaporite sequences containing gypsum and anhydrite but also to the development of gypsum crusts and gypsiferous soils. The formation of gypsum karst, the development of patterned ground on gypsum crusts and the failures associated with many large engineering projects on gypsum-rich sediments can be successfully investigated only when the chemistry of the whole family of calcium sulphate minerals is understood in at least a basic form. The importance of this hinges on two main factors, the variability in the properties of the different minerals in the system $CaSO_4$ - H_2O and the sensitivity of the different mineral phases to environmental changes associated with temperature, water availability and the presence of other minerals.

The chemistry of gypsum, and its dehydration products, has been studied for over two hundred years. The aims and methods of these studies have varied greatly and the observation that "an amount of confusion exists which is almost without parallel in organic chemistry" (Davis, 1907) in some ways still holds true today. It is worthwhile attempting to reassess the results of early workers in the field, particularly in the late nineteenth century. Much purely experimental work was undertaken which was uncoloured by the conclusions reached by later workers who employed theoretical thermodynamic principles which were occasionally erroneous. To this end the following discussions of the binary

system $CaSO_4$ - H_2O, of the solubility of the different minerals in the system and of the effects of other salts in ternary and more complex systems will make reference to some of these early analyses. Studies of solution rates, crystal growth and isotopes have been undertaken only in the last two decades but the chemical grounding is long, if not very firmly, established.

b. The binary system $CaSO_4$ - H_2O.

Gypsum (from the Greek gypsos meaning chalk) is the dihydrous form of calcium sulphate ($CaSO_4.2H_2O$) which occurs widely in nature in various forms, including selenite (clear crystalline), alabaster (massive opaque to semi-transparent) and satinspar (fibrous crystal masses). It has a hardness of 1.5 to 2, hence is easily scratched with the finger nail, and has a specific gravity of 2.3 (Read, 1970, p.281). Despite the fact that gypsum has long been used to produce building plaster, the chemical changes involved in the production of Plaster of Paris by heating gypsum and the setting of the plaster were not understood until the early twentieth century. Although Lavoisier (1765, 1766) appreciated that plaster setting involved rehydration and gypsum crystal growth, it was assumed that unset plaster was anhydrous calcium sulphate. Not until 1838 was Johnston able to show that there was a third distinct mineral in the calcium sulphate-water system. This was a partially hydrated form, the chemical formula of which was reported as either $2CaSO_4.H_2O$ or $CaSO_4.\frac{1}{2}H_2O$, hence the name hemidryate. Further studies revealed that there were different forms of the three hydration states.

The difficulty in identification of the different forms of calcium sulphate may account for some of the apparently anomalous results of experiments conducted on the mineral transitions and solubilities. Early workers generally relied upon chemical methods, such as the amount of water of crystallization or solubility, or crystallography for determining the particular form. These

methods have their shortcomings and to some extent even recent studies based on X-ray diffraction and X-ray crystallographic techniques are not ideal. This summary of the information available on the different minerals in the binary system $CaSO_4$ - H_2O will deal with the three main groups, dihydrate, hemihydrate and anhydrite, separately.

$CaSO_4.2H_2O$ (gypsum)

A number of early workers suggested that there are two forms of gypsum. Davis (1907) and Armstrong (1907), for example, felt that the peak in the solubility curve of gypsum at 38°C represents a change from a monoclinic to an orthorhombic crystal form. Recent crystallographic studies do not support this. The dihydrate crystal structure is monoclinic, C^2/c space group (Nancollas, Reddy and Tsai, 1973) though crystal habits are variable depending on temperature, pressure and the presence of additional ions during the crystal growth (Lambert and Schaffer, 1926). Wooster (1936) used ionization spectrometry to determine the molecular arrangement of this crystal structure. Calcium atoms are surrounded by six sulphate oxygen atoms and two water molecules, while the sulphur atoms are at the centre of sulphate oxygen tetrahedra. The water molecules have two oxygen and one calcium as their first neighbours. This layer-lattice structure accounts for the perfect (010) cleavage of gypsum crystals, parallel to the clinopinacoid, the planes of weakness being the double layers of water molecules separating the calcium and sulphate layers (Atoji, 1959). The implications of this structure are dealt with below.

$CaSO_4.\frac{1}{2}H_2O$ (hemihydrate)

Though Johnston (1838) and Johnson (1848) identified crystals found in the boilers of steam engines as hemihydrate, opinions on whether it exists as a distinct mineral with the formula $CaSO_4.\frac{1}{2}H_2O$ have differed widely.

De Forcrand (1906) suggested there may be three forms with the formulae $2CaSO_4.H_2O$, $4CaSO4.2H_2O$ and $6CaSO_4.3H_2O$ while, in contrast, Shenstone and Cundall (1888), Vater (1900) and Melcher (1910) held that gypsum is the only true hydrate of calcium sulphate. Van 't Hoff and a number of co-workers were the first to attempt to determine the relationship between gypsum and hemihydrate experimentally. It was shown that gypsum loses some of the water of crystallization at 101.45°C in air and 107°C in water to form the hemihydrate Van 't Hoff and Armstrong, 1900; Van 't Hoff and Weigert, 1901; and Van 't Hoff et al., 1903). These experiments conducted by determining the partial pressure of the water of crystallization using dilatometer measurements, as well as thermodynamic treatment of Boyer-Guillon solubility data, were criticized by later workers (Davis, 1907; Patridge and White, 1929) and Armstrong (1907) conceded that hemihydrate and anhydrite may have been confused.

Despite these criticisms most later workers accepted the existence of the hemihydrate phase, although a lower gypsum-hemihydrate transition temperature was proposed between 96° and 102°C (Partridge and White, 1929; Ramsdell and Partridge, 1929; Hall, Robb and Coleman, 1926; Posnjak, 1938). Only D'Ans and Höfer (1937) upheld the 107°C transition point.

Confusion about the crystal structure of hemihydrate is also far from resolved. Linck and Jung (1924), Jung (1925) and Caspari (1934) supported the views of Shenstone and Cundall (1888) when suggesting that the water of crystallization present in hemihydrate is zeolitic. This implies that there is not a true crystal form corresponding to the formula $CaSO_4.\frac{1}{2}H_2O$ but rather a substance $CaSO_4.\ n\ H_2O$ where $0.5 > n > 0$ and hemihydrate may be regarded as a form of anhydrous calcium sulphate. Weiser, Milligan and Ekholm (1936) used X-ray diffraction to show that hemihydrate is not a true zeolite since water occupies a definite place in the crystal lattice. Though Davis (1907) suggested that hemihydrate is dimorphous, he did not expand on this. Kelley, Southard

and Anderson (1941) proposed that α and β $CaSO_4 . \frac{1}{2}H_2O$ exist, the first being formed in the normal process of gypsum dehydration and the other by partial dehydration of gypsum in a vacuum at 100°C. The β form was suggested to be more soluble and less stable than the α form. There is little supporting evidence for this subdivision. MacDonald (1953) and Ostroff (1964) accepted the dimorphism but did not provide any experimental justification for this.

Power, Fabuss and Satterfield (1964) found that the α-form crystallite has an average size of 0.3 um and the β-form 0.09 um. They suggested that the different temperatures of the gypsum-hemihydrate transition in air and water (Van 't Hoff et al., 1903) result from the existence of two forms of hemihydrate. In contrast, the crystallographic studies of Flörke (1952) and Gay (1965 b.) showed that there is only one hemihydrate. It is likely that the confusion has arisen from misidentification of the mineral produced by dehydration and also the fact that the dehydration temperature of hemihydrate to anhydrite tends to be lower each time if the same crystals are repeatedly rehydrated and dehydrated. Anhydrous calcium sulphate may be formed at 100°C in some cases (Gay, 1965 b.) and this is within the gypsum-hemihydrate transition temperature range.

Ramsdell and Partridge (1929) held that hemihydrate does not occur naturally except perhaps associated with volcanic activity. There are, however, a number of reports of hemihydrate from hot desert areas, usually associated with evaporite deposits (Popov and Vorobiev, 1947 in Central Asia; Allen and Kramer, 1953; Moiola and Glover, 1965, and Hunt et al., 1966 in the south-western United States; and Butler, 1969, Kendall and Skipwith, 1969, Kinsman, 1969 and Amiel and Friedman, 1971 in the Middle East; and Nagtegaal, 1973 in the Central Namib Desert). Hemihydrate has a hexagonal crystal structure, a $P\bar{3}m1$ space group and a specific gravity of 2.60 (Nancollas, Reddy and Tsai, 1973). Usually occurring as fine acicular crystals, it may exist as masses forming pseudomorphs

after gypsum. The fibres are parallel to the fossil gypsum cleavage (010) and are therefore a product of the weak structure of the H^+ component of the gypsum water of crystallization (Atoji, 1959).

$\underline{CaSO_4}$ (anhydrite)

Anhydrous calcium sulphate, anhydrite, is a common naturally occurring mineral, frequently found in association with gypsum in evaporite deposits. It occurs as either fibrous, lamellar, granular or compact rock usually white but sometimes with a grey, bluish or reddish tint (Read, 1970, p.279). It has a hardness of 3 to 3.5. Natural anhydrite has an orthorhombic crystal structure, an A_{mma} space group, and a specific gravity of 2.86 (Nancollas, Reddy and Tsai, 1973).

The confusion associated with anhydrite chemistry stems from the difficulty in differentiating the products of gypsum dehydration once this has reached the stage at which hemihydrate dehydration commences. The most important distinction is that of solubility. Potilizin (1894 a.b., 1985) recognized that natural anhydrite was almost insoluble in water as was gypsum heated in air to over 200°C. This material is termed insoluble anhydrite or β -$CaSO_4$. When gypsum is heated to between about 120° and 200°C an anhydrous material is formed which is soluble in water; it is also strongly hygroscopic and will rehydrate to hemihydrate below 120°C (Cloez, 1903; Hammond and Withrow, 1933; Bowers, 1934; Posnjak, 1938) if any moisture is available. This is soluble anhydrite or γ-$CaSO_4$. Potilizin termed this α-$CaSO_4$ but here this designation is applied to an enantiotropic modification at 1194°C (Grahmann, 1914). The existence of this was disputed by Gay (1965 a.) who found that β-$CaSO_4$ decomposed at 1150°C as a result of SO_3 loss. Following Potilizin's identification of soluble anhydrite, Lacroix (1898 a.b.) stated that the mineral was present between 125° and 250°C and Van 't Hoff, Hinrichsen and Weigert (1901)

held that it may form directly from gypsum at temperatures as low as 89°C. All other workers have found that soluble anhydrite forms only from hemihydrate and always rehydrates to hemihydrate (Gaudefroy, 1914 a.). A number of workers (Haddon and Brown, 1924; Linck and Jung, 1924; Biltz, 1925; Jung, 1925; Partridge and White, 1929; Ramsdell and Partidge, 1929) suggested that the soluble anhydrite was merely zeolitically dehydrated hemihydrate and not a distinct mineralogical phase. Jolibois and Lefebre (1923) stated that hemihydrate was stable up to 160°C. X-ray diffraction and crystallographic analyses by Flörke (1952) and Gay (1965 a.) positively identified the soluble anhydrite that many workers had earlier assumed existed (Gaudefroy, 1914 b.; Balarew, 1926, 1927). These studies also showed that there is only one crystallographic polymorph of γ-$CaSO_4$. The suggestion had been put forward by Balarew (1927), Kelley, Southard and Anderson (1941) and Power, Fabuss and Satterfield (1964) that either there were two forms of soluble anhydrite at two different temperatures or that two forms of hemihydrate dehydrated to two types of soluble anhydrite. The different heats of hydration to α- and β-$CaSO_4.\frac{1}{2}H_2O$ were believed to result from the presence of at least 0.9% by weight of zeolitic water in β-$CaSO_4.\frac{1}{2}H_2O$. This water would be lost only at high temperatures to form $CaSO_4$ (β-soluble) which on cooling would similarly rehydrate to β-$CaSO_4.\frac{1}{2}H_2O$ at a high temperature.

Some early workers referred to a number of other forms of anhydrous sulphate (Kraut, 1865; Rohland, 1903, 1910) The justification for differentiating the forms appears to stem from subdivision of the β-$CaSO_4$ (insoluble) and γ-$CaSO_4$ (soluble) according to different modes of formation. Rohland (1910) and Van 't Hoff et al. (1903) also held that natural anhydrite and "dead burnt gypsum", that is insoluble anhydrite, are two distinct forms. There is no justification for this. Insolube anhydrite has the same crystallographic pattern regardless of its mode of origin (Gay, 1965 a.). Posnjak (1938) and Gay (1965 b.) pointed out that although insoluble anhydrite usually forms

above 200°C it is detectable even at 150°C and there is not a discontinuity on the heating curve. Hence the transition of γ- to β-$CaSO_4$ appears to be gradual and they may be monotropic.

In effect, in the system $CaSO_4$-H_2O gypsum will dehydrate to hemihydrate at about 95° - 105°C depending on the medium in which the process takes place, between 100° and 120°C hemihydrate will lose most of its water of crystallization to form soluble anhydrite which itself decomposes to insoluble anhydrite between 150° and 250°C. The dimorphism of hemihydrate and soluble anhydrite is not well established, nor is the existence of the high temperatures polymorph α-$CaSO_4$. The mechanisms involved in the transitions are, however, not well understood. The complexity is epitomized by the fact that insoluble anhydrite (β-$CaSO_4$) appears to be chemically stable at temperatures well below 100°C, perhaps 40°C, while gypsum does not dehydrate completely to insoluble anhydrite until temperatures reach about 200°C (see table 1.1).

Phase transitions

Experimental determination of the transition point between any two solid phases in the $CaSO_4$-H_2O system is complicated by very slow reaction times, particularly of insoluble anhydrite. This applies not only to the rate of change from one mineral to another but also to attainment of solubility equilibria in aqueous and saline solutions. Zen (1965) held that the point of maximum solubility must be approached from over-saturation as well as under-saturation. Below 70°C it may be impossible to determine anhydrite solubility because these two figures do not correspond even when the solutions are left for three months. Other complications arise because the rate of dissolution of gypsum is controlled by film diffusion at the crystal surface (Liu and Nancollas, 1971). Hence the increased solubilities in agitated solutions and finely powdered samples (Hulett, 1901; Hulett and Allen, 1902; Dundon and

TABLE 1.1

MINERALS IN THE SYSTEM $CaSO_4$ - H_2O

MINERAL	FORMULA		STABILITY RANGE (^oC)
GYPSUM	$CaSO_4.2H_2O$		Up to between 38 to 42 or 56 ± 3 to 66 (in water)
			Metastable up to between 95 and 103 (air)
			Metastable up to 107 (water)
HEMIHYDRATE	$CaSO_4.\frac{1}{2}H_2O$	α	Metastable between 95 to 103 (air), 107 (water) and 120 to 160.
		β	Metastable between 100 and 250 (?)
ANHYDRITE	$CaSO_4$	γ (α-soluble)	Metastable between 100 to 125 and 150 to 250
		γ (β-soluble)	Metastable between 250 (?) and ?
		β(insoluble)	Metastable up to between 38 to 42 or 56 ± 3 to 66
			Stable between 38 to 42 or 56 ± 3 to 66 and 1194 (?)
			Transition from γ(soluble)-$CaSO_4$ between 150 and 250
		α	Stable (?) above 1194 (?)

Mack, 1923; Haddon and Brown, 1924). Various approaches have been adopted to overcome these problems. All essentially involve thermodynamic treatment of high temperatures or pressure solubility data to enable extrapolation into the low temperature (0 to 100°C) régime.

Van 't Hoff's experimental work on evaporite minerals included the first detailed analysis of mineral transitions in the $CaSO_4$ - H_2O system. Though Armstrong (1901) had pointed out that calcium sulphate is present in seawater in relatively small amounts compared to even the bittern salts, Van 't Hoff's work was critical in that it stressed the importance of understanding complex mineral assemblages in geochemistry. The experimental work involved the determination of dissociation vapour pressures using dilatometers and thermodynamic treatment of Boyer-Guillon solubility data. The findings have been attacked on the grounds that the theoretical considerations employed to overcome the slow reaction times were incorrect. Van 't Hoff assumed that the dissociation pressure would be equal to vapour pressure at the transition temperature. Hence, "by applying a thermodynamic formula it was then a simple matter to calculate from two determinations the dissociation pressure over the whole range of temperatures" (Posnjak, 1938). It was suggested that gypsum can dissociate in three different ways at three different pressures: to hemihydrate Van 't Hoff and Armstrong, 1900), to soluble anhydrite Van 't Hoff, Hinrichsen and Weigert, 1901) and to insoluble (natural) anhydrite Van 't Hoff and Weigert, 1901). This is illogical. Posnjak (1938) stated that gypsum is stable as long as the partial water pressure of the atmosphere exceeds the pressure of dissociation. On heating, dissociation pressure eventually exceeds the partial water pressure of the atmosphere and dehydration commences. It is also possible, however, to lose water of crystallization by reducing water vapour pressures at a constant temperature. Dissociation resulting from a change in temperature or pressure, or both, can only occur in steps the order of which is constant, gypsum ⟷ hemihydrate ⟷ soluble anhydrite ⟶ insoluble anhydrite.

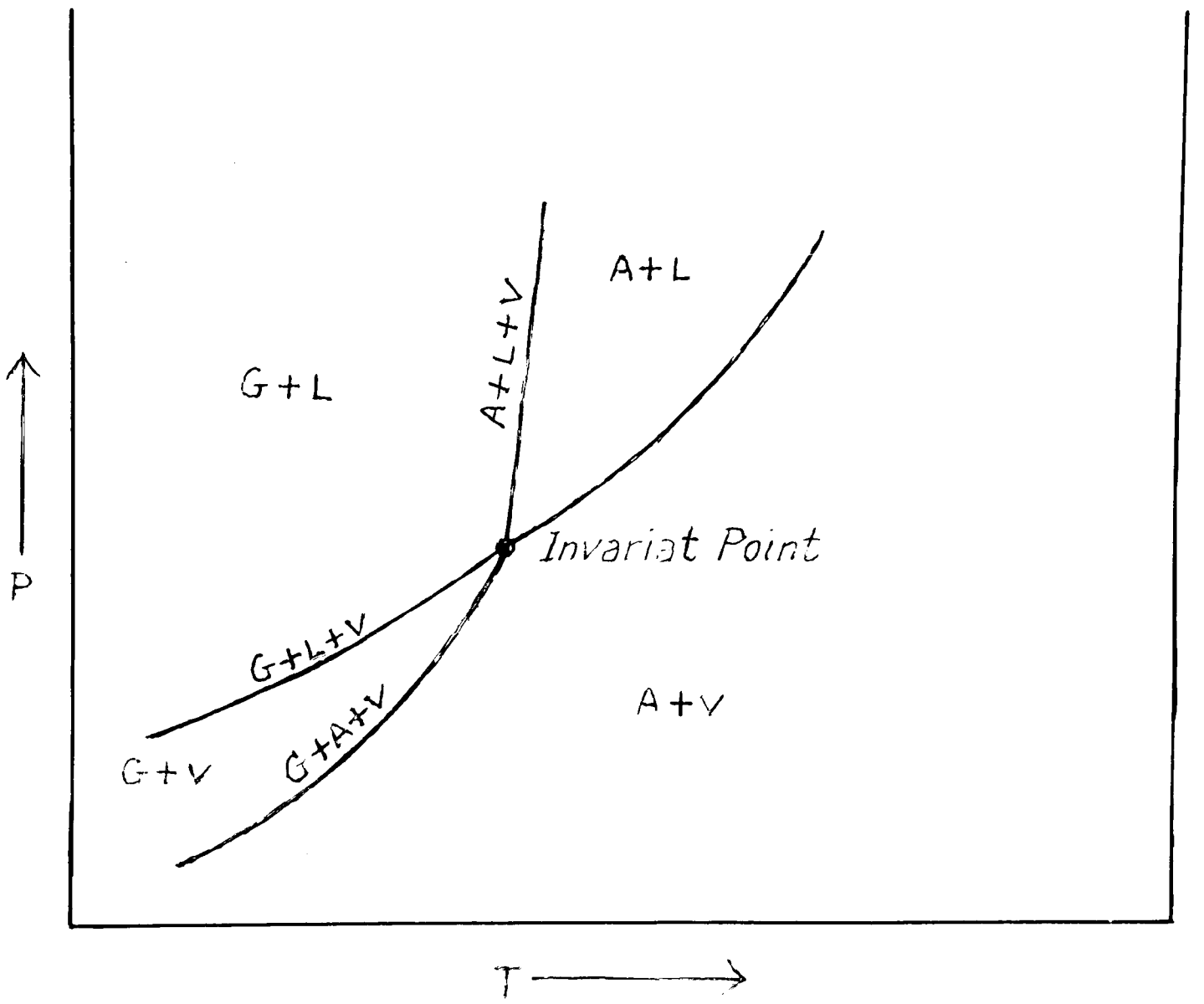
By applying the phase rule to the system $CaSO_4$ - H_2O the problem of transition is simplified. At any transition point in air there are two solid phases and a vapour phase. This is a three-phase system with one degree of freedom, temperature. Since,

$P + F = C + 2$ Equation (1)

(where (P) is the number of phases (F) the number of degrees of freedom and (C) the number of components) it follows that in air at any transition temperature a reaction occurs "representing a univariant two-component system in which one component is volatile" (Posnjak, 1938). For gypsum to dissociate to soluble anhydrite the intermediate hemihydrate phase would have to disappear, while dissociation of gypsum to insoluble anhydrite would require the disappearance of both the hemihydrate and soluble anhydrite phases. The process of hydration of insoluble anhydrite to gypsum must pass through the same steps in reverse order. This, however, does not occur. Insoluble anhydrite is stable certainly above 60°C and possibly as low as 40°C, not only above the soluble anhydrite - insoluble anhydrite transition temperature of about 200°C. Posnjak (1938) explained this in terms of two types of transition point in the system $CaSO_4$ - H_2O. In air dehydration from gypsum to insoluble anhydrite occurs by dissociation through the series of $CaSO_4$ mineral phases. In water, however, though it is possible to dehydrate gypsum to insoluble anhydrite through the intermediate phases (Kraut, 1865), these are all metastable. This is because as temperature rises the least soluble mineral will precipitate. Since the solubility curves of gypsum and insoluble anhydrite intersect between 40° and 60°C in water the transition point in the four-phase gypsum-insoluble anhydrite-liquid-vapour system lies in that range. There is not a true dissociation reaction since dehydration occurs through the medium of solution.

Although Hardie (1967) suggested that in the four-phase system dehydration might be by direct dehydration from gypsum to anhydrite as well as solution-precipitation or through the intermediate phases, this is not feasible.

Figure **1.1**



Schematic phase relations in the system
$CaSO_4$ - H_2O (after Blount and Dickson, 1973)

G, Gypsum
A, Anhydrite
L, H_2O solution
V, Vapour

Blount and Dickson (1973) made the same error in constructing their model of the four-phase system on the temperature and pressure plane (see fig. 1.1). The three-phase univariant equilibrium of gypsum-insoluble anhydrite-vapour at low pressures was constructed from thermochemical data obtained from experiments conducted on sodium chloride-water solutions. Their approach was based on the principle that a phase can be removed from a multi-phase system in equilibrium without destroying that equilibrium. In the case of the gypsum-insoluble anhydrite-NaCl saturated liquid-vapour system, the equilibrium cannot be maintained after removal of the NaCl solution since the liquid phase is the "catalyst" in the transition (Hardie, 1967). The two solids cannot form a univariant system with vapour because there are two intermediate reaction products. If these intermediate phases were metastable in the vapour phase, the model would be valid, assuming long reaction times, but there is no evidence for this metastability. In effect, transition from gypsum to anhydrite can take place in two ways, by dissociation through the intermediate phases in water, or directly through a solutional phase. Some characteristics of the system remain unexplained, however. Posnjak (1940), Zen (1962, 1965) and Hardie (1967) were unable to precipitate anhydrite from saturated solutions in the stability field. This is possibly a result of the very slow reaction times but another important fact was first noted by Van 't Hoff et al. (1903). They found that gypsum could be converted to insoluble anhydrite only when the gypsum was made from Plaster of Paris. This gypsum contains nuclei of anhydrite formed during dehydration above about 150°C. These act as the foci for anhydrite crystallization.

The rate law for the crystal growth reaction (Johnson and O'Rourke, 1954) has been applied to gypsum and anhydrite by Conley and Bundy (1958) and Nancollas, Reddy and Tsai (1973). It takes the general form,

$$-dM/dt = K_c^n (C - C_s)^q \qquad (2)$$

where (M) is the mass of calcium sulphate in solution at time (t), (C) is the concentration of calcium sulphate (monomers) in solution at time (t), (C_s) the solubility value or concentration of monomers at saturation, (K_c) the crystallization reaction-rate constant, (n) is proportional to the number of growth sites and (q) is a contant. Though the number of nuclei of growth is taken into account, the equation is only valid from the point in time when the number of nuclei is constant. Hence, when growth is initiated, all considerations of nucleation are overridden. The rate of nucleation is proportional to,

$$\left[(Ca^{2+})\ SO_4^{2-}\right]^{P/2} \qquad (3)$$

where (P) is the number of ion members required to sustain a crystallite (O'Rourke and Johnson, 1955). If (n) in equation (2) took into account the rate of insoluble anhydrite nucleation from the point in time when insoluble anhydrite becomes the stable phase, this nucleation would represent the anhydrite precipitation rate controlling step at temperatures between the theoretical gypsum - insoluble anhydrite transition point in solution and about 100°C, unless the gypsum present contained some anhydrite nuclei (Turnbull and Vonnegut, 1952). There is a low variation in free energy with temperature change above the gypsum-insoluble anhydrite-liquid transition point (Zen, 1965). This factor, combined with the high activation energy required in the formation of an insoluble anhydrite critical nucleus, precludes rapid nucleation at temperatures below about 100°C (Nancollas, Reddy and Tsai, 1973).

Rehydration of insoluble anhydrite to gypsum below 40° to 60°C requires the presence of gypsum nuclei. Though the reaction is still slow (Gelmroth, 1953; Einfalt and Götz, 1976) this appears to be because anhydrite dissolution rather than gypsum nucleation is the rate controlling step. Gill (1918) showed that while only 34.15% of a sample of anhydrite converted to gypsum over six years in water, 87.46% conversion occurred with anhydrite formed by heating gypsum

to 560°C. Apparently some gypsum nuclei remained, stimulating or facilitating rehydration.

The space group characteristics of the phases may help to account for the stepped dehydration pattern of gypsum to insoluble anhydrite in air. The solid state rearrangement $CaSO_4.2H_2O \rightleftharpoons CaSO_4.\frac{1}{2}H_2O \rightleftharpoons CaSO_4 \longrightarrow CaSO_4$ (insoluble), is not reversible once the completely anhydrous phase has formed, soluble anhydrite probably contains at least 0.9% water by weight (Kelley, Southard and Anderson, 1941) While soluble anhydrite rehydrates to hemihydrate in the solid phase, it has been suggested that hemihydrate will only rehydrate to gypsum through a solutional phase (Cloez, 1903; Jolibois and Chassevent, 1923) and that it is not hygroscopic (Davis, 1907; Hammond and Withrow, 1933) but this is not proven. The large activation energies associated with the rearrangements may preclude solid state rehydration of insoluble anhydrite. The differences between the symmetry and lattice parameters certainly rule out epitaxial growth of anhydrite on hydrate crystals (see table 1.2).

Table 1.2

Crystal structure parameters for gypsum, hemihydrate and insoluble anhydrite (after Nancollas, Reddy and Tsai, 1973).

	Class	Unit Cell Parameter a	b	c	Specific Gravity	Space Group
$CaSO_4.2H_2O$	Monoclinic	6.28	15.15	5.67	2.32	C^2/c
$CaSO_4.\frac{1}{2}H_2O$	Hexagonal	6.82	-	6.24	2.60	$P\bar{3}_1m$
$CaSO_4$ (insol.)	Orthohombic	6.95	6.96	6.21	2.86	A_{mma}

This helps to explain the absence of a gypsum-insoluble anhydrite-vapour equilibrium and also the slowness of direct dehydration through the solutional phase, since independent nucleation must occur. Studies of these mechanisms

of phase transition in the $CaSO_4$ - H_2O system are prone to kinetic difficulties, which result in one case from the high activation energies required for solid state crystal lattice rearrangements and in the other from low film diffusion controlled solubilities and slow nucleation. In air, gypsum dehydrates to hemihydrate at about 95° to about 100°C but may do so at lower temperatures (Shenstone and Cundall, 1888) if the atmospheric partial water vapour pressure falls below the partial pressure of the water of crystallization in the gypsum. The same is true of the hemihydrate-soluble anhydrite transition. This occurs at about 120°C but the exact temperature is indeterminate because transition to insoluble anhydrite commences at about 150°C, though it is not complete until temperatures reach between 200° and 300°C. In a four-phase system containing gypsum, insoluble anhydrite, water and vapour, the equilibrium point is determined by the intersection of the solubility curves. Insoluble anhydrite becomes less soluble than gypsum in pure water between 40°C and 60°C at one bar. Rapid heating in water will dehydrate gypsum through the intermediate phases to insoluble anhydrite, with each phase occurring in about the same temperature ranges as in air, but the other phases are all metastable above 40° to 60°C because anhydrite should eventually nucleate. It is possible for gypsum to be indefinitely metastable between 60° and 95°C in water if insoluble anhydrite nuclei are absent. Ostroff (1964) stated that in pure solutions gypsum will not dehydrate to anhydrite below 97°C.

c. Solubility of the solid phases in the $CaSO_4$ - H_2O system.

Accurate determination of the invariant point at which gypsum-insoluble anhydrite-water and vapour are in equilibrium on the pressure-temperature plane is only possible through analysis of the solubilities of the solid phases. This is because the transition point is where the solubility curves of gypsum and insoluble anhydrite intersect. In order to overcome the problems of slow reaction times many workers have used thermochemical techniques to extrapolate solubility

information from systems which are kinetically more active. Greater activity occurs at high temperatures and pressures and in the presence of some soluble salts (Blount and Dickson, 1973; Eugster, Harvie and Weare, 1980).

The mechanisms which bring this about are dealt with below. Here only the extrapolated results and determinations from pure water, at standard temperature and pressure will be considered.

The rate of dissolution of gypsum is controlled by film diffusion at the crystal-liquid interface. Gypsum crystals dissolve according to first order kinetics (Liu and Nancollas, 1971) so from equation (2):

$$dM/dt = K_S A\ (C_S - C) \qquad (4)$$

where (K_S) is the solution rate constant and (A) is the area of calcium sulphate exposed to the aqueous phase. In contrast, insoluble anhydrite dissolves according to second order kinetics (James and Lupton, 1978), so:

$$dM/dt = K_S A\ (C_S - C)^2 \qquad (5)$$

This dependence of the solution rate upon the square of subsaturation concentration is controlled by dislocation of ions from the crystal lattice. Roller's (1931) observation that the rate of solution of anhydrite particles increases 17.6 times the proportional increase in the surface area when particles are reduced in size from 25 um to 2.8 um, may result from an increased electrostatic charge with decreasing particle size facilitating ionic dislocation (see fig. 1.2). James and Lupton's (1978) experimental results are summarized in table 1.3 and figs. 1.3 and 1.4.

Figure 1.2

Rate of solution of different particle diameters of anhydrite. Ratio of relative rate of solution to relative specific surface against mean diameter (after Roller, 1931).

Figure 1.3

Dissolution rate of gypsum at 23°C. Flow velocity 0.30 m/s, volume in circulation 0.40 l (after James and Lupton, 1978).

Figure 1.4

Dissolution rate of anhydrite at 23°C. Flow velocity 0.09 m/s, volume in circulation 0.40 l (after James and Lupton, 1978).

Figure **1.2**

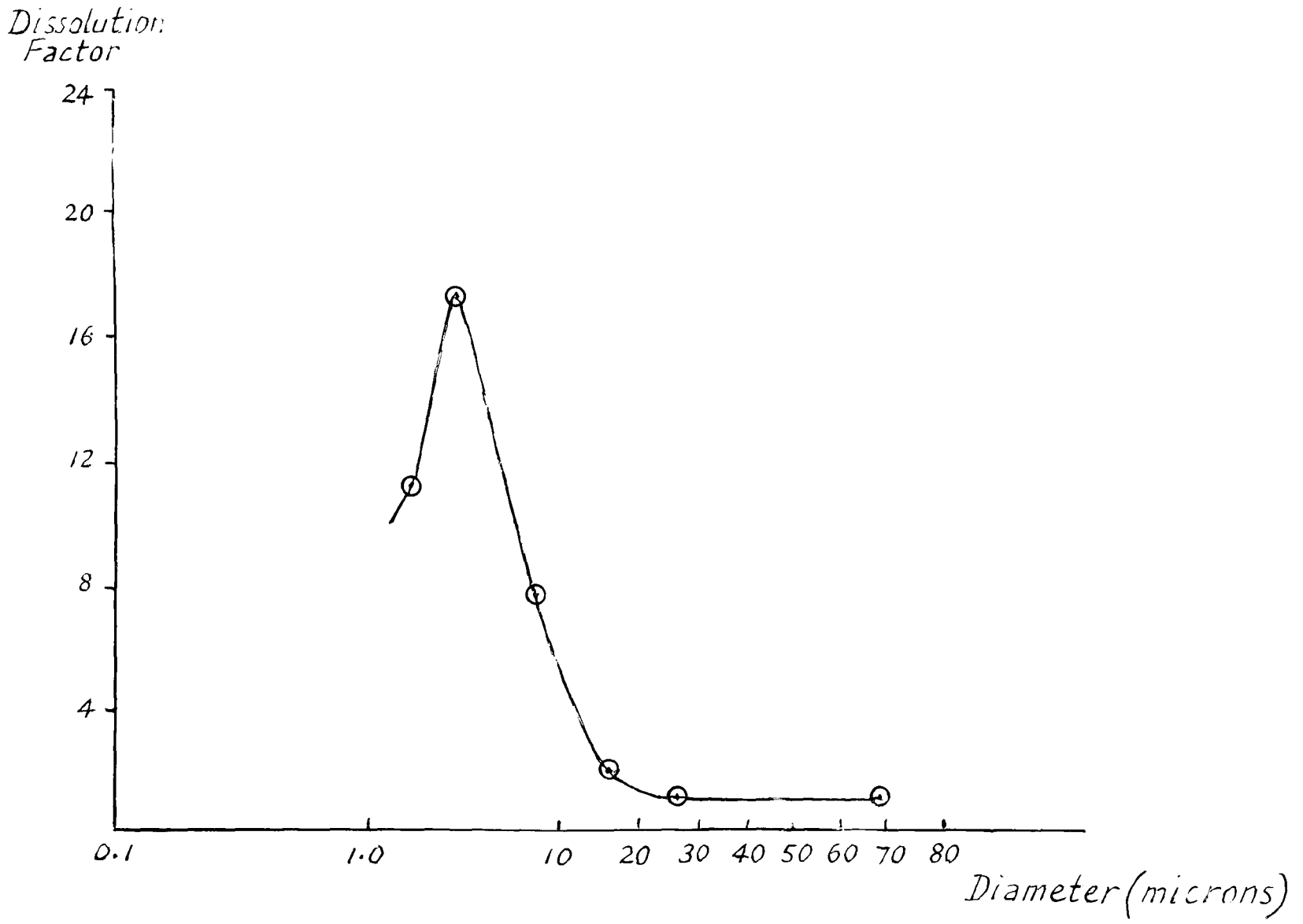

Dissolution Factor
24
20
16
12
8
4
0.1
1.0
10
20
30
40
50
60
70
80
Diameter (microns)

Figure **1.3**

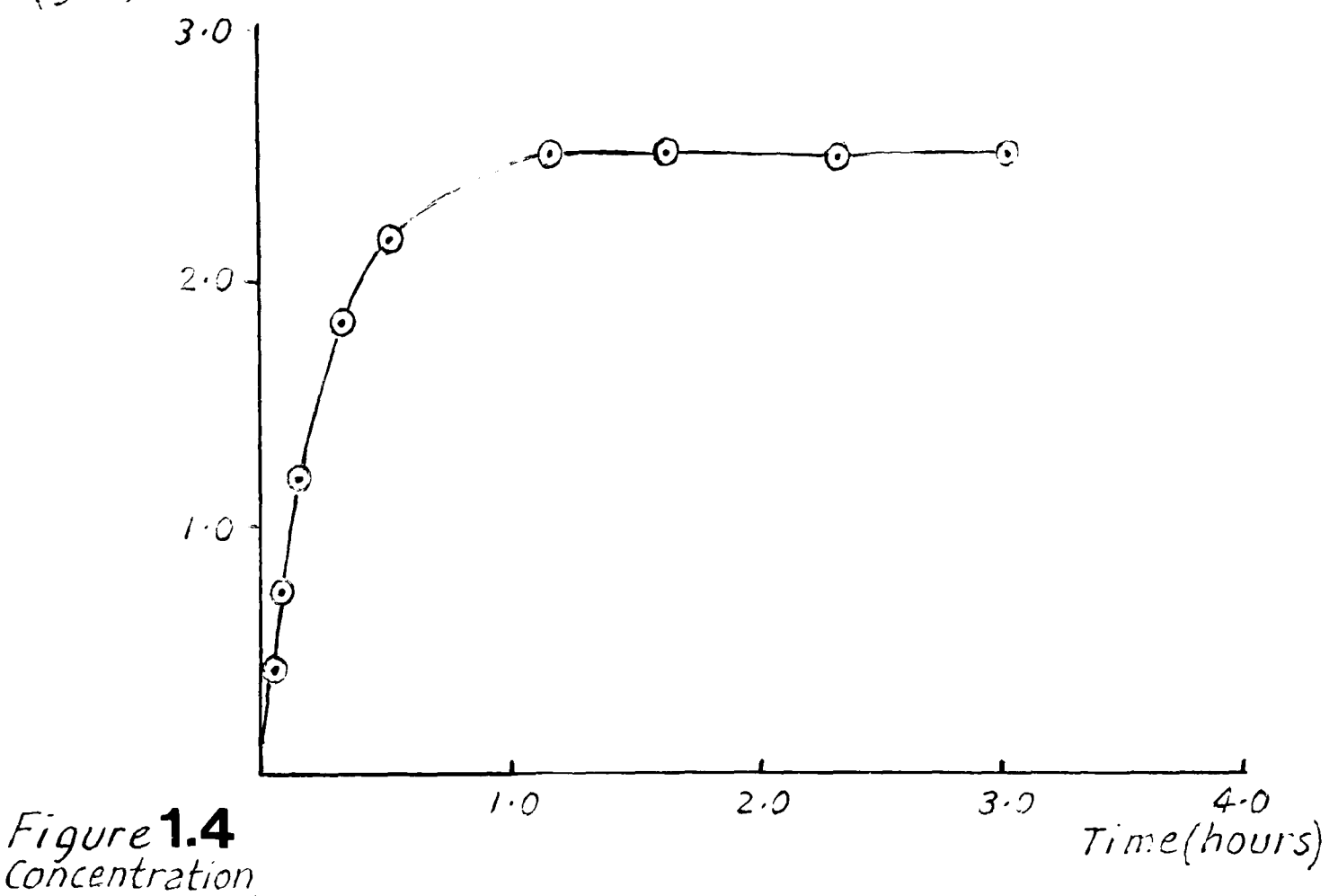


Figure **1.4**

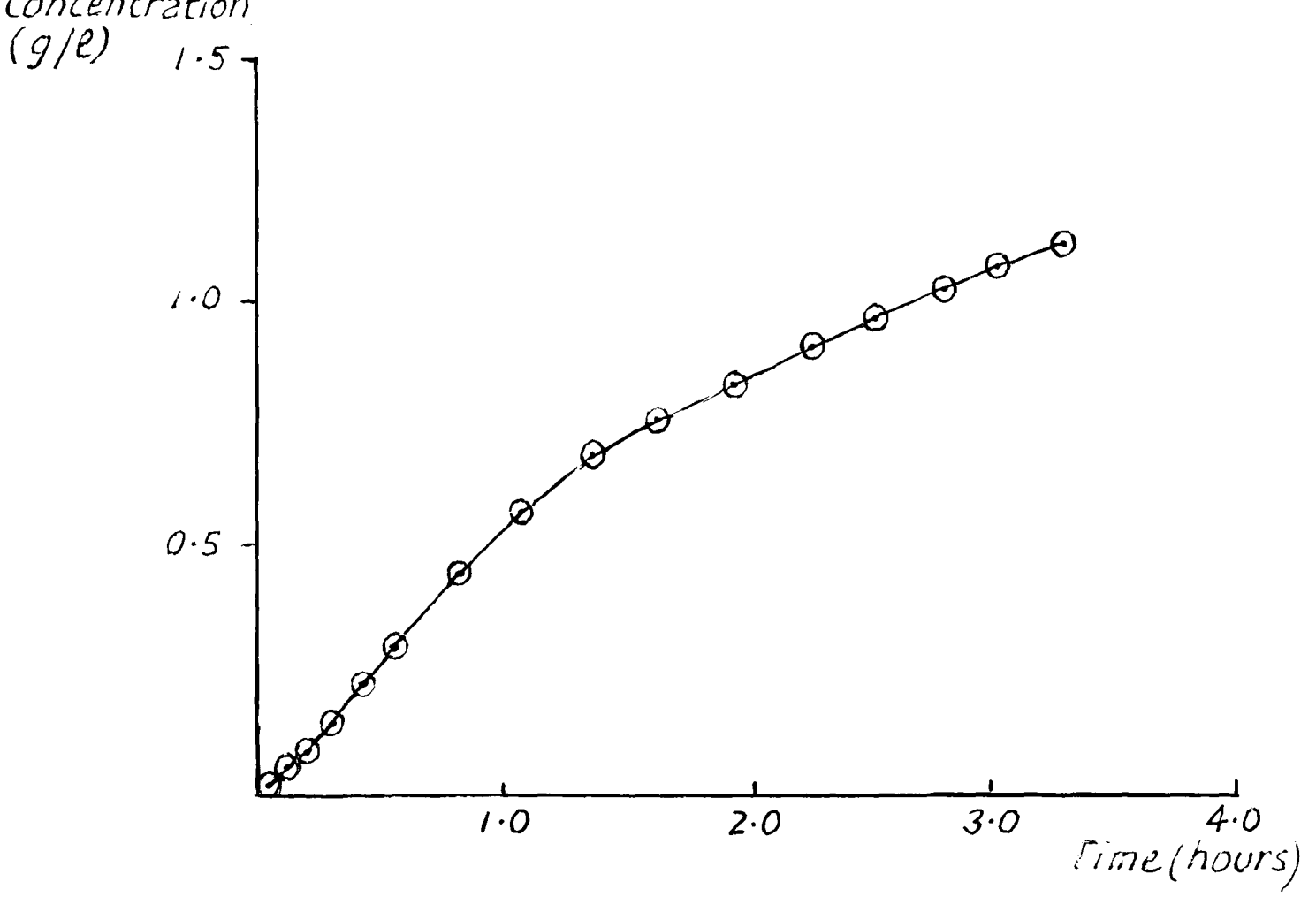


Table 1.3

Variation in dissolution rate constant of massive gypsum with temperature.

Temperature °C	Rate constant (m/s x 10^5)
5	0.8
15	1.7
23	2.6

In any closed system, dissolution rates are determined by the total volume of water available and the velocity at which it moves across the crystal surface as well as temperature and pressure. Most studies have concentrated on absolute solubility not the rates of solution, though a slow rate may cause erroneous results.

Three factors appear to affect the solubility of gypsum and its dehydration products in pure water. These are particle size, pressure and, most important, temperature.

There is no information available regarding the effect of particle size on hemihydrate and soluble anhydrite solubilities. Hulett (1901) and Hulett and Allen (1902) first showed that gypsum solubility could be increased up to 20% when grain size was reduced from 2 μm to 0.3 μm. Dundon and Mack (1923) suggested that this was the result of partial dehydration to the more soluble hemihydrate during powdering. Since later workers (Roller, 1931; Marshall, 1964) did not test for this dehydration their experiments are of little value. With regard to anhydrite only Roller (1931) conducted experiments to compare particle size and solubility but these were concerned with solution rates rather than total solubility.

The effect of pressure upon the solubility of calcium sulphate minerals has also received little attention in terms of experimentation. An understanding

of the effects of pressure is critical when examinating the gypsum-anhydrite transition which occurs when surface sediments are buried by further sedimentation or exhumed through surface erosion. Geologists have assumed either that pressure has a negligible effect on solubility in the $CaSO_4$ - H_2O system (Mossop and Shearman, 1973), or that high pressures will facilitate low temperature dehydration to anhydrite (Newland, 1921) or that high pressures counteract the dehydration effects of high temperatures (Bowles and Farnworth, 1925). Marsal (1952) showed that pressure has little effect on gypsum and anhydrite solubilities. Gypsum solubility doubles between 1 and 1000 bars at about 50°C while anhydrite solubility increases about three times over the same range (Blount and Dickson, 1973). It is interesting to note an empirical example of this. Adams (1936) found that drilling for oil in the anhydrite deposits of the Castile formation of west Texas was difficult because gypsum precipitated from groundwater around the drill tubing on reduction of pressure. The gypsum-insoluble anhydrite solution transition temperature will increase with pressure. The rate of increase has been determined as 1°C/83 bars (Marsal, 1952), 1°C/85.4 bars (MacDonald, 1953), 1°C/71$\pm$1.9 bars (Zen, 1965), and 1°C/78 bars (Blount and Dickson, 1973) (see fig. 1.5).

It has long been known that the maximum solubility of gypsum at 1 bar pressure lies between 35° and 45°C (Poggiale, 1843, Marignac, 1874; Goldammer, 1888). Determination of the precise solubility is, however, difficult owing to the slow achievement of equilibrium, the possibility of supersaturation occurring (Cavazzi, 1905) and the problem of ascertaining the amount of calcium sulphate in solution. For example, Arth and Crétien (1906) found that the amount varied depending on whether calcium or sulphate was determined. Hulett (1903) and Madgin and Swales (1956 b.) advocated conductivity measurements for dissolved gypsum determination while more recently EDTA determination of calcium has been employed (Zen, 1965; Furby, Glueckauf and MacDonald, 1968). Most workers have found that the maximum solubility is between 2.05 and 2.15 g $CaSO_4$/l, that is about 2.6 to 2.7 grams of gypsum per litre of water (Marignac, 1874; Hulett

and Allen, 1902; Hill, 1937; Hill and Wills, 1938; Posnjak, 1938; Kelley, Southard and Anderson, 1941). The figures suggested by Bock (1961) and Zen (1965) are higher, reaching 2.85 g $CaSO_4.2H_2O$/l. The solubility curve is quite shallow between 10° and 70°C. Gypsum solubility at these extremes is about 2.45 g/l and falls below 10°C and above 70°C (see fig. 1.6) Even at the gypsum-hemihydrate transition temperature at about 106 ± 5°C the gypsum solubility falls only to about 2.15 g/l (Hall, Robb and Coleman, 1926; Nancollas, Reddy and Tsai, 1973).

In contrast with the shallow solubility curve of gypsum the intermediate dehydration products, hemihydrate and soluble anhydrite, show considerable variation in solubility with temperature. There are relatively few hemihydrate solubility determinations available since it is unstable in the presence of water below about 95°C and above about 120°C. Between these temperatures it is metastable and will convert to soluble anhydrite and insoluble anhydrite within 48 hours or more. The gypsum-hemihydrate solubility curves intersect at 100± 5°C, where hemihydrate solubility falls rapidly to about 1.06 g $CaSO_4.\frac{1}{2}H_2O$/l at 120°C (Hall, Robb and Coleman, 1926; Partridge and White, 1929; Nancollas, Reddy and Tsai, 1973). Below the gypsum hemihydrate transition temperature hemihydrate solubility rises rapidly. At 80°C it is about 3.2 g/l; at 60°C about 4.25 g/l; at 40°C, 6.4 g/l; and at 20°C about 9.55 g/l (Hulett and Allen, 1902; Partridge and White, 1929). In effect, hemihydrate is considerably more soluble than gypsum at low temperatures.

The soluble anhydrite solubility curve has usually been treated as that of hemihydrate (for example, Melcher, 1910; Robb and Colemann, 1926; Partridge and White, 1929). Above 120°C, the temperature at which soluble anhydrite begins to form from hemihydrate, the joint solubility curve becomes less steep. At 150°C solubility is about 0.5 g/l and at 200°C about 0.15 g/l. These figures, from Melcher (1910) and Hall, Robb and Coleman (1926) agree well with those of

Figure 1.5

The logarithm of gypsum and anhydrite solubility in the system $CaSO_4$ - H_2O plotted against temperature at one bar or the vapour pressure of the solution, 500 bars and 1000 bars. Dashed lines are extrapolations of the calculated anhydrite solubility curves. The line connecting the intersections of the isobaric curves is the projection onto the composition-temperature plane separating G + L from A + L in figure 1.1 (after Blount and Dickson, 1973).

Figure 1.6

Solubility curves in the system $CaSO_4$ - H_2O (after Power, Fabuss and Satterfield, 1964).

Figure **1.5**

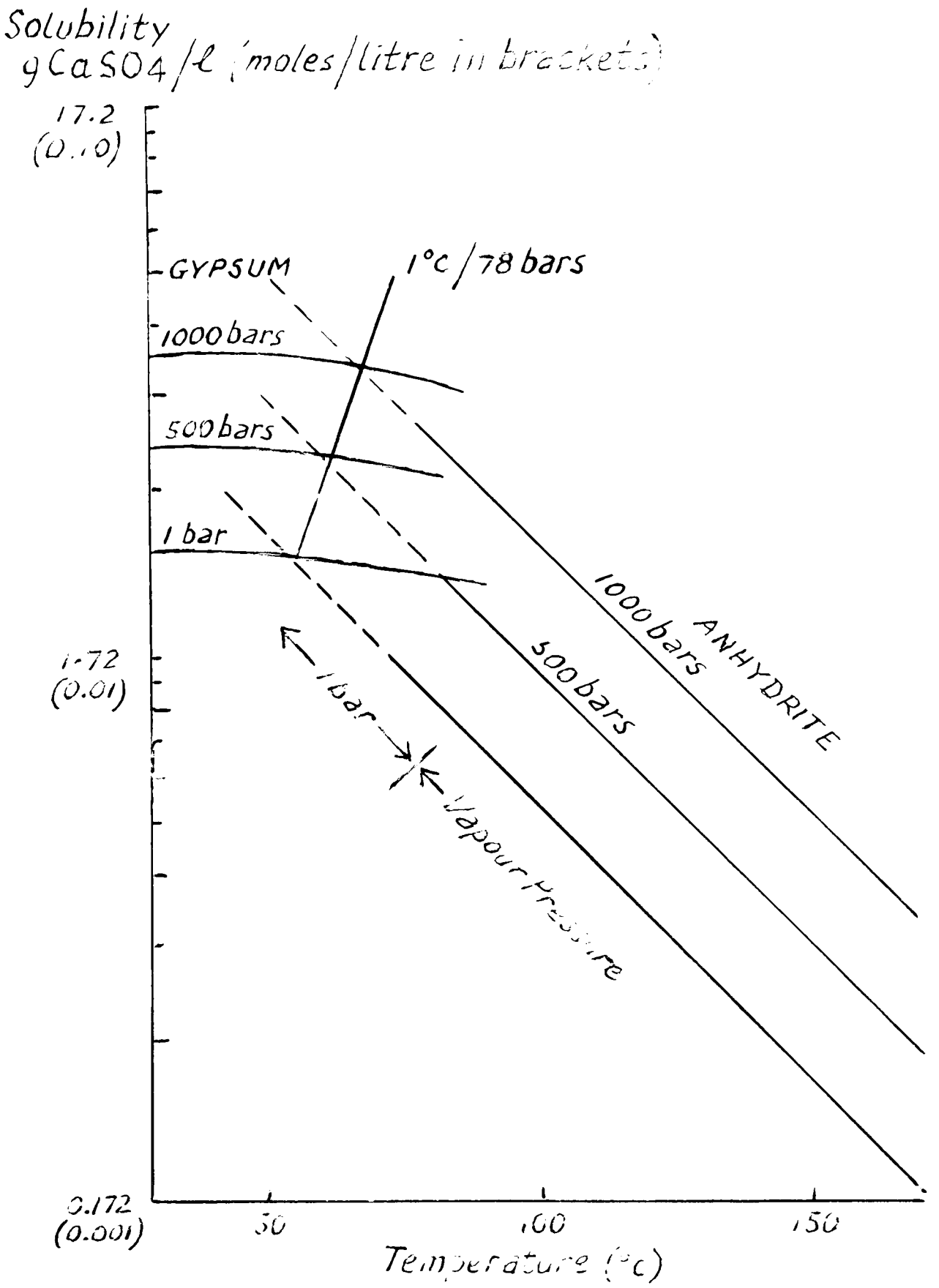
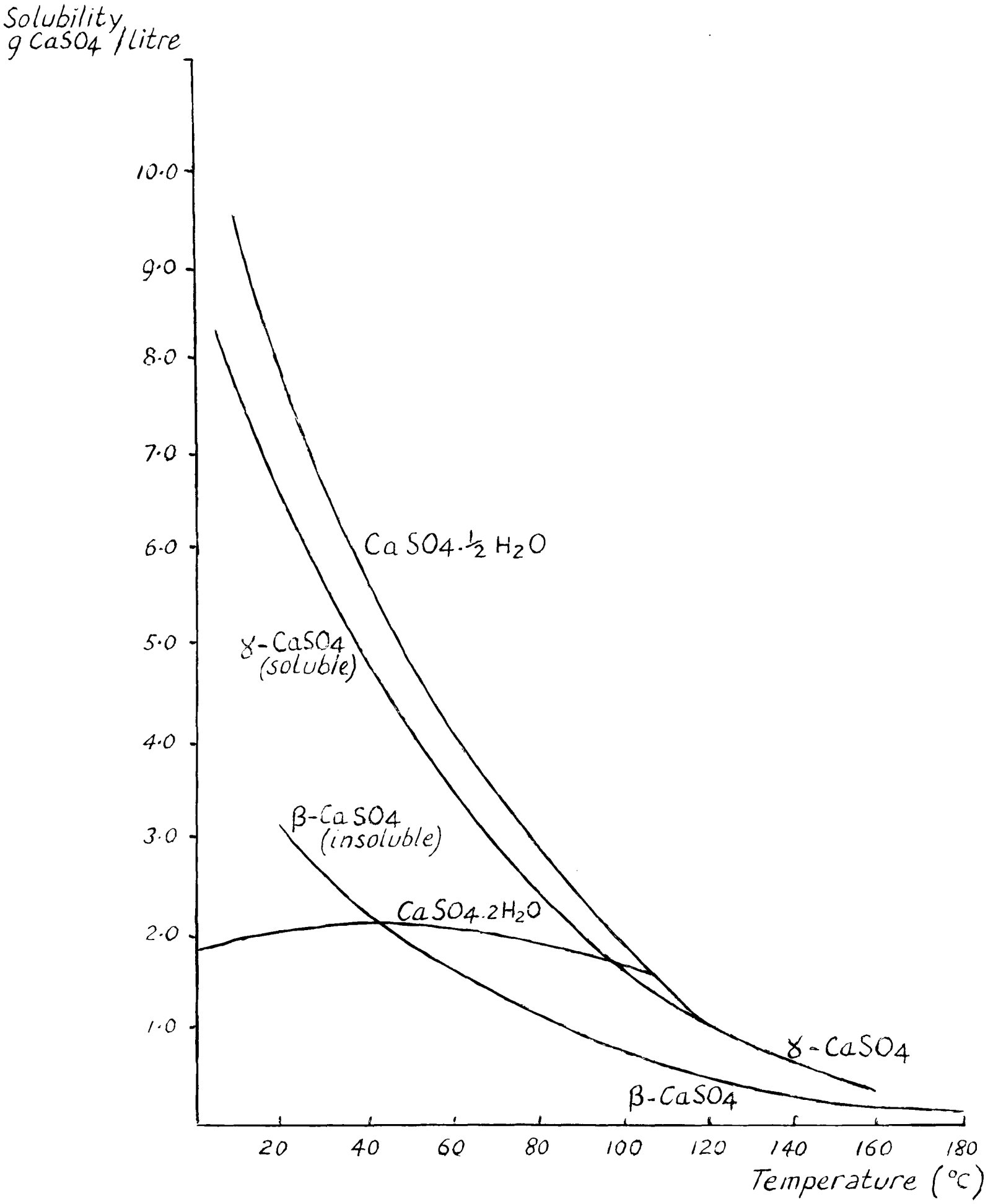

Solubility
g CaSO4/ℓ (moles/litre in brackets)
17.2
(0.10)
GYPSUM
1°C / 78 bars
1000 bars
500 bars
1 bar
1000 bars
ANHYDRITE
500 bars
1 bar
Vapour pressure
1.72
(0.01)
0.172
(0.001)
50
100
150
Temperature (°C)

Figure **1.6**

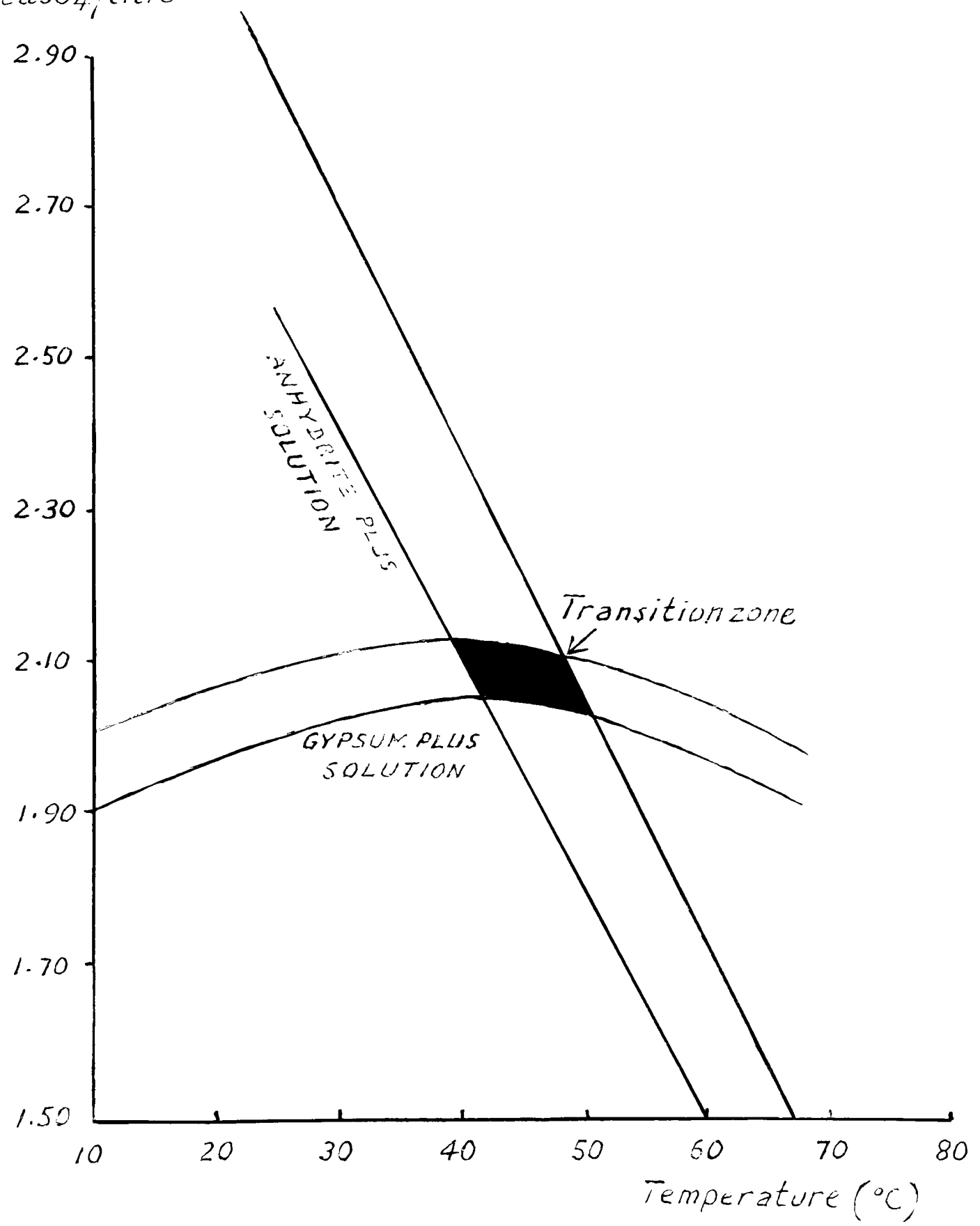

Solubility
g CaSO4 / litre
10·0
9·0
8·0
7·0
6·0
5·0
4·0
3·0
2·0
1·0
Ca SO4·½ H2O
γ-CaSO4
(soluble)
β-Ca SO4
(insoluble)
CaSO4.2H2O
γ-CaSO4
β-CaSO4
20
40
60
80
100
120
140
160
180
Temperature (°C)

Tilden and Shenstone (1884) who found that soluble anhydrite solubility is 0.78 g/l at 140°C. At temperatures above 200°C the solubility line of soluble anhydrite intersects that of "insoluble" anhydrite.

Insoluble anhydrite is in fact slightly soluble, about 0.05 g/l at 200°C (Melcher, 1910). The solubility curve is always lower than hemihydrate and insoluble anhydrite accounting for the metastability of these phases. At 150°C solubility is about 0.3 g/l and at 100°C about 0.85 g/l. The curve steepens as temperature falls and eventually intersects the gypsum solubility curve. At lower temperatures gypsum is the least soluble solid phase and therefore anhydrite is metastable. The precise point of the intersection has received a great deal of attention. Difficulties in achieving solubility equilibria from both under- and over-saturation (Hill, 1937 found 5% variation in the curves at 50°C) have necessitated extrapolation from high temperatures or ternary systems with high activation energies. At 20°C insoluble anhydrite solubility is about 3.0 g/l (Hill, 1937; Bock, 1961; Zen, 1965) though extrapolation of Posjnak's (1938) data suggests the figure may be lower, about 2.8 g/l.

The intersection of the gypsum and insoluble anhydrite solubility curves in pure water has been reported between 38°C and 66°C. The reasons for this variation are apparently a combination of poor experimental techniques and the low angle intersection of the two curves (Zen, 1965), with large variations in the determined transition temperature resulting from minor experimental errors in solubility determinations (see fig. 1.7) Despite the large range of results there is a clear division of opinion concerning the transition temperature. Van 't Hoff and Weigert (1901) suggested that the transition occurs at 66°C, this was modified to 63.5 by Van 't Hoff et al. (1903). These figures were attacked as being too high on the grounds that the extrapolations were based on erroneous thermochemical principles (Hill, 1937). Standard temperature (0 to 120°C) experimental investigations show that the solubility curves intersect at between

Figure **1.7**



Variations in solubility levels of gypsum and anhydrite in the system $CaSO_4$ - H_2O as a function of temperature. From a compilation by Hardie (1967).

38°C and 39°C (Partridge and White, 1929; Ramsdell and Partridge, 1929) or 40°C (Power, Fabuss and Satterfield, 1964) or 42°C (Hill, 1934; Posnjak, 1938; Conley and Bundy, 1958; Bock, 1961 and D'Ans 1968), while extrapolations from experiments at high activation energies tend to suggest an intersection at temperatures close to that proposed by Van 't Hoff et al. (1903). For example, Melcher (1910) 60°C, Hall, Robb and Coleman (1926) 66°C, D'Ans and Höfer (1937) 63.5°C, Hardie (1967) 58 ± 2°C, and Blount and Dickson (1973) 56 ± 3°C. The only thermochemical extrapolation suggesting an intersection in the lower range is that of Kelley, Southard and Anderson (1941) at 40°C. Zen (1965) calculated that the intersection is at 45° to 46°C, similar to that suggested by Furby, Glueckauf and MacDonald (1968). Zen (1962, 1965) suggested that previous insoluble anhydrite solubility measurements below 70°C are unlikely to be accurate, owing to the slow reaction times. He used Debye-Hückel theory to extrapolate the anhydrite solubility curve.

It is impossible to resolve the argument at present. All the figures presented here are based on rigorous experimental procedures but it cannot be determined whether the error lies in the experimental results from low temperatures or in the thermochemical principles used in extrapolation from high temperature/ pressure régimes.

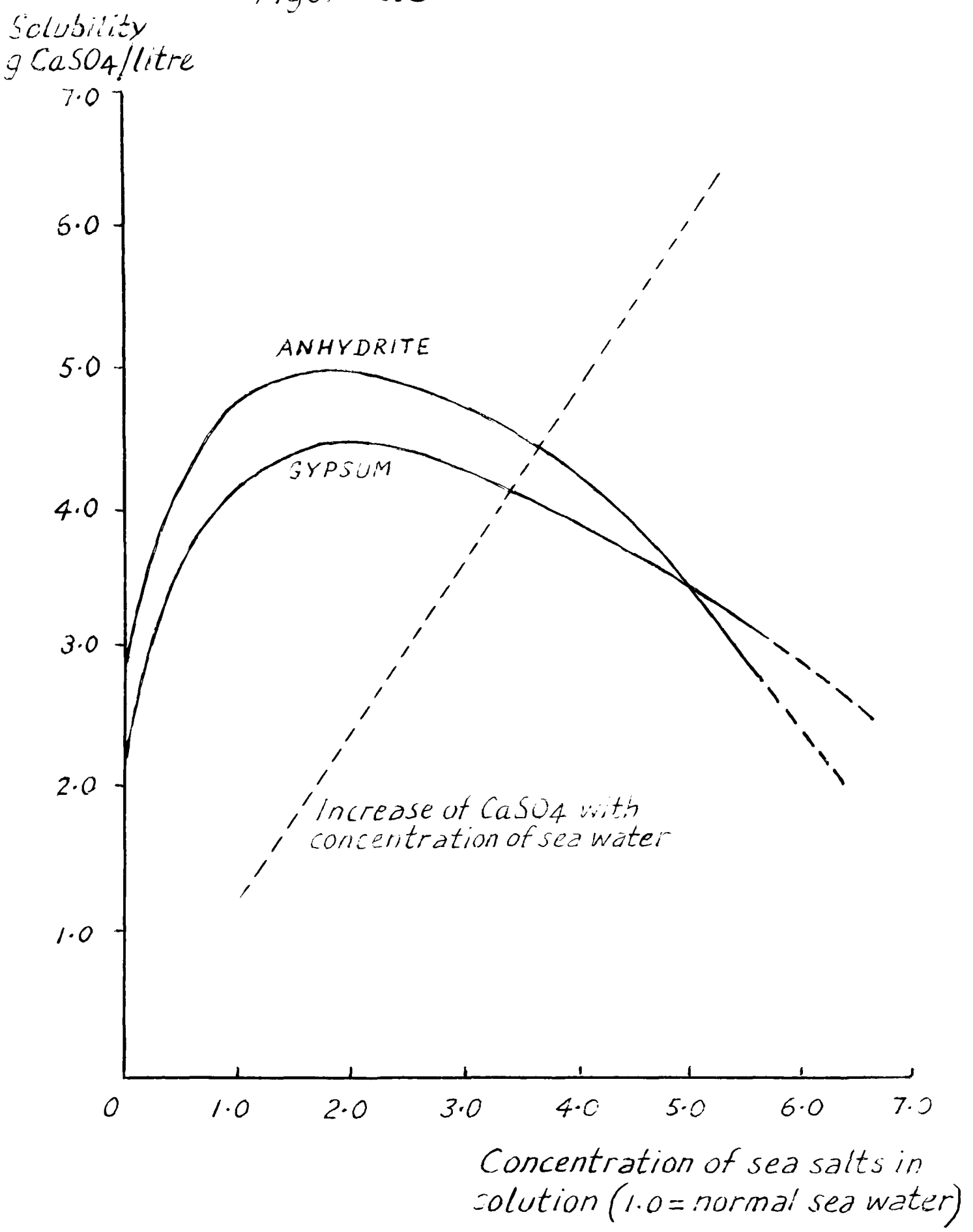
d. The effects of other salts on the solubility of the calcium sulphate minerals in water.

Despite the difficulties of determining solubilities in the binary system $CaSO_4$ - H_2O much attention has been focused on ternary and more complex systems. The main reason for this is the effect of sodium chloride on gypsum solubility and the significance of this to marine evaporite sedimentation. Van 't Hoff et al. (1903) found that in the presence of sodium chloride the solubility curves of gypsum and anhydrite intersect at lower temperatures than in pure

water because anhydrite solubility is not raised as much as that of gypsum (see fig. 1.8). This suggested that anhydrite might be the primary depositional mineral rather than it being a product of gypsum dehydration caused by geothermal heating during deep burial. Though solubility in seawater is the critical factor, the complexity of the solution has necessitated experimentation in simple ternary systems.

Droeze (1877) conducted a series of experiments between 10° and 13°C to determine the effects of soluble salts on gypsum solubility. He found that gypsum is four-times more soluble in saturated sodium and potassium nitrate and ammonium chloride solutions than in pure water. In sodium, potassium and magnesium chloride saturated solutions, solubility increases two or three times. In contrast, in saturated solutions of sodium and ammonium sulphate gypsum is only slightly more soluble than in pure water, while potassium and magnesium sulphates greatly reduce gypsum solubility. Droeze concluded that the presence of sulphate reduces gypsum solubility owing to a common ion effect resulting in complex double salt formation as Sullivan (1859) had suggested. Cameron (1901) and Cameron and Seidell (1901) expanded upon these experimental findings. At temperatures between 23° and 26°C the common ion effect results in reduced gypsum solubility in the presence of up to 17.5 g/l sodium sulphate. Above this figure a soluble double salt forms increasing gypsum solubility up to 2.5 g/l $CaSO_4$ at 230 g/l $NaSO_4$. The presence of sulphate therefore has different effects depending upon the cations available. In contrast, calcium chloride reduces gypsum solubility since a double salt does not form. Cameron and Seidell (1901) noted that although gypsum solubility decreases in solutions with more than 8% calcium carbonate, this effect is negligible when sodium chloride is also present to saturation. Wigley (1973) showed that a solution saturated with gypsum or calcium carbonate is capable of dissolving the other mineral but this will cause precipitation of the mineral at saturation.

Figure **1.8**

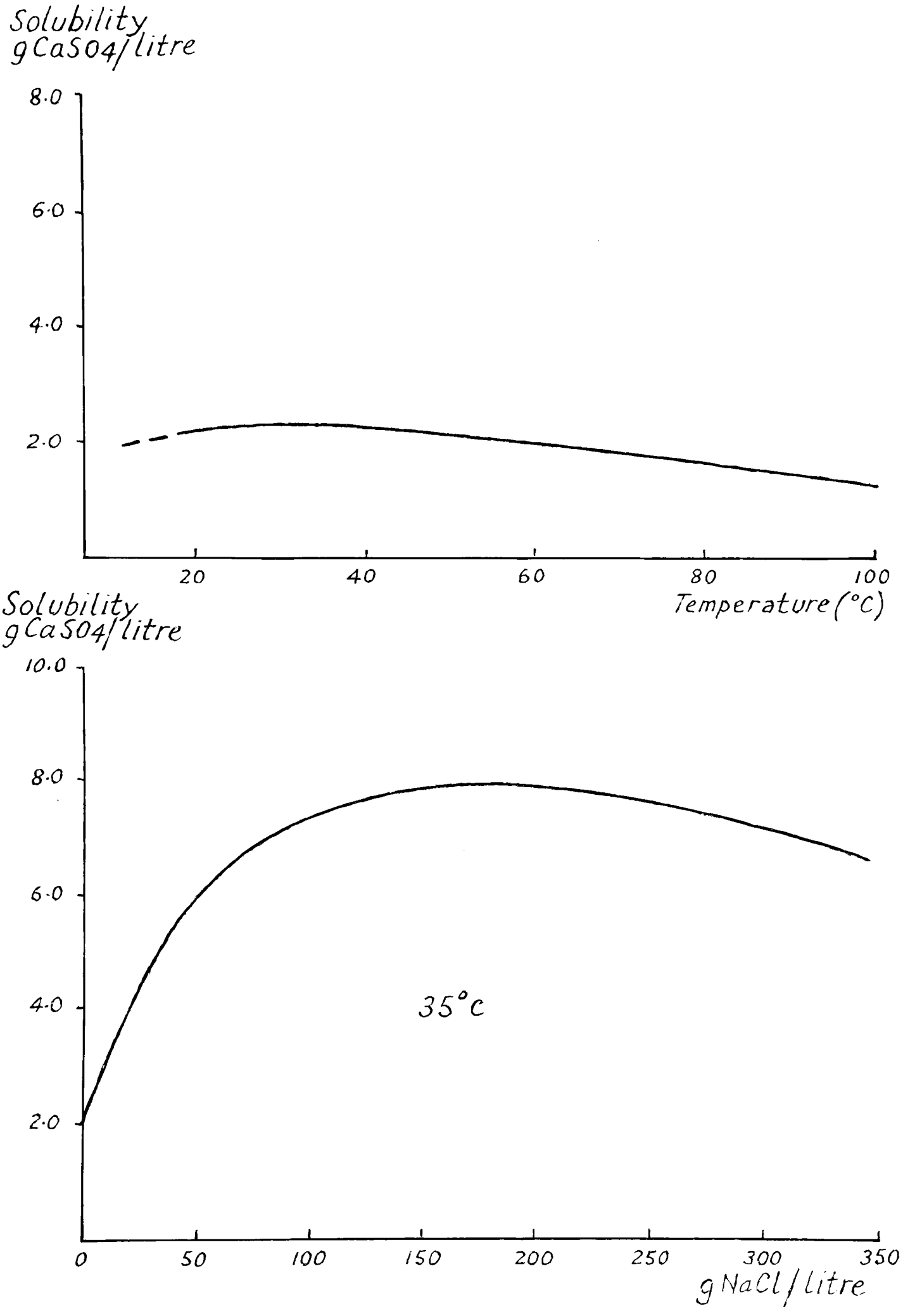


Solubility curves for gypsum and anhydrite in solutions of sea salts at 30 °C. Dashed line shows the amount of $CaSO_4$ that is available when seawater is evaporating (after Posnjak, 1940).

Little systematic work on ternary systems containing calcium sulphate was carried out before the 1930's. Hill (1934) examined the system calcium-sulphate-potassium sulphate-water in an attempt to assess the possibility of obtaining potassium sulphate from polyhalite, $(CaSO_4)_2 . K_2SO_4 . MgSO_4 . 2H_2O$, by solution at high temperatures. Later work by Hill and Yanick (1935) on calcium sulphate - ammonium sulphate - water and Hill and Wills (1938) on calcium sulphate - sodium sulphate - water was also conducted at temperatures between 25° and $100^{o}C$. For the first time the solubility curves from both over-saturation and under-saturation were determined. The results of these experiments are important for understanding the complexities of calcium sulphate precipitation from seawater. Posnjak (1940) attacked Hill (1937) for implying that anhydrite could precipitate from the ternary system $CaSO_4 - K_2SO_4 - H_2O$. This criticism is not justified since Hill (1937) pointed out that syngenite will precipitate at a lower temperature than anhydrite. Zen (1965) also misinterpreted the results of this work when he stated that Hill (1937) had been able to reverse the gypsum-anhydrite transition at 35° and $45^{o}C$ in the system $CaSO_4 - K_2SO_4 - H_2O$. In fact, Hill (1934) was measuring solubility of gypsum and anhydrite seeds. His results show that it is impossible to convert gypsum to anhydrite in this system even at $100^{o}C$. D'Ans and Höfer (1937) showed that it requires 80 g of phospheric acid/100g water to convert gypsum to anhydrite at $25^{o}C$.

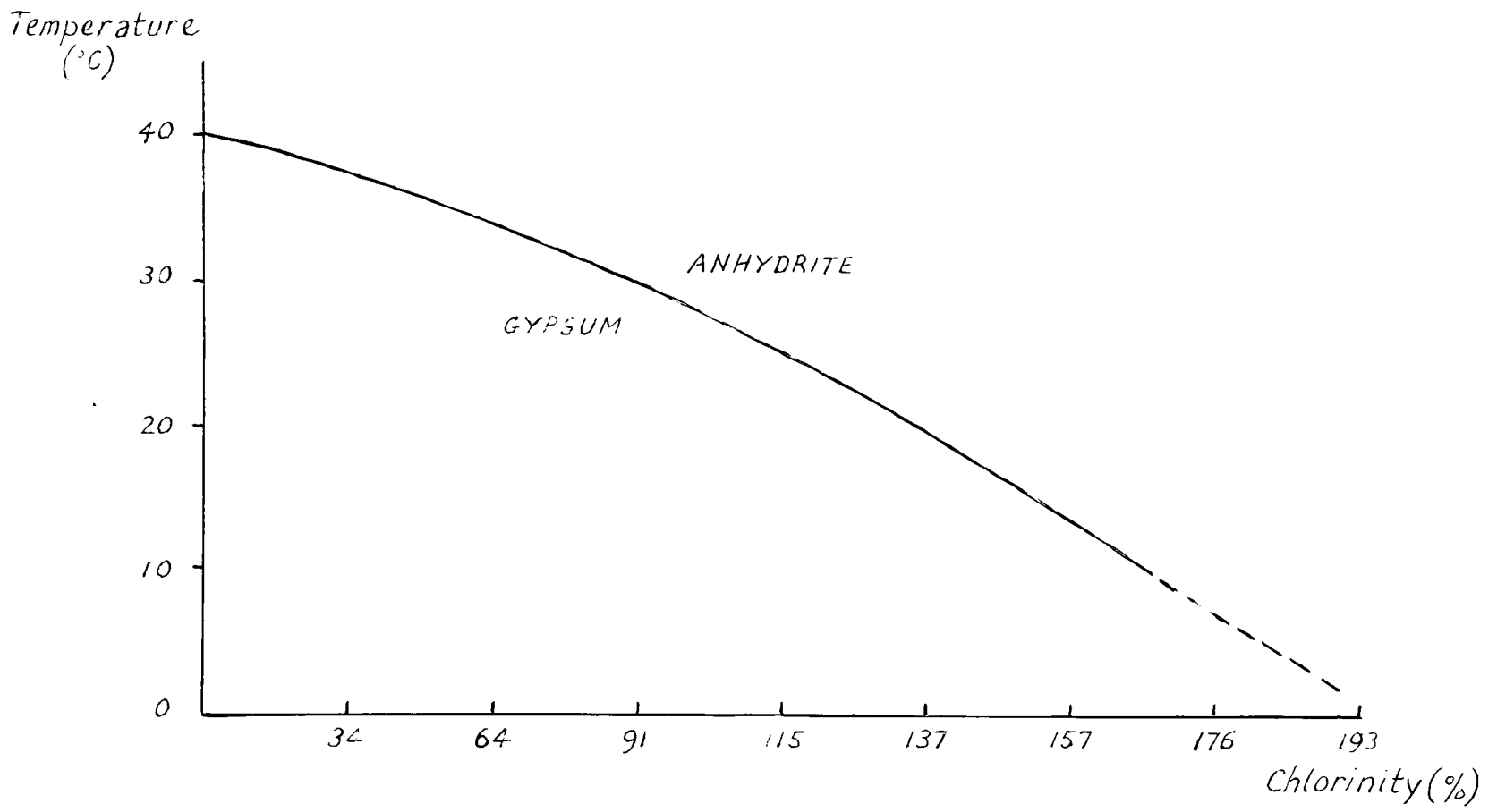
Detailed studies of the ternary system $CaSO_4 - NaCl - H_2O$ confirmed the findings of early workers such as Hill (1937) and Posnjak (1940). Madgin and Swales (1956 a.) found that at $25^{o}C$, 9.86 g of gypsum dissolved in 1.0 l water with 180 g of sodium chloride. This compared with about 2.5 of $CaSO_4.2H_2O$/ pure water (see fig. 1.9). Bock (1961) explained this enhanced solubility in terms of sodium chloride reducing the activity of the water of crystallization of the gypsum. Since insoluble anhydrite has no water of crystallization, it follows that its solubility curve will intersect that of gypsum at a lower temperature in the presence of sodium chloride than in pure water (see fig.1.10).

Figure **1.9**



Gypsum solubility. Above, in pure water (after Hill, 1937). Below, in NaCl solutions at 35°C (after Zen, 1965).

Figure **1.10**



Dependence of the dehydration temperature of gypsum on the chlorinity of solution, at one bar pressure (after Eastaff, no date).

It must be noted, however, that Reitemeier and Ayers (1947) found that the increased solubility of gypsum and calcium carbonate in solutions containing sodium metaphosphate is not a result of reduced ionic activities. With 220 g NaCl/l water the gypsum-insoluble anhydrite transition point is at 25°C (Madgin and Swales, 1956 a.) with 300 g NaCl/l water it is at 20°C (Ostroff, 1964). These findings are based on solubility determinations. Zen (1962, 1965) was unable to accomplish reversible transitions at 35°, 50° or 70°C, even in saturated sodium chloride solutions, over three months. Maximum gypsum solubility was determined as about 10.24 g/l at 35°C with 225 g NaCl (see fig. 1.9) and 10.18 g/l at 50°C with 180 g NaCl, but at these temperatures and chlorinities insoluble anhydrite is stable not gypsum. Gypsum solubility increases with NaCl concentration, only falling slightly before the piercing point, but anhydrite will not precipitate even at 70°C. Zen (1965) reported hemihydrate precipitation at 70°C in sodium chloride saturated solutions.

Zen's (1965) findings imply that anhydrite may not precipitate from seawater solutions and brines at low temperatures (25° - 70°C) where it is stable relative to gypsum. Furby, Glueckauf and MacDonald (1968) examined the solubility of gypsum and anhydrite in solutions containing the main soluble salts present in seawater (0.4861 moles/kg water NaCl, 0.0363 moles/kg water $MgCl_2$, and 0.01813 moles/kg water $MgSO_4$). Though they found that gypsum is more soluble in this solution than in simple sodium chloride solutions of the same ionic strength, presumably because of the enhanced solubility caused by the presence of $MgCl_2$ (Droeze, 1877; Posnjak, 1940), anhydrite precipitation did not occur even at 93°C when hemihydrate is readily precipitated. Furthermore, anhydrite will convert to gypsum within the supposed anhydrite stability field. Posnjak (1940) explained this in terms of salts present in the brines but the precise mechanisms probably involve complicated calcium complexing (Nakayama, 1971).

The findings of Posnjak (1940), Zen (1962, 1965) and Furby, Glueckauf and MacDonald (1968) suggest that in seawater brines anhydrite will neither nucleate nor precipitate owing to the presence of the other salts. Leininger, Conley and Bundy (1957) showed that the ions which enhanced hydration and anhydrite to gypsum are ranked:-

$K^{+} \gg N^{+} \gg NH_4^{+} > Mg^{++} > Fe^{++} > H^{+} > Al^{+++} > Ca^{++}$

and $SO_4^{--} \gg Cl^{-} > OH^{-}$.

The effects of potassium and sodium sulphate in the calcium sulphate-water system, first investigated by Hill (1934) and Hill and Wills (1938), were re-examined by Conley and Bundy (1958). The presence of the double salts syngenite and hemicalcium sulphate or Labile salt ($2Na_2SO_4.CaSO_4.2H_2O$) was confirmed. At 25°C, 60% of seeded anhydrite converts to gypsum in the presence of 0.75M sodium sulphate over two hours. This compares with 0.2% conversion in pure water. The 300 fold increase cannot be explained in terms of solubility differentials. Furthermore, since the activation efficiency of sodium sulphate is 50 times that of sodium chloride, it is apparent that both cation and anion enter into the activation complex. Conley and Bundy (1958) concluded that the interaction at the anhydrite surface between Ca^{++} and SO_4^{--} and the activator produces surface deformities which reduce surface energy releasing more Ca^{++} and SO^{--} per unit time. If the double salt is the least soluble species under the conditions of the system a double salt crystallite forms and the solution is replenished by more anhydrite solution. If gypsum is the least soluble component, saturation of the solution precipitates gypsum but the maximum rate of exchange is limited to 50% since a competing reaction on the gypsum surface is induced.

In seawater and concentrated seawater brines anhydrite solution is surface reaction controlled. Though common ion effects will tend to reduce gypsum solubility in the presence of other calcium salts, calcium complexing may occur in the presence of other sulphates (Nakayama, 1971). This significantly in-

creases insoluble anhydrite solubility relative to gypsum. Hence, even if anhydrite crystallites develop in seawater they would immediately convert to gypsum. It should be pointed out, however, that Cody and Hull (1980) reported primary anhydrite crystallization at 60°C in solutions containing only 50‰ NaCl when organic compounds are present.

e. Gypsum crystal structure, habit and growth.

Gypsum has a monoclinic crystal structure with a C^2/c space group (Nancollas, Reddy and Tsai, 1973), the molecular arrangement has been described above (Wooster, 1936). The crystallography has been summarized by Broughton (1934) (see fig. 1.11) but crystal habit is very variable, consisting of combinations of prism, clinopinacoid and negative hemipyramid, flattened parallel to the clinopinacoid (Read, 1970, p. 280). Cleavage is perfect parallel to the clinopinacoid (010) which represents sheets of water molecules (Wooster, 1936; Atoji, 1959). Masson (1955) reported weak cleavages on (011) and (100) the orthopinacoid. Twinning is common and can take the form of arrow-heads twinned on the hemiorthodome (101), swallow-tails on the orthopinacoid (100) or cruxiform twins also on (100) (Merritt, 1935). It should be noted that there is some confusion regarding the nomenclature of twinned crystal forms. Bright and Ridge (1962), for example, refer to (101) twins as swallow-tails, while Lambert and Schaffer (1926) term them arrow-headed twins.

The reason for the variability in the habit of gypsum crystals is primarily the presence of foreign ions in the precipitating medium regardless of whether this is within a liquid or solid host. Lambert and Schaffer (1926) recognized that acetate ions stimulate cruxiform twinning while apparently irregular and rounded crystals are the product of solvent action on certain crystal faces. The effect of different ions depends on their affinity to the ions of any crystal face. Since the cleavage plane (010), the clinopinaciod, is the water

Figure 1.11

Crystallography of selenite (after Broughton, 1934).
(110) prism.
(111) hemipyramid.
(010) clinopinacoid (main leavage).
($\bar{1}$02) and ($\bar{1}$03) positive hemiorthodomes.

Figure 1.12

Optical and crystallographic orientation of selenite from Laguna Madre, Texas (after Masson, 1955).
A. Cleavage fragments of euhedral crystals.
B. Lenticular crystals.

Figure **1.11**

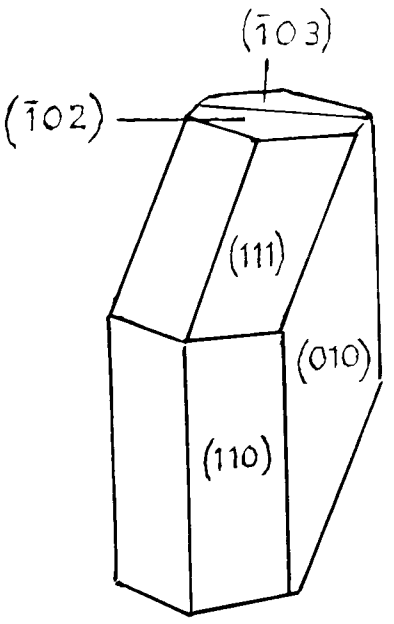
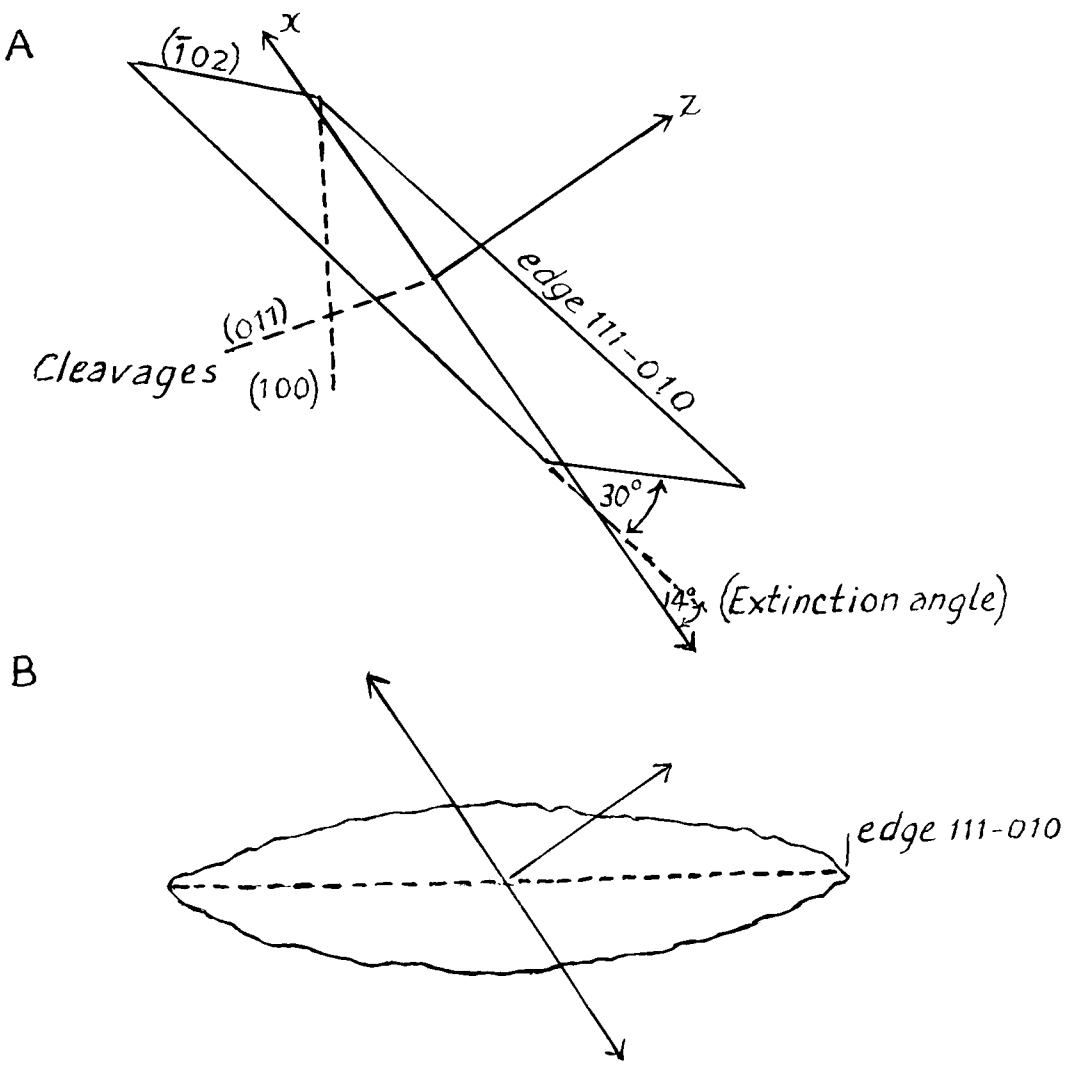


Figure **1.12**



molecule plane, it is affected very little by foreign ions, only OH^- is adsorbed on the surface. In contrast, (111) consists of Ca^{++} ions and adsorbs HSO_4^-, NO_3^- and Na^+ while (110) adsorbs H^+, Sr^{++} and Ag^+ (Edinger, 1973). Of these ions the presence of Na^+, OH^-, $H_2SO_4^-$ and Ag^+ has been found to result in the formation small crystals apparently because their desorbtion from the crystal surface is necessary before further adsorbtion of the appropriate gypsum ions can take place.

These findings may also be applied to the more unusual natural crystal habits. Though selenite crystals are frequently well developed (Broughton, 1934; Eardley and Stringham, 1952; Merritt, 1935) it is more usual to encounter gypsum in the form of rosettes or desert roses. These are intergrowths of lenticular gypsum crystal from several millimetres to 10 centimetres in diameter (Avias, 1942; Bellair, 1954; Masson, 1955; Kinsman, 1964 a., 1969). Masson (1955) described them buried in the mud and sands of Laguna Madre, sout-west Texas. They are the product of $(\bar{1}02)$ growth at the expense of (111) and while (010) is restricted because it has the highest reticulate atomic density or lattice packing (Eardley and Stringham, 1952), the (111) : (010) edge forms the edge of the disc (see fig. 1.12) Masson realized that a degree of corrosion produces the lenticular habit but the size of the crystals might be a result of their age or subsurface brine movement concentrating the gypsum. Deicha (1943) argued that the presence of sodium chloride results in lenticular crystal development by causing curvature of the crystal faces. Cody (1976) found that this was an over simplification because crystals grown in different bentonite muds and silica gels exhibit variable habits regardless of the sodium chloride content. Cody and Shanks (1974) found that sodium montmorillonite (Wyoming bentonite) host sediment produces individual crystals while silicate acid gels produce elongated crystals (Barta, Žemlička and René, 1971). It is likely that chemically complex host sediments affect the cation exchange capacity which is critical to crystal development. Growth patterns are affected by monovalent

salts particularly sodium, potassium, lithium and ammonium which alter the surface energy of the gypsum crystal. These salts retard the growth of (111) and, since the slowest growing face is the largest overall, $(\bar{1}02)$ represents the dominant growth plane.

The other major crystal habit of gypsum, the fibrous, will be dealt with briefly here. Most workers have attributed the formation of satinspar to processes involved in the hydration or rehydration of anhydrite to gypsum several hundred metres below the Earth's surface (Bundy, 1956; Shearman et al. 1972). MÜgge (1899) described fibrous celestite ($SrSO_4$) pseudomorphic after satinspar. Fibrous gypsum consists of masses of acicular crystals less than 0.5 mm in diameter but 0.20 m or more in length. It occurs as a vein fill with the crystals usually perpendicular to the walls, though Shearman et al. (1972) held that the fibres are always vertical. Though the crystals have their own extinction, Shearman et al. (1972) suggested that groups of crystals with approximately the same extinction may define large prismatic aggregates of fibres. Bundy (1956) and Holliday (1970) held that the fibrous beds are produced by crystallization under the pressures resulting from the volumetric expansion caused by anhydrite hydration. The cracks filled with veins and beds of satinspar were enlarged by crystal growth at the crystal/rock interface (Matsuura, 1925). Beds of fibrous gypsum from Nottinghamshire exhibiting casts and moulds of sedimentary structures, such as mud cracks on top and bottom, support this view (Aljubouri, 1971). In contrast, Shearman et al. (1972) held that the crystals grew in cracks created by hydrostatic pressure. They discounted crystal growth as a factor in crack expansion on the grounds that curved crystals, with the same extinction, must have developed under pressure when crystal growth was faster than crack widening. Inclusion of wall material within the vein fill supports the concept of crack expansion owing to crystal growth but Phillips (1974) pointed out that such inclusions are usually restricted to the centre of the vein and not the bulk of it. Bending of the crystals (variable extinction along the fibre) may be

a product of deformation after crystallization has ceased (Matsuura, 1925).

Phillips (1974) stated that fibrous vein fill nearly always has undulose extinction. He put forward a model of vein infilling involving tensile strain crystallization. Some solution may occur on planes perpendicular to the direction of greatest stress but crystallization takes place at points of maximum tensile strain. This generally results in infilling of voids between fibres from the centre of the vein towards the walls. Bundy (1956) and Shearman et al.(1972) also interpreted the fibrous structures as high pressure features. Watts (1978), however, described fibrous calcite textures from Kalahari calcretes which have never been subjected to high pressures. He felt that curved fibres indicated antitaxial growth, that is outward from crack margins with stress increments. Watts held that these features suggest displacive growth of fibrous calcite crystals. This is evident in the case of fibrous gypsum crystals and fibrous halite associated with gypsum crusts in the Central Namib Desert which are described below.

A common feature of naturally occurring gypsum crystals of all habits is the inclusion of fragments of the host material within the crystal. This may take the form of fine clay particles colouring the almost pure gypsum crystals or of large amounts of sand which makes up the bulk of the crystal. The processes which bring about this poikilitic inclusion are related to the process of crystal growth. Nancollas, Reddy and Tsai (1973) held that gypsum crystal growth is controlled by surface reaction rather than bulk diffusion of lattice ions to the crystal surface. This is supported by the aforementioned work on the effects of ions on crystal surface energy (Lambert and Schaffer,.1926; Deicha, 1943; Edinger, 1973; Cody and Shanks, 1974; Cody, 1976). Doss (1897) described sand filled gypsum crystals from the Bogdo Mountains of Kazhakhstan. He attributed the inclusion to "Krystallisationsbestreben". The same crystallization force was advocated by Kastner (1970) to explain hourglass pattern

inclusions in synthetic gypsum crystals. At low temperatures, precipitation of gypsum is reduced and host material is pushed ahead of the growing crystal face. At high temperatures, precipitation is rapid and the "pushing force" is exceeded causing inclusion of host material in the crystal. The experimental findings of Cody (1976) cast doubt on this model. He found that the presence of monovalent salts alters the surface energy, reducing the rate of crystallization but enhancing poikilitic inclusion of clay particles. The process of inclusion is dependent more upon the effect of foreign ions on crystal surface energy than on the crystal growth-rate. In the absence of foreign ions the crystal surface energy will decrease with rising temperature.

f. Gypsum isotope studies.

Since the amount of Oxygen 18 and deuterium in natural waters varies, it can be assumed that $\delta^{18}O$ and δ D of the water of crystallization of any sample of gypsum will reflect the properties of the water it precipitated from. The technique of isolating the isotopes from the water of crystallization of gypsum was perfected in the early 1960's but any interpretation of the data must also be based on thorough knowledge of the isotopic characteristics of natural waters. To some extent this information is still lacking. The same holds true for the isotopic data relating to $\delta^{34}S$ which can be isolated from the sulphate component of gypsum and also the $\delta^{18}O$ of the sulphate. The sulphur and oxygen isotope values of natural sulphates are subject to considerable variation owing to not only fluctuations in their availability over geological time but also environmental variations and post depositional alteration. Despite these problems, isotopic techniques have been applied to gypsum deposits, including gypsum crusts, with some success.

Early studies of $\delta^{18}O$ of gypsum by Gonfiantini and Fontes (1963) and Fontes and Gonfiantini (1967) showed that the $\delta^{18}O$ level of the water of crystallization is 4‰ higher than that of the water which precipitated the gypsum.

The values were reported as -3.3 for $\delta^{18}O$ of the water of crystallization and -7.3 for the mother brine (Fontes, 1965) in relation to standard mean ocean water (SMOW) (Craig, 1961). Fontes (1966) found that $\delta^{18}O$ in relation to SMOW, against the percentage of water, in an evaporating seawater basin, peaks at -7 to -8‰ with 20% remnant water. With further evaporation the $\delta^{18}O$ level of the brine falls. This peak is at the gypsum precipitation point. The higher $\delta^{18}O$ levels of the gypsum water of crystallization account for the decrease in $\delta^{18}O$ of the brine with further evaporation. The heavy isotope is preferentially bonded in the water of crystallization of the precipitating gypsum.

This looping effect at the gypsum precipitation point is also reported for δD (Sofer and Gat, 1975). Relatively little work has been undertaken on δD values of the water of crystallization of gypsum compared with $\delta^{18}O$. Fontes and Gonfiantini (1967) found that the δD value of the water of crystallization is 15‰ more than seawater δD. Sofer (1973) plotted δD against $\delta^{18}O$ in his study of the isotopic characteristics of the water of crystallization of gypsum with a wide variety of origins. Comparison of the two sets of isotopic values overcomes many of the problems encountered when only one of the isotopes is examined. It becomes possible, for example, to determine the origin of the water of crystallization more precisely and assess the degree of meteoric water enrichment since the characteristics can be reported in two dimensions (see fig. 1.13). The slope of the isotopic evaporation lines varies with humidity as well as salinity (Sofer and Gat, 1975) but δD and $\delta^{18}O$ both increase. In contrast, hydration of anhydrite shows increasing δD but decreasing $\delta^{18}O$ (Matsubaya and Sakai, 1975) because the isotopic characteristics of meteoric waters are adopted. Sofer (1978) also pointed out that if gypsum comes in contact with water with isotopic characteristics different to those of the water of crystallization, isotopic exchange reactions occur. It is possible to plot their path if the isotope balance of the exchanging water is known.

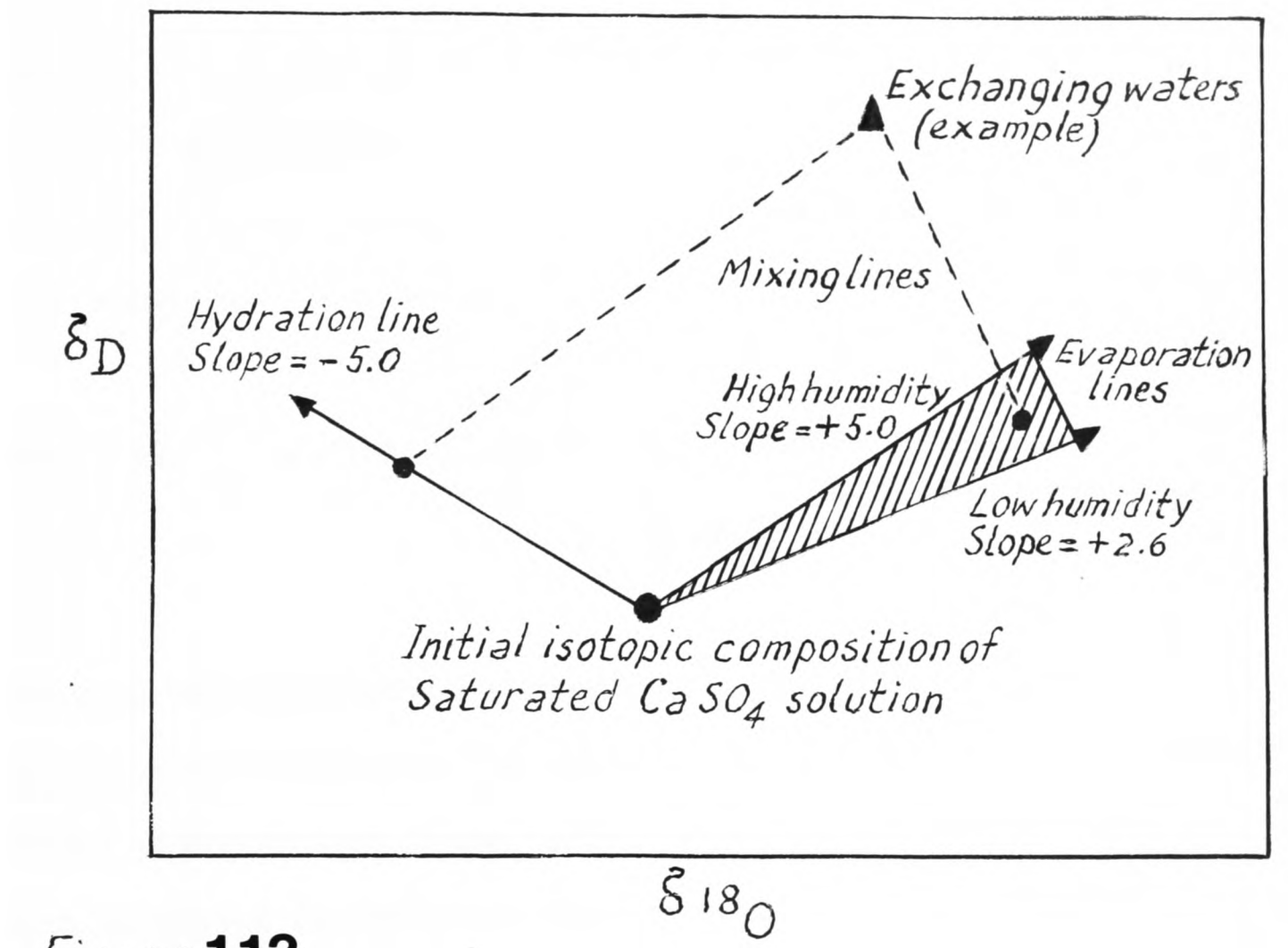


Figure **1.13**

Schematic representation of the evolution of the isotopic composition of crystal water in gypsum during evaporation, hydration and isotopic exchange (after Sofer, 1978).

Owing to the problems of isotopic exchange and the ease with which gypsum may dissolve and reprecipitate, thereby altering the characteristics of the water of crystallization, a number of workers have examined the isotopic characteristics of the sulphate component. Nielsen (1966) showed that recent sediments have a $\delta^{34}S$ content of + 20‰ and , while Cambrian, Silurian and Devonian marine sediments are higher, levels fall to + 11‰ in Upper Permian sediments. $\delta^{34}S$ levels of + 20 to + 21‰ for gypsum sulphate of marine origin and -20 to -30‰ from gypsum of bacterial origin have been reported from the Paris Basin (Fontes and Nielsen, 1966). Schreiber and Kinsman (1975) obtained δS^{32} values of -3.0 to -6.8 for native sulphur from the Montallegro gypsum of Sicily which is apparently produced bacterially. The low levels for the gypsum produced by bacteria result from bacterial reduction of sulphate. Reduction of SO_4^{--} to SO_3^{--} is a two step reaction:

$$SO_4^{--} \text{ soln.} + \text{enzyme} \xrightarrow[\text{STEP I}]{} SO_4^{--} \text{ enzyme complex}$$

$$\text{STEP II} \downarrow \; SO_3^{--} \quad \text{S - O bond break step}$$

If step I is the rate controlling step, ^{34}S reacts faster than ^{32}S and is enriched; if step II controls the rate, the faster reaction of ^{32}S results in its enrichment (Harrison and Thode, 1958). Mizutani and Rafter (1969 a.) showed that although bacterial oxidation of sulphur results in different $\delta^{18}O$ and ^{34}S fractionation the $\delta^{18}O$: $\delta^{34}S$ of 1:4 is neither temperature dependent nor greatly altered during reduction. In effect, it is difficult to identify the extent of bacterial isotope fractionation in a sulphate sample since the two isotopic components are affected similarly. Jensen and Nakai (1961) found that sulphates in the atmosphere are $\delta^{34}S$-rich and suggested that recent sulphates are derived from bacterial enrichment of hydrogen sulphide and also from sulphur dioxide of industrial origin. Husain and Krouse (1978) reported direct reaction between SO_4^{--} ions and hydrogen sulphide species. Holt, Cunningham and Engelkemier (1978) found that sulphates may form by hydrolysis of sulphur dioxide by atmospheric

water followed by oxidation to sulphate in the presence of metal ion catalysts in dust. Photochemical oxidation of sulphur dioxide followed by hydrolysis to SO_4^{--} may also take place. In coastal areas, atmospheric sulphate may be derived from sea spray and/or hydrogen sulphide liberated from seawater (Mizutani and Rafter, 1969 b.).

The $\delta^{18}O$ content of SO_4^{--} has been shown to vary significantly with geological time (Sakai, 1972) but $\delta^{18}O$ and $\delta^{34}S$ of/SO_4^{--} values do not correlate well over time. Similarly, Matsubaya and Sakai (1973) found no correlation between SO_4^{--} $\delta^{18}O$ and water of crystallization $\delta^{18}O$ for the same gypsum samples.

Several isotopic investigations of natural gypsum deposits have been undertaken. Fontes, Gonfiantini and Tongiorgi (1963) examined the evaporite sequence of the Paris Basin. $\delta^{18}O$ levels in gypsum water of crystallization suggest a meteoric and/or continental origin of the deposits. Though this may suggest that the gypsum originated by weathering of pyrite or hydration of anhydrite involving meteoric waters, it is also possible that large scale dissolution and precipitation has occurred. By comparing the $\delta^{18}O$ with δD, Matsubaya and Sakai (1973) showed that the Kuroko gypsum in Japan was anhydrite hydrated by meteoric water. The Japanese Noto gypsum water of crystallization corresponds isotopically to neither seawater nor meteoric water but approximates the isotopic composition of local thermal spring waters. Sofer (1978) successfully differentiated gypsum deposits from Sinai, the Namib Desert, West Germany and Israel on the basis of water of crystallization $\delta^{18}O$ and δ D. To some extent he found it possible to determine the climate under which hydration of anhydrite took place. Successful application of isotope analysis to studies of gypsum origin require analysis of $\delta^{34}S$ and $\delta^{18}O$ of the sulphate component owing to the susceptibility of the water of crystallization and to post-depositional change by solution and reprecipitation and isotopic exchange. This is well exemplified by the work of Lyon (1978) on gypsum from

Miers Valley, Antarctica. The water of crystallization corresponds isotopically to local glacial melt water. $\delta^{34}S$ is the same as Antarctic seawater but sulphate $\delta^{18}O$ is different. Lyon interpreted the deposits as a precipitate from water in a fjord isolated from the sea. Bacterial fractionation effected the sulphate $\delta^{18}O$ while the water of crystallization $\delta^{18}O$ and δD were lowered by inflowing glacial meltwater.

Isotopic analysis of the gypsum making up desert gypsum crusts has been undertaken by Sofer (1978) and Vieillefon (1980). There is considerable potential for further investigation since isotopic exchange is minimal in arid environments, effectively preserving the characteristics of the natural waters the gypsum precipitated from. Sofer (1978 and pers. comm.) determined $\delta^{34}S$ as well as water of crystallization $\delta^{18}O$ and δD for south-west African gypsum crust samples. The results and implications are considered below.

The foregoing review of work pertaining to the chemistry, crystallography and isotopic characteristics of gypsum and the other calcium sulphate minerals has been, out of necessity, very broad. Many of the characteristics of gypsum can be fully understood only in the light of the mineral's relationship with other forms of calcium sulphate. While some aspects of the description are not strictly pertinant to the overall discussion of desert gypsum crusts, it is appropriate to ouline the broader framework before attempting to deal with questions specific to gypsum crust. Much of work on the origins of calcium sulphate minerals in the geological record rests firmly on the chemical considerations which have been outlined here.

CHAPTER 2

ORIGINS OF CALCIUM SULPHATE ROCKS

a. Mechanisms of gypsum and anhydrite deposition.

Gypsum deposits have long been an important focus of the attention of economic geologists in Europe and North America. There are two main reasons for this. Firstly, gypsum is used in the manufacture of many commodities including cement, fertilizers, paper, paint, rubber and, most important, Plaster of Paris. Gypsum workings are often extensive, as is well exemplified in Paris where 65 ha of underground workings, most predating 1813, have caused considerable engineering problems (Leggett, 1973). Secondly, the physical and chemical properties which make gypsum so useful industrially cause severe problems when engineering projects are attempted on gypsum bedrock. Not only the solubility of gypsum but also the volumetric changes associated with anhydrite rehydration, when in contact with meteoric water, have resulted in some catastrophic structural failures of tunnels (Fecker, 1976; Grob, 1976; Henke, 1976 a.b.; Krause, 1976; Müller-Salzburg and Götz, 1976; Wohnlich, 1976) and of dams (Calcano and Alzura, 1967; James and Lupton, 1978). Though gypsum and anhydrite deposits are well mapped and much detail on the stratigraphy is available, the mechanisms of sedimentation are far from well understood.

Some gypsum deposits are formed during weathering and some by biochemical processes but the great majority are the product of evaporite sedimentation. In order to determine the processes of gypsum crust development, it is essential that the immediate source of gypsum is known as well as the chemical and/or biochemical processes involved in the evolution of the crust.

It will be shown that although only a small percentage of desert gypsum crusts are formed by surface evaporite sedimentation, this process is of great importance in the primary deposition of gypsum. Several models of evaporite sedimentation have been proposed since that of Ochsenius (1888). Most early models regarded deep-water basins as essential to account for thick evaporite sequences but the absence of present day analogues has provoked considerable debate particularly since the mid-1960's. Though lagoonal and sabkha environments have produced many evaporite deposits,it is difficult to account for exceptionally thick and extensive sequences, such as the Zechstein and Ochoa, in the same way. Another critical question is the likelihood of primary anhydrite deposition. Chemically this appears to be impossible but stratigraphic evidence suggests it may have occurred frequently and indeed is taking place at present. Regardless of whether gypsum or anhydrite is the dominant primary deposit,the effects of deep burial and exhumation must also be understood since volumetric changes caused by dehydration and hydration may cause considerable damage to man-made structures and alteration of stratigraphic features.

Here the information on authigenic and bacterial production of gypsum will be examined and the fundamentals of evaporite sedimentation and post-depositional alteration briefly reviewed.

b. Authigenic gypsum.

In the nineteenth century and early twentieth century a number of geologists attributed gypsum deposits to volcanic metamorphosis of calcium carbonate (Wilder, 1903; Louderbank, 1904; Ransome, 1908) involving sulphuric acid derived from metal sulphides. The process is discounted now except in the rare cases of contact metamorphosis of gypsum to anhydrite (Rowe, 1905; Harder, 1909; Jones,

1912; Rogers, 1914). A major exception to this is the Kuroko gypsum of Japan. Matsubaya and Sakai (1973) stated that this is hydrated anhydrite which formed on the sea-bed by hydrothermal processes related to submarine and volcanic activity in the middle Miocene.

Authigenic gypsum produced during the weathering of sulphides, and perhaps sulphur, frequently involves bacterial activity. This will be dealt with below. In some cases it has been suggested that the processes do not involve bacteria. Vialov (1946) found that the Palaeogene green Ferchana clays of Turkestan are gypsiferous near the surface because iron pyrite replaced oyster shell fossils (Fatina esterhàzyi Pav., Ostrea (Flemingostrea) schurabica Vial. and Chlamys vialori Korob) are weathering to gypsum. He suggested the chemical equation:

$$FeS_2 + O_2 + H_2O + CaCO_3 \longrightarrow CaSO_4 + Fe_2O_3\ H_2O + CO_2$$

The process is variable, Broughton (1934) found, for example, iron sulphates associated with the gypsum and pyrite in Texan Tertiary strata, indicating intermediate sulphuric acid formation.

$$2FeS_2 + 7O_2 + H_2O \longrightarrow 2H_2SO_4 + 2FeSO_4$$

$$CaCO_3 + H_2SO_4 + 2H_2O \longrightarrow CaSO_4.2H_2O + CO_2 + H_2O$$

Jessup (1960 b.) proposed a similar process to account for gypsum in the pallid zones of lateritic horizons but accepted that leaching of gypsum from the surface also occurs. Isotopic analyses of water of crystallization of gypsum from the Paris Basin (Fontes, Gionfiantini and Tongiorgi, 1963) and the Palaeocene of the Negev Desert (Sofer, 1978) have indicated that iron pyrite has weathered to gypsum in the presence of meteoric water. The association of jarosite (a hydrous iron-potassium sulphate) with some gypsum deposits adds further support to this (Sass, Nathan, and Nissenbaum, 1965 in the Negev; Buurman and Jongmans, 1975 in Limburg).

The weathering of igneous rocks was suggested to be a source of sulphate for

gypsum formation in Antarctica by Gibson (1962) though later work has not confirmed this view (Smith, 1965; Torii *et al.*, 1966; Lyon, 1978).

In desert areas with abundant soluble salts at the soil surface, authigenic gypsum may precipitate through processes of salt reaction with the soil absorption complex. Krupkin (1963) conducted experiments leaching natural soil columns with saline solutions containing combinations of sodium sulphate, chloride and carbonate and calcium chloride. He found that sodium sulphate reacted with the absorption complex producing gypsum.

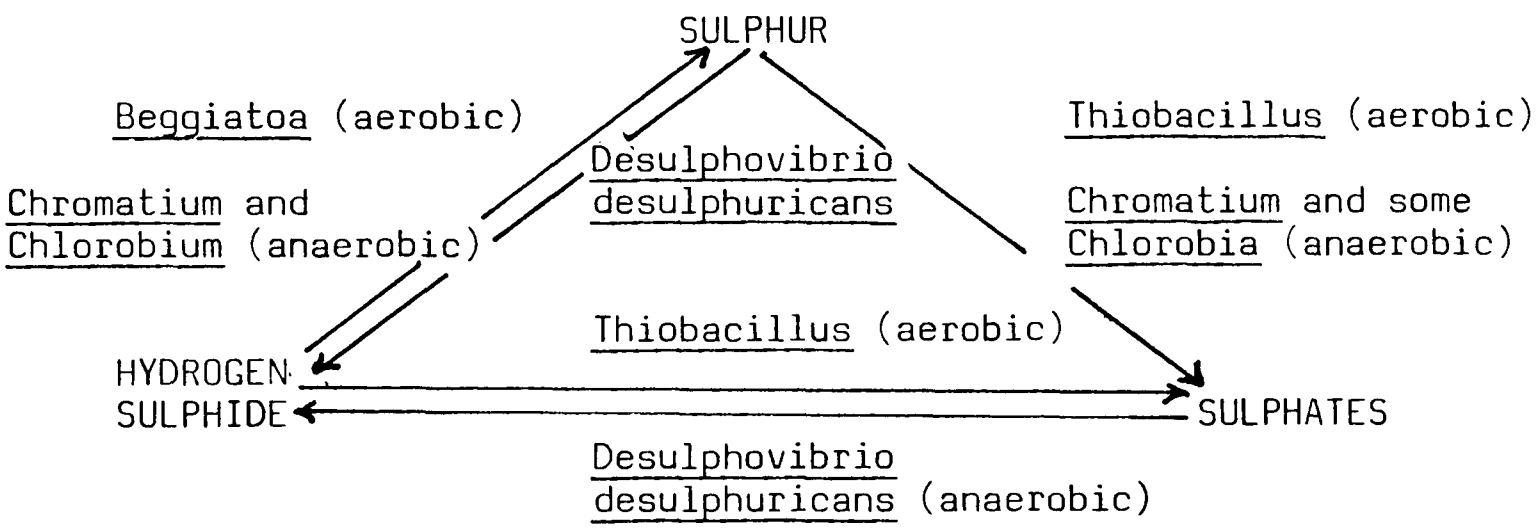
$$(\text{Soil})\quad Ca^{++} + Na_2SO_4 \longrightarrow (\text{Soil})\quad 2Na^{+} + CaSO_4$$

Calcium chlorides are formed by reaction of sodium chloride with the absorption complex and these are in turn converted to gypsum by reaction with sodium sulphate. These reactions are dependent upon the direction of soil water movement and the location in the profile of different salt accumulations but the gypsum content is always equivalent to the amount of displaced calcium. These processes may be very important in gypsum crust development but their significance in the development of thick deposits is uncertain. Shearman (1966) held that evaporite basins with calcium carbonate bedrock may precipitate gypsum in excess of the amount available in seawater through a process of sulphate reaction with calcium freed from the bedrock by dolomitization.

It has been suggested that the gypsum crusts of the Central Namib Desert are the product of *in situ* alteration of calcrete to gypsum (Martin, 1963). Hydrogen sulphide liberated from the waters of Walvis Bay is carried inland in fog moisture and, Martin argued, reacts with calcium carbonate to form gypsum. Besler (1972) held that the reactions involved would have to be considerable more complex than this, possibly involving bacteria. This model of gypsum crust development in Namibia will be dealt with below.

c. Bacteria.

With regard to the gypsum crusts of the Central Namib, Besler (1972) felt that sulphur in dews could produce thiosulphates in the presence of soil chlorides and nitrates. These thiosulphates may form gypsum when subjected to extreme temperatures with calcium carbonate present. It is more feasible, however, that bacterial oxidation of thiosulphates occurs because high temperatures are not required. The complexity of the bacterial sulphur cycle is such that oxidation and reduction may occur freely under aerobic and anaerobic conditions in the presence of different bacteria. Butlin (1953) summarized the cycle as follows:



In effect, the most important bacteria affecting gypsum are the Thiobacilli which produce sulphates from sulphur and hydrogen sulphide and Desulphovibrio desulphuricans which reduce the sulphate to hydrogen sulphide. It is not unusual to find both processes occurring in equilibrium. Zverev (1965) showed that calcium and sulphate ratios in groundwaters are partly determined by the availability of oxygen, slowly moving groundwater has less oxygen and is depleted in sulphate by Desulphovibrio desulphuricans.

Bacterial formation of gypsum is less common than reduction but has been reported from a number of environments. Eardley and Stringham (1952) proposed bacterial formation of selenite crystals in clays of the Great Salt Lake.

Besides causing oxidation of hydrogen sulphide they also suggested that bacteria may affect alkalinity and/or temperatures thereby reducing gypsum solubility resulting in its precipitation from natural waters. Pethuisot and Jauzein (1978) put forward the hypothesis that blue-green algae forming mats in the inter-tidal zone of a lagoon in Qatar caused gypsum precipitation through brine enrichment. Barcelona, Tosteson and Atwood (1976) , however, found that gypsum precipitation occurred on removal of fatty acids and protein hydrolyzates from aqueous solutions, while pH was not altered. Organic substances perhaps inhibit nucleation of gypsum crystallites. Variations in pH and temperature do not affect gypsum solubility as greatly as other minerals likely to be present, especially sodium salts and calcium carbonate. Parker (1951) showed that the reduction of pH by bacteria was critical in the destruction of concrete sewers. Hydrogen sulphide produced in the sewage forms thiosulphuric and polythionic acids lowering pH from 11 or 12 to about 7.5. Below pH9 Thiobacillus X oxidize the thiosulphates to sulphur and sulphuric acid reducing pH still further. Below pH5 Thiobacillus concretivorus multiplies rapidly producing large amounts of sulphuric acid of up to 5% strength (Butlin, 1953) and pH may fall to 1.0. Similar processes may produce the authigenic gypsum found in the sea-bed sediments off the south-west African coast (Siesser and Rogers, 1976) and the Central Pacific (Xavier and Klemm, 1979). The bacteria involved are probably different since the hydrogen sulphide is produced by plankton decay under anaerobic conditions and oxygen is not as freely available for the Thiobacilli as in sewer pipes. The bacteria Sporovibrio desulphuricans probably takes the place of the Thiobacilli under anaerobic conditions (Vinot, 1965).

The Thiobacilli which produce gypsum from sulphur (Thiobacillus thioparus and Thiobacillus thiooxydans) have been reported from Lake Eyre, Australia. Bacteria attack native sulphur producing sulphuric acid which reacts slowly with calcium carbonate to form gypsum (Baas-Becking and Kaplan, 1956; Bonython and King, 1956). Similar bacterial processes probably occur when iron sulphides are oxidized in

the presence of calcium carbonate forming gypsum (Vinot, 1965).

The reduction of sulphates to hydrogen sulphide has been reported on many occasions though the precise process is not agreed upon. Harrison and Thode (1958), in their analysis of bacterial enrichment of S^{34}, suggested that the process involved, firstly, the formation of an enzyme complex, secondly, the break of the sulphur-oxygen bond to form sulphur trioxide and, finally, rapid reduction to hydrogen sulphide. This process is usually associated with the bacteria Desulphovibrio desulphuricans of which Abd-el-Malek and Rizk (1963 a.) identified four strains. In saline waters Microspira desulphuricans and Microspira aestuarii are common (Bavendamm, 1932).

The reduction of gypsum in sea and lake waters was recognized by Bavendamm (1932). He proposed a two-step process:

$$CaSO_4 + 8H \longrightarrow 4H_2O + CaS$$

$$CaS + CO_2 + H_2O \longrightarrow CaCO_3 + H_2S$$

Bodenheimer and Neev (1963) suggested the second step may also involve:

$$CaS + 2H_2O \longrightarrow Ca(OH)_2 + H_2S$$

Both reactions probably take place.

Abd-el-Malek and Rizk (1963 b.) found that in soils both carbonate and bicarbonate increased with the amount of sulphate reduced ($r = 0.98$ significant at 0.01 in clay loams and $r = 0.96$ significant at 0.01 in sandy loams). Similar relationships are apparent in the Wadi NatrOn where sodium sulphate is reduced to sodium carbonate (Abd-el-Malek and Rizk, 1963 c.).

These processes of sulphate reduction are critical to understanding evaporite sedimentation. Bodenheimer and Neev (1963) and Friedman (1972) found that the bottom deposits of the Dead and Red Seas consist of low-magnesium calcite

which is produced when gypsum precipitated at the surface through evaporation is reduced to calcium carbonate as it sinks to the bed. Reduction appears to be complete at two metres below the water surface. Near the shores the process is less effective because oxygen in the waters inhibits the anaerobic bacteria. Yaalon (1964 c.) proposed similar bacterial activity to explain the low sulphate content and abundant hydrogen sulphide in Israeli marsh waters. Since most natural waters contain sulphate reducing bacteria, the process will be of great significance in evaporite sedimentation. It is also important in the post-depositional alteration of gypsum deposits. Ashirov and Sazonova (1962) found that calcium carbonate seals above certain oil pools are the product of gypsum or anhydrite reduction by bacteria in the presence of the hydrocarbons. They suggested the following processes:

$$CaSO_4 + H_2O + 2C_{org.} \longrightarrow CaCO_3 + H_2S + CO_2$$

or

$$CaSO_4 + 8H + CO_2 \longrightarrow CaCO_3 + 3H_2O + H_2S$$

Lattman and Lauffenburger (1974) proposed that the calcretes of Nevada are the result of bacterial alteration of gypsum deposits. They remarked upon the smell of hydrogen sulphide on piercing the calcrete sheet and also upon the greater thicknesses of calcrete around springs where anaerobic conditions prevail.

d. The problem of evaporite sedimentation.

The findings of Friedman (1972), referred to above, indicate that gypsum sedimentation cannot occur in deep water because sulphate reducing bacteria will alter the gypsum to calcite and hydrogen sulphide. This problem is one of many related to thick deposits of evaporites, the origins of which have defied adequate explanation for nearly one hundred years, when they were first examined in detail.

Thick deposits of evaporites are found throughout the world dating from nearly all geological periods. Yet certain deposits have attracted far more attention

than others. The Permian basins of west Texas and New Mexico and Central Europe, particularly the Ochoa and Zechstein Series, have been the subjects of considerable geological investigation owing to their economic importance in terms of potash (Smith, 1933) and oil reserves (Adams, 1936).

It has long been acknowledged that saline deposits are predominantly the product of marine water evaporation though Brayley (1829 a.b.) suggested that the saliferous deposits of North America were of an igneous origin.

Some of the earliest experimental investigations of the origins of saline deposits were carried out by Usiglio who attempted to ascertain the soluble salt content of Mediterranean water by evaporation (Usiglio, 1849). It was apparent that the least soluble salts precipitated first from the evaporating water. Iron oxides were first to appear followed by calcium carbonate, calcium sulphate, sodium chloride and finally the bittern deposits, principally magnesium and potassium salts. These findings helped to explain some evaporite deposits but thick salt beds were problematic. Firstly, seawater 500 km deep would have to be evaporated to produce a gypsum-anhydrite deposit 500 m thick, such as that of New Mexico (Baker, 1920) and, secondly, many deposits contain neither the correct proportions of salt nor the correct sequence to be the product of simple seawater evaporation. To overcome these problems, Ochsenius (1888) proposed a barred basin model. Based upon work in Kara-Bogaz Gol, a large basin adjoining the eastern Caspian Sea, the model involved dilute water influxes over a shallow bar to replace the water evaporating from the basin. While this model may be able to explain thick evaporite deposits,the problems presented by variable sequences remain.

Modifications of Ochsenius' bar theory have been applied widely in order to account for many evaporite deposits. Branson (1915), for example, envisaged a

series of interconnected basins, the waters in each being at different levels of saturation and hence precipitating different evaporite minerals. This model, which could be modified to account for variable salt sequences and great thicknesses, was rejected as more stratigraphic information became available. The Ochoa Series of west Texas and south-western New Mexico consists of the Castile, Salado, Rustler and Dewey Lake Formations of the late Permian (King, 1942) and possibly early Triassic, Comanchean (Udden, 1915). While the Castile is mainly anhydrite,the Salado consists of halite and potash beds apparently representing the final phases of marine evaporation. The Rustler and Dewey Lake Formations are probably lacustrine red-bed sediments (Adams, 1935 and 1936; Kroenlein, 1939; Lang, 1939; King, 1942). Exploratory drilling for oil produced a great deal of information on the nature of these deposits. Lateral changes in the nature of the sediments' are dramatic (Baker, 1920; Blanchard and Davis, 1929) and microstructures such as horizontal laminations (Udden, 1924) and desiccation cracks, ripple marks and mud balls (Lang, 1937) are abundant. These findings stimulated the development of more advanced models of evaporite sedimentation.

One of the problems posed by Ochsenius' bar theory is that continuous influx of dilute water and the precipitation of, for example, gypsum would eventually lead to saturation with sodium chloride. In some basins this has not occurred. King (1947) proposed a reflux model. This still incorporated a bar but the inflow at the surface was compensated for by a counterflow of denser brine at depth. The salinity of the water in the basin would theoretically remain constant so that only one salt was precipitated for long periods (Scrutton, 1953). Emery (1956) and Sugden (1963 a.b.) reported that the Persian Gulf had a warm saline counter-current flowing at depth through the Straits of Hormuz. Sugden (1963 b.) suggested that the model may be relevant to a number of other partially isolated bays including Sivash lagoon on the edge of the Aral Sea, two lagoons in Shark Bay, Western Australia and Lake St.Lucia in Natal, South Africa. It

should be pointed out, however, that these are all examples of shallow water environments. By applying the differential equations of motion of a fluid subject to estimated boundary conditions of the Silurian Michigan Basin, Briggs and Pollack (1967) were able to simulate the observed pattern of the Upper Salina Formation by utilizing a reflux model.

Further refinement of the model was required to explain the lateral variability of basin evaporite deposits. The Capitan Reef of the west Texas Permian Basin appears to have formed a barrier between seawater of normal salinity and saline water precipitating gypsum (Darton and Reeside, 1926; Lloyd, 1929). Yet, other bioherms seem to be contemporary with evaporites (Roth, 1942). Sloss (1953) proposed six different environments of evaporite deposition; brackish, marine, euxinic, penesaline, saline and super-saline. Several of these could be represented in a basin at any one time depending on the nature of the opening to the sea and frequency of influxes of landwater. It is true that in some cases changes in the nature of the sediments may reflect climatic variations (Brückner, 1953) but this is not feasible when the different deposits are interfingered (Carozzi, 1955).

A major problem posed by the barred basins model is that of resolution of precipitated salts by the inflowing dilute water. This would make accumulation of the most soluble salts unlikely. Sloss (1969) advocated a model based on layered solutions in deep basins. If the denser brine is at the bed of the basin the precipitated salts would not be redissolved. The main difficulty arising from this model is that there is no suitable mechanism for salt precipitation. Only if the brine becomes supersaturated with respect to any salt will that salt be precipitated. Saturation may result if ionic exchanges take place at the bed or if temperature fluctuations occur as Douglas and Goodman (1957) proposed. These are, however, unlikely. Sloss (1969) proposed that if the pynocline, the

boundary between the more dilute surface water and the underlying brine, was exposed at the surface by wind blowing the upper water to one side of the basins, evaporation, and thereby precipitation, would occur. Support for this model comes from the Dead Sea which possesses a pynocline and exhibits halite deposits at its bed (Bentor, 1961; Sloss, 1969). Neev (1979) suggested that periodic aeration of deep water bodies may sufficiently hinder the activity of sulphate reducing bacteria to allow gypsum sedimentation.

There is, in fact, very little empirical evidence in support of deep-water evaporite sedimentation. As mentioned earlier in this chapter,there is evidence for deep-sea carbonate sedimentation as a result of surface precipitation. This may involve bacterial reduction of calcium sulphate (Bavendamm, 1932; Neev, 1963; Bodenheimer and Neev, 1963; Friedman, 1965; Gevirtz and Friedman, 1966). Alternatively, siliceous diatoms which cause anigonitic "whitings" in the Persian Gulf (Wells and Illing, 1964) may be involved in the formation of sea-floor high-magnesium calcite deposits (Fisher and Garrison, 1967). Besides the difficulty of explaining the origins of thick evaporite deposits by any process other than deep-water sedimentation, the only support for the various models comes from analyses of the bromine content of saline deposits. Kühn (1953) found that the Stassfurt anhydrite had a bromine content equivalent to that found at a depth of 300 m, though the variation in the amount within varves was enigmatic. Goldsmith (1969) pointed out that the bromine content would increase as the brines became more enriched unless resolution of the salts occurred. Wardlaw and Schwerdtner (1966) rejected the suggestion, based on bromine concentrations, that halite-anydrite varves in the Prairie Evaporite Formation of Saskatchewan were precipitated in less than 0.50 m of water.They proposed a depth of 200 m of water and held that the bromine levels had been affected by post-depositional groundwater movement.

Post-depositional alteration of saline deposits has been advocated to explain many enigmatic features of evaporite sedimentation. The problems associated with gypsum and anhydrite will be dealt with below but the implication of Shearman's (1966) thesis that all halite and potassium salts represent replaced calcium sulphate deposits will be considered here.

Work undertaken by Van't Hoff on the Stassfurt salt deposits was based on the assumption that the minerals were produced by seawater and concentrated brine evaporation. Post-depositional changes were largely ignored despite evidence for these in the German deposits (Rose, 1972; Zimmerman, 1909) and elsewhere (Lacroix, 1897, 1898 c.). Later workers suggested that many of the more complex salts, there are over thirty saline minerals in the Stassfurt beds, are the products of super-saturated brine movement long after the normal processes of evaporite sedimentation have ceased (Gale, 1916; Clarke, 1924; Nolan, 1927). Schaller and Henderson (1932) and Mansfield (1937) held that glauberite ($Na_2Ca(SO_4)_2$) and polyhalite ($K_2Ca_2MG(SO_4)_4.2H_2O$) could be produced by alteration of anhydrite in the presence of other minerals in solution. Despite these indications that concentrated brines may alter the primary depositional character of evaporites, Kroenlein (1939) and Lang(1935, 1939) based their classification of the formations of the Ochoa Series on potassium content of the salts. More recently, the importance of post-depositional modifications has been acknowledged. While Bonte (1955) showed that significant changes in the crystal structures of saline minerals may occur, Stewart (1963) and Goldsmith (1969) pointed out that many of the rare evaporite minerals are the product of complex chemical processes. Minerals such as kieserite ($MgSO_4.H_2O$) and sylvite (KCl) are more common than predicted by the chemical ratios of seawater because they replace salts such as epsomite ($MgSO_4.7H_2O$), bloedite ($Na_2Mg(SO_4)_2.4H_2O$), kainite ($KCl.MgSO_4.3H_2O$), polyhalite and carnellite ($KMgCl_2.6H_2O$). Tanaka, Takasago and Komatsu (1950) did not use gypsum in their experimental de-

termination of equilibria in sea-salt systems because it caused excessive kieserite crystallization. Excessive, that is, in terms of the theoretical mineral speciation. Some workers have gone so far as to propose that all evaporite minerals can be accounted for in terms of primary calcite, gypsum and halite assemblages (Borchert, 1963 a.; Jones, 1963). Shearman (1966) argued that the enrichment of calcium carbonate and calcium sulphate in evaporite deposits relative to seawater and the depletion in magnesium and potassium toward the "top" of the sequence call for a greatly modified model of evaporite sedimentation. He proposed that in a calcium carbonate floored basin, excess sulphate after gypsum precipitation would combine with calcium liberated during dolomitization of the bedrock. Gypsum and anhydrite would be the only primary minerals and all the more soluble salts, including halite, represent the products of post-depositional diagenesis by replacement of the calcium sulphate minerals.

The model proposed by Shearman (1966) and refined by Shearman and Fuller (1969) is not based on deep-water sedimentation. Though the deep-water models have many disadvantages, such as not being able to account for certain evaporite assemblages, especially gypsum and anhydrite, they do provide a possible mechanism for the accumulation of thick salt deposits. This is particularly important when there is no evidence of tectonic subsidence at a rate capable of keeping pace with shallow-water sedimentation, as in the case of the Zechstein System (Borchert, 1963 b.). It has become necessary to develop an alternative model, preferably one which not only incorporates the increasing amount of detailed information available on the structure of evaporite deposits but also a model which does not depend on abnormal conditions for sedimentation to occur.

Varve structures have been recognized in many evaporite deposits. An under-

standing of their origin is critical to any interpretation of the mechanisms of evaporite sedimentation. The laminated anhydrite of the Castile Formation was first described by Udden (1924). Layers of anhydrite 0.50 mm to 2.0 mm thick are separated by dark bands of calcite admixed with bituminous material. The possible interpretations are that they represent fluctuations between chemical precipitation and organic sedimentation owing to annual temperature fluctuations (Udden, 1924; King, 1947) or periodic filling of a basin with seawater followed by evaporation depositing first calcite and organic matter and secondly gypsum and/or anhydrite (Cartwright, 1930). Lang (1950), Wardlaw and Schwerdtner (1966) and Richter-Bernburg (1968) identified similar varving in halite and sylvite deposits and found evidence of rhythmic oscillations suggesting a relationship with the 11.3 year solar cycle. These workers all interpreted the varves as deep-water deposits because of the uniformity of the strata over large areas. The only contemporary analogues for this pattern of sedimentation come from shallow-water environments. It is these which have increasingly been the focus of attention as potential environments of evaporite deposition.

Despite evidence from Lake Bonneville (Gilbert, 1890) and the Great Salt Lake (Talmage, 1889) that salt deposits may be the product of shallow water evaporation, little importance was attached to the process in terms of major evaporite sedimentation. Talmage (1889) showed that less than 1.5 m of water in the Great Salt Lake would precipitate over 0.20 m of salts. The main difficulty of applying this process to thick salt deposits is that incremental deposition may only be achieved through terrestrial subsidence. Morris and Dickey (1957) found that evaporation of seawater in a shallow estuary in Peru had precipitated over 10 m of thinly interbedded gypsum and halite though the depth of water in the estuary rarely exceeded 0.30 m.Similar lagoon sedimentation had been reported from the Persian Gulf (Bramkamp and Powers, 1955) and Perthuisot and Jauzein (1978) described gypsum precipitation below and within algal mats in shallow lagoon

water in the Khour el Aadid in Qatar. The Montallegro gypsum of Sicily seems to have originated in environments similar to these (Schreiber and Kinsman, 1975). At Pupuri Salina, Mexico, Moore and Hayes (1958) recorded similar sedimentary processes but without halite since this was redissolved during the rainy season. At Ojo de Liebre lagoon, also in Mexico, Phleger (1969) found that marine inundation followed by evaporation had built up halite and gypsum deposits over 3.0 m thick. Varving was well developed but represented rare floods, perhaps only one in twenty years. He speculated that thick salt deposits might build up with a steady rise in sea level. Even with a twenty year flood,300 m of salt could be deposited in 200,000 years, providing sea level rose that amount. This compares with the figure of 300,000 to 360,000 years proposed by Udden (1924) and Cartwright (1936), on the basis of annual varves, to account for 500 m to 600 m of Castile Formation anhydrite in Texas. The shallow-water model has many advantages over the various models of deep-water evaporite sedimentation. The greatest being that it has modern analogues. The varves of the Michigan Basin Salina Formation of halite and 0.50 mm of anhydrite, can be explained in terms of 1.15 m of seawater evaporating (Briggs and Lucas, 1954). This sequence is almost identical to the recent deposits of Ojo de Liebre (Phleger, 1969).

Since the mid-1960's the shallow-water evaporite model has been modified considerably. Many major structures common in evaporite deposits may be interpreted as products of shallow-water environments. Yet some features such as anhydrite nodules, festoon structures and certain mineral assemblages are suggestive of supratidal facies.

e. Sabkha sedimentation

'Sabkha' (various spellings have been adopted) is an Arabic term for a basin which is occasionally filled with water but is dry for most of the year. The name is applied to both inland seasonal lakes and to coastal lagoons. Here the

two will be referred to as continental and coastal sabkhas, the distinction being necessary because the processes of sedimentation associated with each have important differences.

Coastal sabkhas have been described in detail from many parts of the world. These include the lagoons of New Caledonia (Avias, 1952) and of Bonaire in the Netherland Antilles (Deffeyes, Lucia and Weyl, 1964; Lucia, 1968). The bulk of early work, however, was undertaken in a relatively small area on the Trucial Coast of the Persian Gulf. Here the sabkhas are developed in an area where marine water circulation is partially restricted by islands offshore (Evans, Kendall and Skipwith, 1964; Evans, Kinsman and Shearman, 1964). The coastal plains may be 8 km to 10 km wide, being made up of a wide range of marine sediments from quartz sands and milolite limestones to evaporite sediments (Kinsman, 1969). They cover 6,000 km^2 to 8,000 km^2 along 100 km of coastline in Abu Dhabi.

Shallow-water evaporites have been recorded in this area (Bramkamp and Powers, 1955) but offshore sedimentation consists primarily of calcium carbonate precipitation. Organic processes appear to influence the precipitation of aragonite in bottom sediments which are 75% carbonate minerals (Wells, 1962; Sugden, 1963 a.; Kinsman, 1964 b.; Wells and Illing, 1964) while beach minerals contain abundant coral debris (Kinsman, 1964 c.) and oolites (Kendall and Skipwith, 1969) though the composition is very variable (Pilkey, 1965). The presence of reefs offshore indicate that salinites are not high enough to precipitate evaporite minerals, generally 38‰ to 48‰ (Kinsman, 1964 c.; Evans, 1966). The evaporites are confined to an area from the upper intertidal zone to an area well above the highest tide zone (Butler, 1969; Kendall and Skipwith, 1969; Kinsman, 1969). It is in this zone that groundwater, originating both as seawater and continental water (Patterson and Kinsman, 1978, 1981) is concentrated through evaporation to reach salinities up to eight times that of seawater (Butler *et al.*, 1965) precipitating gypsum and halite in the subsurface zone as further evaporation

occurs (Shearman, 1963; Evans, Kendall and Skipwith, 1964; Kinsman, 1964 a.).

Characteristic mineral assemblages result from these processes of evaporite sedimentation. Besides halite and gypsum, dolomite is common locally. Dolomite was reported by Wells (1962) as being possibly a primary precipitate from saline groundwater but later work suggested that it was an alteration product of calcite and aragonite (Patterson and Kinsman, 1982). In areas with magnesium-rich groundwater,the saturated zone is extensively dolomitized at the expense of aragonite with dominates above the shallow water table (Curtis et al., 1963; Shearman, 1963; Kinsman, 1964 a.). The process of magnesium enrichment is closely related to gypsum precipitation since once this has occurred the Mg^{++}/Ca^{++} ratio may reach 11 over large areas and locally 35, seawater being 5 (Kinsman, 1965; Butler 1969). During dolomitization any excess sulphate will combine with the liberated calcium to precipitate more gypsum (Illing, 1963; Shearman, 1966).

$$2CaCO_3 + Mg^{++} + SO_4^{--} + 2H_2O \longrightarrow CaMg(CO_3)_2 + CaSO_4.2H_2O$$

Essentially the same processes were advocated by Deffreyes, Lucia and Weyl (1964) to explain extensive dolomitization in coastal lagoons in the Caribbean.

Other minerals found in sabkha evaporite sequences include celestite ($SrSO_4$) (Evans and Shearman, 1964) and huntite ($Mg_3Ca(CO_3)_4$) (Kinsman, 1965). Perhaps the most important mineral in sabkha environments in anhydrite. Despite the widespread occurrence of anhydrous calcium sulphate in thick evaporite sequences, usually at a depth of more than several tens of metres, there were, before the early 1960's, very few reports of anhydrite in near-surface sediments of recent age. Indeed experimental and theoretical considerations had suggested that anhydrite could not precipitate from seawater (Posnjak, 1940; Conley and Bundy, 1958). Evidence of anhydrite in sabkha sediments was first presented by Shearman (1963) from the Trucial Coast. It occurs as continuous beds up to 0.10 m thick (Butler et al., 1965) and as horizons of nodular anhydrite up to 1.5 m thick

containing over 50% anhydrite (Kinsman, 1964 a.). The nodules,which are composed of lath-like elongated crystals, are from 1.0 mm to 0.20 m in diameter. Although early workers suggested that the anhydrite was altered gypsum (Curtis et al., 1963; Evans and Shearman, 1964; Butler,1969),this is not always the case. Certainly, Kinsman (1969) reported anhydrite as discoid crystals which were pseudomorphic after gypsum but the nodular horizons were primary precipitated anhydrite (Kinsman 1964 a., 1965). The anhydrite is located in the supratidal zone, generally above the water table. Remobilization of calcium and sulphate ions in a highly saline environment at temperatures averaging 34°C at 1.0 m below the sabkha surface has enabled anhydrite nucleation. Though initial nucleation is probably associated with dehydration of gypsum, when sabkha surface temperature reach their maximum of about 60°C, these crystallites later act as nuclei for direct precipitation of anhydrite from groundwater brines. Anhydrite appears to be stable only where chlorinity exceeds 145‰ (Butler, 1969) or 150‰ (Hunt, et al., 1966 in Death Valley, California). On the inland fringe of the sabkha, anhydrite is undergoing hydration to gypsum when it comes into contact with terrestrial waters of low salinity (Kendall and Skipwith, 1969). Anhydrite has also been reported from Clayton Playa, Nevada (Moiola and Glover, 1965) while hemihydrate has been recorded there and from two basins in south-eastern California (Allen and Kramer, 1953). Hemihydrate is formed by topotactic alteration of gypsum at surface temperatures above about 45°C under conditions of low relative humidity. It has been reported from the surface of Trucial Coast sabkhas by Kendall and Skipwith (1969) and Kinsman (1969).

These sabkha mineral assemblages are of great significance since they provide analogues for some of the complex stratigraphic associations found in thick evaporite sequences. Yet the validity of the analogy can only be proven if the sabkha facies can be shown to be, firstly, potentially as thick as major evaporite sequences and, secondly,stratigraphically and structurally similar to them. Evans

et al. (1969) recorded Quaternary evaporite sequences near Abu Dhabi 8.5 m thick and Wood and Wolfe (1969) identified nine cycles of sabkha facies in a 53.7 m section of the Arab/Darb Formation off the Trucial Coast. Tectonic subsidence is not necessary for considerable thicknesses of evaporites to build up (Falcon, 1965). Pleistocene sea-level changes may account for thicknesses of about 200 m of evaporites. More significantly, the petrographic structures and textures associated with sabkha facies are being recognized in many thick evaporite sequences.

Gypsum is found commonly in three different zones of the coastal sabkha. In the intertidal,zone masses of small discoid crystals form a subsurface "mush" up to 1.0 m thick. A rubbery sheet of algae several centimetres thick usually covers the surface in this zone. The crystals grow interstitially but poikilitic inclusion of quartz grains and carbonate is not unusual (Kinsman, 1969). In the high-water zone and the zone above this,large lenticular gypsum crystals are found below the water table. These may be up to 0.25 m in diameter, flattened normal to the c axis (100) (Butler, et al., 1965) and form desert roses which may be exposed by deflation in the upper sabkha zone. Besides their large size, these gypsum crystals are characterized by the large number of poikilitically included sand grains. In the area of the sabkha bordering the continental deposits, the assemblage of gypsum crystals is again different. Here large sandy lenticular crystals up to 0.10 m in diameter are flattened parallel to the a-c plane (010) (Butler, et al., 1965). This gypsum may be the product of anhydrite hydration though Patterson and Kinsman (1978, 1981) stressed the role of continental waters precipitating gypsum in supra-tidal areas of coastal sabkhas. Kinsman (1964 a.) found that lenticular crystals, of all types, were randomly orientated. Masson (1955) suggested that the lenticular crystals from Laguna Madre, Texas, showed a tendency to vertical elongation and flattening. These structures have been recognized in many thick evaporite deposits but gypsum deposits frequently exhibit mottled and nodular textures in these sequences.

These textures are not represented in modern sabkha environments by gypsum deposits but rather by anhydrite deposits.

Several different textures of anhydrite have been identified in ancient evaporite sequences. Riley and Byrne (1961) recorded layered, mosaic or chicken-wire, flocculent, contorted or banded enterolithic, boudinage or lens and nodular forms Various models have been proposed to explain these differences in structure. While Phillips (1947) and Riley and Byrne (1961) argued that dehydration of gypsum deposits at depths and/or later rehydration, might produce mosaic and enterolithic structures,they did not discount the possibility of primary anhydrite forms or that the textures might be the result of flows of semi-fluid sediments under pressure. In contrast, Kerr and Thompson (1963), Murray (1964) and Rich and Pernichele (1965) felt that all these anhydrite structures are the product of gypsum dehydration during deep burial at about 300 m to 600 m below the Earth's surface.Chicken-wire textures may be formed by metamorphosis of beds of gypsum rosettes (Kerr and Thompson, 1963) and nodular anhydrite frequently exhibits pseudomorphs after gypsum or dolomite, again suggesting a replacive origin (Murray, 1964; West, 1967; Rooney and French, 1968). It must be acknowledged that much of this work was carried out when the evidence of primary anhydrite in recent sediments was very limited and experimental and theoretical chemical investigations cast doubt on the likelihood of primary anhydrite precipitation from natural waters. In effect, while geologists recognized that nodular and mottled textures in ancient anhydrite and gypsum deposits were most probably primary anhydrite structures, the lack of present day analogues of the depositional environments severely weakened the model (Bundy, 1956; Withington, 1961).

Detailed work in the sabkhas of the Trucial Coast has provided a wealth of information relating to primary anhydrite structures. Anhydrite is found ex-

tensively in the area of the sabkhas above the high-water line. The lath-like crystals elongated paralled to the c axis and flattened parallel to (100) are 2 μm to 3 μm thick and have a good pseudo-cubic cleavage (Kinsman, 1964 a.). The crystals are aggregated on nodules which grow by displacing the host sediment in the zone above the water table. Bedded anhydrite also occurs but the nodular form is more extensive. When well developed, the masses of nodules form a "series of interlocking saucer-shaped structures which are polygonal in outline" (Butler et al., 1965; Kendall and Skipwith, 1969) this may be the product of displacive growth (Walton, 1965). In cross-section, these small diapirs of nodular anhydrite have a festoon structure which has been recognized in many ancient evaporite deposits (Illing, 1965) including the Keuper halite beds of Cheshire (Hollingsworth, 1965). As the anhydrite layer thickens to 2.0 m or 3.0 m the lower layer becomes contorted under pressure, forming enterolithic bands characterized by tight folds and crenulations which Kendall and Skipwith (1969) likened to ptygmatic quartz veins. An important property of the anhydrite nodules is the thixotropy resulting from the crystal lath structure. With increasing pressure the nodules become more rigid as the crystals form a structure resembling a house of cards. In the presence of water, the loss of frictional adhesion at the crystal contacts caused complete collapse. In ancient deposits, this produces layers with orientated lath features (Shearman and Fuller, 1969).

The different textures listed by Riley and Byrne (1961) may be explained in terms of their diagenesis and to some degree post-depositional metamorphosism. Nodular, enterolithic and chicken-wire textures are all found in sabkha environments (Evans et al., 1969). They represent diagenic development of nodular anhydrite horizons. Flocculent and boudinage structures probably represent modifications of nodular beds under conditions of deep burial. Though Shearman and Fuller (1969) interpreted laminar anhydrite in the Middle Devonian Winnepegosis Formation of Saskatchewan as nodular collapse forms on the basis of orientated crystal laths, laminae may result in a number of ways. They may represent

pseudobedding of the host sediment before nodular growth. The dark bands may be the remnants of plankton influxes into a shallow-water environment, as early workers had suggested (Udden, 1924; Cartwright, 1930), or they may be algal mats which were surface stabilization layers from the intertidal or infra-tidal zone of the sabkha (Perthuisot and Jauzein, 1978). Buried algal mats have been reported from many sabkha environments (Kinsman, 1964 a.; Butler, 1969; Phleger, 1969; Hagan and Logan, 1974; Logan, 1974). Illing and Taylor (1967) argued that sabkha sediments showed very poor lamination but Shearman (1967), Lucia (1968) and Phleger (1969) have described well stratified sabkha sediments. West (1964, 1967) objected to the primary nodular anhydrite model on the grounds that pseudomorphic structures, for example in the Purbeck Beds of Dorset, suggest that remobilization of calcium sulphate after deep burial may produce nodular structures Another objection has been raised by Stewart (1967) who argued that thick sequences of polyhalite, kieserite and sylvite could not be the product of sabkha sedimentation. The textures associated with these minerals in some evaporite sequences suggest that they may be the products of post-depositional alteration of gypsum and anhydrite (Shearman, 1966). Furthermore, Shearman (1967) reported polyhalite from the sabkhas of the Trucial Coast. In effect,coastal sabkha facies are valuable modern analogues for the facies found in many thick evaporit deposits though a large amount of post-depositional alteration must be envisaged to account for assemblages of highly soluble evaporite minerals.

Continental sabkhas are quite different from coastal sabkhas. Not only do they originate in a different way but the sediments are also different (Kinsman, 1969 Though comparatively little information is available on continental sabkha formation, they are very extensive features. The great depressions of the northern Sahara, including the Chotts of Algeria and Tunisia and the salt lakes of the United States, Australia, Ethiopia, India and Central Asia are all examples of large continental sabkha environments. Though some of these basins are the

products of tectonic subsidence, the characteristics of the sedimentary processes result from two dominant processes. Firstly, flooding of the basins, usually to a depth of no more than 2.0 m, by rising groundwater and rarely inflowing surface waters. Secondly, aeolian deflation of the precipitated salts onto the surrounding terrain maintaining the depth of the basin (Blackwelder, 1931). The water table rarely falls more than 1.0 m below the sabkha floor thereby limiting deflation. Without deflation sediments would accumulate until the groundwater no longer rose above the floor and then surface sedimentation would cease. A low aeolian sediment input is a necessity for continental sabkha development.

Continental sabkhas are characterized by gypsum and halite deposits and frequently by a lack of calcium carbonate (Kinsman, 1969; Amiel and Friedman, 1971). The salts are usually well bedded, the strata being produced when a standing water body evaporates precipitating the minerals in solution. Amiel and Friedman (1971) identified three concentric zones in the Yotvata sabkha in Sinai. The outer vegetated zone where gypsum is the only evaporite mineral and an inner barren zone, where halite dominates about 3:1 over gypsum, are separated by a transitional zone with sparse vegetation and gypsum and halite crusts. The decrease in gypsum and sulphate content toward the centre of the sabkha was interpreted as being the result of bacterial sulphate reduction or precipitation of all of the gypsum in the transition zone (Amiel and Friedman, 1971). Concentration of sodium chloride in the central zone as the water body evaporates will lower the gypsum-halite ratio, increase gypsum solubility preventing precipitation (Risacher, 1978) and possibly stimulate bacterial reduction of gypsum, as in the Dead Sea (Bodenheimer and Neev, 1963). This pattern of salt distribution is also typical of Lake Eyre in Australia (Madigan, 1930; Bonython, 1956; King, 1956; Dulhunty, 1975) though in some areas the halite crust thins out, apparently where groundwater reaches the surface for most of the year. Similar features, called aïoun, are found on Chott Djerid in Tunisia (Bellair, 1957; Coque, 1962, pp. 136-139; Meckelein, 1977).

It is generally accepted that the saline water which precipitates the salts is groundwater which floods the sabkhas, though Bellair (1958) argued that capillary rise produced surface salt efflorescences similar to those described by Russell (1885) in Utah. There is considerable dispute over the origin of the minerals in solution. Holland (1912) suggested that the salt in the sabkhas of Rajasthan originated in the Rann of Kutch and had been transported inland by wind. Auden (1952) supported this opinion. Similarly, Madigan (1930) and a number of soil scientists (Downes, 1954; Bonython, 1956; Jessup, 1960 c.; Jessup, 1961 b.) proposed that the soluble salts in Lake Eyre and in the soils of South Australia were of oceanic origin, being deposited as dust or in rainwater. This may be the case in a number of basins, notably Lake Eyre, but Krishnan (1963) pointed out that the Rajasthan salt is probably derived from underlying Tertiary sediments. The Algerian and Tunisian Chotts are still more complex. Gautier (1953) pointed out that the catchment area for rainfall contributing to groundwater in the Chott region was made up of thick evaporite sequences as well as having a thin surface veneer of saline minerals deflated from the Chotts. Hence, much of the salt circulates within the catchment. In addition to this salt and any airborne and fluvial inputs (Meckelein, 1977),there is also the possibility that these sabkhas were flooded with marine waters during the Pleistocene (Milokhoff, 1954), as was the Sabkha el Melah near Zarzis (Busson and Perthuisot, 1977). This question will be dealt with below.

Despite the limitations on salt supply to the groundwater, the necessity for a certain amount of deflation to maintain the basin and the susceptibility of the salts to dissolution by rainwater influx (Atfeh, 1965; Krinsley, 1970),thick evaporite sequences can develop in continental sabkhas. Gypsum deposits over 10 m thick were deposited in the Lake Eyre Basin during the Pleistocene (King, 1956) and evaporite deposits have been recorded to a depth of 120 m in Californian playa basins (Allen and Kramer, 1953). As early as 1907 Lomas (a.b.) suggested that many of the evaporites of Triassic age found in Britain probably

originated as precipitates in desert pans and lakes.

f. Post-depositional alteration of gypsum and anhydrite.

Having examined the mechanisms of evaporite sedimentation including the models which involve post-depositional alteration of primary deposits in the development of the highly soluble salts, it is now possible to assess the processes which affect calcium sulphate deposits during burial and exhumation. These may be conveniently divided in two. Firstly, the effects of deep burial on the hydrous forms of calcium sulphate and, secondly, the effects of unloading on the anhydrous forms.

Though many geologists (Grimsley, 1904; Rogers, 1904, 1912, 1915; Wallace, 1914; Newland, 1921; Bowles and Farnworth, 1925; Brownell, 1931; Dunham, 1948; MacDonald, 1953; Ogniben, 1955) assumed that anhydrite could precipitate directly from seawater, a lot of evidence indicates that much of the anhydrite in ancient sequences is secondary after gypsum (Borchert and Baier, 1953). Pseudomorphic structures after gypsum have been identified in many anhydrite deposits (Rose, 1972, from the German Zechstein; Lacroix, 1897, in the Paris Basin; Jones, 1963 and Kerr and Thomson, 1963, from the Permian deposits of New Mexico; West, 1964, 1979, from the Purbeck Beds of Dorset; Rich and Pernichele, 1965, from North Dakota). Newland (1921) and Phillips (1947) pointed out that theoretically gypsum should convert to anhydrite as a result of pressure and temperature increases. A 30% to 40% reduction in volume will occur during dehydration as it is buried beneath surface sediments. Wallace (1914) had argued that this was not feasible since the dehydration process would result in a net increase in volume if the liberated water of crystallization is not removed. This factor was also stressed by Bundy (1956). He found that the anhydrite deposits of south-western Indiana show considerable evidence of recrystallization but the primary rock type is indeterminate. Goldman's (1952) examination of a complete core from the sulphur-

salt dome of Louisiana shows that all gypsum has been converted to anhydrite below about 350 m. MacDonald (1953) suggested that at depths greater than this, with a pore water chlorinity of 61‰, gypsum will dehydrate. This argues against McCormack's (1926) and Marsal's (1952) findings that increased pressures raise the gypsum-insoluble anhydrite transition temperature but MacDonald (1953) held that pressure effects on the solid phase would be 2.4 times greater than the liquid phase and so at 500 bars pressure the dehydration temperature would be 27°C. Unfortunately, there is no experimental verification of this. Nevertheless, MacDonald(1953) pointed out that temperatures are raised by geothermal flux to 42°C at about 1400 m and at 800 m pressure and temperature would be great enough to facilitate dehydration (see fig. 2.1). Even if MacDonald's theoretical consideratons of the effect of pressure on the solid states are rejected, the temperature increases occurring during deep burial would appear to be sufficient to overcome the effect of increased dehydration temperature that increased pressure has been proven to have (Marsal, 1952; Zen, 1965; Blount and Dickson, 1973). Taking Zen's (1965) over-estimate of the effect of pressure on the transition temperature of gypsum-insoluble anhydrite (+1°C/70.9$\pm$1.9 bars) the increase in geothermal temperature with depth will be greater than the increase in the transition temperature with the equivalent overburden. At depths of over 700 m gypsum will dehydrate to insoluble anhydrite even if the interstital water is not saline (Murray, 1964), if sodium chloride is present in the groundwater the conversion will take place nearer the surface. The time periods involved in these processes of burial and dehydration are so long (at least 10^{5} - 10^{6} years) that the slowness of anhydrite nucleation and solid-state dehydration of gypsum is unimportant.

The processes involved in the hydration or rehydration of anhydrite to gypsum as the overburden is stripped from the buried deposits are as complex as those that occur during burial. Bowles and Farnworth (1925) pointed out that in 1892

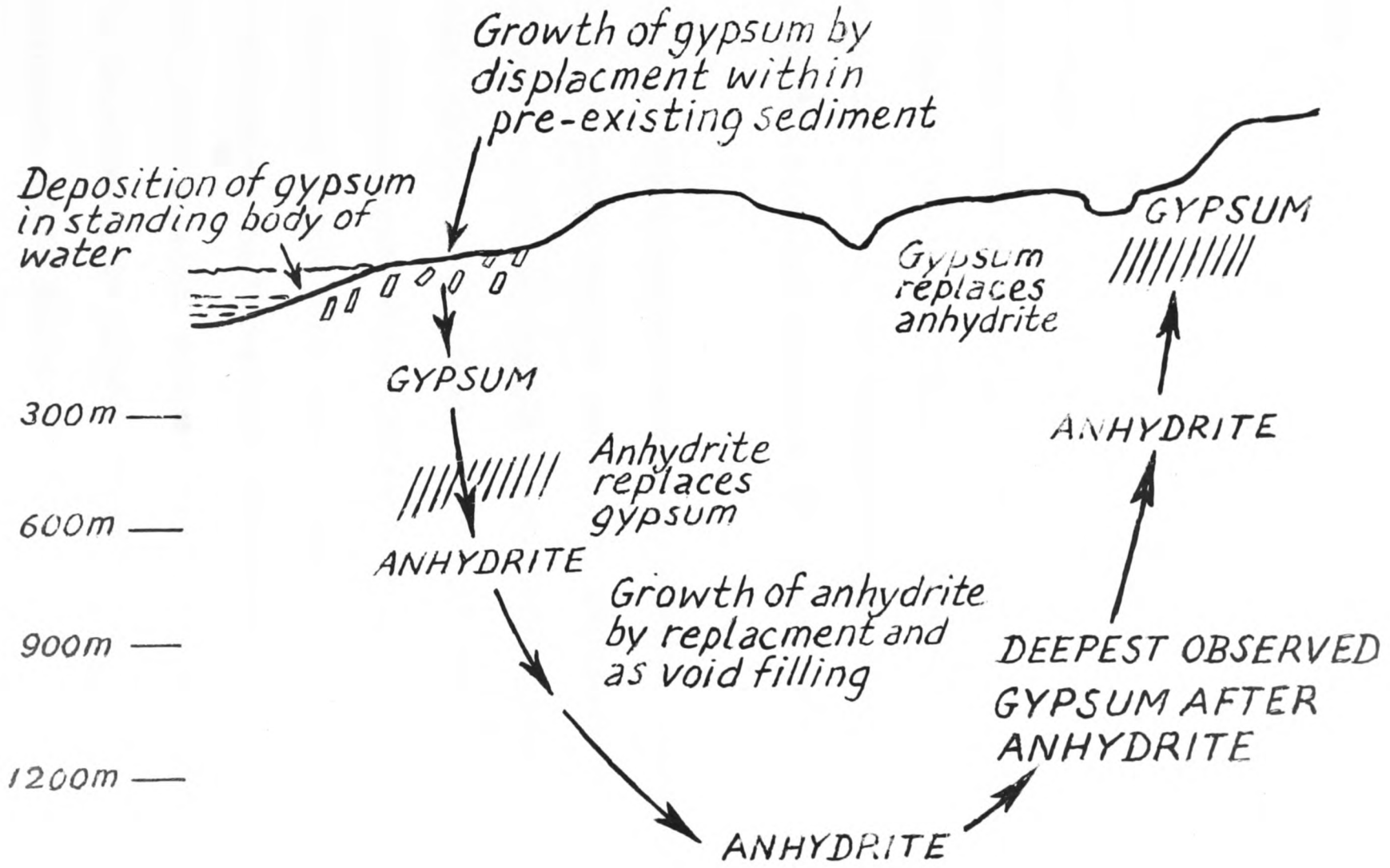


Figure 2.1
Gypsum - anhydrite diagenetic cycle (after Murray, 1964).

only three occurrences of anhydrite were known in the United States. By 1914, 60 were recorded (Rogers, 1915), largely as a result of deep mining of gypsum for the plaster industry. It became apparent that anhydrite hydrated to gypsum when it comes in contact with meteoric waters. Murray (1964) reported maximum depths of hydration at 1200 m (see fig. 2.1), but it is most marked at less than 100 m (Rogers, 1913; Wallace, 1914; Ogniben, 1955; Sund, 1959). Rarely, anhydrite is found at the surface with only a thin crust of hydrated gypsum (Wallace, 1914,at Gypsumville, Manitoba; Warden and Pallister, 1958, in the Somali Republic; Holliday, 1970, from Arctic areas). The importance of meteoric water in the hydration process is well exemplified in the case of the Amaranth deposit in Manitoba where an anhydrite zone over 1.0 m thick is sandwiched between 8.0 m of gypsum above and 3.0 m below (Brownell, 1931). Hydration has taken place on either side of the anhydrite layer where groundwater is available in limestones which lie above and below the calcium sulphate zone.

Secondary gypsum is characterized by several crystal textures including automorphic outlines of gypsum against anhydrite, porphyroblastic anhydrite remnants in gypsum and vein structures showing an anhydrite-gypsum transition through a solutional stage (Bundy, 1956). Gypsum porphyroblasts may develop when slow hydration in solution occurs while alabastrine (grains less than 50 μm) to granoblastic textures result when rapid hydration occurs with abundant water at temperatures well below the transition temperature (Holliday, 1970; Mossop and Shearman, 1973). Chicken-wire, enterolithic and nodular structures, associated with anhydrite development in sabkha environments, have been reported in gypsum deposits (Withington, 1961; Rooney and French, 1968). Hydration through solution is the most likely mechanism since solid-state crystal rearrangements involve extremely high activation energies (Nancollas, Reddy and Tsai, 1973). Step-wise hydration, through soluble anhydrite and hemihydrate may occur to a limited extent; Bundy (1956) reported hemihydrate from the anhydrite-gypsum hydration zone in Indiana but there is the danger that this derives from high temperature

dehydration of gypsum during coring or analysis (Milton, 1942). Whatever the process, it necessarily involves an apparent solid-state volume increase. It is this which has posed many questions to geologists and problems for civil engineers.

A theoretical volumetric increase of 50% to 60% during hydration was suggested by Newland (1921) and Henke (1976 a.), while Shearman et al. (1972) and Mossop and Shearman (1973) proposed 63%. The disruption to sedimentary structures such expansion would cause has been reported from numerous areas (Brownell, 1931, Manitoba; Goldman, 1952, Louisiana; Ogniben, 1955, Sicily; Sund, 1959, New Brunswick). Brune (1965) found evidence of explosive surface heaving of up to 7.0 m resulting from hydration of anhydrite in Texas. Notwithstanding this evidence, expansive structures are not associated with all secondary gypsum deposits. Bundy (1956) suggested that volumetric expansion may be limited by removal of some gypsum in solution. Furthermore, if the volume of water needed to affect hydration is taken into account, there is a 13% reduction in total volume (Dal Vesco, 1976).

1000 g $CaSO_4.2H_2O$ = 431 cm^3

Equivalent $CaSO_4$ (insoluble) + water = 473 cm^3

Wallace (1914) went so far as to suggest that this volume increase on dehydration might preserve gypsum at greater depths than would be expected in terms of the transition temperature/pressure relationship. It has been pointed out, however, that increased temperatures will overcome this barrier effect.

A certain amount of anhydrite solution is necessary to allow groundwater circulation and, thereby, gypsification. It is this which Bundy (1956) suggested results in the penecontemporaneous development of gypsum veins. Fibrous vein-fill originating during gypsification may be non-displacive, that is, it represents crystal growth from saturated solutions in hydraulic fractures produced during geological unloading (Shearman et al., 1972) though Bundy (1956) and Holliday

(1970) suggested it might be a displacive feature resulting from minor volumetric expansion during hydration. The volume of connate brine present in association with the anhydrite strata will determine the degree of hydraulic fracturing during unloading. Volumetric expansion will occur only if insufficient water is present for hydration, thereby necessitating water influx. Even if this is the case,solution during the influx may provide the additional space required without large-scale deformation.

Heaving of anhydrite-rich sediments has been reported not only by geologists from evaporite sequences, as mentioned above, but also by civil engineers. Yet the theoretical considerations suggest that expansion will only rarely be the result of hydration. Kaiser (1976) found that although pulverized anhydrite expanded 90% in two days after wetting, solid anhydrite showed no change over 100 days. Swelling of the powdered material was probably the result of colloid swelling. When the sample is loaded under a pressure of 10 kg/cm^2 settlement occurs on wetting and though hydration to gypsum occurs more slowly than at atmospheric pressure it is effected without volumetric expansion (Einfalt and Götz, 1976). This experimental data appears to contradict the evidence from road and railway tunnels in southern Germany and Switzerland where floor heaving and even collapse are associated with anhydrite bearing strata of Triassic age (Krause, 1976; Wohnlich, 1976). Krause (1976) recorded 1.03 m of floor heave in 27 years in the Wagenburg tunnel in Germany and Grob (1976) reported deformation rates of 11 mm annually in the Hanenstein railway tunnel and 1.0 mm per annum in the Bozberg tunnel in Switzerland. The Belchen motorway tunnel experienced heave of 0.90 m in a few months. Though the stresses occurring are high, 150 to 200 kg/cm^2, they do not reach the theoretical levels for anhydrite hydration of 400 to 500 kg/cm^2 (Grob, 1976). A critical factor, however, is the linear increase in stress over time which has been recorded (Müller-Salzburg and Götz, 1976). The fact that deformation and heave are most pronounced in strata where anhydrite is interbedded with Keuper marls and mudstones has prompted the

suggestion that the expansion is not the result of anhydrite hydration. Indeed, Krause (1976) pointed out that in the Gaildorf-Kappelsberg Tunnel, anhydrite has been completely converted to gypsum over 100 years without any heaving.

Using test tunnels in interbedded Keuper anhydrite-mudstone rocks, Henke (1976 b.) found that swelling was confined to the mudstones. The swelling pressures were directly correlated to the corrensite,
$Mg_2(OH)_2.SiAlO_{10}.Mg_2Al(OH)_6.Mg_3(OH)_2Si_3AlO_{10}.\frac{1}{2}Ca(H_2O)_x$, content.
Lippmann (1976) explained this feature in terms of properties of corrensite akin to sillicate montmorillonite and other smectites. The colloids of clays from 0.01 μm to 1.0 μm in size are coated with a 1Å thick layer of water molecules which causes little volumetric expansion. The smectites, however, have weak interlayer forces and intercrystalline expansion occurs on wetting. Corrensite molecular layers are 14Å thick but on wetting increase to 33Å. Even if the dry corrensite is a hydrous form with 29Å thick layers, the volume increase on wetting is 14%. Mudstones with 40% corrensite have swollen by 25% over 250 days (Henke, 1976 b.).

Hydration of anhydrite to gypsum may involve volumetric expansion if water is introduced into the system but in a closed system swelling is neither normal nor necessary during hydration. Several alternative mechanisms have been proposed to account for distortion, disruption or even explosions associated with anhydrite hydration. Besides the presence of potentially expansive clay minerals (Henke, 1976 b.; Lippmann, 1976) some sulphides may also cause dramatic volumetric increases. Dougherty and Barsotti (1972) pointed out that the oxidation of pyrite or marcasite to gypsum or malenterite ($FeSO_4.7H_2O$) may involve up to an eightfold increase in volume. A number of workers have pointed out that forces produced by preferential crystal growth in certain directions may cause expansion when hemihydrate (Lea and Nurse, 1949; Bundy, 1956; Chatterji and Jeffery, 1963) or anhydrite plasters (Gill, 1918) set to form

gypsum. The net volume decrease during hemihydrate hydration should be about 7% but acicular gypsum crystal growth or possibly colloidal swelling (Traube, 1919; Budnikov, 1928) results in a 0.5% gross volume increase which can crack a test-tube. It is feasible that similar processes occur during hydration of anhydrite to gypsum and they may be involved in the formation of gypsum bubbles or blisters (Quellungshohle) at the land surface (Von Gaertner, 1932; Breisch, 1978). These are hollow domes of gypsum usually 1.0 m to 2.0 m in diameter and 0.20 m to 0.30 m high, being perhaps a buckled stratum of hydrated anhydrite.

g. Some Karstic features associated with gypsum terrain.

Gypsum karst has been described from many parts of the Earth. It usually takes the form of sink holes and dolines associated with the collapse of underground cavities (Olive, 1957, in New Mexico; Warden and Pallister, 1958, Somali Republic; Wigley et al., 1973, in British Columbia). The gypsum content of Keuper marl appears to be an important factor in the development of the mardellen of Lorraine (Colin and Godard, 1962) and the Gutland of Luxembourg (Slotboom, 1963) Gypsum in the Keuper marl of the English Midlands (Shotton, 1968) may be dissolved to form water-filled depressions in the Delamere Forest area of Cheshire (Watson, 1976). Only rarely have well developed cave systems been associated with gypsum but Breisch (1978) described extensive systems in the USSR, which may be over 213 km in length and have well developed speleotherms. Gypsum caves may be formed by vadose or phreatic water. Since gypsum deposits are often relatively impermeable, compared with limestones which are well jointed, caves will develop to a lesser extent. The greater solubility of gypsum in pure water compared with calcite may not be a significant factor in gypsum karst development since calcite is more soluble than gypsum in water with a pH of less than about 4, though temperature and water chemistry are important rate-controlling factors.In effect, in areas with well developed soils and vegetation cover limestone may be more prone to solution than gypsum. In semi-arid areas,

however, the shortage of organic material at the surface will result in generally high soil and groundwater pH, reducing calcite solubility, high temperatures may partially counteract this. In dry areas, soluble salts, such as sodium chloride, may be abundant near the surface. Their presence in groundwater will enhance the solubility of gypsum. Gypsum karst may, therefore, be better developed than limestone karst in semi-arid and arid regions though the shortage of water in these areas will necessarily limit the extent of karstic landform evolution.

A major characteristic of gypsum caves is their susceptibility to collapse. This may be, at least in part, attributable to stresses associated with hydration of anhydrite in the sequence. It is also significant that gypsum deposits have a low density of joints compared with many limestone sequences. This results in a concentration of water flow along a few paths where the gypsum is subjected to accelerated solution. In effect, the cave systems are extremely active. James and Lupton (1978) pointed out that solution rates will vary greatly according to not only the rates of flow, temperature and chemistry of groundwater but also the mode of occurrence of the gypsum. While water may flow through fissures, interstitially or through an adjacent aquifer, the gypsum may be massive, detrital, secondary or laminar. James and Lupton (1978) applied first and second order kinetic rate law equations (Equations 4 and 5 above) of dissolution, for gypsum and anhydrite respectively, to expressions of particle geometry and changes in solute concentration of groundwater in order to determine the likely course of bedrock solution. Theoretical and experimental findings indicated that while gypsum dissolves at a steady rate, anhydrite can dissolve producing uncontrollable "runaway" conditions in which seepage flows accelerate. These properties apply to massive rock where solution occurs along fissures. If gypsum is a cementing agent in a conglomerate or in particulate form, the loss of cohesion resulting from relatively minor amounts of solution may cause severe loss of strength. In contrast, particulate or cementing an-

hydrite is less prone to catastrophic failure owing to dissolution than either massive anhydrite or particulate gypsum.

Calcano and Alzura (1967) described many instances of dam collapse in the United States which were caused by gypsum solution. Seepage across abutments increases rapidly after the reservoir is filled. The San Fernando Dam in California experienced leakage of 240 l/minute in 1964 despite regular grouting The Rattlesnake Dam in California leaks 800 l/minute. The McMillan Dam on the Pecos River in New Mexico was built in 1893, within twelve years the whole river flowed through caves which had developed after construction (Calcano and Alzura, 1967; Breisch, 1978). These engineering problems are very difficult to overcome. Notwithstanding the need for an extensive geological survey before dam construction (Fecker, 1976), waterproofing of the bedrock and cement injections into the fissures can alleviate the problems though they can never be completely successful since some fissures will always remain. Calcano and Alzura (1967) advocated the use of cement with a 5% to 6% iron oxide content and a low tricalcic aluminium content. Even so there is a danger that gypsum crystal growth (Fookes and Collis, 1975), expansion of sulphide minerals (Dougherty and Barsotti, 1972; B.R.E., 1977) or bacterial attack associated with sulphate reduction (Parker, 1951) will weaken the concrete and allow further gypsum or anhydrite solution.

An understanding of the geochemical processes involved in the deposition of gypsum and in its post-depositional alteration is essential for any examination of the origin of gypsum crusts. In a number of cases, specific types of gypsum crust are formed by processes closely akin to those active in sabkha environments. Sabkha gypsum deposits may in fact be classified as gypsum crusts. Furthermore, dehydration of gypsum and hydration of anhydrite and the potential physical changes associated with them are important to the development of many of the structural characteristics of the crusts.

CHAPTER 3

GYPSUM CRUST DISTRIBUTION

a. Definition of gypsum crusts.

Though gypsum crusts have been reported from all the continents few workers have attempted to define them. In chapter five the classification of different types of gypsum crust will be examined. It is necessary here to outline the limits of the investigation in terms of the types of gypsum accumulation which will be included therein. The definition employed here is based on chemical, structural and genetic factors. Only surface or near-surface (generally at less than 5.0 m depth) gypsum accumulations will be recorded. Descriptions of structures in bedrock gypsum deposits analogous with those found in surface sediments of recent age will be dealt with in chapter seven.

The term 'gypsum crust' implies a cemented deposit of calcium sulphate dihydrate but for the purposes of this investigation cementation will not be considered in the definition. This is necessary because extensive areas in Australia and north Africa, for example, are covered with powdery gypsum accumulations which French workers have termed encroûtements. These deposits appear to be closely related to indurated gypsum crusts, frequently overlying them. They are physically and chemically distinct from gypsum dunes even though it is probable that they are aeloian in origin. Gypsum dunes will also be included here since they are important features representing major redistributions of surface gypsum. Beyond this they are frequently indurated or have relic stabilization layers within them.

Another structural feature is the thickness of the gypsum deposit.

Though Buringh (1968) proposed a minimum thickness of 0.15 m for calciorthids, essentially calcium carbonate or gypsum indurated soils, this will not be adhered to here. The 5.0 mm thick sheets of gypsum crystals which lightly cement the surface materials in many areas in the Chott region of Tunisia (Douglas, 1912, p.206), may be regarded as gypsum crusts. Though these are physically insignificant features, they play an important role in limiting aeolian erosion and soil moisture movement and therefore will be included.

Although this definition is very broad in terms of the structural parameters the inclusion of all gypsum sediments will be avoided by employing genetic and chemical parameters. Surface exposures of gypsum deposits consisting of bedrock will not be considered except where the bedrock significantly influences the soils developed upon it. Buringh (1968) suggested that the regolith, or gypsum accumulation zone within it, should have 5.0% more gypsum than the C horizon in order to qualify as a gypsic calciorthid.

The exclusion of gypsum deposits on the grounds that they are bedrock may be unjustified in some cases where lagoonal or lacustrine evaporates are involved. The classification of thick gypsum beds associated with sabkha environments, such as those around the Sabkhah el Melah near Zarzis in Tunisia (Busson and Perthuisot, 1977) as either bedrock or gypsum crust is arbitrary. Similarly, some accumulations may contain less gypsum than underlying bedrock but may be classified as true crusts on the grounds that the consolidating cement is reprecipitated gypsum. The chemistry of the deposits is of great importance yet it is difficult to draw a hard and fast rule as to what percentage of gypsum merits the inclusion of the material in the category of gypsum crusts. Some of the dune sand deflated from Chott el Fedjedj in Tunisia contain little more than 20% gypsum yet it plays an important role in the development of crusts in the area. D'Hoore (1964) classified all soils containing more than 15% gypsum as gypsiferous. However, some crusts may be only partially

gypsiferous with an equal amount of calcium carbonate also acting as a cementing agent (Despois, 1935; Cointepas, 1971). The problem arises whether to define a crust with 40% gypsum and 40% calcium carbonate as a gypsum crust or a calcrete, while a crust containing 30% gypsum cementing quartz sand is a true gypsum crust.

For the purposes of this examination of the world-wide distribution of gypsum accumulations, only surface and near-surface materials which contain 5.0% more gypsum than the bedrock they overlie or have secondary gypsum as the principal cement will be considered. Only those deposits with more than 15% gypsum will qualify for inclusion. Deposits less than about 0.20 m thick will only be recorded if the gypsum horizon plays a significant geomorphological or pedalogical role.

b. Historical background to the investigation of gypsum crusts.

Geologists, geomorphologists and pedologists have reported and examined gypsum crusts from most deserts of the world. Perhaps because of the wide variety of interests represented by these fields, as well as the different nationalities of the workers, a broad spectrum of opinion exists. The earliest reports of gypsum crusts come from the Sahara, Piccard in 1868 (Agafonoff, 1936 b.) and Pomel and Pouyanne (1882) in Algeria, Duveyrier in 1860 (Maunoir and Schirmer, 1905), Pomel (1878), Bruun (1898, p. 39) and Pervinquière (1903, pp.234-239) in Tunisia.

Although lacustrine gypsum deposits were recorded elsewhere, for example by Russell (1855) in Utah, there were few reports of non-lacustrine gypsum deposits of recent age before the turn of the century. Some notable exceptions are the reports of German explorers in the Namib Desert of south-west Africa

(Stapff, 1887; Von François, 1893; Fleck, 1899). These early records were made either by explorers who recorded diverse details of the natural environments of the inhospitable regions through which they travelled, or by geologists conducting the first surveys of the new colonial lands of France and Germany. In the twentieth century as geomorphology and pedology evolved, more attention was paid to desert crusts which previously had merely been noted in passing.

French geomorphologists and soil scientists concentrated their efforts in the northern and central Sahara. French geomorphologists have been particularly interested in the role of crusts, calcrete and gypsum, in landscape evolution, notably the glacis, or pediment, and the value of the crusts as palaeoenvironmental indicators for the Quaternary (Jaranoff, 1937; Despois, 1940, 1954; Coque, 1951, 1955 a.b., 1958, 1960, 1962; Dalloni, 1953; Castany, 1955; Boulaine, 1958, 1961 a.b.; Awad, 1963). In contrast, German geomorphologists in the Namib, South America, Egypt and Iran were concerned more with the processes of weathering, erosion and sedimentation, (Kaiser 1923, 1926, 1927, 1929; Maack, 1924, 1963; Kaiser and Neumaier, 1932; Knetsch, 1937, in the Namib, Blanckenhorn, 1921; Passarge and Meinardus, 1933, in Egypt; Mortensen, 1927, 1933; Weischet, 1959, in Chile and Peru, Sedlacek, 1955; Bobek, 1959, 1963; Gabriel, 1964, in Iran). In the last twenty years this diversity in the approach to geomorphology and field areas has become less marked with, for example, many German workers in north Africa (Mensching, 1958, 1964, 1968; Meckelein, 1959, 1977; Beckmann, Scharpenseel and Stephan, 1972; Stephan, 1972; Brunnacker, 1973; Kulke, 1974; Brosche and Molle, 1975, 1976; Besler, 1977; Schwenk, 1977) and French geomorphologists in, for example, Afghanistan (Pias, 1972), South America (Risacher, 1978) and the Namib Desert. There remain, however, underlying differences in the aims and methods of desert geomorphological research in France and Germany though they are by

no means as schismatic as Tricart and Cailleux (1961) suggested when they subdivided hypotheses of crust formation into German per ascensum and French per descensum models.

Within the field of pedology these differences are not as pronounced largely because German soil scientists have undertaken little work in arid and semi-arid areas compared with the French, though the works of Ganssen (1960), Ganssen and Moll (1961) and Scholz (1963, 1968, 1971, 1972) in Namibia are notable exceptions. French pedologists have undertaken much work in Algeria and Tunisia on gypsum crusts largely related to their categorization in terms of the French soil classification system and their agricultural potential (Agafonoff, 1936 b.; Yankovitch, 1936; Gaucher, 1948 a.b.c., 1958; Gigout, 1958; Le Houérou, 1960, 1969; Bureau and Roederer, 1961; Pouget, 1968; Cointepas and Gaddas, 1971; Gaddas, 1971; Vieillefon, 1976). This system differs from that developed by Soviet pedologists in that it is based largely on the physical properties of soils, for example, their degree of development, calcium carbonate, gypsum and sodium chloride content. The Soviet system, though aimed at assessing agricultural potential, is predominantly a genetic classification. The work of Kovda (1958), Shumakov and Mikhovich (1960), Rozanov (1961), Akhvlediani (1962), Tolchel'nikov (1962), Kondorskaya (1967) and Petrov (1976) in Soviet Central Asia and the laboratory experimentation of Krupkin (1963) has provided considerable information on the nature of gypsum soils and, indeed, the processes involved in their development.

In comparison with the considerable amount of literature relative to gypsum crusts in German, French and Russian there is relatively little in English. Perhaps owing to the dominantly geological training of North American geomorphologists, the crusts of the United States of America and Mexico have only received detailed scientific attention in connection with work

relating to evaporite sedimentation or economic exploitation of the material (Hess, 1909; Hunt et al., 1966). The same is true of the gypsum crusts in Australia (Jack, 1921; De la Hunty and Low, 1958; Jessup, 1960 a.) and the Indian subcontinent (Auden, 1952; Srivastava, 1969). In the last two decades, however, soil scientists have undertaken more investigations of gypsum soils in the semi-arid areas of the United States of America and Australia as the demand for agricultural land, and with it the problem of soil salinty, increases.

Other areas of the Earth, such as the Middle East, parts of southern Europe and the Horn of Africa have gypsum crusts, some of which have been examined in detail. The result of all this work, in a wide variety of areas by scientists of different nationalities and from a range of disciplines, is a wealth of information relating to gypsum soils and crusts and an equal profusion of hypotheses proposed to explain the origin of the accumulations.

c. General distributional characteristics.

Here the work relating to gypsum soils, crusts and other forms of accumulation will be summarized. Some preliminary comments on the nature and possible origin of these deposits will be made but the details of these observations will be dealt with in chapters five and six. For the purposes of this chapter the types of gypsum accumulation will be subdivided, on a structural basis, into crusts, soils, bedded evaporites and sands. This may result in some confusion since these categories are not mutually exclusive. For example, some crusts may be bedded evaporites or consolidated sands and some soils may be essentially sands. Some of the large number of names of different types of crusts are listed in table 3.1.

TABLE 3.1

GENERAL	NORTH AMERICA	AUSTRALIA	CENTRAL ASIA	MIDDLE EAST	NORTH AFRICA	FRENCH	GERMAN
SURFACE	GYPSITE JESO	KASO	ARZYK GANDZH, GANJ BOZYNGAN BOZYNGEN BOZYGEN	JUSSE	DEB-DEB (?) GASSI (?)	CROÛTES-GYPSEUX	GYPS-KRUSTEN
POWDERY		FLOUR GYPSUM COPI, KOPI, KOPAI, CASO, KASO			FECH-FECH	ENCROÛTEMENT GYPSEUX	
INDURATED	GYPCRETE	ROCK (?) GYPSUM			DEB-DEB (?)	CROÛTES PAVI-MENTAUX	
SUBSURFACE CRUSTS (DESERT ROSE)	ROSETTE SAND-ROSE		REPETEKSKII		LOUS DEB-DEB (?) GASSI (?)	CROÛTES DE NAPPE ROSES DE SABLE	
OTHER SUB-SURFACE FORMS		SEED GYPSUM (?)	GAZHA, GAJA		TERCH		
NODULAR					RAS-KELB		

i. North America.

In the semi-arid and arid zone of the United States of America and Mexico there are large accumulations of gypsum at the surface in the form of playa and lagoon evaporites, dunes and various types of gypsum crusts and soils. Despite Glincka's (1927) brief reference to gypsum soil crusts, jeso, in North America, little work has been carried out on their nature or origin. Gypsum deposits associated with playas and the Quaternary lakes (Russell, 1885; Talmage, 1889; Gilbert, 1890; Allen and Kramer, 1953; Moiola and Glover, 1965; Davis, 1968 and many others) have been examined in some detail but largely to determine their economic potential for the plaster industry or the relevance of their mode of formation to the origin of thick evaporite sequences. The same is true of the gypsum deposits in coastal lagoons (Masson, 1955; Phleger, 1969). Similarly, the large areas of gypsum dunes in the south-western United States, both recent, such as those of the White Sands area of New Mexico (Talmage, 1932; Jaeger, 1957; McKee and Moiola, 1975) and the Great Salt Lake of Utah (Jones, 1953; Eardley, 1962) and ancient, such as the Upper Jurassic gypsum dunes of the Entrada and Todilto Formations of New Mexico (Tanner, 1965), have been examined in terms of their gypsum content only as potential economic resources or their role in evaporite sedimentation.

Much of the work relating to desert soils and crusts that has been undertaken in North America has concentrated on caliche (calcrete) rather than gypsum crusts. This does not seem to be because the latter are of restricted distribution in the area since Nikiforoff (1937) described examples throughout the semi-arid and arid zone. Bretz and Horberg (1949) and Brown (1956) stressed the importance of pedogenic processes in the formation of calcrete and more recently Arkley (1963, 1967) and Gile, Peterson and Grossman (1966) attempted to develop quantitative models of crust genesis and evolution based upon a

combination of climatic and edaphic factors, particularly temperature, actual evapotranspiration and soil moisture storage capacity. Though dealing specifically with calcium carbonate crusts, these models are also relevant to gypsum accumulation in the soil zone.

Specific description of surface gypsum deposits of recent age in North America are confined to the semi-arid and arid zone, particularly California and Nevada in the Death Valley and Mojave Desert areas where mean annual rainfall is less than 100 mm. Page (1972) mentioned gypsum soils in southern Arizona, New Mexico and in the Chihuahua area of Mexico. Hess (1909) described "gypsite" deposits over 3.0 m thick in the Mojave which he interpreted as products of capillary rise and surface evaporation of groundwater. The general description of soils in the area provided by Glincka (1927) and Nikiforoff (1937) suggested that gypsum soils are restricted to the zone with less than 100 mm to 150 mm of rainfall annually, mean temperatures above about $10^{o}C$ in January and $30^{o}C$ in July and that they are the product of capillary rise of groundwater into the soil zone. Hunt et al. (1966) proposed a similar origin for the gypsum and anhydrite crusts of Death Valley but also suggested that some may be the product of spring water evaporation, hence essentially tufaceous. In his analysis of uranium minerals associated with gypsum crusts and calcretes, Carlisle (1978) reported "gypcrete" 3.7 m to 4.0 m thick from Gypsum Wash, Nevada. Detailed descriptions were not provided, other than that the material is conglomeritic, therefore it is not possible to suggest how the deposit originated. Also in Nevada, Lattman and.Lauffenburger (1974) suggested that calcretes may form directly from underlying Permo-Triassic gypsiferous beds through the action of sulphate reducing bacteria. Table 3.2 summarizes the information available on near-surface gypsum deposits in North America and South America.

TABLE 3.2

Type	Form	Author(s)	Location	Thickness (m)	Gypsum (%)	Origin
Surface	General	McCauley et al. (1977)	Peru	-	-	Fog moisture evaporation
		Nijensohn (1960)	Argentina	-	50	-
		Nikiforoff (1937)	California, Utah		-	Lacustrine, capillary rise or illuvial
	Columnar	Hunt et al. (1966)	California	1.0 - 2.0	-	Spring water evaporation
	Powdery	Van Everdingen (1969)	Alberta, British Columbia	-	-	Spring water evaporation
		Hess (1909)	California	3.0	-	Spring water evaporation
Subsurface	Desert rose	Masson (1955)	Texas	-	-	Subsurface evaporation of lagoon water
		Risacher (1978)	Bolivia	-	-	Groundwater evaporation
Evaporitic	Massive or crystalline	Allen & Kramer (1953)	California	At depth	97.3 hemihydrate	-
		Carlisle (1978)	Nevada	3.7 - 4.0	-	Playa deposit
		Deffeyes et al. (1964)	Netherlands Antilles	-	-	Lagoonal evaporite
		Lucia (1968)	Netherlands Antilles	0.1 - 0.8	-	Lagoonal evaporite
		Moiola & Glover (1965)	Nevada	-	hemihydrate & anhydrite	-
		Moore & Hayes (1958)	Mexico	0.10	-	Playa deposit
		Morris & Dickey (1957)	Peru	10	-	Shallow water evaporite (estuarine
		Phleger (1969)	Mexico	3.0 with halite	-	Supra-tidal/tidal evaporite
		Risacher (1978)	Bolivia	2.0	-	Groundwater or shallow-water evaporite
		Russell (1885)	Utah	2.0	-	Capillary rise of groundwater

TABLE 3.2 (cont.)

Type	Form	Author(s)	Location	Thickness (m)	Gypsum (%)	Origin
Dunes		Eardley (1962)	Utah	-	66	Playa deflation
		Jaeger (1957)	New Mexico	10 - 15	94	Playa deflation
		Jones (1953)	Utah	10 - 30	-	Deflation of selenite from clays
		McKee & Moiola (1975)	New Mexico	7.0 - 10	-	Playa deflation
		Talmage (1932)	New Mexico	-	-	Playa deflation
		Tanner (1965)	New Mexico	40	-	Fossil dunes from deflated coastal deposits

ii. South America.

Very little is known of the distribution of gypsum soils, crusts and other surface deposits in South America. The work of Lucia (1968) in the Caribbean and Morris and Dickey (1957) on the coast of Peru was concerned solely with lagoonal and estuarine sequences. Dregne (in McGinnies, Goldman and Paylore, 1968) reported extensive gypsum beds in the Atacama Desert of Peru but they were neither described in detail nor were their origins alluded to. Mortensen (1927) described sodium chloride crusts up to 1.0 m thick, apparently of pedogenic origin, in northern Chile but the only other gypsum crusts reported in the area are thin coatings of gypsum on the ground surface probably deposited by evaporating fog moisture (McCauley, Grolier and Breed, 1977). Similarly, in the Monte-Patagonia area, soils with up to 50% gypsum have been reported (Nijensohn, 1960) but their extent and context are not recorded. The scarcity of references to gypsum deposits in South America would appear to reflect more the aims of geomorphologists and soil scientists working there than an absence of the features. The Atacama desert has a climate very similar to the Namib, where there are extensive gypsum crusts, and coastal gypsum deposits exist in both areas. In the case of the Monte-Patagonia, though rainfall levels are about the same as those in southern Tunisia, for example, the mean annual temperature is lower. This may significantly affect the development and preservation of gypsum deposits. It is interesting to note, however, that there are gypsum crusts in Bolivia around the Jayu Kkota salina (Risacher, 1978).. Though annual rainfall is less than 300 mm, mean monthly temperatures are low, between about 10° and 15°C throughout the year. While some of the crusts are fossil lacustrine evaporites, some examples up to 2.0 m thick are the product of groundwater mixing causing gypsum precipitation. The gypsum is derived from local sandstones.

iii. Australia.

As in North America the earliest references to recent gypsum deposits in Australia were made in connection with resource evaluation for the building industry. Jack's (1921) survey of the gypsum deposits of South Australia subdivided them into three types, namely, flour gypsum, also called kopi, caso, or "gypsite", seed gypsum and rock gypsum. Rock gypsum is essentially bedrock, while seed gypsum consists of small crystals which have been deposited in evaporating pans. Sometimes the seed gypsum is interbedded with laminar calcretes, deposited when terrestrial water first began to evaporate, but when exposed at the surface the crystals break down in the heat and are deflated. The redeposited material is flour gypsum. Old deposits of flour gypsum are consolidated by material Jack (1921) termed travertine. This may be a type of gypsum crust. The origin of the gypsum in lake basins, such as Lake Frome in Western Australia or Lake Eyre in South Australia is uncertain. Seed gypsum deposits may be over 20 m thick in Lake Eyre (Madigan, 1930; King, 1956) while the surrounding Pleistocene land surface has well developed gypsum crusts (Wopfner and Twidale, 1967; Jones and Segnit, 1972). Though marine inundation of the Lake Eyre Basin is discounted, Bonython (1956) pointed out that the soluble mineral content of the basin supports the hypothesis that the salts are marine in origin but have been transported from the sea (Termier and Termier, 1963). Jennings (1954) estimated a deposition rate of soluble salts in rainwater of 125 to 750 kg/ha in Western Australia and 12.5 to 7.5 kg/ha in the interior. The salts accumulate in pans which are the foci of local drainage and seepage. The salts are deflated during dry periods and deposited as discrete lunettes (Hills, 1940) or mantle the surrounding land surface. This model is supported by evidence from Western Australia (De la Hunty and Low, 1958) where dry pans have a seed gypsum lunette on their downwind side and further downwind from this there is a lunette composed of

flour gypsum.

Cyclic salts have also strongly influenced the soils of a large area of Australia where there is insufficient rainfall to flush the soluble minerals out of the soil zone (Downes, 1954). The effects of high sodium chloride contents of soils during Quaternary arid periods would not be removed by leaching the salt through the soil in a later wet phase because clays and calcium and magnesium salts would also be lost owing to the mobilizing effect sodium chloride has upon them. In the present day semi-arid zone of Australia, gypsum accumulations are common in a number of soil types. Though Hallsworth, Gibbons and Darley (1951) argued that Australian desert soils differ only according to the degree of deflation they have experienced, Jessup (1960 a.b.c.) stressed the importance of aeolian deposition of gypsum and clays deflating from the playas. In the south-eastern part of the arid zone of Australia, essentially South Australia, a subsurface zone 1.0 m to 2.0 m thick composed of 75% to 86% gypsum (Jessup, 1960 a.) is overlain by 0.10 m to 0.30 m of largely montmorillonitic clay (Ollier, 1966). Though some of the gypsum may originate during the weathering of pyrite in Cretaceous soils (Jessup, 1960 c.), Jessup (1960 a.) suggested that juxtapositioning of gypsum and clay has resulted from two different phases of deflation from dry lakes. The first, removing evaporated gypsum and the second, clays. Jessup (1961 a.b.) pointed out that the hypothesis of aeolian origin was supported by the characteristics mantling of the surface with gypsum, calcium carbonate and clay, regardless of the topography. The presence of coarse gravels and cobbles on top of the gypsum horizon was difficult to explain. It has been suggested that they were emplaced by later fluvial deposition or upward movement of coarse material as a result of montmorillonite volume changes during wetting and drying (Jessup, 1960 a.; Ollier, 1966; Cooke, 1970).

TABLE 3.3

Type	Form	Author(s)	Location	Thickness (m)	Gypsum (%)	Origin
Surface	General	Jones & Segnit (1972)	S Australia	-	-	Weathering profile (?)
		Madigan (1930)	S Australia	-	-	Aeolian deposit
	Powdery	De la Hunty & Low (1958)	W Australia	-	-	Playa deflation, deposit
		Jack (1921)	S Australia	-	-	Deflated lacustrine evaporite
Subsurface	Mesocrystalline	Jack (1921)	S Australia	-	-	Aeolian deposit
		Jessup (1960a.b., 1961a.)	S Australia	0.6 - 2.0	75 - 86	Aeolian deposit
		Jessup (1960c.)	S Australia	0.15	-	Weathering profile (replacive?)
		Ollier (1966)	S Australia	-	-	Precipitation from groundwater
		Wopfner & Twidale (1967)	S Australia	0.6 - 1.5	-	Weathering profile
Evaporitic	Massive or crystalline	Avias (1953)	New Caledonia	-	-	Supra-tidal evaporite
		Bonython (1956)	S Australia	2.5	-	Lacustrine evaporite
		Carlisle (1978)	W Australia	-	-	Lacustrine evaporite (?)
		Jack (1921)	S Australia	0.5	-	Lacustrine evaporite
		King (1956)	S Australia	1.5	-	Lacustrine evaporite
		Madigan (1930)	S Australia	15 - 20	-	Lacustrine evaporite (deflated?
Dunes		De la Hunty & Low (1958)	W Australia	-	-	Playa deflation
		Jack (1921)	S Australia	-	-	Playa deflation
		Termier & Termier (1963)	S Australia	-	-	Playa deflation

Gypsum soils are, in effect, extensive features of the South Australian and Western Australian landscape (see table 3.3). Though their origin appears to be closely related to deflation of the gypsiferous source material from playas and dry lake basins, there is not an adequate model to explain the position of the gypsum horizons within the soil profiles and the blanket distribution of these soils over the land surface.

iv. Central Asia and the Indian sub-continent.

Kondorskaya (1967) described areas with soluble salt accumulation in the soil zone in many areas of the U.S.S.R. stretching from Kiev to Yakutsk. The main areas of gypsum soils and crusts are restricted to the driest part of the U.S.S.R., excluding the Arctic zone, namely Azerbaizhan, south of the Caucasus bordering the Caspian Sea, and the east of the Caspian as far as the Mongolian border. In Azerbaizhan, subsurface accumulations of gypsum up to 1.0 m thick, called gandzh (ganj) or gazha (gaja), may contain 95% gypsum (Kovda, 1958; Petrov, 1976) on average the figure is lower, 50% to 85% (Akhvlediani, 1962). Though Petrov (1976) maintained that many of these accumulations are the product of groundwater fluctuation and evaporation precipitating the soluble salts, Shumakov and Mikhovich (1960) found evidence of considerable soluble salt deposition at the surface by fog. Near Elista, north of the Caucasus, they estimated an annual deposition of 475 kg/ha. Tolchel'nikov (1962) argued that this process of impulverization was of equal importance in the development of gypsiferous soils as precipitation from groundwater and in pans (see chapter six).

To the east of the Caspian there are extensive areas of gypsum soils in Kazakhstan, Tadzhikistan, Turkestan and Uzbekistan. Fersmann and Wlodawetz (1926) described gypsum crusts in association with silcrete in the Kara-Kum

and large areas of gypsum soils have been reported from the northern slopes of Kopet-Dag (Kondorskaya, 1967). The Ust-Urt plateau between the Caspian and Aral Seas has well developed gypsum crusts which may be accreting at present (Kovda, 1958; Kondorskaya, 1967; Dregne, 1976). A large area of Kazakhstan, from the Syr-Dar'ya river and the northern slopes of Talasskiy Ala-Tau in the south to the Turgay river in the north and to the Dzungarian Ala-Tau in the east, has gypsiferous soils (Kondorskaya, 1967). The gypsum accumulations occur in an area where mean annual rainfall does not exceed 200 mm and is less than 100 mm over large areas. Rozanov (1961) and Kondorskaya (1967) suggested that the bacterial weathering of pyrite bearing limestone bedrock may lead to gypsum accumulation in the soil zone as Vialov (1946) postulated. Lacustrine evaporation is responsible for some of the surface deposits (Fersmann and Wlodawetz, 1926; Kovda, 1958) but only Rozanov (1961) and Dregne (1976) put forward the proposition that aeolian or atmospheric deposition may be important. This model may be particularly valid in Ust'Urt, Uzbekistan and Tadzhikistan where present day and late Quaternary lacustrine evaporite deposits are undergoing deflation.

Beyond Soviet Central Asia there are relatively few reports of gypsum soils or crusts. Dregne (in McGinnies, Goldman and Paylore, 1968) stated that the surface materials of the Gobi are predominantly non-gypsiferous but McCauley, Grolier and Breed (1977) reported gypsum deposits in the Lop Nor Basin of Takla Makan. These appear to be lacustrine evaporites. The gypsum crusts of the Dacht e Margo of Afghanistan (Pias,. 1972) are not lacustrine in origin. They are subsurface features up to 0.80 m thick below 0.10 m to 0.15 m of sand. The proximity of the Hassun Helmand suggests that the gypsum may have been deflated from the evaporating lake waters. Srivastava (1969) described similar deposits 1.0 m to 2.0 m thick with up to 87% gypsum in the Pali, Nagaur, Chura, Bikaner and Barmer districts of Rajasthan. He termed these gypsite. Though

TABLE 3.4

Type	Form	Author(s)	Location	Thickness (m)	Gypsum (%)	Origin
Surface	General	Kondorskaya (1967)	USSR	-	-	Groundwater; evaporation or weathering product
		Kovda (1958)	USSR	0.5 - 1.0	60 - 95	Groundwater or lacustrine evaporation
		McCauley et al. (1977)	Tarim Basin	-	-	-
		Petrov (1976)	USSR	-	80 - 95	Possibly capillary rise of groundwater
		Rozanov (1961)	USSR	0.1 - 0.6	-	Possibly bacterial activity
		Tolchel'nikov (1962)	USSR	-	20	Aeolian deposit
		Vialov (1946)	USSR	1.0	-	Weathering of pyrite
	Columnar	Dregne (1976)	USSR	-	-	Lacustrine evaporation (?)
		Fersmann and Wlodawetz (1926)	USSR	0.5	-	Relic lacustrine evaporite
Subsurface	Desert rose	Tolchel'nikov (1962)	USSR	0.4 - 0.6	-	Lacustrine evaporation (?)
	Mesocrystalline	Akvlediani (1962)	USSR	0.5 - 2.0	50 - 85	Precipitation from groundwater
		Petrov (1976)	USSR	0.2 - 0.25	-	Precipitation from groundwater
		Pias (1972)	Afghanistan	0.6 - 0.7	-	Precipitation from groundwater
		Rozanov (1961)	USSR	0.1 - 0.6	-	-
		Srivastava (1969)	Rajasthan	up to 2.0	up to 87	Groundwater evaporation
Evaporitic	Massive or crystalline	Auden (1952)	Rajasthan	-	-	Lacustrine evaporation
		Glennie (1970)	India	-	-	Coastal shallow-water evaporite
		Holland (1912)	Rajasthan	1.5	-	Lacustrine evaporation
		Krishnan (1963)	Rajasthan	-	-	Lacustrine evaporation
		Srivastava (1969)	Rajasthan	up to 2.0	up to 87	Lacustrine evaporation
Dunes		Glennie (1970)	India	-	-	Deflation of coastal sediments

many of these deposits are associated with pans, Srivastava (1969) felt that they are often precipitated from groundwater and are not invariably lacustrine evaporites. The primary source of the gypsum in the Thar Desert was discussed in chapter two. Suffice it to repeat here that while Holland (1912) advocated aeolian transport of salts from the Rann of Kutch, from where Glennie (1970) reported gypsum dunes, Krishnan (1963) held that the gypsum was derived from Tertiary bedrock.

All of the near-surface gypsum accumulations described from Soviet Central Asia, Afghanistan, Takla Makan and the Thar are found in areas with less than 300 mm of rainfall annually and often less than 100 mm. Though mean temperatures may be below freezing in the mountainous areas in January, in July mean temperatures are above 25° to 30°C. the characteristics of the deposits that have been reported are summarized in table 3.4.

v. The Middle East.

For the purposes of this survey of the distribution of gypsum crusts, the Middle East is defined as the countries bordering the Arabian Peninsula from Iran in the east, Turkey to the north and Egypt in the west.

Though most of Turkey receives more than 300 mm of rainfall annually the area around the Konya Basin in the centre of the country receives less and gypsum soils have been reported (De Meester, 1970; Driessen, 1970). They appear to be associated with poorly drained, marshy soils in which gypsum is precipitated as water evaporates. In neighbouring Iran, gypsum soils and crusts have been described in the Lut and Baluchestan (Sedlacek, 1955; Bobek, 1959, 1963; Gabriel, 1964). Extensive gypsum crusts are found in the south-east of the country, some with well developed polygonal structures (Gabriel, 1964).

Rainfall in this area is below 300 mm annually and locally below 200 mm. Temperatures are high throughout the year, mean temperatures for January are above 15°C and for July above 30°C. The Hamun-e Jaz Murian is a possible source of gypsum for deflation but there are also extensive evaporite deposits as bedrock (Bobek, 1959). In northern Iraq, Smith and Robertson (1962) described gypsum soils developed on "gypsum rock"; this was probably a well indurated gypsum crust containing over 50% gypsum. Tucker (1978) suggested that the gypsum crusts were aeolian deposits dating from the late Pleistocene subsequently covered by alluvium. The Iraqi crusts are probably the same as those found in neighbouring Syria (Dregne, in McGinnies, Goldman and Paylore, 1968). The rainfall in the area is 200 mm to 300 mm per annum.

On the Arabian Peninsula few examples of gypsum soils and crusts have been reported. Considering there are extensive gypsum deposits in coastal and continental sabkhas throughout the area (Johnson et al., 1978) it seems unlikely that other types of gypsum accumulation are absent. Philby (1933) recorded numerous examples of surface gypsum crusts throughout the Jafura and northern Rub'al Khali. While some were desert rose forms (Campbell-Smith p. 382 in appendix to Philby, 1933), others were surface crusts possible of lacustrine origin. Brunsden, Doornkamp and Jones (1976) reported evaporite gypsum crusts in Bahrain. These crusts contain about 83% gypsum. Gypsum dunes and crusts not of evaporitic origin were also described. Rainfall in the area is between 100 mm and 200 mm a year, the same as a large part of the Peninsula, therefore if gypsum soils are absent on the mainland the most likely explanation is a paucity of calcium sulphates in groundwater and wind blown material.

In the areas of the Middle East bordering the Mediterranean, rainfall is less than 300 mm per annum from Gaza westwards and falls sharply inland to less than 25 mm in central Egypt. Dan et al. (1972) described gypsum deposits

TABLE 3.5

Type	Form	Author(s)	Location	Thickness (m)	Gypsum (%)	Origin
Surface	General	Blanckenhorn (1921)	Egypt	-	-	-
		Bobek (1959)	Iran	-	-	-
		Dregne (1968)	Syria	-	-	-
		Glennie & Evamy (1968)	Trucial Coast	-	-	-
		Passarge & Meinardus (1933)	Egypt	-	-	Alteration of calcrete
		Sedlacek (1955)	Iran	-	-	-
	Columnar	Brunsden et al. (1976)	Bahrain	-	-	-
		Butzer & Hansen (1968)	Egypt	-	-	Gypsum-rich fluvial sediment
		Gabriel (1964)	Iran	-	-	Cemented aeolian deposit
		Tucker (1978)	Iraq	0.3	-	Cemented aeolian deposit
Subsurface	Desert rose	Brunsden et al. (1976)	Bahrain	-	40	-
		Butler (1969)	Abu Dhabi	-	-	Precipitation form groundwater
		Evans et al. (1969)	Abu Dhabi	total 8.5	-	Supra-tidal evaporite
	Mesocrystalline	Dan et al. (1972)	Israel	-	-	-
		Driessen (1970)	Turkey	-	-	Capillary rise of groundwater
		Fathi et al. (1973)	Egypt	0.05 - 0.45	-	Capillary rise of groundwater
		Smith & Robertson (1962)	Iraq	-	50	Alteration of bedrock
Evaporitic	Massive or crystalline	Amiel & Friedman (1971)	Jordan Valley	0.6	-	Sursurface groundwater evaporite
		Brunsden et al. (1976)	Bahrain	-	80 - 90	Lacustrine evaporation
		Butler (1969)	Abu Dhabi	0.3	-	Supra-tidal/tidal evaporite
		Evans (1966)	Trucial Coast	0.6	-	Supra-tidal evaporite

TABLE 3.5 (cont.)

Type	Form	Author(s)	Location	Thickness (m)	Gypsum (%)	Origin
Evaporitic (cont.)	Massive or crystalline (cont.)	Johnson et al. (1978)	Saudi Arabia	0.2	-	Subsurface groundwater evaporit
		Kendall & Skipwith (1969)	Abu Dhabi	0.2	-	Supra-tidal/tidal evaporite
		Kinsman (1964a.b., 1965)	Trucial Coast	1.0 - 1.25	50	Supra-tidal evaporite
		Perthuisot & Jauzein (1978)	Qatar	-	-	Shallow-water lagoonal evaporite
		Shearman (1963)	Trucial Coast	1.2 anhydrite	-	Supra-tidal/tidal evaporite

overlain by calcareous serozem soils in Israel and Sofer (1978) reported coastal and continental sabkha gypsum from northern Sinai. In Egypt, similar lacustrine or sabkha gypsum accumulations have been described by Butzer and Hanson (1968) in the Aswan areas but a number of workers have described extensive surface crusts which are not surface evaporites. Blanckenhorn (1921) reported soil gypsum deposits from Cairo to Thebes along the Nile Valley. Passarge and Meinardus (1933) observed that while in Palestine calcretes predominate, Egypt, being drier, has gypsum crusts. Their explanation that aridity results in the alteration of calcrete to gypsum seems unjustified in terms of work carried out in other areas, except the Central Namib Desert where Martin (1963) proposed a similar model of gypsum crust genesis. Glinka (1927) mentioned gypsum crusts associated with poorly drained soils in Egypt and Fathi et al. (1972) described massive gypsum layers up to 0.25 m thick in soil horizons south of Manzala Lake. West, Ali and Hilmy (1979) reported nodular gypsum from coastal sabkhas on the Mediterranean coast of Egypt. These workers felt that the gypsum precipitated from groundwater and soil moisture.

Though there are large areas in the Middle East which are apparently devoid of gypsum crusts, even in the arid zone, in some areas there are extensive accumulations which apparently originated in several different ways, in sabkhas by evaporation, by precipitation from groundwater and perhaps by aeolian deposition or some other process in areas where surface water or groundwater is not available. The existing information is summarized in table 3.5.

vi. North Africa.

In Mediterranean north Africa the distribution of gypsum crusts and soils is enigmatic. While they are rare in Libya and Morocco, the southern part of

Tunisia and the Souf of Algeria have the most extensive and best developed crusts reported anywhere.Table 3.6 provides a summary.

Depois (1935) reported soils with high gypsum and calcium carbonate contents in the Djebel Nefousa of Libya. The accumulation zone is beneath the surface and the soils cover rolling terrain and therefore do not seem to be sabkha or groundwater types. Near the coast, Glennie (1970) identified gypsum dunes up to 10 m in height. The material originated in coastal lagoons and sabkhas. Other than these minor accumulations there appear to be few near-surface gypsum deposits in Libya though this may simply reflect the nature of geomorphological and pedological research undertaken there.

In Morocco, though there are extensive areas of calcrete, there do not appear to be any gypsum crusts or soils (Jaranoff, 1937; Awad, 1962). This may simply be because the climate of Morocco is too wet, nearly every part of the country receives more than 300 mm of rainfall annually. An alternative explanation for the change from gypsum crusts in Algeria to calcretes in Morocco is that the bedrock differs. It being the source of the gypsum in Algeria (Dalloni, 1953). This, however, may be only a secondary factor since there are Triassic evaporite deposits in Morocco (Tortochaux, 1968).

The gypsum soils and crusts of Algeria are very extensive (Bunting, 1965) notably in the Souf where Bellair (1954) estimated there are 10,000 km^2 of subsurface crust. Gypsum crusts have been reported from Biskra in the north (Horta, 1979) to the northern slopes of the Hoggar in the south (Bellair, 1954). This large area is centred on the Chotts Melrhir and Merouane and the broad trough of the Oued Rhir extending south from them to Ouargla and El Golea. The Chotts are 27 m and 35 m below sea level respectively and act as basins of inland drainage, concentrating the soluble salts introduced by wind,

TABLE 3.6

Type	Form	Author(s)	Location	Thickness (m)	Gypsum (%)	Origin
Surface	General	Agafonoff (1936a.b.)	Tunisia	-	-	Subsurface accumulation of gypsum leached from surface
		Beckmann et al. (1972)	Tunisia	-	60	Capillary rise of groundwater
		Boulaine (1958, 1961a.)	Algeria, Tunisia	1.5	70 - 90	Cemented aeolian deposit
		Bunting (1965)	Algeria, Tunisia	-	-	-
		Bureau & Roederer (1961)	Tunisia	0.1 - 0.2	50+	Superficially cemented dune sand
		Coque (1958, 1960, 1969)	Tunisia	0.6 - 1.0	70 - 90	Cemented aeolian deposits
		Coque & Jauzein (1966, 1967)	Tunisia	-	75 - 95	Cemented aeolian deposits
		Despois (1935)	Libya	-	-	-
		Douglas (1912)	Tunisia	0.05	-	-
		Durand (1963)	Algeria	1.5	-	Capillary rise of groundwater
		Gobert (1948)	Tunisia	-	-	Capillary rise of groundwater
		D'Hoore (1964)	Algeria, Tunisia	-	15+	-
		Horta (1979, 1980)	Algeria	-	-	Lacustrine, aeolian or groundwater deposits
		Pervinquière (1903)	Tunisia	-	-	Evaporation of surface or groundwater
		Pomel & Pouyanne (1882)	Algeria	-	-	-
		Quézel (1965)	Algeria	-	-	-
		Stainier (1912)	Algeria	0.3	60	Capillary rise of groundwater
		Yankovitch (1936)	Tunisia	-	-	-
		Zimmermann (1963)	Tunisia	-	10 - 80	Various possibilities
	Columnar	Besler (1977)	Tunisia	-	-	-
		Bureau & Roederer (1961)	Tunisia	1.0	-	Capillary rise of soil moisture
		Coque (1955a.b., 1962)	Tunisia	0.5 - 1.0	70 - 85	Cemented aeolian deposit
		Le Houérou (1960)	Tunisia	1.0 - 2.0	50	Cemented aeolian deposit

TABLE 3.6 (cont.)

Type	Form	Author(s)	Location	Thickness (m)	Gypsum (%)	Origin
Surface (cont.)	(cont.)	Meckelein (1959)	Libya	2.0	-	Alteration of calcrete (?)
		Page (1972)	Tunisia	0.5 - 1.5	20 - 98	Subsurface accumulation of gypsum leached from aeolian deposits
		Schwenk (1977)	Tunisia	1.0	35	Capillary rise of groundwater to surface
		Tricart & Cailleux (1961)	Tunisia	-	-	Subsurface accumulation of gypsu leached from aeolian deposits
		Vieillefon (1976)	Tunisia	0.5	up to 95	Various
	Powdery	Bureau & Roederer (1961)	Tunisia	1.0	-	Capillary rise of soil moisture
		Durand (1963)	Algeria	-	95	Deep (?) water evaporite
		Le Houérou (1960)	Tunisia	-	-	Aeolian deposit
		Vieillefon (1976)	Tunisia	1.0+	10 - 80	Various
Subsurface	Desert rose	Bellair (1954)	Algeria Tunisia	1.0	50+	Lacustrine or groundwater evaporation
		Bureau & Roederer (1969)	Tunisia	-	-	Subsurface evaporation of groundwater
		Coque (1962)	Tunisia	-	-	Precipitation from groundwater
		Durand (1949)	Algeria	0.6 - 1.5	-	Evaporation of groundwater
		Horta (1979, 1980)	Algeria	-	-	Precipitation from groundwater
		Kulke (1974)	Algeria	5.0	40 - 95	Evaporation of groundwater
		Pouget (1968)	Tunisia	1.0 - 2.0	20 - 80	Precipitation from groundwater
		Vieillefon (1976)	Tunisia	-	60 - 65	Precipitation from groundwater

ABLE 3.6 (cont.)

Type	Form	Author(s)	Location	Thickness (m)	Gypsum (%)	Origin
	Mesocrystalline	Boulaine (1961a.)	Tunisia	1.0	60 - 80	Subsurface accumulation of gypsum leached from the surface
		Bureau & Roederer (1961a.)	Tunisia	-	-	Evaporation of groundwater
		Cointepas (1965)	Tunisia	-	13 - 30	Precipitation from groundwater
		Coque (1962)	Tunisia	-	-	Precipitation from groundwater
		Coque & Jauzein (1966, 1967)	Tunisia	-	-	Precipitation from groundwater
		Dalloni (1953)	Algeria	2.5	-	Capillary rise of groundwater
		Gaddas (1971)	Tunisia	-	70	-
		Le Houérou (1956, 1960)	Tunisia	up to 2.0	55	Evaporation of groundwater
		Pouget (1968)	Tunisia	0.06 - 0.18	60.5	Capillary rise from groundwater (?)
		Stainier (1912)	Algeria	0.5 - 1.0	-	Capillary rise or lacustrine evaporation
		Stephan (1972)	Tunisia	-	-	Capillary rise from groundwater
Evaporitic	Massive or crystalline	Busson & Perthuisot (1977)	Tunisia	-	90+	Lacustrine evaporation
		Coque & Jauzein (1966, 1967)	Tunisia	-	-	Lacustrine evaporation
		Le Houérou (1960)	Tunisia	-	-	Lacustrine evaporation
	Coarsely crystalline	Bellair (1954)	Tunisia	0.1 - 0.2	-	Lacustrine evaporation
Dunes		Besler (1977)	Tunisia	-	19	Chott deflation
		Coque (1962)	Tunisia	-	50+	Chott deflation
		Glennie (1970)	Libya Tunisia	-	-	Deflation of coastal sediments
		Termier & Termier (1963)	Algeria	50	-	Chott deflation
		Trichet (1963)	Tunisia	1.0 - 2.0	60	Chott deflation

rain and bedrock dissolution. The gypsum crusts are either precipitated from groundwater (Durand, 1949) or deposited by wind, though there is a wide variety of forms suggesting equally diverse genetic processes. Durand (1963) suggested that some deposits may be deep or shallow-water lacustrine evaporites. The Tidikelt area centred on Chott Mekerrhane also has large areas of gypsum crusts (Dalloni, 1953; D'Hoore, 1964). Though Stainier (1912) held that the crusts of the Souf were usually 1.0 m to 2.0 m thick, Kulke (1974) reported examples reaching thicknesses of 5.0 m in some oases. The close association of gypsum crusts with areas of shallow groundwater and gypsiferous bedrock supports Gaucher's (1958) thesis that in most of north Africa the soils owe their nature to the geology and related hydrogeology.

While acknowledging the presence of these large areas of gypsum crusts in the Souf and Tidikelt, Boulaine (1961) argued that Algerian soils are characterized by calcrete development whereas gypsum crusts predominate in Tunisia. North of Sfax on the east coast of Tunisia, there are a few areas of gypsum soils associated with inland sabkhas, particularly Kelbia and Sidi El Hani east of Kairouan, Ech Cherita and El Rhar west of El Djem (Cointepas and Gaddas, 1971). Calcretes predominate from Tunis (Pimienta, 1950) to a line roughly between Gafsa and Sfax where gypsum crusts replace them (Burollet, 1956). Schwenk (1977) held that the transition was related to climatic factors. Pervinquière (1903) had suggested this but also stressed the importance of gypsiferous bedrock as a source of gypsum in groundwater and chott water. Calcretes are found south of the Gafsa-Sfax line, for example, in the Matmata Highlands, on the Island of Djerba and as far south as Ghadamés (Bellair and Jauzein, 1952). These may be fossil features dating from periods in the Quaternary wetter than today though the calcretes of Matmata and Djerba are

at present in relatively wet zones. Most of the Tunisian gypsum soils and crusts lie within a zone bounded by the 100 mm isohyet, for the year 1st September 1960 to 31st August 1961, a drier than average year, and within the 300 mm isohyet for the year 1st September 1958 to 31st August 1959, a wetter than average year (Mensching, 1968, see figs. 4.2A and B). The importance of gypsum in the bedrock is difficult to assess in Tunisia because nearly all the deposits whether of Triassic, Jurassic, Cretaceous or Tertiary age have gypsum beds (Burollet, 1967; Page, 1972).

The best developed Tunisian gypsum crusts, which are generally 1.0 m to 2.0 m thick, are found in the area immediately surrounding the Chotts el Fedjedj, Djerid and el Rharsa (Bellair, 1954; Beckmann, Scharpenseel and Stephan, 1972; Stephan, 1972) (see fig. 5.1). They mantle the pediments or glacis which are found around the Chotts. Most workers have identified three glacis (Coque, 1951, 1955 a.b., 1958, 1960; Bureau and Roederer, 1961; Steiner, 1963; Mensching, 1964) though locally there may be up to five (Coque and Jauzein, 1966, 1967; Coque, 1969). The three upper glacis are frequently covered by gypsum crusts, though in some areas the uppermost may have a calcrete layer. The two lowest glacis, if present, do not have crusts. Several different types of gypsum crusts have been identified in Tunisia, though workers have categorized them in various ways. Vieillefon (1976), for example, differentiated seven main types or a total of thirteen sub-classes. In terms of a structural rather than genetic classification, four types have been identified. These are, firstly, powdery accumulations which French workers call encroûtements and are locally called fech-fech (Le Houérou, 1960; Bureau and Roederer, 1961; Coque, 1962; Coque and Jauzein, 1966, 1967). Secondly, there are crusts which are either precipi-tated from groundwater called croûtes de nappe by French workers and gassi,

terch or louss in Arabic, or those which are apparently pedogenic, the croûtes gypseux or pavimentaux or deb-deb in Arabic (Coque, 1951, 1955 a.b., 1958, 1960, 1962, 1969; Coque and Jauzein, 1966, 1967; Le Houérou, 1960; Bureau and Roederer, 1961; Pouget, 1968; page, 1972). Deb-deb is an onomatopoeic name which imitates the sound of a donkey walking on the hard gypsum crust, it is occasionally applied to exposed croûtes de nappe. Thirdly, some workers have differentiated gypsum beds of lacustrine origin found in small pans in the Grand Erg Oriental and around Chott el Rharsa (Le Houérou, 1960; Coque and Jauzein, 1966 and 1967). Finally, Bureau and Roederer (1961) and Pouget (1968) identified a nodular form of gypsum accumulation which is called ras-kelb in Arabic.

Beyond the area of the Chotts gypsum crusts are found along the coast of the Golfe de Gabès from La Skhirra to the Libyan border. The Djeffara, the plain lying between the Matmata Highlands in the east and the coast to the west, has extensive gypsum crusts which are frequently overlain by calcretes. This same juxtaposition occurs in the north between La Skhirra and Sfax (Page, 1972). The implications of this and the structural and distributional characteristics of the Tunisian gypsum crusts will be dealt with in detail below.

Elsewhere in Africa north of the Equator gypsum crusts have not been described in detail. Munzinger (1869) described extensive gypsum crusts, apparently lacustrine or continental sabkha deposits, in Ethiopia. Gypsum soils are found in the Somali Republic from Hordio to Obbia on the southern side of the Horn of Africa, stretching 200 km to 300 km inland in some areas (D'Hoore, 1964; Dregne, 1964 and in McGinnies, Goldman and Paylore, 1968). Though some are directly associated with bedrock gypsum (Warden and Pallister,

1958) others are probably of more complex origin. Rainfall in these areas is less than 300 mm annum. Quézel (1965) held that the lack of gypsophytic vegetation in the western central Sahara indicated that there were few gypsum deposits in southern Algeria, Mauritania, Mali, Niger and Chad. Meckelein (1959), however, described gypsum crusts up to 2.0 m thick on river and lake terraces in southern Libya near Wāw el Kabīr, 300 km south-east of Sabbah, where rainfall is less than 25 mm per year.

vii. Southern Africa.

The gypsum crusts of Africa south of the Equator are confined to the Namib Desert, generally within 50 km of the coast between Namaqualand in South Africa and Moçamedes in Angola. Ganssen (1960) reported some pan sediments with high gypsum contents from south-eastern Namibia. The best examples of crusts are found between the mouth of the Kuiseb River near Walvis Bay and Cape Cross in Namibia (Spreitzer and Matznetter, 1961). This part of the Central Namib Desert receives less than 25 mm of rainfall annually. The gypsum soils and crusts rarely extend further inland than the 50 mm isohyet, possibly because the area is a cool desert with temperatures averaging about 15°C in July and in January about 20°C. Alternatively, the distribution pattern may be related to other factors such as the source of gypsum making up the crusts. The available information is summarized in table 3.7.

The first workers to report gypsum crusts in the Namib Desert pointed out that they are found only in the driest zones while calcretes are more extensive in the less arid parts (Jaeger, 1927; Range, 1927; Kaiser, 1929). Kaiser (1923) and Jaeger (1927) stressed the importance of fogs carrying soluble salts from the sea into the desert. In the driest areas, there are sodium chloride, barytes, manganese and silica crusts (Jaeger, 1927; Kaiser and

TABLE 3.7

Type	Form	Author(s)	Location	Thickness (m)	Gypsum (%)	Origin
Surface	General	Besler (1972)	Namib	1.0	-	Alteration of calcrete or authigenic
		Carlisle (1978)	Namib	1.0	-	Alteration of calcrete (?)
		D'Hoore	Somalia	-	15+	-
		Fockema & Fraser (1963)	Namib	4.0	70 - 80	Alteration of calcrete
		Jaeger (1927)	Namib	-	-	Capillary rise or fog deposits
		Kaiser (1923, 1929)	Namib	-	-	Capillary rise or fog deposits
		Martin (1963)	Namib	4.0	80	Alteration of calcrete
		Scholz (1963, 1968)	Namib	-	-	Alteration of calcrete
		Spreitzer & Matznetter (1961)	Namib	-	-	-
		Stapff (1887)	Namib	-	-	Marine evaporite (?)
		Wieneke & Rust (1973a. b., 1975, 1976)	Namib	-	39	-
	Columnar	Goudie (1972)	Namib	4.0	-	-
Subsurface	Desert rose	Besler (1972)	Namib	-	-	-
	Mesocrystalline	Gevers & Westhuyzen (1931)	Namib	-	-	Lagoonal evaporite
		Kaiser & Neumaier (1932)	Namib	-	84	Capillary rise of groundwater
		Knetsch (1937)	Namib	3.0 - 3.5	-	Capillary rise of groundwater
		Rust (1975)	Namib	1.0+	-	Precipitation from groundwater
		Rust & Wieneke (1973, 1976)	Namib	3.14	80+	Polygenetic
		Scholz (1972)	Namib	0.90	-	Alteration of calcrete
		Walter (1937)	Namib	-	-	Fog deposit or cemented dust
Evaporitic	Massive or crystalline	Gevers & Westhuyzen (1931, 1937)	Namib	-	-	Lagoonal evaporite
		Munzinger (1869)	Ethiopia	-	-	Lacustrine evaporite
		Nagtegaal (1973)	Namib	-	8.3	Supra-tidal evaporite

Neumaier, 1932). Most of this work was undertaken in the south of the Central Namib when the Lüderitz diamond deposits were first being exploited. Kaiser and Neumaier (1932) and Knetsch (1937) held that the crusts with high gypsum contents, up to 84%, are probably precipitated from groundwater in vleis (pans). Though the salts were originally deposited in fog moisture, the high potassium content of the sodium chloride crusts suggests that there is a strong terrestrial influence. Kaiser and Neumaier (1932) reported gypsum crusts from beneath calcrete horizons, as Page (1972) found in Tunisia, and sodium chloride crusts from beneath gypsum horizons. The significance of these sequences will be explained below. Sodium chloride crusts and gypsum deposits have been reported from coastal pans north of Swakopmund (Gevers and Westhuyzen, 1931; Gevers, 1936). Fibrous gypsum overlying sodium chloride crusts are found in areas which do not appear to have been flooded or to have had near-surface water tables (Gevers and Westhuyzen, 1931).

North of the Kuiseb River the gypsum crusts are generally 0.50 m to 0.10 m thick though they may reach 4.0 m (Martin, 1963). These deposits have attracted attention as a possible source of gypsum for the plaster industry because they are very pure, often containing 70% to over 90% gypsum (Fockema and Fraser, 1963; Martin, 1963). Carlisle (1978, 1980) identified the uranium minerals carnotite ($K_2(UO_2)_2(VO_4)_2.3H_2O$) and soddyite ($((UO_2)_5(SiO_4)_2(OH)_2.5H_2O$) in the gypsum crusts. In the Tubas Valley, a U_3O_8 content of 0.1% was reported. Though these uranium bearing deposits are found in river valleys, many of the gypsum crusts mantle the plains and hills north of the Kuiseb River and appear to be the product of pedogenic processes (Scholz, 1963, 1968, 1972; Goudie, 1972). These may involve alteration of calcium carbonate to gypsum by chemical processes involving the hydrogen sulphide present in fog moisture in this part of the desert (Martin, 1963; Besler, 1972). The possible sources of gypsum are limited owing to the absence of evaporite deposits of

any kind in the bedrock (Wagner, 1916; Korn and Martin, 1937; Martin, 1950, 1963, 1973). The fact that groundwaters near the coast are saturated with gypsum but not sodium chloride (Nagtegaal, 1973) prompted Rust and Wieneke (1976) and Wieneke and Rust (1976) to suggest that the crusts are produced by capillary rise and evaporation of these waters. Several ^{14}C dates are available for gypsum crusts along the coast. These suggest that they have developed during several different periods (Wieneke and Rust, 1976).

The significance of these dates and the modes of formation of these crusts will be assessed below. Suffice it to point out here that although the Namib crusts are less extensive and less diverse in form than those of Tunisia, the local geology does not complicate an examination of the origins of the crusts.

viii. Other Areas.

There are only two other areas of interest in terms of the world distribution of gypsum crusts. These are the drier areas of southern Europe and parts of Antarctica.

Since no part of Europe receives less than 300 mm of precipitation annually, the pattern of gypsum crust distribution elsewhere suggests that there are unlikely to be any significant gypsum accumulations in the soil zone. Smith and Robertson (1962) mentioned gypsum soils in Spain but gave no details. Grandjacquet and Haccard (1975) suggested that some gypsum deposits on the southern French Alps may be Mio-Pliocene pedogenic features but Vernet (1976) argued they were Triassic deposits modified by tectonic activity.

Several workers have reported surface gypsum accumulations in Antarctica, especially Victoria Land. Gibson (1962) described gypsum deposits 0.30 m to

0.75 m thick in dry kettle holes and on moraines. The fact that the deposits are 50 km from the sea suggested that they were not marine evaporites. While Gibson (1962) held that the gypsum is the product of bacterial reduction of pyrite in bedrock, Lyon's (1978) isotopic analyses suggested precipitation from glacial waters in the case of the Miers Valley deposits. Gypsum concentrations may reach 92%. Other salts found in association with the gypsum include calcite, halite, thenardite (Na_2SO_4), mirabilite ($Na_2SO_4.10H_2O$) and epsomite ($MgSO_4.7H_2O$) (Smith, 1965; Torii et al., 1966). Whatever the origin of the salts, whether evaporitic, atmospheric (Evans, 1970) or weathering products, their preservation at the surface results from the low rainfall, less than 150 mm annually, and temperatures above 0^oC in summer when katabatic winds cause rapid evaporation of surface moisture precipitating the soluble salts. The crystal habits are variable but in general lenticular crystals are found at the surface while columnar forms develop under boulders and within the moraines.

It is possible that similar deposits exist in high latitudes in the northern hemisphere where north of the Arctic Circle much of Alaska, Canada, Greenland and Siberia east of Novaya Zemlaya receive less than 300 mm precipitation annually and less than 100 mm in some areas. Little altered surface exposures of gypsum and anhydrite bedrock have been reported from Svalbard (Holliday, 1970).

d. Environmental characteristics.

Though factors such as climate and topography appear to have different effects on the distribution of gypsum crusts and soils in the different areas examined above, it is valuable to attempt to generalize some of the environmental characteristics of the distributional pattern. As well as climate, topographic

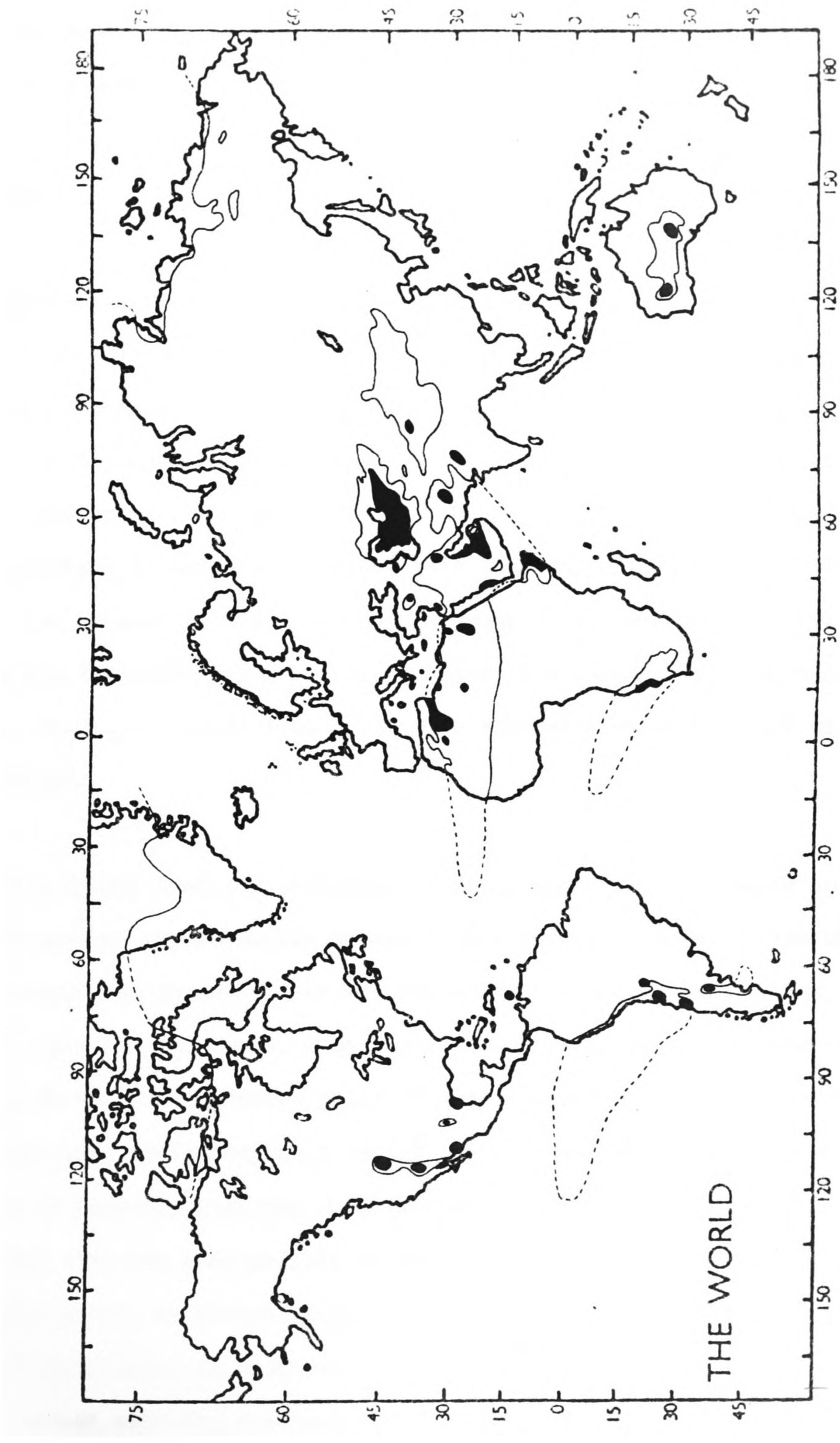


Figure 3.1 Distribution of gypsum crusts and soils (shaded) in relation to the 200 mm mean annual isohyet.

and geomorphic features, the possible sources of gypsum making of the soils, crusts and dunes will also be summarized as will some of the characteristics vegetation assemblages found in association with gypsiferous sediments.

i. Climate.

It is apparent from north Africa and the Namib Desert that gypsum crusts and soils predominate in the areas where rainfall is less than a specific amount. In Tunisia this amount is about 200 mm annually and in the Central Namib it is about 50 mm annually. In adjacent areas, where rainfall exceeds these amounts, calcretes are found, often to the exclusion of gypsum accumulations. It is important to ascertain whether this limiting relationship reflects a direct link between rainfall and gypsum crust distribution or whether other factors are involved. Furthermore, if there is a close relationship between the two, does this result from a climatic control on crust genesis or simply preservation?

There is a danger that the distribution of gypsum crusts and soils in an area reflects several environmental characteristics. For example, in the Namib Desert though the gypsum crusts are found only in the zone with less than 50 mm of rainfall per annum, this isohyet runs parallel to the coast about 50 km to 80 km inland. Hence, while rainfall may be an important factor, the crusts are also restricted to a narrow coastal belt which experiences frequent fogs. It is feasible that the distribution reflects not only the rainfall régime but also the availability of calcium sulphate which may be distributed by onshore winds, as Ulrich et al. (1959) proposed in California, or in fog moisture. Similarly, in southern Tunisia the coincidental alignment of the 200 mm isohyet with the northern edge of the Chott el Fedjedj and Djerid region complicates any accurate determination of the climatic limitations on

gypsum crust distribution. The distribution may result from the availability of gypsum deflated from the Chotts or in groundwater rather than predominantly climatic factors.

The amounts of precipitation experienced in the main areas of gypsum crusts and soils throughout the world, though variable, is generally less than 300 mm (see fig. 3.1). The areas with gypsum crusts which receive between 250 mm and 300 mm rainfall are limited to parts of the Thar Desert, Iran and Iraq and the coast of Somalia. In Australia, Central Asia, north Africa and around the Persian Gulf the upper rainfall limit is between 150 mm and 250 mm annually. The areas where the upper limit is less than 150 mm are North and South America, the Central Namib and Antarctica. As might be anticipated, these broad divisions in terms of precipitation may be explained, at least in part, by the temperature régimes in the different areas.

Rainfall is not in itself an important factor since it is the availability of the water at the surface and within the soil zone that affects gypsum deposits. The most important control on rainwater availability in arid and semi-arid regions is temperature because this determines the evaporation rates. Even if crude temperature data are used it becomes apparent that the areas of gypsum crusts and soils which receive the highest amounts of rainfall also experienced the highest mean temperatures. The mean monthly temperature for January and July may be compared, here they will be termed winter and summer to simplify comparison between the northern and southern hemispheres. The Thar Desert, southern Iraq and Iran and coastal Somalia have winter temperatures of about 15°, 10° and 20°C respectively and temperatures in summer of 30°, 30° and 25°C. These are notably higher than the corresponding figures for the peripheral zones of the gypsum crust areas of Australia, Central Asia, the Persian Gulf and north Africa, winter temperatures of 10°, 0°, 10° and

10° respectively and summer temperatures of $25^{o}C$ in the four areas. In the Namib and Atacama coastal deserts, winter and summer mean monthly temperatures are about 10° and 15°, though Besler (1972) quotes an average annual temperature of 16° to $20^{o}C$ for the Central Namib. Apparently, the more moisture available as precipitation the higher the temperatures must be to allow gypsum crust formation or preservation. Besler (1972) also pointed out that the high relative humidity experienced in the Namib, between 50% and 85% is important in the evolution of gypsum crusts. Her argument that the climate of the Namib may produce crusts essentially different from those in hot deserts is not strongly supported. There are two noteworthy exceptions to the generalized patterns outlined here. In North America, gypsum crusts and soils appear to be limited to the zone with less than 150 mm of rainfall despite the high mean temperatures, $10^{o}C$ in winter and $30^{o}C$ in summer. This may be because of low availability of gypsum or feasibly because the arid zone which also experiences high temperatures is very limited there being an abrupt change to montane areas with lower temperatures and higher rainfall. The other anomalous areas is Antarctica where gypsum precipitates have been described from a zone where temperatures rarely rise above freezing point. Though precipitation levels are low, 150 mm, this would nevertheless appear to be too high to allow gypsum preservation at such low temperatures. The explanation is that evaporation of meltwater in summer is intense owing to warm winds and the precipitated salts are rarely redissolved because atmospheric precipitation is in the form of snow.

On the basis of these findings it may be possible to develop a model of gypsum crust genesis and preservation on the basis of climatic data. It would not be possible to determine the mechanisms involved in crust development but at least to define the areas where climatic conditions preclude their formation. The data used above are not sufficiently detailed to construct

accurate potential evapotranspiration rates and soil water balances. Arkley (1963) attempted to show that the depth of calcareous accumulation horizons in soils was related to rainfall, in that the development of the horizon or its preservation was related to the soil moisture conditions. He proposed that:

$$D = 2.5\ (P - 12) \qquad (6)$$

where (D) is the depth of the horizon and (P) is the amount of precipitation, both in inches. This may be modified to provide a more accurate assessment of the likelihood of soluble salt accumulation in the soil zone. By determining the "leaching effectiveness" on a monthly basis, rather than annual precipitation and evaporation, it is possible to ascertain the soil water balance. When this is compared with the soil moisture storage capacity of any particular soil, the probability of salt accumulation or removal from the soil zone may be evaluated. Further implications of this will be considered in chapter seven. Though Arkley (1963, 1967) worked on the principle that salts are introduced form the surface, essentially the same conditions apply to any salt accumulation in the soil zone. Salts within the soil zone, whatever the mechanism of accumulation and regardless of whether they are in the form of an indurated calcrete or diffuse gypsum or sodium chloride crystals, will only remain there as long as there is not a net downward movement of undersaturated moisture through the soil or any individual horizon. The less soluble minerals, such as calcium carbonate, will persist longer than the more soluble salts when equilibrium is destroyed owing to wetter or cooler conditions. Yet, regardless of the solubility of the mineral, once leached out of the soil horizon it cannot be reintroduced until the climatic conditions change either seasonally or in the long-term.

This model also helps to explain the variation in the distribution of the crusts made up of different minerals. Calcium carbonate crusts will have a

greater stability range than gypsum crusts and these will be stable over a greater range of moisture conditions than sodium chloride accumulations solely because of the differences in solubility. It is still enigmatic, however, why calcretes, for example, are generally absent from areas where gypsum crusts are common. Precipitation levels low enough to allow gypsum accumulation near the surface are certainly within the zone of calcium carbonate stability. It is possible that the mechanisms of salt accumulation are of primary importance in explaining this. As pointed out above, this model does not necessarily explain the mode of accumulation only the preservational characteristics.

ii. Geomorphic/topographic locations.

For the purposes of this survey, the geomorphic locations of gypsum crusts and soils will be subdivided into four categories, coastal areas, inland basins, lowlands and undulating terrain. The purpose of this subdivision is as much to show the diversity of landscapes which are covered by gypsum accumulations, as to suggest some of the processes which may account for the extensive deposits. The different categories are not mutually exclusive but the dominant characteristic of the landscape is usually expressed by one of the subdivisions.

Coastal areas which have extensive gypsum crusts and soils are restricted to the Central Namib, the Somali Republic and the Chilean and Peruvian deserts. Of these areas, only the Namib has received any detailed attention in terms of the gypsum deposits. In this case, it would appear that the sea and the littoral zone are the sole sources of gypsum making up the crusts or providing the sulphur or sulphate components. There are no gypsum deposits in the bedrock of the area, hence the gypsum originates either in coastal pans, in

fog moisture or by chemical alteration of calcium carbonate in the presence of hydrogen sulphide or sulphates carried inland in the fog water. Other fog deserts, such as the Atacama and parts of California, experience deposition of soluble salts from sea mists (Ulrich et al., 1959; McCauley, Grolier and Breed, 1977). Coastal lagoons and sabkhas are important sources of gypsum which forms dunes and possibly contributes to crust formation in a number of areas. On Bahrain and along much of the Trucial Coast, precipitated gypsum is blown inland (Brunsden, Doornkamp and Jones, 1976; Kinsman, 1964, 1969; Butler, 1969; Kendall and Skipwith, 1969). It is likely that similar processes provided the material for gypsum deposits in many other areas including the Djeffara of Tunisia (Mensching, 1964) and parts of the Kara-Kum bordering the Caspian and Aral Seas (Dregne, 1976). These deposits, whatever the precise mechanisms of emplacement, cover undulating terrain without topographic preference.

Though the Caspian and Aral Seas are continental basins, their size and depth of water distinguish them from other inland basins such as the Algerian and Tunisian Chotts, Lake Eyre and numerous smaller pans in Australia, Death Valley and the dry lakes of the Thar, Afghanistan, Iran and Iraq. The gypsum associated with these is often in the form of bedded evaporites deposited when the lake waters evaporate after a wet period. Deflation may occur producing lunettes (De la Hunty and Low, 1958) or large dune fields such as the White Sands area of New Mexico (McKee and Moiola, 1975). The dunes may be degraded or consolidated when they age, creating gypsum crusts quite distinct in form from those found on the shores of the dry lakes. The processes involved are similar to those in coastal environments. Gypsum crusts composed of aeolian deposits often mantle a great variety of terrain as is well exemplified in the Tunisian Chott region. The gypsum crusts cover not only the lower glacis, which slope gently toward the Chotts, but also the flanks of

the surrounding hills on slopes of over 20° (Coque, 1962).

Many lowland gypsum deposits originate as lake deposits but some do not appear to be directly related to lacustrine processes. In the Algerian Souf, in Iraq and lower Egypt and over extensive areas of Uzbekistan and south and western Australia, gypsiferous soils and crusts are found at or just below the surface. Some probably originated as precipitates at the top of a shallow, fluctuating water table, for example, those of the Algerian Souf and Tunisian Nefzaoua. Others, however, do not appear to have a simple origin. Tucker (1978) argued that the crusts of Iraq are aeolian deposits later covered by alluvium. In Australia, similar models have been proposed (Jessup, 1960 a., 1961 b.) though the material mantling the gypsic horizon is suggested to be of aeolian origin. Besides the problem of accounting for the emplacement of coarse materials above the supposed aeolian deposit, the models do not adequately explain the uniform thickness of the gypsum horizons over very large areas. It is true that many loess deposits, the aeolian origin of which is well supported, show uniform horizontal bed forms but the deflated gypsum would be expected to behave similarly to quartz sand rather than silt and clay-sized loess. This is the case with the material deflated from pans in Australia (Jack, 1921; De la Hunty and Low, 1958), North America (Talmage, 1932; Jones, 1953; Jaeger, 1957; Eardley, 1962; McKee and Moiola, 1975) and the sediments deflated from the Tunisian Chotts (Trichet, 1963; Besler, 1977). In all these areas, dunes, rather than blanket deposits, are formed. It is possible, however, that some of the powdery accumulations, the encroûtements or fech-fech of north Africa, are the product of aeolian deposition of fine-grained sediments.

Finally, gypsum crusts found on hill sides and mantling undulating terrain are also common. There are good examples in Tunisia, the Central Namib Desert and

Iraq, where hill slopes and crests have gypsum crusts at the surface or at a shallow depth. The most common explanation for these deposits is that they are of aeolian origin (Coque, 1962, in Tunisia; Tucker, 1978, in Iraq) and are of essentially the same origins as the extensive crusts found covering lowland plains in Australia and Central Asia. Thougn in some areas the terrain may consist of consolidated dune material, as in parts of the Tunisian Nefzaoua, in many cases the crusts cover a landscape shaped by the local bedrock. Gypsum crusts are found on the crest of the Djebel Tebaga in Tunisia and on top of the Swartbankberg in the Central Namib. While the model of aeolian emplacement is more acceptable than those involving evaporation of surface or groundwaters in these situations, it does have drawbacks. Aeolian bedforms are not encountered in the deposits and they often form a blanket cover. The aeolian model may require modification or, alternatively, the source of the gypsum making up the deposits may differ from that in coastal and inland basins. It is appropriate to examine here some of the alternative sources of gypsum in the context of the large scale distributional characteristics of gypsum crusts and soils.

iii. Sources of gypsum.

Though the methods of emplacement of gypsiferous sediments are very diverse, including processes of aeolian, fluvial and lacustrine deposition, these frequently involve local mobilization of "primary" source material. These may occasionally represent authigenic gypsum but most frequently they are marine evaporites of recent age or more ancient deposits occurring as bedrock. For example, the gypsum soils and crusts of the Trucial Coast are either sabkha evaporites or material originating in the coastal sabkhas which has been redeposited after being deflated by wind or dissolved by groundwaters. In contrast, much of the gypsiferous material found in the Souf and Tidikelt

areas of Algeria is probably derived from bedrock through dissolution by groundwaters and subsequent evaporation when they reach the surface in the chotts and sabkhas. Aeolian and lacustrine deposits cannot usually be considered as primary sources of gypsum. It would, however, be incorrect to assert this in all cases. While aeolian deposits are frequently of local origin, as in the case of the White Sands of New Mexico, airborne gypsum travels great distances in some cases. Eriksson (1979) stated that gypsum is found in deposits of Saharan dust in Europe, indicating that it is carried over 2000 km in some instances. The distances involved in such relocations may justify the classification of these deposits as primary.

Atmospheric salts may similarly be carried great distances as either dust or in moisture droplets. The nature of atmospheric salt deposition shows considerable variation, in both amounts of material and types of salts, with distance from oceanic water (Junge and Werby, 1958; Fanning and Lyles, 1964; Yaalon, 1964 b., 1968, 1971; Yaalon and Lomas, 1970). Sulphates and sulphur have different characteristics in terms of their distribution over land and sea from most other components of atmospheric salts (Eriksson, 1958). The signification of these relationships to gypsum deposition and crust development will be assessed in chapter six. Suffice it to point out here that the chemical processes occurring while the salts are within the atmosphere and during their deposition on the land surface may produce authigenic gypsum (Martin, 1963; Besler, 1972). It is interesting to note that Yaalon (1964 b.) working in Israel and a number of workers in the Central Namib (Walter, 1936 b., 1937; Boss, 1941; Nagel, 1962) found that the affects of atmospheric salt deposition were limited to a 50 km to 100 km coastal strip. Local atmospheric circulation as well as climatic factors, especially rainfall are probably strongly influenced by the proximity of the sea. While there is a possibility that the gypsum forming the crusts and soils of coastal regions, such as South

and Western Australia, the Namib and Atacarma Deserts and Somalia, originates as atmospheric salt, other sources such as bedrock and coastal sabkhas may also be important.

The role of gypsum bedrock in the formation of gypsum soils and crusts is difficult to assess in many areas. It was pointed out above that in Tunisia nearly all of the geological formations represented have gypsum horizons. This is also true of large areas of Central Asia, Afghanistan, Iran, Iraq and the Trucial Coast. While in some cases soils with a high gypsum content may be developed directly on gypsum bedrock, as in the Le Kef and Foum Tathouine areas of Tunisia (Cointepas and Gaddas, 1971), parts of Iraq (Smith and Robertson, 1962) and Somalia (Warden and Pallister, 1958), other examples do not necessarily have such a direct relationship. Colluvial and alluvial sediments derived from gypsiferous bedrock frequently show evidence of crust development. Bureau and Roederer (1961) suggested that many of the gypsum crusts on high glacis around the Tunisian Chotts originated through concentration of gypsum in partially weathered bedrock essentially colluvial in nature. Butzer and Hansen (1968) held that the gypsum crusts of the Aswan areas of Egypt were the product ofsuperficial gypsum accumulation on gypsiferous alluvium. In all of these areas, the gypsum soils develop when bedrock material is transported over relatively short distances and the liberated gypsum is concentrated by processes such as soil moisture evaporation. Despite the fact that the transport of solid material is important, it is the concentration of the gypsum in solution that is critical to the development of crusts (Pouget, 1968, in Tunisia; Risacher, 1978, in Bolivia).

This process of solutional concentration occurs on a very large scale in basins of inland drainage. Meteoric waters introduced into the catchment will dissolve gypsum as they move through the bedrock. When the waters evaporate,

either at the top of the water table within the soil zone or in chotts and sabkhas flooded when the water table rises, the gypsum precipitates. The chemistry of the water in a chott may be determined by the bedrock it has been in contact with a considerable distance away. The Algerian Souf is a catchment with an area of about a quarter of a million square kilometres. The lowest point of the basin lies in the north near Chott Melrhir (De Lapparent, 1907) where extensive gypsum crusts are found (Durand, 1949; Bellair, 1954) but where the local bedrock does not contain a great deal of gypsum. Indeed, neighbouring small basins generally have non-gypsiferous sediments (Dalloni, 1953; Horta, 1979). The importance of these hydrogeologic factors was pointed out by Gaucher (1958) who argued that they might be more important than climate in determining the distribution of gypsum crusts and calcretes. Taking into account the distributional characteristics in other areas of the world, Gaucher's statement requires qualification. While it is true that gypsum crusts will only develop in areas where gypsum is available, their extent is also limited by climatic factors influencing moisture availability within the soil. Hence, calcretes may predominate over large areas of, for example, Algeria, owing to an absence of gypsum in quantity. It is equally feasible that in some areas, where gypsum is available and climatic conditions are conducive to accumulation, calcretes may be absent owing to a shortage of mobile carbonates, possibly resulting from high alkalinity. Examples of calcretes overlying gypsum crusts, and vice versa, have been described in Algeria and Tunisia (Page, 1972; Kulke, 1974), Australia (Jack, 1921) and the Central Namib Desert (Kaiser and Neumaier, 1932; Martin, 1963; Scholz, 1963, 1968). Gypso-calcareous crusts are found in some areas (Despois, 1935; Cointepas, 1971). It is incorrect, therefore, to suggest that the two types of accumulation are mutually exclusive. Though Page (1972) proposed that the different horizons were not coeval in the Sfax area of Tunisia, they are contemporaneous in the Namib. Climate and mineral availability determine the dis-

tribution and type of crusts on the regional scale, though different processes of accumulation influence local distribution.

Of all the major areas of gypsum crusts and soils only parts of Western Australia and the Central Namib Desert have no bedrock gypsum deposits, though this is debatably the case in Rajasthan (Auden, 1952). In these areas, the possible gypsum sources may be assessed without having to take into account the added complication of bedrock sources. Even in these relatively uncomplicated areas the source of gypsum is not well understood. Despite the absence of gypsiferous bedrock, the possibilities remain that there are aeolian, atmospheric or groundwater inputs. Moreover, authigenic gypsum formation cannot be ruled out (Martin, 1963).

iv. Vegetation associated with gypsum crusts and soils.

Gypsum crusts and soils are frequently characterized by distinct floral assemblages. It has been suggested that not only does the soil chemistry affect the vegetation but also that the flora may play a critical role in the evolution of gypsum crusts (Boulaine, 1958, 1961 b.; Vieillefon, 1976). Gypsified rootlets are found in some areas (Glennie and Evamy, 1968) but they are only rarely in evidence in well developed crusts. The role of bacteria and organic acids associated with vegetation may be significant in terms of gypsum precipitation. Cody (1979) showed that under alkaline conditions certain types of dissolved organic material promote growth of lenticular gypsum crystals. Plant materials such as fatty acids and protein hydrolyzates apparently increase gypsum solubility slightly (Barcelona, Tosteson and Atwood, 1976), while the effect on crystal habit is significant. The effects are independent of pH changes in the case of gypsum but not calcium carbonate. This may help to explain the variability in the distribution of calcretes and gypsum crusts.

The solubility of calcium carbonate in water is enhanced when organic acids lower the pH. In warm desert areas with less than about 300 mm of rainfall annually, there is little vegetation to provide these acids and, furthermore, the absence of resistance fractions, such as lignin, results in rapid decomposition of organic matter with little effect on soil pH. Calcium carbonate mobility will be greatly reduced in this arid zone, while the potential mobility of gypsum is increased with the likelihood that salts such as sodium chloride will be present at or near the surface. Pouget (1968) suggested that vegetation may promote gypsum crust development through the liberation of carbon dioxide. Though this will not directly effect gypsum solubility it will increase calcium carbonate solution. This in turn will result in some gypsum precipitation if the soil water is saturated with gypsum since the common ion effect will lower gypsum solubility (Shternina and Frolova, 1945; Wigley, 1973).

A considerable amount of work on vegetational characteristics has been undertaken in north Africa and the central Sahara but the relationship between gypsiferous soils and vegetation has received little attention elsewhere. In Tunisia, a great deal of floral information is available for the Gabès area and the region bordering the Chotts el Fedjedj and Djerid. Le Houérou (1956) based the floral classification on the soil types and plant associations. For example, gypso-calcareous and gypsiferous soils of eluvial origin are characterized by two associations, Zygophyllum album / Anarrhinum brevifolium and Gymnocarpos decander / Atractylis serratuloides while gypsum-rich colluvial and alluvial sediments are characterized by the Ziziphus lotus and Nitraria tridentata associations. Slightly saline gypsiferous alluvial soils in the semi-arid zone commonly support a Nitraria tridentata / Suaeda vermiculata / Salsola cruciata association. One species which is common to most of these associations is the grass Lygeum spartum which is particularly charac-

teristic of gypsum soils with low chloride contents. Commonly called sparte, or gdin, sennagh, (Ozenda, 1958, Quézel and Santa, 1962) or Halfa maboul, in Arabic to distinguish it from the superficially similar Halfa (Stipa tenacissima L.) formed on calcrete covered plains (Maunoir and Schirmer, 1905, p. 64), this grass is found throughout Tunisia wherever there are gypsum soils. Le Houérou (1969) pointed out that the associations differed in central Tunisia, in the region of Sidi Bou Zid, Maknassy and Kairouan, from the coastal area from Sfax to Mahdia and Agareb to Souassi (see fig. 4.2 A for locations). In the coastal zone, the following groups were identified within the Lygeum spartum /Aristida ciliate association.

Gypsophiles.

LYGEUM SPARTUM

ERODIUM GLANCOPHYLLUM

LYCIUM ARABICUM

HELIANTHEMUM LIPPII var. INTRICATUM

ERODIUM ARBORESCENS

LAUNAEA QUERCIFOLIA

SISYMBRIUM CORONOPIFOLIUM

CHRYSANTHEMUM TRIFURCATUM

ASTRAGULUS TENNIFOLINS

ARISTIDA CILIATA

LAUNAEA ANGUSTIFOLIA

PITURANTHOS CHLORANTHIS
ssp COSSONIANUS

PTERANTHUS DICHOTOMUS

SCABIOSA THYSDRUSIANA

ANABASIS OROPEDIORUM

DIPLOTAXIS HARRA

ANARRHINUM BREVIFOLIUM

PLACODIUM GYPSACEUM

On skeletal soils and gypsum crusts.

ATRACTYLIS SERRATULOIDES

HERNIARIA FONTANESII

STIPA PARVIFLORA

HELIANTHEMUM KAHIRICUM

GYMNOCARPUS DECANDER

(LINARIA AEGYPTIACA
ssp FRUTICOSA)

Psammophiles.

LAUNAEA RESEDIFOLIA

DIPLOTAXIS SIMPLEX

SCHISMUS CALYCINUS

KOELERIA SALZMANN

ERAGROSTIS PAPPOSA

ATRACTYLIS FLAVA

STIPA LAGASCAE

CUTANDIA DICHOTOMA

ECHIOCHILON FRUTICOSUM

HELIANTHEMUM LIPPII var. SESSILIFLORUM

ANACYCLUS CYRTOLEPIDIOIDES

RHANTHERIUM SUAVEOLENS

ECHINOPS SPINOSUS

ERAGROSTIS BARRELIERI

ONOPORDON ARENARIUM

LOTUS PUSILLUS

In the more arid zone.

FAGONIA KAHIRINA

PICRIS CORONOPIFOLIA

DEDESMUS BIPINNATUS

PARONYCHIA ARABICA spp ANNUA

ARISTIDA PLUMOSA

ENARTHROCARPUS CLAVATUS

Where groundwater is available at a shallow depth.

ZIZIPHUS LOTUS

Several other sub-groups were identified including relic woodland communities and those found on loessial deposits. In the Gabès area, many of these species are represented but additional species occur at the expense of others (Le Houérou, 1956; Novikoff, 1961). There is considerable variation even over relatively small areas, notably between the Djeffara and the Chott Djerid area (Le Houérou, 1969). Schoenenberger (1967) listed the following species

as the primary gypsum indicators, these being essentially gypsophytes rather than gypsoclines which tolerate gypsum in the soil but also grow in non-gypsiferous soil (Vieillefon, 1976).

LYGEUM SPARTUM L.

ERODIUM GLANCOPHYLLUM l'Her

MORICANDIA ARVENIS (L.) D.C.

BROTEROA AMETHYSTINA (Spach) P.K.

VELLA ANNUA L. (= CARRICHTERA VELLAE D.C.)

LIMONIUM THOUINI Viv. (= STATICE T.)

Between Gabès and Chott Djerid, a distance of about 100 km there are three major floristic zones, the Mediterranean, sub-tropical Saharan and Saharo-Arabic (Frankenberg, 1978). Hence, any such list of indicator species may include only one or two species represented in any area. In the Algerian Souf, floral associations very different from those in southern Tunisia are encountered. The best indicators of gypsum soils are the gypsophiles Arthro-phytum scoparium and Traganum nudatum. Of fifteen species listed by Quézel (1965) as characteristic of the Stephanochilus omphalodes / Arthrophytum schmittianum association found on gypsum soils in Algeria, only Launaea quercifolia, Anacyclus cyrtolepidioides and Launaea angustifolia were recorded in coastal Tunisia (Le Houérou, 1969).

Relatively little information on the flora associated with gypsum soils is available from other areas of the world. This perhaps reflects the fact that in north Africa the bulk of the work on gypsum crusts has been undertaken by pedologists rather than by geologists or geomorphologists. It is possible to outline some broad interrelationships between flora and gypsum soils in the Central Namib Desert. In some areas it is difficult to ascertain whether the

species are gypsophytes or gypsoclines since there are several factors besides soil conditions which effect the distribution. Walter (1937) showed that the distribution of Welwitschia mirabilis, an ancient gymnosperm found only in the Central Namib, was limited to the coastal fog belt. He held that in this area the flora could be divided into those species which utilize fog moisture, which never penetrates the soil to a greater depth than about 20 mm, and those which tap deep groundwater (Walter, 1936 a.). This moisture is invariably saline in the coastal strip, with fogs depositing up to 100 g of sodium chloride/m^2/year near the coast (Walter, 1937) where fog precipitation may reach 4.3 mm/month (Walter, 1936 b.; Boss, 1941). Information is not available as to whether the flora in this area are gypsophytes to simply halophytic gypsoclines. Many species are halophytes and show variable tolerance to less saline conditions. Welwitschia mirabilis does not extend more than 50 km inland while Anthraerua leubnitziae reaches 55 km and Zygophyllum Stapfii 60 km. The grass Stipagrostis gonotostachys which colonizes the gypsum crusts between Walvis Bay and Gobabeb does not appear to be a gypsophyte since it is also found extensively in the interdune valleys south of the Kuiseb River growing on calcrete and dune sand (Ollier and Seely, 1977; Watson, 1981). Owing to the great differences in edaphic conditions north and south of the Kuiseb, few species of plants are found on both sides of the river valley. Hence, their relationship to gypsum soils cannot be gauged. Acanthosicyos horrida, the nara melon, is a good example being found almost exclusively south of the river bed. The superimposition of edaphic, climatic and floral boundaries makes identification of the determinate factors difficult.

Rozanov (1961) listed the main species found on the gypsum-bearing serozems of Soviet Central Asia. He did not provide any detailed information about the character and distribution of the flora but pointed out that the associations are very variable and many of the species are gypsophytes. He listed the

following as the most common species and genera.

ARTEMISIA HERBA ALBA

HALIMOCHAMIS

GAMANTHUS OVINUS, G. KELIFI

SALSOLA SUBAPHYLLA, S. ARBUSCULA, S. SCLERANTHA, S. SOGDIANA, etc.

GIRGENSOHNIA OPPOSITIFLORA

STATICE SOGDIANA, S. RIGIDA, etc.

FRANKENIA

REAUMURIA TURKESTANICA, R. FRUTICOSA, etc.

KOCHIA PROSTRARA

HAPLOPHYLLUM BUCHARICUM, H. ROBUSTUM, H. AFFINE, etc.

CLEOME NOEANA, C. GORDJAGINII, etc.

COUSINIA BADGHYSI, C. DIMOANA, C. ROTUNDIFOLIA, etc.

ASTRAGULUS KELLER, A. SQARROSUS, A. KABADIENUS, etc.

GYPSOPHILA

GAILLONIA BUCHARICA, G. DUBIA, etc.

SISYMBRIUM

Grasses include:-

POA BULBOSA

HORDEUM MURINUM

AGROPYRUM

Sedges include:-

CAREX PACHYSTYLIS, C. PHYSODES

Despite the great variety of species and genera found in association with gypsum soils, only rarely does the total vegetational cover exceed 10% (Rozanov, 1961). Only during wet seasons or during abnormally wet years is the plant cover increased when ephemeral flora appears.

This review of the distribution of gypsum crusts and soils is intended to provide a global picture. This has necessitated making broad generalizations which do not take into account many exceptions and anomalies. In the following chapter, the two areas in which gypsum crusts have been examined in detail are described. On the basis of this description, the distributional and environmental characteristics of gypsum encrusted landscapes will be examined more closely.

CHAPTER 4

THE FIELD AREAS

In the foregoing chapter the distribution of gypsum soils and crusts in southern Tunisia and the Central Namib Desert was described in broad terms. These areas will now be examined in greater detail because it is from them that the information on the physical and chemical structure of gypsum crusts has been drawn. It will be shown that although the gypsum crusts in both areas have some striking similarities and probably originate in broadly the same way, the physical environment of the areas are very different. This difference is most evident in terms of climate and geology but extends to the geomorphology and palaeoenvironment of the areas. These characteristics will be described here.

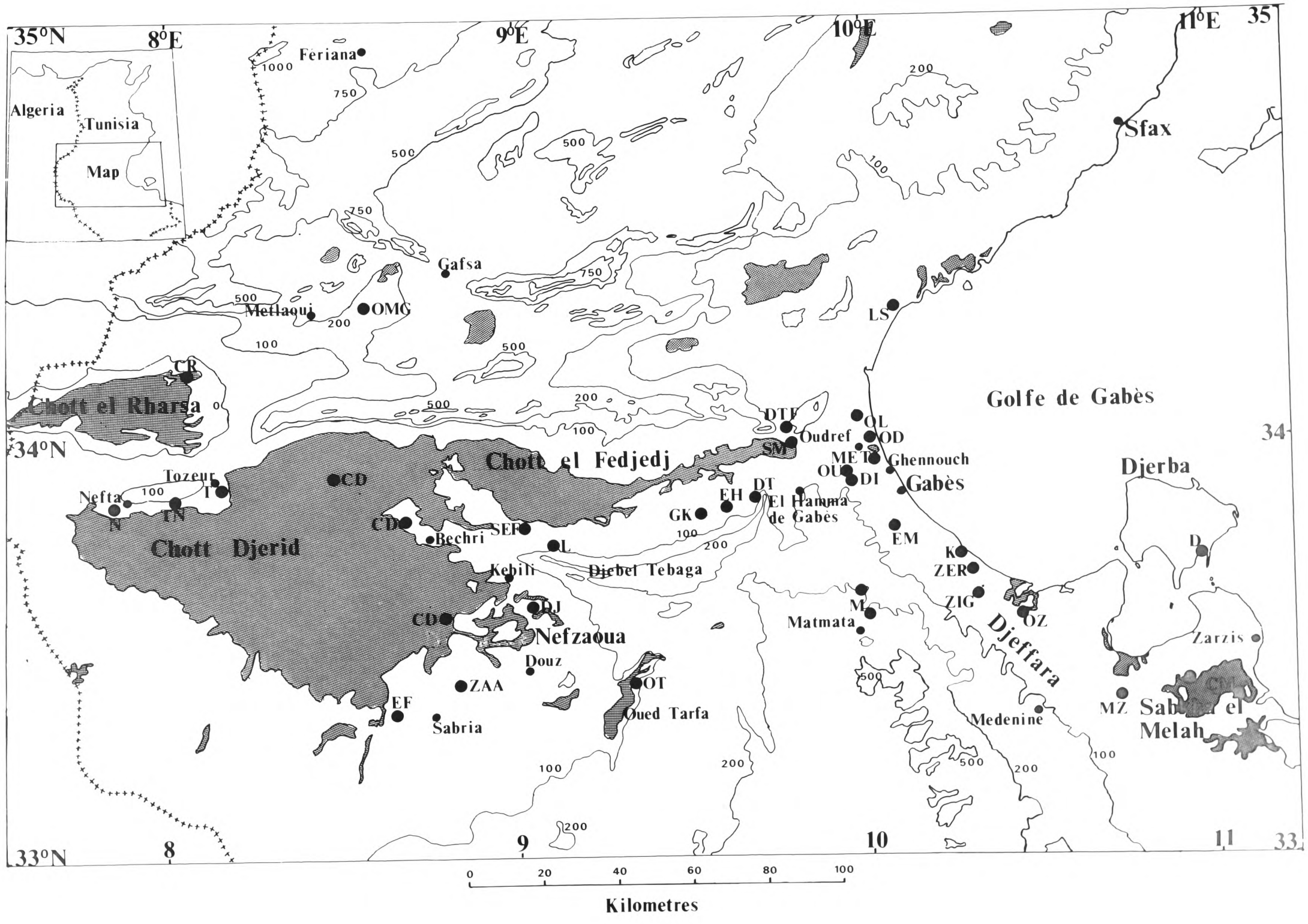
a. Tunisia.

i. Location of field area.

The field region is in the most northerly part of southern Tunisia, an area extending from Gafsa in the north (34° 25' 0" N) to Medenine in the south (32° 45' 30" N), from Nefta in the west (7° 43' 50" E) to Zarzis in the east (11° 2' 20" E) an area of about 3500 km^{2} (see fig. 4.1). It comprises the coast of the Golfe de Gabès from La Skhirra to Djerba and the Mediterranean coast almost to the Libyan border. Inland the area is dominated by large seasonally flooded salt lakes, Chotts Djerid, el Fedjedj and el Rharsa, and by the serrated crests of the escarpments which lie to the north and south. To the east of Chott Djerid and south of the Djebel Tebaga escarpment, is a flat, lowlying area, the Nefzaoua. Small dune fields in this area form the northern tip of the Grand Erg Oriental. East of this area are the Matmata Highlands which reach just over 700 m above sea level. Between these hills and the coast there is a broad

Figure 4.1

The field area in southern Tunisia. Shaded areas are chotts and sabkhas. The contours are in metres. Sample localities are designated by the abbreviations used throughout this work.


35°N
8°E
9°E
10°E
11°E
35
Algeria
Tunisia
Map
Feriana
Sfax
Gafsa
Metlaoui
OMG
LS
CR
Chott el Rharsa
Golfe de Gabès
34°N
34
Chott el Fedjedj
DTF
Oudref
OL
OD
SM
ME
Ghennouch
OU
DI
Gabès
Djerba
Tozeur
Nefta
TN
N
CD
Chott Djerid
Bechri
SEF
L
GK
EH
DT
El Hamma de Gabès
EM
K
ZER
ZIG
D
Kebili
Djebel Tebaga
M
Matmata
OZ
Djeffara
Zarzis
DJ
Nefzaoua
Douz
ZAA
OT
EF
Sabria
Oued Tarfa
Medenine
MZ
Sabkhat el Melah
33°N
33
8
9
10
11
0 20 40 60 80 100
Kilometres

undulating plain, the Djeffara, which extends eastwards to the sabkhas and lagoons along the Mediterranean coast south of the island of Djerba.

ii. Climate.

The field area may be subdivided into four climatic zones (Mensching, 1968). The bulk of the area may be classified as arid sub-tropical Saharan, but there are marked differences in the climate with distance from the coast, as was noted above in relation to the vegetation. The coastal strip, including the Djeffara receives between 150 mm and 200 mm of rainfall annually (Gabès, 178 mm; Ben Gardane, 175 mm), falling over 40 days on average, 69% in autumn and winter. Djerba and the adjacent mainland around Zarzis experience a wetter climate (Homout Souk, 208 mm). Mean annual temperature is 19.2^{o}C at Gabès, with a range in mean monthly temperatures of 16.5^{o}C. Gabès experiences maximum temperatures of over 30°C on 60 days of the year. Potential evaporation at Gabès is 1410 mm annually (Pouget, 1968). On a monthly basis Potential Evapotranspiration (P.E.) shows little variation throughout the field area. In January, the whole area except the interior (Remada, Tozeur and Ghadamès in the extreme south), experiences P.E. less than the mean monthly rainfall. Throughout the rest of the year there is a water deficit everywhere except Matmata and Ben Gardane where in Feburary and December there is rainfall in excess of P.E. (Le Houérou, 1960).

Inland from this coastal belt lies the sub-tropical Saharan transition zone. It extends from Gafsa to the Libyan border as a broad band with a small wetter enclave in the Matmata Highlands (Matmata, 221 mm). Rainfall is between 100 mm and 150 mm, falling over 30 to 40 days and 60% to 70% in autumn and winter. Temperatures are slightly higher than on the coast, with Medenine having a maximum mean monthly temperature of 29.1^{o}C in August.

The sub-tropical Saharan region of the interior may be divided in two. To the north of Chott Djerid, west of Gafsa, the annual precipitation is 100 mm to 160 mm (Gafsa, 156 mm), 59% of which falls in autumn and winter. The climate is hotter and more extreme than near the coast, with a range in mean temperature monthly of 20.5°C at Gafsa and 21.2°C at Metlaoui. At Metlaoui, 140 days per year have maximum temperatures above 30°C. South of Chott Djerid rainfall is less than 100 mm annually (Tozeur, 94 mm; Remada, 70 mm). There are no more than 30 rainy days per year and 55% of the rain falls in autumn and winter. The range of 22.8°C in mean monthly temperatures at Kebili further reflects the continentality of the climate. At Remada the mean annual temperature is 20.5°C with the range of absolute temperature being from -6°C to 55°C.

These figures have been calculated by Mensching (1968) from meteorological data for the period 1901-1960. Dubief (1963) held that the rainfall records for the Port of Algiers showed no major variation over the period 1840 to 1960, hence it may be assumed that the means quoted are representative of the long-term figures. Nevertheless, Mensching (1968) showed that in the sub-tropical Saharan zone rainfall totals may vary greatly from year to year. During the dry year of September 1960 to August 1961, Gafsa received only 88 mm of rainfall and the Saharan climatic zone extended to the coast of the Golfe de Gabès (see fig. 4.2 A). In contrast, the twelve months from September 1958 to August 1959 were wetter than usual with Gafsa receiving 200 mm of rainfall (see fig. 4.2 B). The significance of this variability in relation to gypsum crust development will be dealt with in chapter seven.

iii. Solid geology.

Most of southern Tunisia is composed of Upper Cretaceous sediments overlain with Quaternary deposits (see fig. 4.3), as in the adjacent Algerian Souf (Renou, 1848). To the north and south of Chott el Fedjedj and to the south of Medenine,

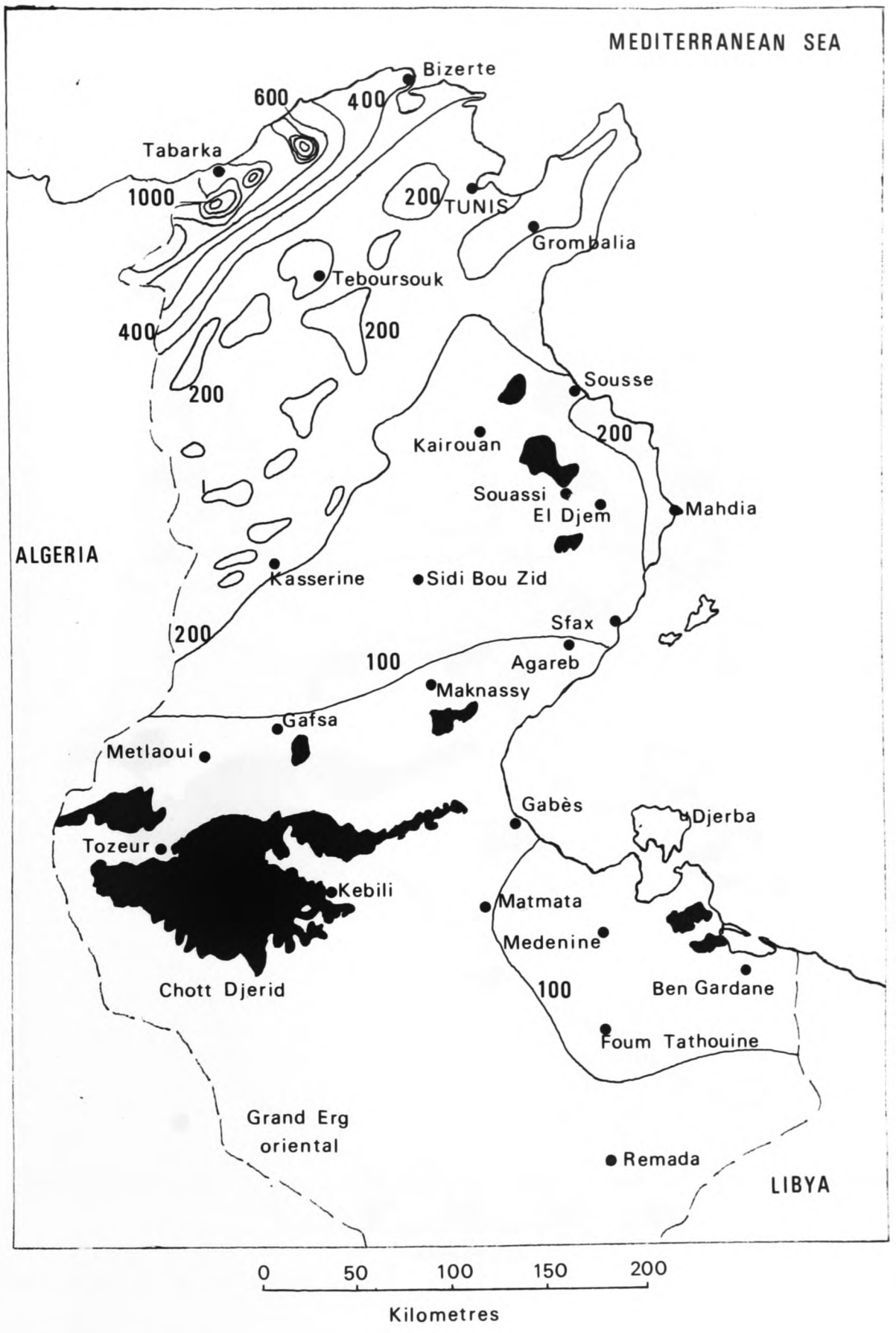


Figure 4.2 A

Rainfall in Tunisia during the wetter than average year, September 1958 to August 1959, isohyets are in millimetres (after Mensching, 1968).

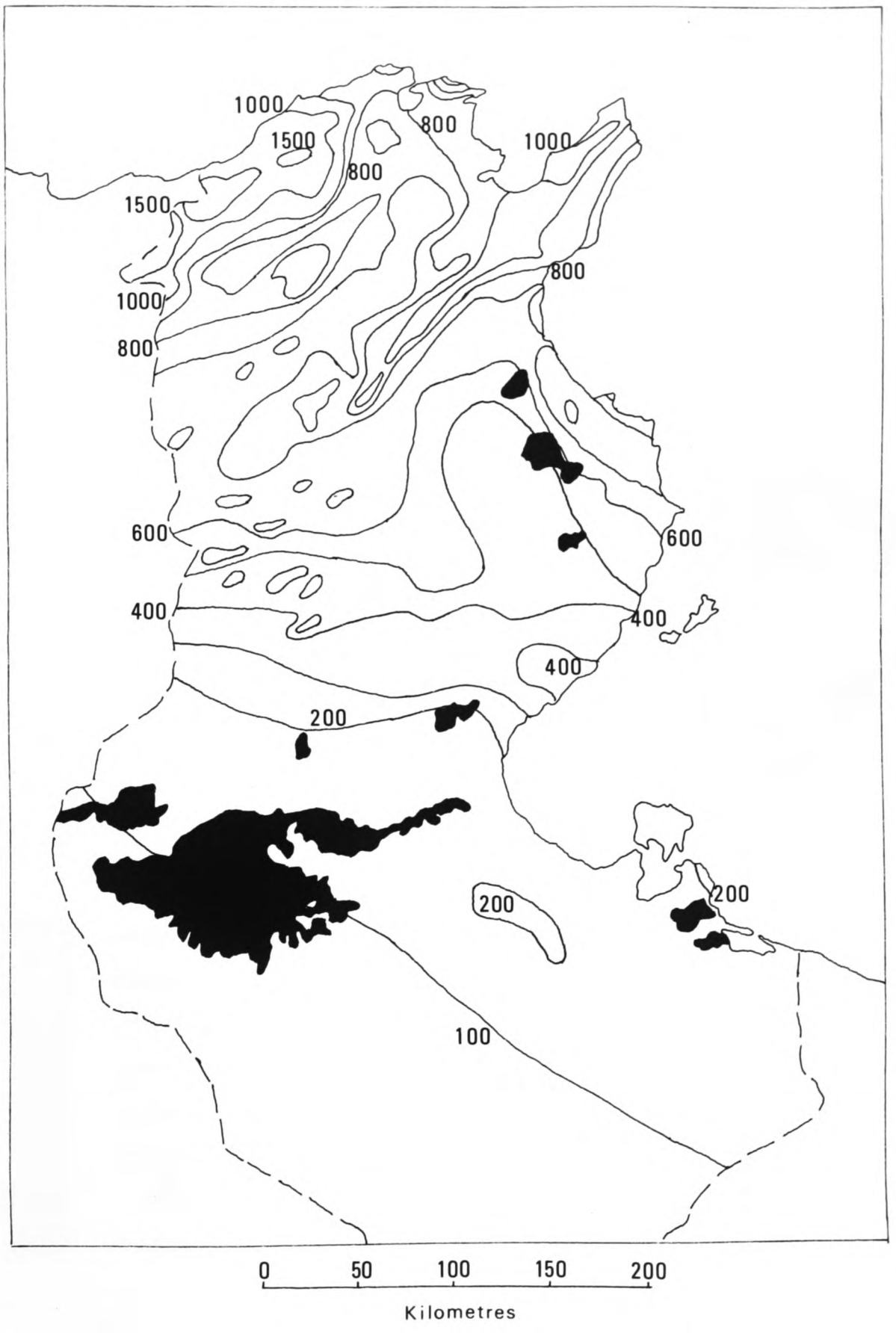


Figure 4.2 B

Rainfall in Tunisia during the wetter than average year, September 1958 to August 1959, isohyets are in millimetres (after Mensching, 1968).

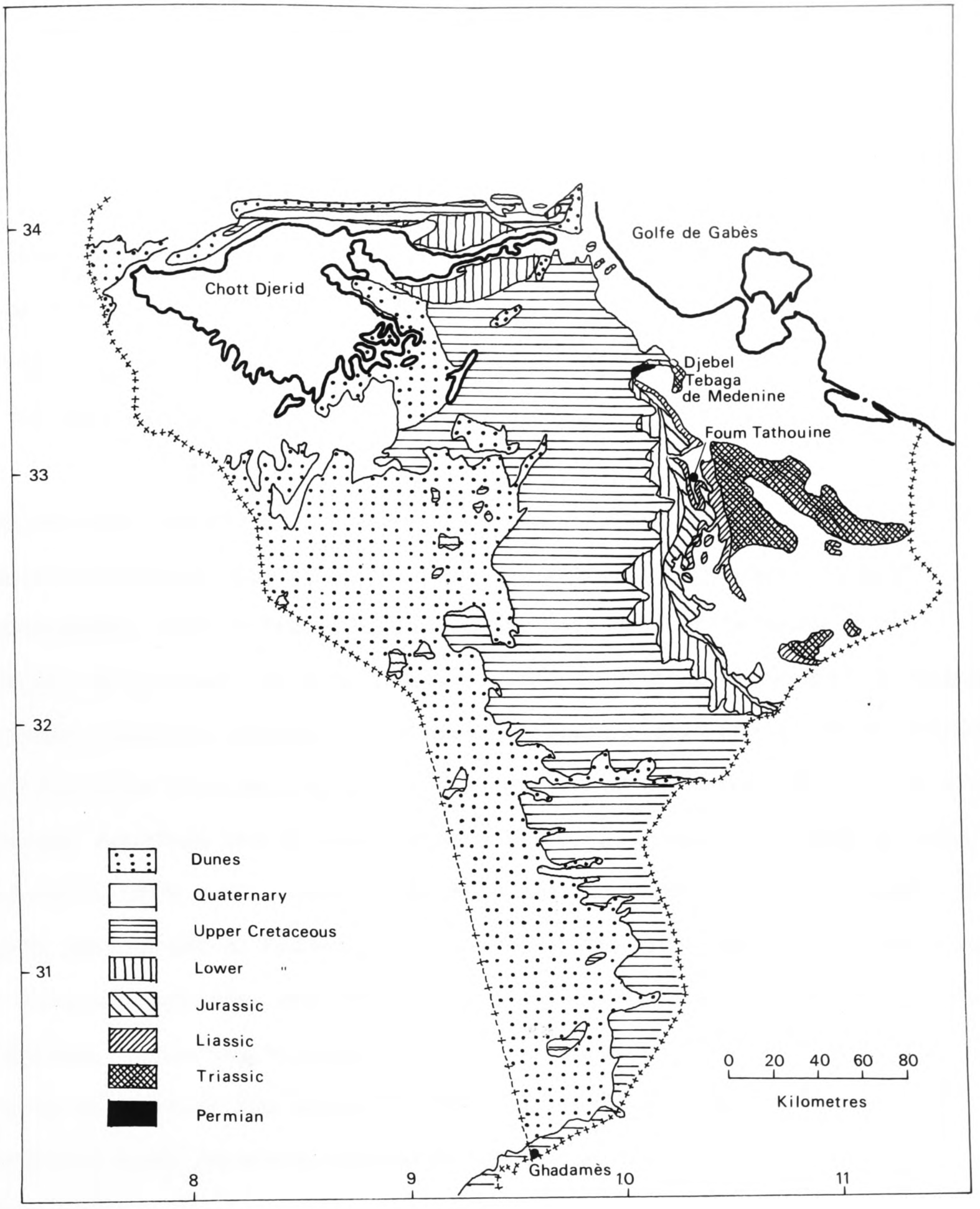


Figure 4.3

The geology of southern Tunisia (after Baird in Martin, 1967).

Lower Cretaceous rocks are exposed and south of the Djeffara there is a major zone of Triassic bedrock (Joleaud, 1919). In this area there are minor exposures of Carboniferous (Domergue et al., 1952), Permian and Jurassic sediments (Baird 1967). In the vicinity of Metlaoui there is a small pocket of Eocene rocks important economically because of their phosphate content.

The small deposits of Carboniferous and Permian age are restricted to the Djebel Tebaga de Medenine (not to be confused with the Djebel Tebaga south of Chott el Fedjedj). They consist of limestones, sandstones, marls and schists. These are the only pre-Mesozoic rocks in southern Tunisia.

There are only two areas of Triassic rocks in the field region though there are extensive evaporite diapirs of this age in central and northern Tunisia (Pervinquière, 1903; Tortochaux, 1968). The main deposit is south of the Djeffara, where Lower Triassic limestones and sandstones occur west of Medenine and Middle Triassic limestones and Upper Triassic gypsum and anhydrite east of Foum Tathouine (Domergue et al., 1952). The anhydrite deposits reach 500 m in thickness and there are extensive gypsum soils and crusts developed on them (Vieillefon, 1976). A second Triassic deposit is found inland from Gabès in the Fedjedj area. Chott el Fedjedj is a collapsed dome of Triassic salt and gypsum. The Triassic evaporites are buried under Miocene and Pliocene sediments (Gaucher and Burdin, 1974). The Tunisian Trias is difficult to interpret owing to the diapiric movement of the deposits (Burollet, 1967) but Bourcart and Ricour (1954) advocated a model involving sedimentation within mud deposits rather than water filled basins.

The Jurassic deposits of southern Tunisia are found adjacent to the Triassic exposures between Medenine and Foum Tathouine. They become thinner towards the south and disappear near Remada. The Lias, Bajocian and most of the Bathonian consists of gypsum and anhydrite with beds of sands, clays and dolomite, the

whole sequence reaching over 400 m in thickness (Domergue, 1952; Busson, 1967). The Upper Jurassic sequence contains increasing amounts of continental deposits and is largely lacking in gypsum. Jurassic deposits have been identified in cores from the El Hamma de Gabès area south of Chott el Fedjedj.

The Lower Cretaceous, south of Medenine, consists of Wealdian clays and marls with beds of limestone and gypsum which are exposed to the west of the main Jurassic outcrops. The Upper Cretaceous is much more extensive underlying most of the area south of the Chotts, though masked by the sands of the Grand Erg Oriental along the Algerian border. The Cenomanian and Turonian are represented by limestones and are separated from Campanian limestones by the green clays of the Lower Senonian. The upper most members of the Cretaceous represented in the south are Maestrichtian siliceous limestones and gypsum beds. Some Danian deposits are reported from the extreme south (Domergue et al., 1952).

North and south of Chott el Fedjedj the Lower Cretaceous is very well developed, forming the escarpments of Djebel Tebaga and the Gafsa chain. The basal Neocomian is made up of dolomite and green clays with increasing amounts of gypsum and anhydrite toward the top of the sequence. Up to 2,000 m of sediments may be present. The overlying Aptian and Albian deposits, respectively 80 m and 150 m in thickness, are made up of the same rocks as the Neocomian. The Upper Cretaceous caps the Djebel Tebaga south of Chott el Fedjedj and is well represented to the north of the Chott. The lowest member, the Cenomanian, 500 m to 600 m thick, consists of dolomitic limestones, green clays and gypsum beds. The overlaying Turonian limestones and dolomites are up to 250 m thick. The uppermost members are the Lower and Upper Senonian, consisting of dolomitic limestones clays and marls with decreasing amounts of gypsum up the sequence. It reaches 1,250 m in total thickness.

In the Metlaoui area some thin Danian deposits are found overlying the Upper Senonian conformably. Also in this area are found the aforementioned Eocene phosphates.These are located near to top of the 100 m thick sequence, the bulk of which consists of interbedded limestones, chalks and grey and yellow gypsiferous clays.

In the Djeffara and to the south-west and north-west of Chott Djerid, there are extensive Miocene and Pliocene deposits. In the Chott region, they consist of aeolian and lacustrine deposits which reach 1,000 m in thickness in the Segui Formation shales (Page, 1972). The top of the sequence is frequently gypsiferous clay and marl up to 100 m thick. The coastal Mio-Pliocene consists of Miocene marine sediments underlying Neogene continental deposits with some lagoonal facies.

Pleistocene sediments are ubiquitous. Besides the calcretes and gypsum crusts of the area there are also thick lacustrine sequences (Seurat, 1927). The most significant of these are the beds of Cardium edule Lin. shells which reach 30 m in thickness near Tozeur. These deposits provide valuable information about the palaeoenvironment of the Chott region during the Pleistocene.

The whole of the area has bedrock which was laid down either in or peripheral to lagoons or on the continental shelf. Gypsum and anhydrite are common throughout the sequence making the bedrock a likely source of gypsum in surface materials, especially those associated with the Chotts.The origin of the Chotts is of great importance because of their influence on the sedimentary geology of the area.

iv. Tectonics.

Chott el Fedjedj is a collapsed anticline which originated through diapiric movement of Triassic evaporites beneath Cretaceous sediments (Gaucher and Burdin,

1974). The movements may be contemporaneous with the Atlas formation in central and northern Tunisia (Castany, 1955). Though Coque (1958) held that the Chotts marked the most southerly influence of the folding of Tunisia, Joleaud (1919) pointed out that the Nefzaoua is a major syncline.

Tectonic movements have occurred during several different periods throughout Tunisia. The Gulf of Tunis, the Plains of Grombalia and Kasserine and the Kairouan and Sidi Bou Zid Basins are all late Tertiary (Lafitte and Dumon, 1948; Castany, 1952) or Quaternary geosynclines, some of which formed lake basins during the early Pleistocene (Berkaloff, 1931). The suggestion that the folding of the Chott region and the Atlas was also Pleistocene (Castany, 1949, 1953 a.b., 1955) is based on the evidence of deformation of Acheulian deposits but not Mousterian. Owing to the great differences of opinion regarding the ages of these archaeological industries, it is impossible to date the tectonic activity more accurately than somewhere between 750,000 and 50,000 years B.P. There is evidence of a second phase of tectonic warping in the late Pleistocene (Castany and Domergue, 1947; Lafitte and Dumon, 1948) in the form of distorted Pleistocene lake shore-lines. Meckelein (1977) held that the aïoun (springs) of Chott Djerid were attributable to present day tectonic movement. Castany (1953 b.) went so far as to suggest that late Pleistocene tectonics have resulted in the uplift of the Oudref bar and isolation of the Chotts from the Mediterranean. Coque and Jauzein (1965, 1967) argued that the most recent Cardium bed of Chott Djerid showed only minor deformation which suggests a cessation of tectonic activity in the Chott region in the early Pleistocene.

There is evidence of pronounced sinkage of the Golfe de Gabès during the Pleistocene. The three main Mediterranean shore-lines at +45 m, +30 m and +15 m to 20 m (Zimmermann, 1963) show considerable variation along the eastern coast. At Monastir, two high raised beaches have been identified (Joleaud, 1919;Castany, Gobert and Harson, 1956). In this tectonically relatively stable area, the

Monastirian beach lies at about 30 m, while at 45 m there is a more continuous dune covered beach dating from the Tyrrhenian (Castany, 1953 c.), or pre-Tyrrhenian if the true Monastirian is a Tyrrhenian transgression (Burollet, 1956, pp. 223-240). The lowest beach has been attributed to the Ouljien of the late Quaternary, possibly early Holocene (Burollet, 1956). Near Gabès, the upper beach which may be Monastirian or pre-Monastirian, is about 5.0 m above sea level (Coque, 1958). On the island of Djerba, it is at 10 m above sea level and in the Biharet el Biban, south of Zarzis, 3.0 m to 5.0 m below present sea level, Slob Rharbi and Slob Chergui representing old coastal dune belts (Bellair, 1954 b.). The sabkhas along this coast, such as the Sabkha el Melah, are submerged river valleys (Bureau, 1960; Perthuisot, 1980). This area has apparently experienced downwarping of nearly 40 m since the Tyrrhenian (? pre-Monastirian) transgression.

The overall pattern of tectonic movement in southern Tunisia since the mid-Tertiary is of folding and subsidence along the Fedjedj-Djerid-Rharsa axis and subsidence of the Golfe de Gabès causing major tilting toward the south-east. While the main period of tectonic activity in the Chott region appears to have been pre-Tyrrhenian, the coastal zone has experienced considerable tilting since then.

v. The Chotts.

The chain of Tunisian and Algerian Chotts extends from a few kilometres inland of Gabès, at Chott Fedjedj, to well south of Ouargla in the Souf forming a dog-leg nearly 1,000 kilometres long. The depressions are predominantly of tectonic origin. In some areas, subsidence has lowered the floors to below sea level, the Chott el Rharsa is the closest to the sea below sea level (Pomel, 1878) and Chott Melrhir is 31 m below present sea level (De Lapparent, 1907). Perhaps more important than the origin of the depressions is the origin of the Mio-Pliocene

and Pleistocene deposits within them (see plate 4.1). The question as to whether they are terrestrial or marine has been debated for over 100 years. A marine origin would imply considerable tectonic activity at the eastern end of Chott el Fedjedj where the Oudref Bar rises to about 35 m to 45 m above sea level. A terrestrial origin would, however, signify considerable fluctuations in the volume of water in the basins during the Quaternary since fossil beaches are found up to 40 m above the present shore of Chott Djerid. These oscillations would in turn suggest major climatic changes during the period.

The debate on past marine inundation of the Chotts was first stimulated by De Lesseps' (1883 a.b.) proposal to construct a canal from the Mediterranean, just north of Gabès, to Tozeur or Kriz and from there through the ridge which separates the Chotts el Rharsa and Djerid. Besides the advantages for trade, he felt that an inland sea would ameliorate the climate of the interior and enable greater agricultural development. Despite the vehement arguements that ensued, the project commenced. The canal excavations, however, never proceded further than the Oasis de Lesseps, a few kilometres from the Golfe de Gabès, owing to financial and engineering difficulties. Pomel (1878) argued that Chotts el Fedjedj and Djerid could not have been flooded by the sea because they are above sea level, 25 m and 15 m respectively (Mensching, 1968, p. 207). He did not take high sea levels in the past into account. Joleaud (1919) did, but still argued that the Tyrrhenian sea level rise was insufficient to over-spill the barrier between the Golfe de Gabès and Chott el Fedjedj at Oudref. The question is still far from resolved, the arguments hinging on the fossil floral and faunal evidence and the stratigraphic information in the form of raised beaches and evidence of tectonic activity.

Most discussion has centred on the beds of Cardium shells admixed with Melania tuberculata Muller, Hydrobia sp. and Corbicula fluminalinsis Muller (Coque and Jauzein, 1967) found around the Chotts (see plate 4.2). Castany,

Plate 4.1 (overleaf, above)

The centre of Chott Djerid between Bechri and Tozeur. A thin crust of sodium chloride overlies black mud and silt. In February 1977 the water table was very near the surface.

Plate 4.2 (overleaf, below)

Horizons of Cardium shells separated by lenses of cross-bedded sand on the shore of Chott Djerid in the vicinity of Tozeur.





Coque and Domergue (1953) identified them as Cardium edule L. var. glauca Brug (Cardium edule L. is an Atlantic species while Cardium glauca Brug is Mediterranean). The beds are up to 3.0 m thick and the main one is at 15 m to 25 m above the present lake level, the variability being caused by differential tectonic subsidence in the region during the late Pleistocene and Holocene (Castany and Domergue, 1947). Castany (1953 a.b., 1962), Burollet (1956, pp. 223-240 and Zimmermann (1963) proposed that the Cardium bed represents a feature contemporaneous with the 32 m Monastirian beach and that the sea overflowed the Oudref Bar in the Tyrrhenian (?). Tectonic uplift of the bar in the late Tyrrhenian apparently isolated the Chotts. Castany (1954) argued that the asymmetry of the shells indicates either a deep-water or highly saline environment. He rejected the possibility of transport by birds on the grounds that fossil foraminfera are also present (Miliolidae, Nonionidae and Rotalidae families (Milokhoff, 1954)) and these could not be introduced in the same way. Castany (1954) identified beds at 40 m to 45 m above present sea level in the Nefzaoua. This higher beach was suggested to be Mio-Pliocene (Castany and Coque, 1956; Arambourg and Coque, 1965) or Villefrancian (Coque and Jauzein, 1966) in age.

The arguments against marine inundation of the Chotts are based largely on the problems of breaching or overflowing the Oudref Bar. Coque (1958) pointed out that since the Cardium beds show evidence of late Pleistocene downwarping, it is unlikely that the Oudref Bar was uplifted, cutting off a link with the sea. Furthermore, the pre-Monastirian beach is at only 5.0 m above sea level at Gabès so the correlation in the heights of the beaches at Monastir and in the Chotts is inappropriate. Bellair (1958) went so far as to argue that the Chott sediments did not originate through lagoonal evaporation but rather subsurface precipitation of salts through capillary rise from groundwater. The Cardium beds do, however, prove that the basins were flooded by water possibly several tens

of metres deep during the Quaternary. Castany's (1954) suggestion that the Chotts Melrhir and el Rharsa were filled during Roman times has not been verified. Mensching (1964) held that the two Cardium beds were deposited during two wet phases, possibly contemporary with the Riss and Würm Glacials in Europe, when groundwater levels rose and streams flowed into the basis filling them with terrestrial and meteoric waters not seawater. This model compares closely with that proposed by Faure (1966) for the Central Sahara where the lakes were flooded from 22,000 to 7,500 years B.P. and possibly as late as 3,000 years B.P.

vi. Quaternary landforms.

The Chotts and their associated landforms, as well as many of the coastal landforms of southern Tunisia, are largely tectonic features. There are also extensive terraces in central and southern Tunisia which give the area a characteristic morphology. Though these features are often deformed by Quaternary movements,their origin is not strictly related to tectonics, nor solely to erosion or deposition but both. Pleistocene sediments and depositional landforms will be dealt with below.

The term glacis has been applied by French geomorphologists to a pediment which has been modified by erosion and/or deposition (Mensching, 1958). Glacis are best developed in semi-arid areas which experienced large variation in climate during the Quaternary. In Tunisia, the different glacis form stepped features along broad river valleys and around the Chotts, particularly el Fedjedj. The glacis are concave in profile reaching an inclination of 10° to 15° where the topmost meets the scarp or valley side (Coque, 1958), decreasing to about 3° on the lowest of the three glacis represented in southern Tunisia (Coque and Jauzein, 1967). These are cut into Lower Cretaceous and Mio-Pliocene sediments (Coque and Jauzein, 1966) and form terraces at about 23 m, 14 m, and 3.0 m (Coque, 1960). There is evidence of a fourth high terrace in some areas around

Chott el Fedjedj. In the Gafsa area, four levels are discernable at about 44 m to 45 m, 30 m to 35 m, 10 m to 25 m and 4.0 m to 6.0 m (Coque and Jauzein, 1966, 1967) and in the Tunis areas there are five (Coque, 1969). Each level is characterized by deposits of different ages, calcretes in central and northern Tunisia and gypsum crusts in the south. Bureau and Roederer (1961) described sequence of three glacis in the Matmata Highlands with calcrete on the highest level and gypsum crusts on the lower two levels.

The process of glacis formation involves cyclic climatic change. Initial lateral planation produces the erosional slopes. These are fossilized during the ensuing more humid phase, when calcretes or gypsum crusts develop on the surface. Later erosion will inset a new generation of slopes (Coque and Jauzein, 1967) but the crusts will preserve the older topography. "Grâce à cette action conservatrice, nous disposons aujourd'hui de vestiges suffisants pour reconstituer une ancienne topographie" (Coque, 1955 b.). Unfortunately, difficulties remain. The ages of the different glacis are uncertain. Coque and Jauzein (1967) suggested that the approximation of the heights of the middle glacis (2) of the el Fedjedj area and the lower Cardium shore-line, which is usually covered by the glacis (2) gypsum crust, suggested a post-late Tyrrhenian age. Several major questions are posed by this model the two most significant being whether gypsum crusts would survive the wet periods when the Chotts were flooded, let alone develop during these phases and, secondly, whether the correlation of high sea level (Tyrrhenian) with flooding of the Chotts is appropriate.

In central and northern Tunisia, though the glacis are present, considerably more attention has been paid to the river terrace sequences. While Dresch, Rondeau and el Aouani (1960) identified only two fossil terraces Jauzein (1959) and Mori (1967) constructed the following more comprehensive Quaternary sequence for the area.

1.) 10 m to 13 m. Sandy terrace without crusts

2.) Red soils with calcretes locally (Jauzein, 1959).

3.) 10 m to 25 m. Non-deformed white calcrete.

4.) 25 m to 35 m. Terrace.

5.) 40 m to 55 m. Poorly consolidated calcrete terrace or glacis.

6.) High calcrete (Villefranchian?).

Working in the Tabarka area of the north Tunisian coast, Brosche, Molle and Schulz (1976) identified a broadly similar sequence developed around the Oued Malah. A dune sand terrace at 6.0 m to 12 m and calcium carbonate cemented dune sands at 12.5 m to 13.5 m, probably correlate with (1) and (3) above. Between these levels, Brosche, Molle and Schulz (1976) recognized a high Chott level at 6.0 m to 10 m. They correlated this with the Würm (Soltanian) 30,000 to 10,000 years B.P. on the basis of ^{14}C dates from the lower sand terrace and data from central Tunisia which will be referred to below. If the comparison with southern Tunisia is valid, it would appear that the three main glacis or terraces with crusts are represented by partially calcretized terraces, all of which are pre-Würm. On this basis the Tyrrhenian age proposed for the middle glacis may be appropriate. It must be stressed, however, that until more fixed points are available, in the form of geochronometric dates, most of the Quaternary climatic sequence for the area will remain speculative.

vii. Quaternary sediments.

For the purpose of this brief account of the Pleistocene deposits of Tunisia the sediments will be subdivided into fluvial, marine and colluvial, aeolian and edaphic.

Reference has already been made to some of the terrace deposits of northern

Tunisia and the coastal sediments south of Monastir. Seurat (1927) mentioned calcareous fluvial conglomerates and Quaternary marine clays in the area north of Gabès but these sediments are probably Mio-Pliocene rather than Pleistocene in age. Thick Pleistocene fluvial deposits, such as those described by Alimen (1965) from the north-western Sahara, are scarce in Tunisia. Colluvial sediments are more extensive and in some cases particularly revealing in terms of the palaeo environments they characterize. Coque (1951, 1960) described colluvial material near the crests of the Djebel Tebaga south of Chott el Fedjedj. The bouldery material is cemented by a gypsum crust and was suggested to be the product of salt weathering of the exposed limestones and dolomites. Gruet (1954) held that fissures forming polygonal surface patterns up to 10 m in diameter in this area are the result of periglacial activity and frost wedging. The arguments against this will be presented below, suffice it to point out here that though there is no strong evidence of periglacial conditions in southern Tunisia, there is in the centre of the country. In the Djebel Chambi and Djebel Semmama areas near Kasserine, solifluxion deposits and calcretized screes (grèzes litées) have been identified (Dresch, Rondeau and el Aouani, 1960; Guillen and Rondeau, 1966). Two scree deposits are superimposed, as in Cyrenaica (Hey, 1963), the lower being more heavily cemented. While Dresch, Rondeau and el Aouani (1960) suggested they are of Tensiftian (Riss) and Soltanian (Würm) age, Hey (1963) proposed early and late Würm scree formation in Libya. Whatever the interpretation of the ages, it is significant that the features are indicative of not only cold conditions but also probably wet. It must be pointed out that even in the northern part of Tunisia these features are restricted to areas over 600 m above sea level, so it is unlikely that southern Tunisia experienced such cold since south of Gafsa only a few isolated hills reach above 650 m.

Arid climate indicators are also abundant in Tunisia, largely in the form of sand dunes. Other aeolian deposits include extensive aeolianite (miliolite) along the coast and loess deposits in the Matmata Highlands. There are active

dune fields in the vicinity of Nefta, in the Nefzaoua and in the extreme south of the country. These are peripheral to the Grand Erg Oriental, which probably extended further north-eastwards into Tunisia during arid phases in the Pleistocene. Many of the dunes in southern Tunisia are fossilized by a calcrete cap (Bellair and Jauzein, 1952) or sometimes a gypsum crust, as in the Nefzaoua (Coque, 1962, p. 80). The age of this cementation is uncertain. It may be coeval with that of dune sand in northern Tunisia (Gottis, 1953; Brosche, Molle and Schulz, 1976). Unlike the sand dunes of southern Tunisia these dunes are all coastal or chott deflation products (lunettes or bourrelets). Though rarely extensive, these deposits may form dunes up to 50 m in height (Coque and Jauzein, 1966). The dunes are pre-Arterian, more than 30,000 years old, suggesting that the calcium carbonate cementation is Soltanian (Würm). It must be stressed that the justification for this is not based on any well-established chronology.

Loess deposits up to 20 m in thickness were reported from the Matmata area by Coque and Jauzein (1967). Brosche and Molle (1975, 1976) proposed their depositior between 30,000 and 7,000 years B.P. on the basis of ^{14}C dates of shells within calcretes (Hv 5402 - 21385 $\pm$ 235; Hv 6603 - 13850 $\pm$ 170; Hv 5395 - 10800 $\pm$ 85). Whether the deposits represent a desert sediment (Goudie, Cooke and Doornkamp, 1979) or not, is uncertain. Brunnacker (1973) and Brosche and Molle (1975) held that the calcrete, and occasionally gypsum beds, are indicative of a late Würm or early Holocene wet phase.

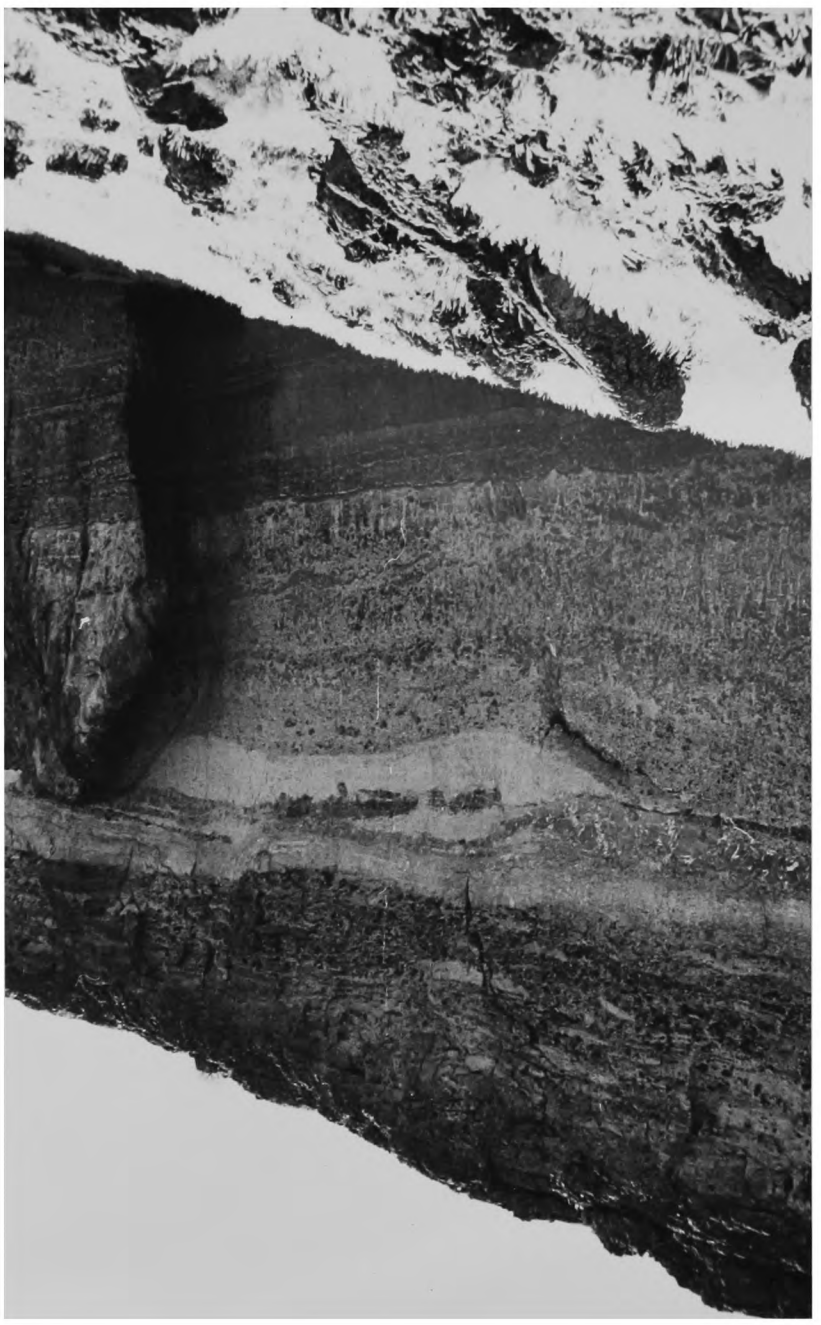
Aeolianite is found locally along the full length of the Tunisian coast with particularly good examples near Cap Bon, Monastir and on Djerba. Castany, Lucas and Reyre (1954) described aragonite oolites from the deposits of Djerba, La Skhirra and Mahdia. It is probable that aragonite formed the initial cement (Stoddart and Cann, 1965). These structures are evident in a dominantly micritic sample collected near Aghir (D) (see plate 4.3). X-ray diffraction revealed aragonite as well as calcite and quartz. The mineralogy is shown in table 4.1.

Plate 4.3 (overleaf, above)

Sample D 1, X 64 . Aeolianite composed of micritic calcite cementing quartz grains, shell fragments and aragonitic oolites (top centre).

Plate 4.4 (overleaf, below)

Carp Cliff. 47 m of calcrete incised by the Kuiseb River.



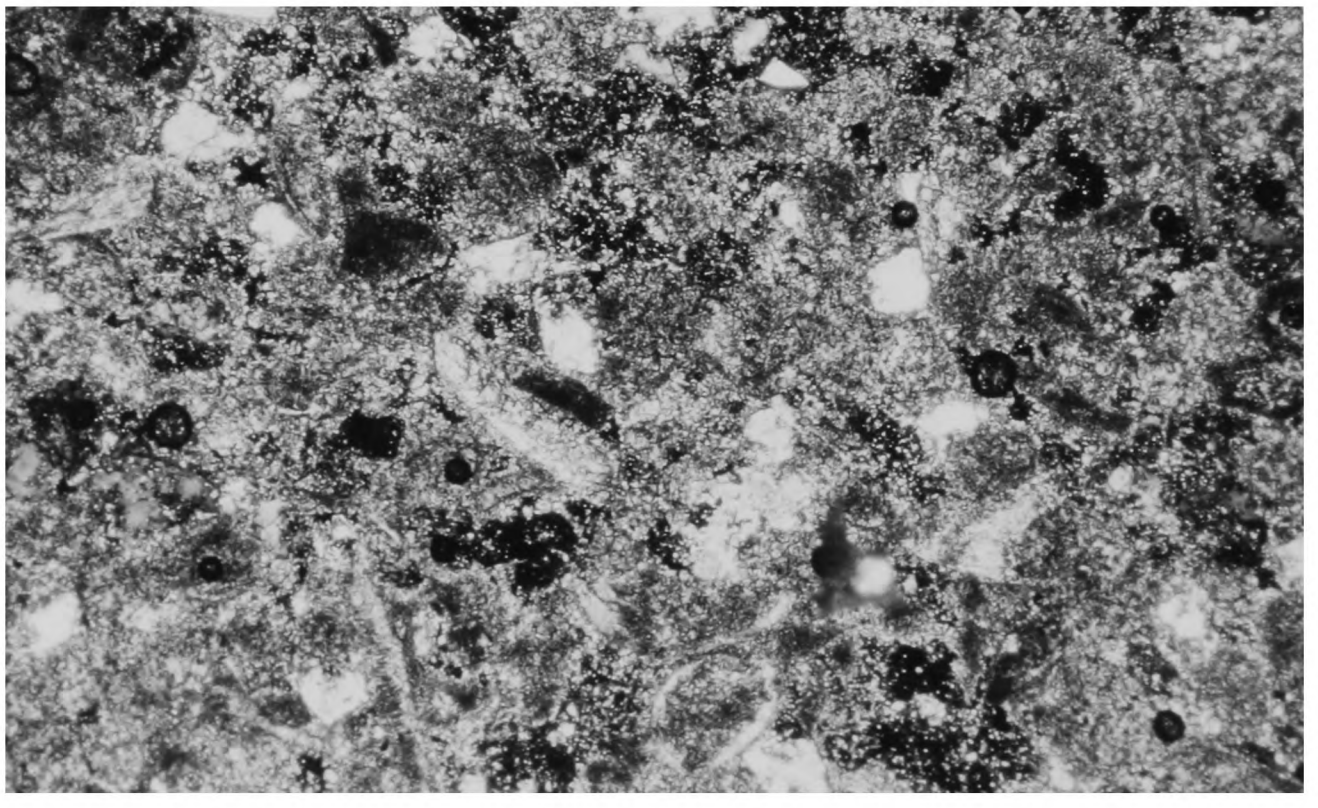


TABLE 4.1

Insoluble residue	24.14%
$CaSO_4.2H_2O$	2.79%
$CaCO_3$	68.97%
NaCl	2.36%
	98.26%

Others, K^+ 860 ppm
Sr^{++} 500 ppm
Mg^{++} 1960 ppm
Fe^{tot} 140 ppm
Al^{+++} 60 ppm

The aragonite content suggests that the material is of recent age, possibly late Pleistocene. The fact that tectonic subsidence in this area is considerable lends further support to this since deposits near present sea level are either Tyrrhenian or recent.

The most extensive Quaternary deposits in Tunisia are soil crusts, consisting of calcretes predominantly north of Gafsa and gypsum crusts to the south (see fig. 5.1). The geomorphic significance of these features is considerable. Not only do they preserve relic landforms, such as the glacis and sand dunes, but they also provide valuable indicators of past climatic régimes. Though there is a wide variety of forms of both calcretes (Durand, 1963; Goudie, 1973) and gypsum crusts these will not be dealt with here. The age and palaeoenvironmental significance of calcretes has received considerable attention in north Africa. While Pervinquière (1903) held that the Tunisian calcretes are still forming at present Mensching (1964) held that those of the Djeffara are fossil features. ^{14}C dating of calcretes and included material has shown that many Tunisian calcretes are

late Pleistocene and early Holocene (Brosche and Molle, 1975, 1976) but the possibility of present day accretion cannot be discounted. Certainly, some thick calcretes are of great age, Jaranoff (1937) and Awad (1963) suggested that Moroccan crusts were of Pliocene or Villefranchian age. In Tunisia, many calcretes are overlain by Tyrrhenian deposits (Despois, 1940, pp. 83-85), suggesting they may be late Villefranchian as Conrad (1969) proposed in Algeria. Calcrete development during cool and wet phases of the Quaternary has been advocated by a number of workers (Burollet, 1956; Guillen and Rondeau, 1966) on the basis of increased mobility of calcium carbonate under these conditions. Agafonoff (1936 a.b.) and Yankovitch (1937) pointed out that many Tunisian crusts are pedogenic features which would be destroyed by leaching under very wet conditions. Calcretes are developing at present in the Tunis area (Pimienta, 1950) therefore it is probable that the climate there approximates that in the past when calcretes fossilized earlier arid and semi-arid landforms. Average annual rainfall is 441 mm, the mean annual temperature is 18.0^{o}C and more than six months of the year have a net moisture deficit (Mensching, 1968, p. 35).

In the case of gypsum crusts, Agafonoff (1936 b.) suggested that they are also products of a prior more humid climate since if the climate became even more humid, the salts would be leached of the soil zone. Coque (1951, 1955 a.b., 1958) similarly argued that the gypsum crusts of the Chott region developed at the end of a wet period and protected the landscape from erosion during an ensuing arid phase. His suggestion, however,that the crusts are of the same ages as the glacis they mantle poses some problems. The upper (3) and middle glacis (2) of the Djebel Tebaga el Fedjedj area have gypsum crusts but the lowest (1) does not. Hence the youngest gypsum crust would be dated as late or post-Tensiftian (post-Riss) or Tyrrhenian (Zimmermann, 1963) with the post-Soltanian (post-Würm) not having experienced crust development. Le Houérou (1960) placed all the Tunisian gypsum crusts in the period between the Villefranchian and Tensiftian. In contrast, Page (1972) placed three phases of gypsum crust development in the post-

Tyrrhenian, the main phase being the most recent, in the early Holocene. The evidence for this will be examined in chapter seven, as will be the evidence for contemporary gypsum crust development.

In order to capitalize on the value of gypsum crusts in palaeoenvironmental reconstructuion, it is essential that the processes of crust genesis and preservation are understood. This is best achieved through detailed examination of the physical structure and chemical composition of the crusts. It is futile to attempt reconstructions before the characteristics of different forms of crust as well as different ages are appreciated and before the genetic processes are understood at least in part.

viii. Palaeoclimatic information.

Landform studies such as those referred to above have provided generalized information on climatic change during the Quaternary. The evidence for numerous changes of considerable magnitude is provided by erosional and depositional landforms frequently preserved by surface crusts which themselves are valuable palaeoclimatic indicators. Nevertheless, the time scale of the events and their correlation with glacial and interglacial phases in Europe is far from clear.

Most workers have favoured close correlation of wet phases in Tunisia and north Africa as a whole, with glacial periods in Europe (Awad, 1963; Coque and Jauzein, 1966). It has been suggested that the climate has become progressively drier since the Villefranchian and has been interrupted by several wet phases which have decreased in intensity in terms of their length, coolness and wetness. Coque (1958) held that the pattern is best shown in the glacis of Tunisia with sedimentation during the wet phases and erosion during the dry. Whether it is correct to correlate these periods with events in Europe on the basis of glacial corresponding with wet, is complicated by the fact that southern Tunisia is

peripheral to the Mediterranean zone. In the central Sahara, however, the largest lake basins appear to have been filled from about 22,000 to 7,500 years B.P. (Faure,1966) and in Algeria thick sequences of semi-arid to humid phase sediments were deposited (Conrad, 1963; Alimen, 1965). After this period, the central Sahara became more arid and the Mediterranean zone similarly experienced a drier climate for a short period (Hey, 1963).This was followed by a wet phase which was identified by Gobert (1948), Hey (1963) and Brosche, Molle and Schulz (1976) in the Mediterranean zone and by Conrad (1963) and Alimen (1965) in Algeria. In effect, the central and northern Sahara appear to have experienced similar changes during the Quaternary.

Palynological evidence largely supports these suggestions (Quézel, 1963). In Tunisia, Van Campo (1958) found that pollen associated with Mousterian artefacts (Würm?) indicates a lack of xerophytes and halophytes, suggesting a wetter climate from today. The presence of palms argues that the conditions were not very cold. Van Campo and Coque (1960) held that the Soltanian climate was essentially similar to present winter conditions. This would result in flooding of the Chotts. Present winter weather at Gafsa, for example, is not characterized as much by high rainfall as low temperatures relative to those in summer. Only 59% of the rainfall falls in autumn and winter but the range of mean monthly temperatures is 20.5^{o}C. In effect, if the Soltanian climate was similar to the present winter climate, the major difference in the annual régime would be a reduction in summer temperatures rather than any major change in the rainfall régime. A cooler climate would increase the availability of moisture in the soil and thereby the groundwater zone, prolonging the period of Chott inundation. The drop in temperature commensurate with that which produced periglacial conditions in the Kasserine area, may have been sufficient to allow perennial flooding of the Chotts with an absolute decrease in the amount of precipitation. Kazakhstan may be an appropriate present day climatic analogue for Soltanian southern Tunisia.

b. Central Namib Desert, Namibia.

i. Location of field area.

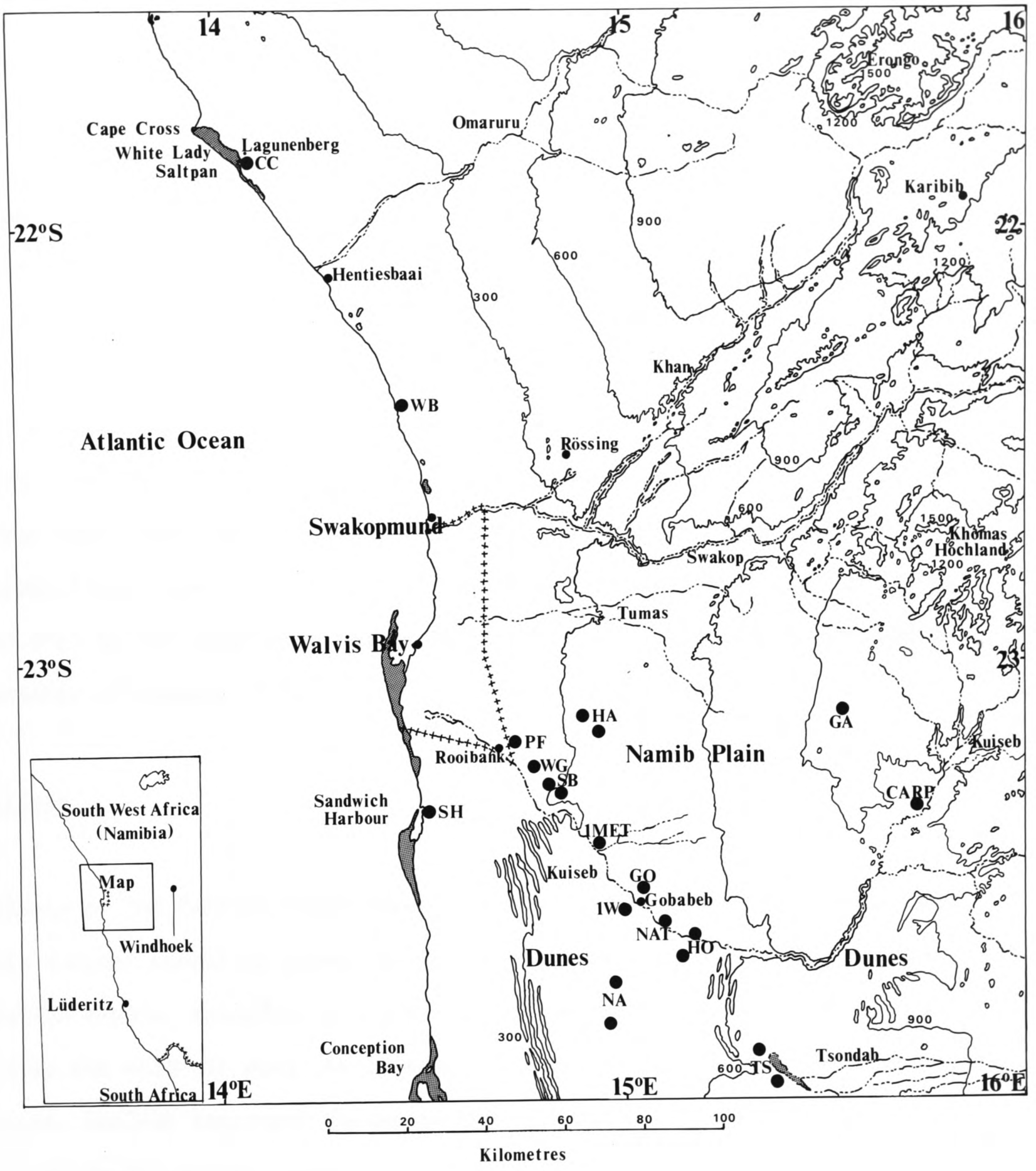
The Namib Desert forms one of the three geographic regions of Namibia, the other two being part of western Kalahari and the southern plateaux area. The Namib Desert stretches over 2,000 kilometres along the Atlantic coast of sub-Equatorial Africa, from Moçamedes in southern Angola to the Olifants River in the Cape Province of the Republic of South Africa. The desert is rarely more than 200 kilometres wide. Several different subdivisions of the desert have been proposed. Kaiser (1921) and Meigs (1966, pp. 94-98) listed three zones, the Trough Namib or Sandveldt south of the Orange River, the Plain Namib north of the Kuiseb River and the Dune Namib between the two rivers. Logan (1972) added the Skeleton Coast and the Kaokoveld, inland north of the Ugab River, though Spreitzer (1965) held that the Huab River, some 35 kilometres further north, formed the northern boundary of the Plain Namib. The field area lies within the Namib Plains and partially within the Dune Namib, on both sides of the Kuiseb River (see fig. 4.4).

The Namib plain from the Huab to the Kuiseb is 350 kilometres long and about 100 to 200 kilometres wide. It is a coastal plain, up to about 1250 m above sea level abutting the Great Escarpment which forms the Randberge, Hakosberge and Khomas Hochland in the south and outliers, such as the Brandberg, which reaches over 2,500 m, and the Erongo Mountains, further north. Though Jaeger (1923) held that the coastal plain was a feature resulting from marine erosion, Spreitzer (1965) argued that it is a product of terrestrial erosion and tectonic arching contemporaneous with uplift of the Escarpment. This interpretation agrees with the conclusions reached by Abel (1954, 1959) in the Kaokoveld, though Martin (1973 a.) held that the Escarpment was wholly erosional.

The Escarpment bordering the plains is drained by the Rivers Huab, Ugab, Messum,

Figure 4.4

The field area in the Central Namib Desert. Shaded areas are coastal pans and vleis. The contours are in metres. Sample localities are designated by the abbreviations used throughout this work.


14
15
16
Eronqo
1500
1200
Cape Cross
White Lady
Saltpan
Lagunenberg
CC
Omaruru
Karibib
22°S
900
600
300
Hentiesbaai
1200
Khan
WB
Atlantic Ocean
Rössing
900
600
1500
Khomas
Hochland
1200
Swakopmund
Swakop
Tumas
Walvis Bay
23°S
HA
GA
PF
Rooibank
WG
SB
Namib Plain
Kuiseb
CARP
Sandwich
Harbour
SH
1MET
Kuiseb
GO
Gobabeb
1W
NAT
HO
Dunes
Dunes
NA
900
300
Conception
Bay
600
TS
Tsondab
South West Africa
(Namibia)
Map
Windhoek
Lüderitz
South Africa
14°E
15°E
16°E
0
20
40
60
80
100
Kilometres

Omaruru, Swakop/Khan, Tumas and Kuiseb, none of which have surface flowing water for more than two or three days in a normal year. The only lacustrine features in the area are coastal salt pans which stretch from Swakopmund north to the Kunene River on the Angolan border. In the Dune Namib,south of the Kuiseb,there are several fossil lake beds and some which are flooded for a week or two each year when rivers flow off the highlands into the dunes. These features, Tsondab Vlei, Sossus Vlei and Koichab Vlei, are the end-points of rivers which do not have sufficient discharge to flow through the dune field (Stengel, 1966). Irregularities along the coast such as Sandwich Bay, Conception Bay, Moeb Bay and St. Francis Bay may be relic estuaries or present day, subterranean river mouths.

Along the coast the field area extends from the most northerly of these bays, Sandwich, (23° 20' 20"S; 14° 30' 0"E) to the Lagunenberg (21° 49' 40"S; 14° 5' 0"E) near Cape Cross (Kaap Kruis) in the north. Inland the field area follows the Kuiseb River from near the head of the canyon at Carp Cliff (23° 20' 20"S; 15° 45' 0"E) to the coast near Walvis Bay, most work having been undertaken in the vicinity of Gobabeb (23° 33' 40"S; 15° 3' 0"E) (see fig. 4.4).

ii. Climate.

The climate of the Central Namib Desert is dominated by the effect of the cold Benguela current upwelling along the coast. The water is so cold that little evaporation occurs, moisture is precipitated from the southerly trade wind as it rises over the cold air mass. As it subsides over the land no further precipitation occurs. Another important factor in the climate is the predominance of off-shore winds in the winter months. Further south, the Cape Province experiences on-shore winds which bring the bulk of the region's precipitation. In the Central Namib, föhn-type "berg" winds create dust storms and occasionally raise the temperatures above the summer maxima. The little rain that does fall is from southerly air masses which penetrate the off-shore wall of cold air in the summer

before the berg winds commence (Gevers, 1936; Seely and Stuart, 1976).

Meteorological data are plentiful for the Central Namib though individual stations rarely have continuous records. Despite the shortage of continuous records for the area, it may be assumed that there have not been many major climatic changes in southern Africa over the last 130 years (Hofmeyer and Schulze, 1963). This contradicts Obst's (1936) subjective observation that rainfall decreased in the early twentieth century. Some of the earliest summaries, published by Gülland (1907), provide temperature, rainfall and Relative Humidity data as well as some seawater temperatures. Sufficient data were available for Van Reenan (1925) to propose cyclic fluctuations in wet and dry years over a 35 year period. Most early records are from the coastal towns of Walvis Bay and Swakopmund with intermittant data from Rössing about 40 kilometres inland from Swakopmund. Since 1962, full meteorological records are available for Gobabeb. These were summarized by Schulze (1968) for the period 1962 to 1967 and by Seely and Stuart (1976) for 1962 to 1972. The mean annual rainfall at Walvis Bay is 22.9 mm (Kendrew, 1961) falling mostly in February and March. Boss (1941) and Jackson (1941) quoted a figure of 12.7 mm. At Gobabeb the ten year average is 16.8 mm though Schulze (1968) quoted an average of 23.8 mm in the first five years of this period. Between 1962 and 1972 annual rainfall varied from none in 1969/70 to 26.2 mm in 1966/67, the coefficient of variation being 54%. Gellert (1971) pointed out that the instability in the rainfall régime of Namibia increases toward the coast and Andreae (1974) argued that this was one reason for the lack of agricultural exploitation in the belt receiving 100 to 250 mm annually. Rainfall increases rapidly with altitude and distance from the sea. Windhoek, at about 1,700 m above sea level and 250 kilometres from the coast, receives over 360 mm annually. Hail has been recorded at Gobabeb on two occasions.

The temperature at Walvis Bay reaches its mean monthly maximum in Feburary. The mean annual temperature is 16.7^{o}C and the range in mean monthly temperatures

5.6°C (13.8°C in August and September to 19.4°C). At Gobabeb the mean annual temperature is 21.2°C and the range 6.5°C (18,3°C in July to 24.8°C in March). The absolute maximum temperature is 42.5°C in February and the absolute minimum 1.8°C recorded in July and August. The lowest daily maximum is 12.5°C in July and the highest daily minimum 25.9°C in March. The evaporation potential may reach 220 mm to 260 mm per day at Walvis Bay when easterly winds blow off the Escarpment (Boss, 1953). Potential evaporation averages 3505 mm/year at Gobabeb, ranging from 219 mm in July to 356 mm in December. Relative Humidity is very variable. When the east winds blow, Relative Humidity may be as low as 5% at Gobabeb (Schulze, 1968) and 4% at Rössing (Walter, 1937).

Perhaps the most significant aspect of the climate of the Central Namib Desert is the prevalence of fog (Nagel, 1962). On the coast, there may be 150 days with fog or mist (Walter, 1937). Walter (1936 b.) recorded an average fog precipitation of 4.3 mm/month at Swakopmund. Boss (1941) pointed out that the annual fog precipitat ion of 35 mm to 45 mm at the coast falls to about 20 mm to 30 mm 40 kilometres inland. At Gobabeb the figure is 35.2 mm annually, occurring on average 38.6 days per year. The importance of this source of moisture has been referred to in relation to the vegetation of the Central Namib. The soluble salt content of the fog has been determined by Walter (1937) and Boss (1941) and will be discussed in chapter six.

Wind directions at Gobabeb are very variable on an annual basis. Of sixteen directional sectors the frequency never exceeds 12% (SW). Average monthly velocities of winds from the different directions vary from 8 km/h (SE) to 15 km/h (SW) It is interesting to note that an average velocity of 17 km/h is exceeded only by winds from the east-north-east and east. These figures are particularly important to an understanding of the sand dune orientation south of the Kuiseb River.

iii. Solid geology.

The Central Namib Desert is composed mainly of Pre-Cambrian misa-schist with intrusions of Palaeozoic (Damara System) Khomas and Hakos Series granites and feldspar pegmatites (Gevers, 1942; Martin, 1965,1973 a.). Between Walvis Bay and Gobabeb, marble with dolerite intrustions crop out at Swartbankberg (SB). Granite intrusions form the characteristic inselberg landscape of the Namib Plains (Rust, 1970: Selby, 1977). The Permo-Carboniferous is represented by basalt covered tillites. Gevers (1929-31) identified U-shaped valleys dating from the Dwyka. In this area, Mesozoic Etjo sandstone and Kaoko lavas overlie the Dwyka. These were originally correlated with the Stormberg Series (Gevers and Frommurze, 1929) but Martin (1973 a.) argued that they were the South American Serra Géral. Cretaceous deposits are restricted to a small exposure of an exogyra bed, north of the Orange River, near Lüderitz. In the same area, there is a deposit 1.0 m to 2.0 m thick composed of silicified terrestrial pediment debris, the Pomona Beds. Kaiser a. argued that these were pre-Middle Eocene in age. Selby (1977) and Martin (1950) proposed they were Cretaceous. On the basis of a K-Ar date of 37^{+}_{-} 1 million years B.P., Martin (1973 a.) revised this to pre-Oligocene. Tertiary marine deposits are found along the coast (Haughton, 1931) but are generally heavily eroded except where consolidated as silcrete or calcrete. Phosphate deposits on the continental shelf are probably Tertiary sediments (Summerhayes et al., 1973). Late Tertiary and Pleistocene sediments consist of consolidated alluvial fan and terrace materials (Wagner, 1916; Martin, 1950; Abel, 1954) which include the diamond bearing conglomerates along the coast south of Lüderitz, as well as extensive deposits of consolidated dune sand underlying the present Dune Namib (Martin, 1965).

The bedrock within the field area is predominantly Pre-Cambrian and Palaeozoic in age and does not contain evaporite minerals. The nature of Quaternary deposits will be discussed below.

iv. Tectonics.

The earliest major tectonic event, the effects of which are still relevant to the geology of the Central Namib, was the rifting of South America from southern Africa. Though Maack (1963) held that the Dwyka tillite of the Kaokoveld was contemporaneous with that of the Paraña Basin, Martin (1973 b.) pointed out that there are no Kaokoveld erratics in South America. Hence, the rift may have been water-filled before the Dwyka. Emery et al. (1975) postulated initial rifting about 165 million years B.P. in the early Jurassic but separation about 125 million years B.P. on the basis of K-Ar dating of Kaoko and Serra Géral lavas. The presence of an enclosed oceanic basin off the coast before separation, would have a considerable effect upon the climate of the area. The submarine ridge extending westward from Walvis Bay is probably the product of faulting and volcanism resulting from plate adjustment about 80 million years B.P. rather than a scar produced as the African continent passed over a hot-spot (Kröner, 1972; Emery et al., 1975).

The uplift of the Great Escarpment and arching of the Namib Platform predates the erosion of the Central Namib Plain which itself must precede the deposition of the oldest sediments thereon. These sediments are the Eocene or Oligocene (?) Pomona Beds, hence the tectonics are of considerable age. Korn and Martin (1957) argued in favour of Tertiary uplift of the Namib Platform. This is supported by evidence of apparent coastal sabkha deposits 300 m to 350 m above present sea level (Gevers and Westhuyzen, 1931). These have been misinterpreted, probably being terrestrial rather than littoral sediments. Though Martin (1973 a.) held that the area was tectonically stable during most of the Tertiary, considerable evidence of movement is provided by deformed raised beaches along the coast. Knetsch (1938) first remarked upon the variable heights of apparently contemporaneous coastal deposits and Davies (1959 b., 1971) pointed out that high raised beaches (Tertiary?) are absent between Lüderitz and the Kunene. This, it was suggested, was the result of Tertiary and Quaternary downwarping of the Walvis

Bay area amounting perhaps to 200 m. Pleistocene beaches also show evidence of deformation between Lüderitz and Walvis Bay (Davies, 1973). Though the evidence from these coastal features is not conclusive, owing to the difficulties associated with correlating and dating the beaches, it appears that coastal subsidence has been the dominant tectonic phenomenon since the early Tertiary.

v. Geomorphic History.

Considerable differences of opinion exist regarding the ages of the various planation surfaces found in Namibia. Abel (1957) identified six main planation surfaces and terrace levels, the summit level of the Escarpment, a high level planation surface, a 1,250 m level which appears to be contemporaneous with the Pomona Beds and therefore is late Cretaceous or early Tertiary, a late Tertiary 1,050 m high level on the Namib Platform and, finally, two river terraces. This sequence compares almost exactly with that proposed by De Blij (1959) in Swaziland. Since the Cretaceous, which Korn and Martin (1937) held to be extremely arid, the Central Namib has experienced considerable climatic fluctuations. These have resulted in alternating phases of erosion and sedimentation producing an erosional platform in places covered by up to 200 m of sediments. Rust's (1970) thesis that the granite inselbergs of the Namib Plains are the product of desert weathering was attacked by Selby (1977) who argued that they are purely erosional features. Selby held that cold Antarctic Intermediate Water, flowing into the embryonic South Atlantic after the rifting of South America and southern Africa, has influenced climate since the Jurassic. The late Cretaceous and early Eocene were humid according to Korn and Martin (1937) and the Eocene is represented by evidence of marine transgression (Emery et al., 1975). Oligocene regression and Miocene transgression, laying down lagoonal sediments, appear to have been associated with aridity marking the onset of the present desert conditions (Korn and Martin, 1937; Selby, 1977). It was during the Miocene that the Kuiseb, Swakop, Ugab and other Namib rivers stopped flowing all the year round (Kaiser, 1929)

hence Ollier's (1977) middle to late Tertiary date for the Tsondab Planation.

Tectonic warping and subsidence makes identification of the beaches associated with these marine transgressions difficult (Knetsch, 1938). In southern Africa as a whole, there is a well developed sequence of raised beaches (Davies, 1971). They range from high beaches at 155 m, 110 m, 82 m and 73 m in Natal to the more extensive features at 60 m, 30 m, 18 m, 8.0 m to 9.0 m, 6.0 m, 3.5 m and 1.5 m (Davies, 1973). Between Cape Cross and Swakopmund, Davies (1959 a.b.) identified shore-lines at 27 m to 45 m, 13 m to 16 m and 2.0 m to 3.0 m and Rust and Wieneke (1972) and Wieneke and Rust (1973) at 17 m and 20 m. There are, however, submerged shores at -2.94 m and -3.38 m which may reflect recent tectonic subsidence rather than marine regressions (Wieneke and Rust, 1975). While the 3.5 m and 1.5 m beaches at Lüderitz are dated at 4,930 $\pm$ 60 years B.P. and 1,170 $\pm$ 50 years B.P. respectively (Pta 419 and 417, Davies, 1973), the 2.0 m beach at Walvis Bay has been dated 25,250 $\pm$ 1,150 years B.P. and 26,100 $\pm$ 1,835/1,495 years B.P. (Hv 5958 and 6825, Wieneke and Rust, 1975) and 27,100 $\pm$ 1,050 years B.P. (Hv 5957, Wieneke and Rust, 1979).

South of Lüderitz at Oranjemund, Hoyt (1967) and Hoyt, Oostdam and Smith (1969) recorded raised beaches at 8.0 m, 5.0 m and 2.0 m and tentatively suggested a Riss-Würm Interglacial age on the basis of a correlation with Bahaman beaches. Davies (1973), however, obtained dates of 35,000 $\pm$ 630 years B.P. and 38,100 $\pm$ 500 years B.P. (GrN 4572 and 4572) for a 9.0 m beach at Oranjemund. It is impossible to correlate the beaches at Walvis Bay, Lüderitz and Oranjemund assuming the published dates are correct and applicable to the ages of the features. The 17 m beach at Walvis Bay has been dated to 35, 950 $\pm$ 2,730/2,170 years B.P. (Hv 5231, Wieneke and Rust, 1976) and may be equivalent to the 9.0 m beach at Oranjemund. It is possible that the 5.0 m and 2.0 m beaches at Oranjemund correlate with the 3.5 m and 1.5 m Holocene beaches at Lüderitz and the submerged notches at Walvis Bay. Davies (1959 b.) suggested that Lüderitz may

represent the axis of tilt along the Namibian coast. The converse motion suggested by the positioning of the circa 35,000 years old beaches at Walvis Bay and Lüderitz remains enigmatic.

Pleistocene sea levels appear to have fallen to between 90 m (Spreitzer, 1965) and 200 m (Emery et al., 1975) below present sea level. At this low stand the Benguela Current cut a trench parallel to the coast.

vi. Quaternary landforms.

There are a wide variety of Quaternary landforms in the Central Namib but here the discussions will be confined to the river terraces and coastal lagoons. Fluvial, aeolian and edaphic sediments will be described below.

Though Abel (1957) listed only two main river terrace levels in his planation chronology of the Namib, individual rivers have a varying number of terraces with little correlation among the systems. A major reason for this is the variable periodicity of flow depending on the size of the catchments. The Kuiseb reached the Atlantic on only 15 occasions between 1837 and 1963 (Stengel, 1964) but has flowed at Gobabeb almost every year for one or two days. Over long periods (10^3 years) this has the effect of producing multiple and convergent terraces (Stengel, 1966). Spreitzer (1963) and Marker (1977) pointed out that the rivers of the Namib Plain all have convex profiles. This is unlikely to be the result of coastal uplift (Marker, 1977) since Holocene beaches appear to be submerged. It is possible that the rivers have not flowed to the coast often enough to modify their profiles since a high sea level stand when last they did. The terraces of the Kuiseb are at 40 m to 42 m, 20 m, 15 m and 6.0 m (Goudie, 1972; Wieneke and Rust, 1972; Rust and Wieneke, 1974; Ollier, 1977; Marker, 1979). The 40 m terrace is mantled by a heavily calcretized gravel, unlike the lower uncemented levels, and may be Late Pleistocene (Korn and Martin, 1957) though an

earlier age seems more probable on the basis of the magnitude of the events which deposited the sediments within the 40 m incision (Rust and Wieneke, 1974). The terraces of the lower Ugab at 25 m to 30 m, 12 m to 15 m and 5.0 m to 6.0 m (Spreitzer, 1966) reflect a sequence broadly similar to that of the Kuiseb, though the two middle terraces of the Kuiseb are cut into bedded silts up to 15 m thick (Marker, 1979). The significance of these will be discussed below.

Along the coast of the Central Namib there is a chain of salt pans which is best developed north of Swakopmund (Von Francois, 1893). Though some, such as Cape Cross Pan, are occasionally flooded at present (Gevers, 1936), Wieneke and Rust (1972) suggested they were relic features. This seems unlikely as the evidence from fossil beaches supports coastal subsidence in the area. In effect, the pans are either contemporary (Gevers and Westhuyzen,1931) or date from a transgression which is not represented by any other geomorphic features. The pans probably represent supra-tidal sabkhas developed inland from sand and shingle ridges built up by the south to north flow of the Benguela Current. This is well illustrated by the sand spit which forms Sandwich Harbour and Pelican Point which protects Walvis Bay.

vii. Quaternary sediments.

There is a wide range of opinion regarding the age of the calcretes and aeolian sands of the Central Namib Desert. Many late Pleistocene deposits are re-worked early Quaternary material and these are themselves probably derived from late Tertiary or Eocene sediments. Some of the earliest deposits still represented in the area are calcrete conglomerates which abut the foothills of the Escarpment at about 1,000 m above sea level. These calcretes are cemented fluvial and fan deposits up to 50 m thick near Carp Cliff (CARP) and near Tsondab Vlei (Ollier, 1977). Though Marker (1979) agreed with Goudie (1972) that the deposit may be Pliocene in age, a distinction may be drawn between the calcrete which predates

a cemented dune sand and that which post-dates it. The earlier calcrete conglomerate may be coeval with the Kalahari limestone which Ollier (1977) held to be early Tertiary.

The early fanglomerate (basement conglomerate, after Ollier, 1977) was sampled at Carp Cliff where the sequence is 47 m thick (see plate 4.4). The base of the sequence consists of calcrete brecciating the underlying mica-schist bedrock. Overlying this are beds of calcrete exhibiting nodular and root structures as well as some beds of pure to sandy calcrete. The sequence is capped by up to 10 m of conglomerate consisting of rounded quartzite cobbles and calcrete boulders cemented by calcrete (see plate 4.5). The micropetrographic and chemical analyses show that the texturally variable but dominantly sparry calcrete is admixed with magnesium-rich minerals in certain horizons and that the amount of insoluble residue increases up the sequence (see table 4.2).

TABLE 4.2

CARP	1.1	1.2A	1.3	1.4	1.5
I.R.%	11.10	17.86	12.02	25.24	42.80
$CaSO_4.2H_2O$%	0.47	0.40	0.33	0.41	0.33
$CaCO_3$%	48.45	78.07	45.95	40.45	50.60
Others (ppm)					
Na^+	2200	2880	2600	2500	2500
K^+	1550	1330	3650	2400	2750
Sr^{++}	1400	100	600	300	--
Mg^{++}	100000	6400	101000	78000	7900
Fe^{tot}	3600	4100	7800	5600	10800
Al^{+++}	2600	3500	5400	4400	5100
xs Ca^{++}	--	--	--	--	--

Plate 4.5 (overleaf, above)

Carp Cliff. Upper conglomeritic calcrete forming a flat-iron landscape in an area deeply dissected by the Kuiseb River in the background.

Plate 4.6 (overleaf, below)

The northern boundary of the Dune Namib. The vegetated bed of the Kuiseb separates the dunes to the south and the gravel plains to the north.





The high magnesium levels exhibited by samples 1.1, 1.3 and 1.4 as well as the high sodium, potassium iron and aluminium levels are a product of the high biotite mica content, this being derived from the local bedrock. The magnesium content through the section does not correlate with the other mineral constituents of biotite ($K(Mg,Fe)_3(AlSi_3)O_{10}(OH,F)_2$). This probably reflects breakdown of the mica (Rust, 1970) and liberation of magnesium which has accumulated in discrete horizons. The absence of a correlation between Ca^{++} determined by atomic adsorption spectrophotometry and CO_3^{--} determined chemically may indicate some dolomitization.

The increase in insoluble residue up the sequence is the converse of the trend exhibited by the second deposit of calcrete conglomerate, that which caps a relic dune sand deposit. Blümel (1976) described fanger calcretes up to 5.0 m thick in the Gaub Valley, a tributary of the Kuiseb, which were 73.1% calcium carbonate near the surface and 55.3% near the base. While the upper crust had 47.9% of the insoluble residue in the 2.0 mm to 0.063 mm size range and 20.7% silt and clay (less that 0.063 mm), the lower crust had respective percentages of 63.5% and 11.0% reflecting the deposition of the calcrete on top of the aeolian sand. The process probably involved lateral movement of calcium carbonate-rich sediments and water and possible upward movement of water by capillary rise and evaporation (Blümel, 1976). Hüser (1976) described three distinct calcretes on different pediments in the Ubib Valley and Tsondab area. He proposed a palaeoclimatic interpretation essentially similar to that of Coque (1958) with glacis formation during wet phases, consolidation with gypsum crusts or in this case calcretes during a transition and incision during the ensuing arid period. He did not, however offer any opinion as to whether these phases occurred during the Pleistocene or since the early Tertiary. Though Scholz (1971) pointed out that there are perhaps seven different genetic forms of calcrete, most of the well developed examples in the Central Namib are either fanglomerates on bedrock or those composed of cemented fluvial deposits over

dune sand. Pedogenic, lacustrine, sintric and other evaporitic forms are also represented but are usually comparatively insignificant features.

Other types of duricrusts which have been described in the Central Namib include silcretes (Reuning, 1913; Range, 1927; Jaeger, 1927; Kaiser, 1929; Knetsch, 1937) and deposits cemented by iron, manganese, barytes and celestite (Jaeger, 1927; Kaiser and Neumaier, 1932). In the arid zone, where rainfall is less than about 50 mm, the most important crust forming salts are gypsum and sodium chloride. These will be described in detail below.

Owing to the long period of aridity or semi-aridity in the coastal Namib, soil development is very limited. Pedogenic crusts have developed (Scholz, 1971) extensively near the coast but these are simply gypsum or halite cemented rock debris or fluvial or aeolian sediments. Scholz (1972, 1973) described fossil reddish-brown soils in Namibia. These occur predominantly in the east of the country bordering the Kalahari but some are found in the Namib. Frequently found beneath gypsum crusts these soils may be relic Tertiary features. Contemporary soils are characterized by pH between 7.5 and 8.5, low organic content and halloysite and attapulgite clay minerals on granite bedrock. Pan soils are more clayey than other desert soils, often 30% clay, frequently montmorillonite in granite areas (Scholz,1973), 30% silt and 40% to 60% sand. Soluble salts are abundant (Ganssen and Moll, 1961) and pH may reach 10 (Ganssen, 1960).

South of the Kuiseb River between the coast and the Kuiseb Canyon there is almost no soil development because the ground surface is dune sand. The present dunes are large longitudinal features up to 300 m high, about 1 kilometre wide and separated from neighbouring dunes by 1 to 2 kilometres of flat interdune plain. Near the Kuiseb and in the vicinities of the Tsondab and Sossus Vleis, the interdune surface is fanger calcrete, with a dominance of either sparry or micritic

calcite cement, elsewhere it is dune sand cemented with sparry calcium carbonate which makes up to about 20% of the "sandstone". The present dunes extend from the Orange River in the south to the Kuiseb in the north though at the coast they have crossed the Kuiseb and reach the Swakop. The south-north longitudinal pattern is interrupted in the vicinities of the end-points of the rivers which disappear in the dune field and also near the coast. Within 10 to 15 kilometres of the coast the dunes are more broken and trend roughly west-north-west to east-south-east. As many of the earliest expeditions in the area travelled along the coast (Seydel 1910; Trenk, 1910; Cloos, 1911; Reuning, 1913), it was held widely that the dunes all originated through coastal deflation. The fact that a continuous longitudinal dune, the Langewand, parallels the coast added credance to this. The Langewand is probably developed on or against the 17 m beach south of Walvis Bay. Besler (1976, 1977 a.) showed that the quartz grains inland and those near the coast were different. At Gobabeb, the sand is reddish in colour, while on the coast it is light yellowish-brown. She held that the interior dunes are derived from the relic dune sand which they overlie. Though it may be argued that the sand may be coloured red because the wetter climate has liberated more iron oxides from feldspars, which make up about 5% of the sand (Spreitzer, 1965), thereby giving a false impression of greater age, several other factors lend support to Besler's thesis. The morphology of the dunes and of the field as a whole supports the helical roll vortex model of seif dune formation (Hanna, 1969; Hastings, 1971) with dominant winds parallel to the longitudinal orientation, that is from the north or south in the Namib. The present wind régime shows slight directional dominance from the south-west but average monthly velocities never exceed 17 km/h. This appears insufficient to move large amounts of sand northwards, suggesting that the dune field was shaped by palaeowinds. At present the dunes are being modified by strong easterly winds in winter which have produced asymmetric cross-profiles through lee-eddying (Hoyt, 1966).

The palaeowind direction was almost certainly from the south. Cloos (1911),

Besler (1976) and Selby (1977 b.) referred to the ventifacts and dreikanter of the interdune valleys and Swartbankberg which indicate directionally constant winds of high velocity. While Selby (1977 b.) held that the faceting suggested a north-westerly palaeowind, Rand (1920) described "terrain which has been swept by an ice sheet" and McCauley, Grolier and Breed (1977) identified yardangs near Lüderitz which testify to southerly winds. Selby (1977 b.) argued that of Goudie's (1972) five hypotheses for the northern boundary of the dune field being at the Kuiseb (see plate 4.6) the most likely was that the sand originated in the river bed and moved southwards. Of the other four hypotheses two are unlikely, these being that the present boundary is fortuitous, it has been at the Kuiseb certainly since Alexander (1840) noted it, and, secondly, that the valley is too large to fill with sand quickly, downstream from Gobabeb it is not a dramatic feature. Two further hypotheses may be valid. It is feasible that the river is being deflected northwards by the advancing dunes. The lack of a well defined valley has been mentioned. While Marker (1977) suggested that the Kuiseb may have flowed to Conception Bay in the past, Wieneke and Rust (1972) held that Sandwich Harbour or Walvis Bay have always marked the Kuiseb Mouth. If this is the case, Goudie's (1972) final hypothesis is most likely to be appropriate, that is that the Kuiseb has always flowed sufficiently to remove the aeolian deposits that have accumulated in its valley since the previous flood. In the cases of the Koichab, Tsauchab, which terminates at Sossus Vlei, and the Tsondab, the discharge has not been sufficient to prevent blockage. These river valleys are not blocked by the present dunes but by cemented sandstone capped by calcrete (Maack, 1924; Besler, 1976; Ollier, 1977), though in most cases the river end-points have advanced during wet periods and incised the sandstone. Snail shells in sediments 6.0 km west of Tsondab Vlei (TS) (Lynmea natalensis and Biomphaleria pfeifferi) have been dated to 10,000 $\pm$ 2,000 years B.P. (Seely and Sandelowsky, 1974). Some 30 km further west at Narabeb (NA), lacustrine carbonates have been dated using $^{234}U/^{230}Th$ to between 210,000 to 240,000 years old (Selby, Hendy and Seely, 1979) The presence of Early Stone Age (Acheulian) artefacts supports this (Seely and

Sandelowsky, 1974). Gravels at Narabeb contain dolomite limestones which are characteristic of the Tsondab and are absent from Kuiseb gravels. Hence, it is unlikely that the Kuiseb flowed to Conception Bay but possible that the Tsondab flowed to Sandwich as Wieneke and Rust (1972) suggested.

At Narabeb, the underlying sandstone is not incised as it is at Tsondab Vlei (Maack, 1924) and Sossus Vlei (Fleck, 1899) therefore this sand is greater than 250,000 years old. The remobilization of the sand into the present dunes is a Pleistocene and Holocene event, though the breaks in the longitudinal alignment between the present vleis and the coast suggest either that the rivers have flowed without incising or that aeolian activity has not been sufficient to move the sand across the former water courses. The sandstone has a significant geomorphic effect which is best seen along the Kuiseb River. Though there is a calcrete cap on both sides of the valley, to the north it is much dissected as is the underlying mica-schist creating a "gramadulla" landscape (Rust and Wieneke, 1974) (see plates 4.7 and 4.8). To the south it is intact except for some micro-karstic features (Hüser, 1976). Apparently the fossil dune sand acts as a permeable medium through which surface water can drain, whereas on the Basement Complex, to the north of the river, this is not the case and erosion has been pronounced.

In the same stretch of the Kuiseb as the gramadulla relief, between Homeb and Ossewater, there are two river terraces developed on fine bedded silts. This material may be the same as that near Gobabeb described as "Flutlehm" (Scholz, 1968) containing pockets of gypsum and up to 17% calcium carbonate. Near Natab, the silt is pale yellow (5Y 8/3) but in places is interbedded with orange (7.5 YR 6/6) sand. Horizons of irregularly-shaped concretions occur within the silts. Their mineralogical composition is shown in table 4.3 (NAT 1.3).

Plate 4.7 (overleaf, above)

Gramadullas on the north bank of the Kuiseb River near Homeb. Note incised silt deposits adjacent to the vegetated river bed.

Plate 4.8 (overleaf, above)

Gramadullas near Homeb.



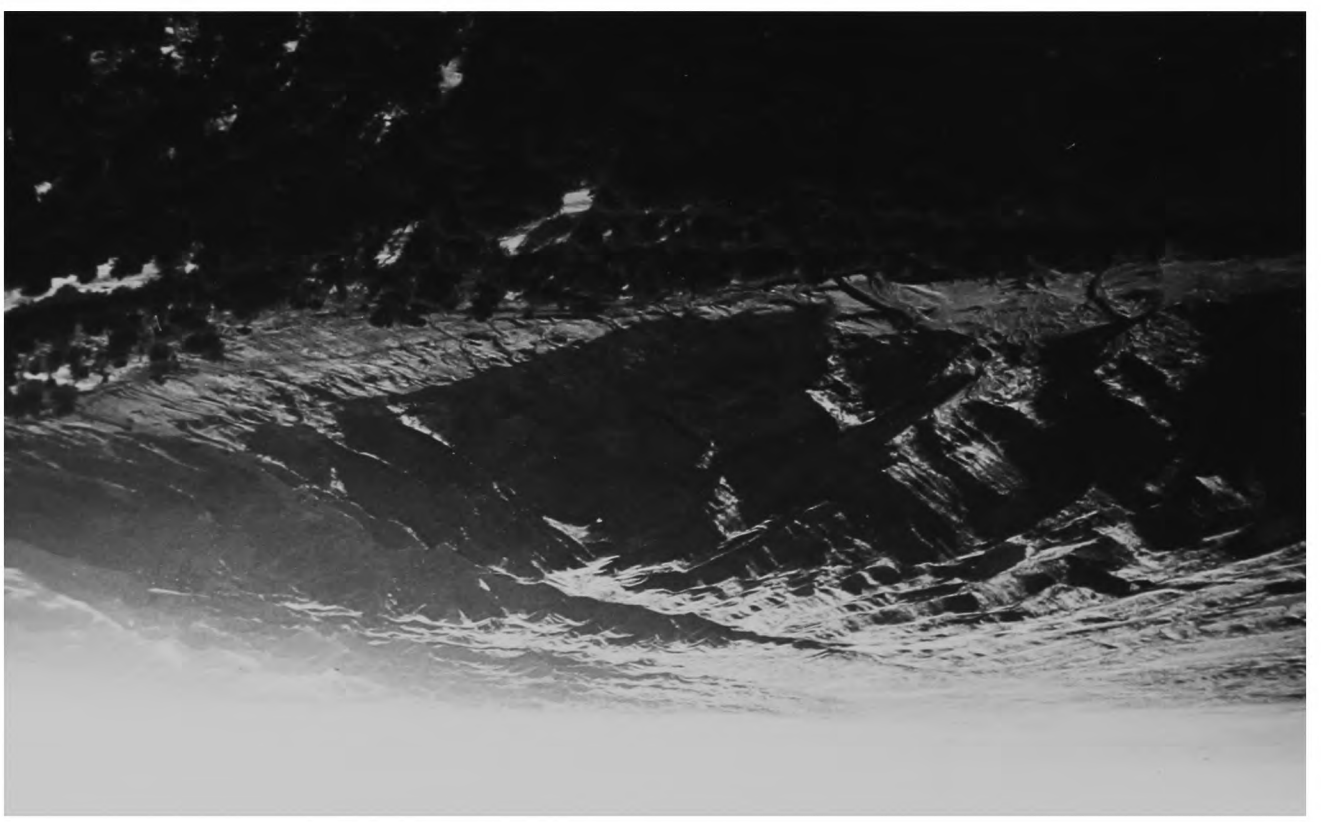


TABLE 4.3

	NAT 1.3	NAT 1.4	NAT 1.1
I.R.%	35.34	58.92	39.46
$CaSO_4.2H_2O$%	2.25	3.34	8.20
$CaCO_3$%	53.10	2.53	4.14
Others (ppm)			
Na^+	2600	76000	120000
K^+	4700	5000	11000
Sr^{++}	700	100	--
Mg^{++}	27300	9400	9500
Fe^{tot}	11000	19500	23400
Al^{+++}	7500	12900	15000

They are essentially calcium carbonate cemented silt, micropetrographic analysis shows abundant sparry and micritic calcite. The silt itself is derived from the bedrock upstream which is mica-schist. The weathering of biotite account for the accessory cations except sodium and the small amount of strontium. The origin of the sodium is evident from the analysis of samples NAT 1.1 and 1.4. These consist of white efflorescences on the side and at the base of the silt terrace which were identified as sodium chloride (see plate 4.9). As chloride, the sodium in samples NAT 1.1 and 1.4 would respectively account for another 30.50% and 19.32 % by weight. In NAT 1.3 the figure would be 0.66%. The salt is probably introduced in fog moisture, is precipitated at the surface and later leached through the sediment.

This material at Natab was sampled from the lower of two terraces at about 5.0 m and 10 m to 12 m above the present river bed. Silt from the upper terrace at Homeb, about 10 km further upstream, is mineralogically very similar (see table 4.4) though it is from a deposit over 20 m above the river bed. The potassium,

magnesium and iron derived from the weathering of biotite show a close relationship to the theoretical concentrations. Sample HO 1.1 contains 11000 ppm potassium which would be equivalent to 17.52% by weight biotite. The magnesium and iron concentrations equate with theoretical biotite contents of 13.24% and 13.49% respectively. The fact that the aluminium concentration equates with a much higher biotite content (51.94%) may reflect either differential natural breakdown of the mica, differential decomposition on boiling in 1.0 N hydrochloric acid before analysis, or the different solubilities of the weathering products leading to differential leaching through the deposit. The sediment contains some small gypsum roses, the age of which is indeterminate, as well as fossil desiccation cracks, root casts, gastropod shells (Bulinus (Bulinos) tropicus and Biomphalaria pfeifferei) and coprolites (Marker and Mueller, 1979). Though Rust and Wieneke (1974) interpreted the silts as lacustrine deposits, this is not necessarily the case.

TABLE 4.4

	NAT 1.2	HO 1.1	HO 1.3
I.R. %	75.28	67.56	72.28
$CaSO_4.2H_2O$%	0.53	0.30	0.61
$CaCO_3$%	0.92	2.38	trace
NaCl%	0.38	0.36	0.51
Others ppm			
K^+	105000	11000	105000
Sr^{++}	--	--	--
Mg^{++}	14400	15500	13900
Fe^{tot}	33900	36300	32500
Al^{+++}	19800	22500	20700
xs Ca^{++}	1100	7800	8600

The faunal evidence is inconclusive regarding the environment but the desiccation

Plate 4.9 (overleaf, above)

Silt terrace abutting mica-schist bedrock in the Kuiseb Valley near Natab. Note the white efflorescences of sodium chloride on the sides of the terrace.

Plate 4.10 (overleaf, below)

Thin sheet of Kuiseb silt overlying aeolian sands near Rooibank. Cracking is a result of desiccation.





cracking as well as interbedding of sands with the silts suggests intermittant flooding and silt deposition (Marker, 1977, 1979; Ollier 1977). While Rust and Wieneke (1974) proposed blocking of the Kuiseb by northward moving dunes and ponding behind the barrage, Marker and Mueller (1979) held that an ephemeral body of water, similar to that found at Sossus Vlei at present (Grobbelaar, 1976) was more probable. It is not essential, however, to invoke dune blockage as do both of these models. It is possible that reduced discharge or frequency of flooding prevented river water reaching the coast and this in turn allowed dune encroachment. This is occurring downstream from Rooibank at present. Whether owing to reduced flooding or increased exploitation of Kuiseb groundwater to supply Walvis Bay with drinking water (Spreitzer, 1963), the river rarely flows past the Rooibank pumping station and the dunes have advanced northwards threatening Walvis Bay and Swakopmund. In the Kuiseb delta area, relic silt sheets provide evidence of river flow where now there are dunes up to 50 m in height (see plate 4.10). Similar aeolian-fluviatile interbedding has been reported from Pennsylvanian, Permian deposits in Colorado (Fryberger, 1979).

viii. Palaeoclimatic information.

This may be subdivided into two categories, information based upon findings from within the Namib and that extrapolated from other areas of Africa.

From the Namib there is relatively little information. It was pointed out by Kaiser (1923, 1926, 1927) that many landforms in the Namib are the products of climatic fluctuation. Troughs of the southern Namib which were formed by deflation under extremely arid conditions now form basins of sedimentation. This concept was developed by Wieneke and Rust working in the Kuiseb Valley and on the coast around Walvis Bay and Swakopmund. Their approach was based on the premise that the landforms of any one area could be classified as products of either humid conditions and active erosion, arid conditions and active aeolian

erosion and sedimentation, or arid conditions and stability which must intervene between the first two. In the Kuiseb Valley, for example, the terraces are products of humid active environments as are the gramadullas. Arid active conditions saw the advance of dunes and the deposition of silts. Arid stability is characterized by incision of the Kuiseb Canyon upstream from Homeb (Wieneke and Rust, 1973 a.; Rust and Wieneke, 1974). Along the coast they examined raised beaches, the grain size and shape of marine and fluvial sediments and gypsum crusts (Wieneke and Rust, 1973 b., 1975, 1976; Rust and Wieneke, 1973, 1976; Rust, 1975; Wieneke, 1975). High sea levels correlate with dry periods and low levels with wet. Gypsum crusts developed during arid stable conditions as sea level was falling and climate becoming less arid. They did not, however, clarify their views on the mode of crust development. Rust and Wieneke (1974) proposed twelve distinct phases inland following the cutting of the 40 m Kuiseb terrace. On the coast thirteen different phases were identified post-dating the beginning of the development of the 17 m beach. This was dated to about 35,000 years B.P. but Rust and Wieneke (1976) accepted that this did not imply that the Kuiseb 40 m terrace was of the same age. Areas even closer to each other, within 50 km, would not necessarily have the same chronology because arid stable conditions may persist owing to the absence of processes indicative of humid or arid active conditions.

The ^{14}C and $^{234}U/^{230}Th$ dates referred to above for the end-points of the Tsondab River do not necessarily indicate wet phases over the last 250,000 years since they may also reflect a gradual eastward migration of Tsondab Vlei with progressive desiccation. A preliminary age of 12,000 years obtained from gastropod shells in the Kuiseb silts suggests late Pleistocene aridity, if it is valid to attribute the silt deposition to dune encroachment (Marker and Mueller, 1979). If Wieneke and Rust's ^{14}C dates and palaeoclimatic interpretations are correct, the last major wet phase in the Central Namib would have been during the period of low sea level between about 35,000 and 25,000 years B.P. Whatever the dates

of Pleistocene wet phases, they appear to have been too minor to allow river flow through the Dune Namib (Korn and Martin, 1957).

Though parts of the interior of Namibia, for example, the western Kalahari and the Fish River, have experienced considerable climatic change during the late Pleistocene and Holocene, the effect of the cold Benguela Current off the coast has maintained aridity in the Namib Desert. The wettest phases may be represented by calcrete development but it is difficult to ascertain the precise climatic implications of these deposits because it is uncertain whether they are currently active or relic. Elsewhere in southern Africa, many workers have identified cool wet climatic phases which appear to correlate with the glacial periods of the temperate regions of the northern hemisphere. For example, Van Zinderen Bakker (1962, 1963, 1967) and Maud (1968) correlated African Pluvials with European Glacials and Butzer (1966) and Van Zinderen Bakker (1969) held that there was a close climatic correlation between the areas in the early Holocene. More recent work has shown that such gross comparisons on continental scales are erroneous. Though parts of South Africa, Lesotho, Swaziland and Zimbabwe have experienced periglacial conditions which correlate with Würmian events in Europe (Alexandre, 1962; Sparrow, 1967; Harper, 1969; Linton, 1969) the interior and sub-tropical zone was more arid in the late Pleistocene (CLIMAP, 1976; Gates, 1976). Whether or not the coastal Namib became wetter or drier at this period is still open to debate. As yet there is no conclusive evidence. Simulations of late Pleistocene climatic conditions have added little to the state of knowledge in the area. While temperatures appear not to have fluctuated greatly (CLIMAP, 1976) it is indeterminate whether climate was more arid owing to a colder Benguela Current or whether winter rainfall brought to the Cape Province by westerly winds extended far enough northwards to affect the Namib.

The ensuing description of the physical and chemical caracteristics of the various types of gypsum crusts and the discussion of their origins may in it-

self add to the understanding of Namibian palaeoclimates. It must be reiterated here that their value as palaeoclimatic indicators may only be assessed when their mode of origin is understood.

CHAPTER 5

THE NATURE OF GYPSUM CRUSTS

The physical structure and chemical composition of gypsum crusts in southern Tunisia and the Central Namib Desert will be examined in detail. The results of these analyses will be compared with the information available from other areas. To facilitate this comparison it is necessary to establish a preliminary classification of the different types of gypsum crusts. Additional categories may be required as the information becomes more detailed and extensive but initially it is possible to include most of the types of crusts in one of three categories. The development of this system will be explained here.

a. Classification systems for gypsum crusts and soils.

For the purposes of describing the distribution of gypsum accumulations in chapter three, the features were limited to those at or within 5.0 m of the land surface, containing at least 5% more gypsum than the underlying bedrock and a minimum of about 15% gypsum. A number of classifications of gypsum crusts have been proposed but owing to the problems of definition few of these have been based wholly on either structural or genetic characteristics. Most early classifications were based on the structure of the features and were often 'borrowed' from calcrete classification. While Durand (1949 a.) identified only one major type of gypsum crust in the Algerian Souf, the desert rose or croûte de nappe type, his general classification of crusts suggested that gypsum crusts could have several forms similar to those exhibited by calcretes in north Africa. The main types were croûtes zonaires (laminated crusts), encroûtements (massive crusts), formations pulvérulents (powdery accumulations), petrified roots and nodular forms (Durand, 1963). The earlier classification included efflorescences, desert varnish and travertine (Durand, 1949 b.). Though the classification

is structural, the distinctions were seen to be the result of genetic factors. For example, while encroûtements were suggested to be the products of capillary rise or subsurface evaporation of groundwater, croûtes zonaires were held to be formed by chemical precipitation from slope wash (Durand, 1951). These subdivisions have not been upheld by workers concerned solely with gypsum crusts. Stainier (1912) identified surface and subsurface crusts and proposed a genetic distinction. A similar classification was put forward by Bellair (1954) who suggested that the subsurface rose de sable (sand rose) type, which is the same as Durand's encroûtement, was produced by precipitation from groundwater while the surface type, 'en plaques', formed when shallow surface waters evaporated. Though the physical structure of this evaporitic crust is unusual, with crystals 'de grandes dimensions (plus de 0.10 m) parfaitement limpides et transparents', the only similar forms being those described by Gibson (1962) in Antarctica, the category of crusts formed by evaporation of standing water bodies is a distinct one.

While many of the gypsum crusts that have been described fall into one of these two broad categories, subdivisions are necessary. Kulke (1974), for example, identified two forms of water table crust in Algeria. The term croûte de nappe (water table crust) (Pouget, 1968) will be adhered to here in preference to encroûtement or rose de sable since it implies a subsurface crust formed by precipitation of salts from the groundwater table. Fathi et al. (1972) also recorded two forms of croûte de nappe in Egypt. There are, however, many powdery accumulations and crusts of gypsum which have the structure of neither the croûte de nappe nor evaporitic types. They include the flour and seed gypsum of Australia (Jack, 1921; De la Hunty and Low, 1958), many of the Tunisian crusts (Coque, 1955 a.; Page, 1972), as well as those in Central Asia and the Namib. These may be placed in Coque and Jauzein's (1966, 1967) third category of surface accumulations which are neither lacustrine nor sabkha evaporites. A subaerial origin provides the distinction from the croûtes de

nappe (Coque, 1955 a.b., 1958, 1960). This third type is difficult to define structurally because of the wide variety of forms. Not only are indurated crusts with 80% to 90% gypsum included but also powdery accumulations which may have as little as 20% gypsum (Pouget, 1968; Page, 1972).

Le Houérou (1960) classified the surface accumulations as calcimorphic palaeosols, the term croûtes pavimenteux (pavement crusts) being used to describe the indurated crusts as distinct from the powdery forms. Further subdivision was proposed by Bureau and Roederer (1961); Calcimorphic soils :

A. Subset (calcium carbonate-rich)

B. Subset (gypsiferous)

1) Contiguous accumulations.

i Crusts

ii Powdery accumulations

iii Nodular accretions

2) Dispersed accumulations

i Microcrystalline

ii Diffuse

iii Macrocrystalline

The same subdivision is possible in the case of hydromorphic soils, essentially the croûtes de nappe, and possibly also halomorphic soils which the evaporitic sabkha deposits exemplify (Cointepas, 1971). A detailed subdivision based on the French soil classification system is not particularly useful in this examination of the nature and origin of the forms. The dispersed accumulations are merely incipient forms, while the subdivision of contiguous forms may represent different temporal stages in the development of a crust (Pouget, 1968). For this reason the work of pedologists such as Cointepas and Gaddas (1971) and Vieillefon (1976) in Tunisia will be re-evaluated within a simpler framework consisting initially of surface accumulations, croûtes de nappe and

surface evaporite deposits. This is similar to the system used by Tolchel'-nikov (1962) in Central Asia and to that proposed by Coque and Jauzein (1966, 1967) in Tunisia. Vieillefon (1976) recognized these three types and added gypsic soils associated with gypsum-rich bedrock and subclasses according to pedologic development and depth of the gypsum horizon. Similarly, though Cointepas and Gaddas (1971) included only two main types of gypsum soils in their Tunisian soil map, a number of other soil types were acknowledged as having significant gypsum concentrations (Cointepas, 1971) (see fig. 5.1).

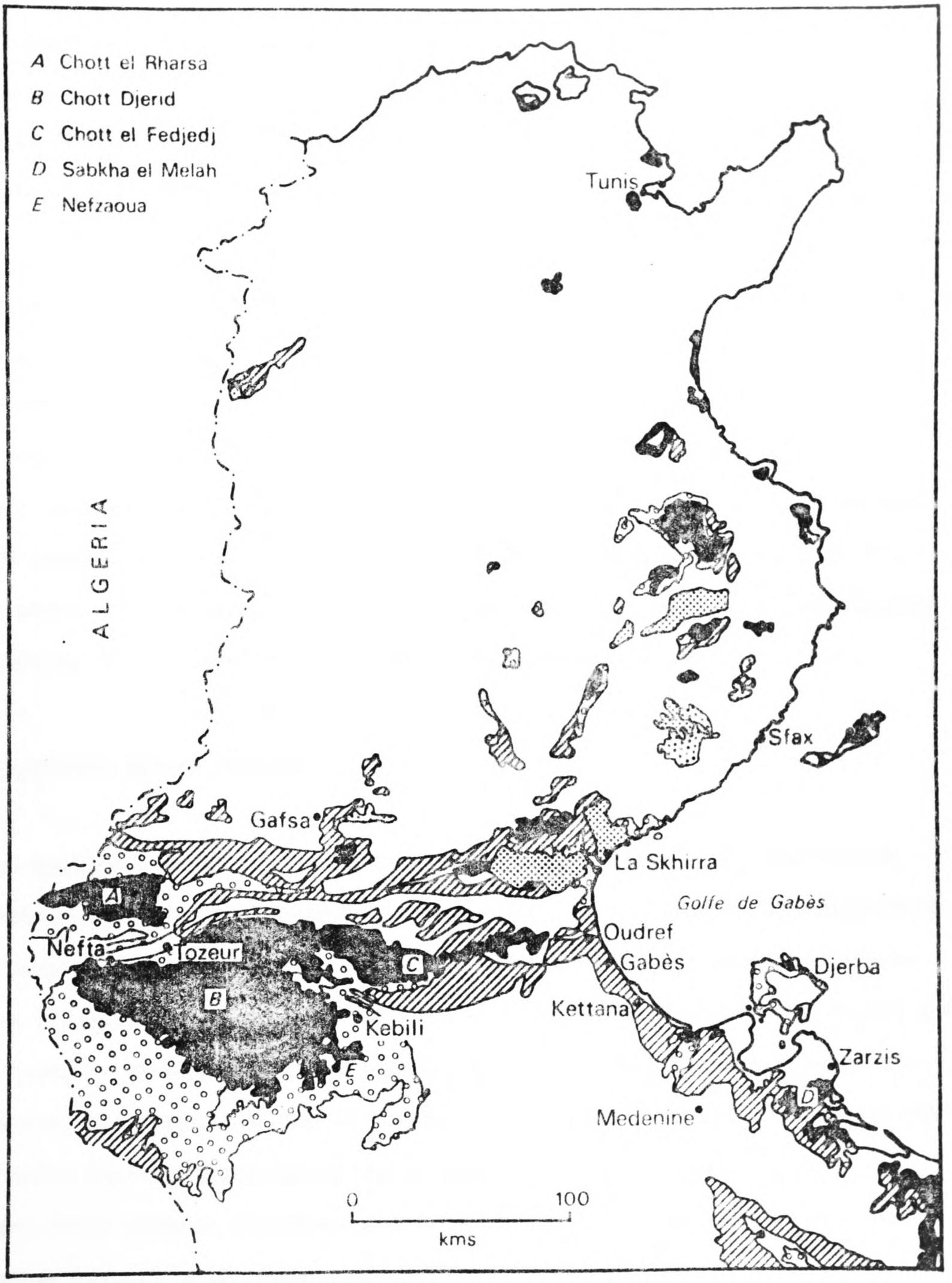
Nodular gypsum accumulations have not been described widely but Pouget (1968) held that they represented an early stage of croûte de nappe development. Bureau and Roederer (1961) used the Arabic term ras-kelb but did not propose a particular process for its development.

Gypsified root casts have also been described in some areas (Glennie and Evamy, 1968, Trucial Coast; Pouget, 1968, Tunisia; Fathi, 1972, Egypt; and Kulke, 1974, Algeria) though again their significance is uncertain. Crusts up to 0.60 m thick developed upon gypsum bedrock have been reported in Tunisia (Bureau and Roederer, 1961). Their importance and extent is difficult to gauge in this area where gypsum is extremely abundant in the bedrock. The soil map of Cointepas and Gaddas (1971) (see fig. 5.1) shows some of the major accumulations of this type but it is likely that other surface crusts are of the same origin. This hypothesis will be examined further in chapter six, for the purpose of this chapter the crusts on bedrock may be included within the category of surface accumulations.

Though the simplified classification system which will be adopted here has the advantage of being able to accommodate most of the types of crust described by other workers, in that it is a compromise between structural and genetic systems, some problems arise. For example, some croûtes de nappe have been exposed by deflation and have become surface features (Coque, 1962, pp. 91 - 93;

Figure 5.1 (overleaf)

Distribution of gypsum crusts and gypsum soils in Tunisia, adapted from Cointepas and Gaddas (1971).
1. Chotts and sabkhas. 2. Surface gypsum crusts.
3. Gypsum crusts and soils associated with lithosols.
4. Gypsum crusts associated with saline soils.
5. Gypso-calcareous crusts.


A Chott el Rharsa
B Chott Djerid
C Chott el Fedjedj
D Sabkha el Melah
E Nefzaoua
ALGERIA
Tunis
Sfax
Gafsa
La Skhirra
Golfe de Gabès
Oudref
Gabès
Djerba
Nefta
Tozeur
Kettana
Kebili
Zarzis
Medenine
0
100
kms
1
2
3
4
5

Pouget, 1968). For some time after exposure the crusts may have the characteristic structure of their genetic type preserved. In time, however, owing to the susceptibility of gypsum to solution and dehydration, these structures may be destroyed creating apparently anomalous or intermediate forms. Some good examples of degrading croûtes de nappe are found in the Central Namib Desert near Wlotzkas Baken.

b. Large scale structural characteristics.

The three major crust types and several of the less extensive types will be examined. Characteristics such as thicknesses of the deposits, their geomorphic locations and their physical and crystalline structures will be described combining evidence from the field areas and from other parts of the world where the features have been examined. The micro-structure and chemical composition of the crusts will be examined in detail below.

i. Evaporitic gypsum crusts.

As was pointed out in chapter three there are some problems related to the definition of evaporitic gypsum crusts, notably the difficulty of subdividing them into recent crusts and bedrock deposits. Furthermore, some confusion exists in the literature regarding the differentiation of these evaporitic crusts and some forms of groundwater crust (croûte de nappe). This is particularly important in the classification of supratidal evaporite deposits in coastal sabkhas. In chapter two the characteristics of some sabkha gypsum deposits were described. Their laminar structure and crystal habits are apparently the result of gypsum precipitation from super-saturated brines within beach deposits, not at the surface. In this context these gypsum accumulations are perhaps more akin to croûtes de nappe than surface evaporite crusts. These crusts are generally 0.10 m to 1.25 m in thickness but may reach 8.50 m (Evans et al.,

1969) and contain about 50% by weight gypsum, though some alteration to anhydrite may occur (Shearman, 1963; Kinsman, 1964 a.b., 1965, 1969; Butler, 1969; Kendall and Skipwith, 1969). Surface water evaporation, precipitating gypsum and sometimes halite as well, also produces bedded deposits but the crystal habits are often markedly different from the subsurface, supra-tidal evaporitic form. They will be described below. Furthermore, the gypsum content is usually higher in the case of the subaerial evaporites. In Bahrain, pan deposits contain 80% to 90% gypsum (Brunsden, Doornkamp and Jones, 1976) and concentrations of over 90% have been reported from coastal sabkhas in southern Tunisia (Busson and Perthuisot, 1977). While the deposits may be very thin (0.10 m to 1.0 m) in small desert pans (Moore and Hayes, 1958 in New Mexico; Le Houérou, 1960; Coque and Jauzein, 1966, 1967 in Tunisia; Krishnan, 1963 in Rajasthan; Risacher, 1978 in Bolivia), large coastal lagoons and continental sabkhas may exhibit sequences of considerable thickness. Srivastava (1969) reported deposits 2.0 m thick containing up to 87% gypsum in India, Phleger (1969) 3.0 m sequences of gypsum and halite in Mexico, Morris and Dickey (1957) 10 m of gypsum in a Peruvian coastal lagoon and Busson and Perthuisot (1977) 30 m of bedded evaporites from the Sabkha el Melah near Zarzis in Tunisia. Other examples of evaporitic pan and lake deposits are found in Australia around Lake Eyre (Bonython, 1956; King, 1956; Wopfner and Twidale, 1967) and in Central Asia (Fersmann and Wlodawetz, 1926; Tolchel'nikov, 1962) (see tables 3.1 to 3.6).

Samples were collected from four sites for the purposes of this work. In Tunisia, evaporitic crusts from the Sabkha el Melah (CM) near Zarzis and from near Chott el Rharsa (CR) were sampled. In the Central Namib, a coastal lagoon near Cape Cross (CC) and a small, dry pan north of Narabeb (NA) provided samples of gypsum of evaporitic origin (see figs. 4.1 and 4.4 for locations).

The Sabkha el Melah lies about 10 km south-west of Zarzis and is cut off from the Mediterranean by a ridge of dune covered aeolianite, calcium carbonate

cemented wind blown quartz sand and foraminifera tests. The sabkha is about 15 km long and 10 km wide (Busson and Perthuisot, 1977) probably representing and oued (a seasonal watercourse) drowned owing to tectonic subsidence (Bureau, 1960, see chapter four). Two areas of sabkha were examined, the first near a chain of aïoun (salt springs) (see plate 5.1) in the north-west the second near the south-eastern shore. The first site exhibited a surface layer of large, prismatic gypsum crystals transparent except for poikilitically enclosed rounded pebbles up to 20 mm in diameter (sample CM2, table 5.7). The pebbles are probably of fluvial origin, while the sheet of selenite is evaporitic. Transparent, prismatic gypsum layers have been described by Bellair (1954) in the Chott el Rharsa area but they have not been recorded elsewhere in Tunisia. Large crystals are unusual in surface evaporite deposits suggesting that these crusts may be shallow-water deposits or possibly subsurface. The deposits to the south-east of the sabkha are much more typical of surface evaporites. The gypsum beds form a small island raised 1.0 m to 1.50 m above a surrounding halite crust, the present sabkha floor (see plate 5.2). The surface of the islet is composed of microcrystalline gypsum which has been considerably distorted by doming. These 'Quellungshohle' are up to 0.50 m in height and about 1.0 m to 1.50 m in diameter. The disruption is probably the result of volumetric changes during dehydration and rehydration of gypsum and anhydrite (see chapter two) but other processes, such as hydrogen sulphide eruption (Meckelein, 1977), may be significant in some cases. The size of the gypsum crystals increases down through the 1.0 m crust. At the base there are well developed arrow-head twins (see plates 5.3 and 5.4), twinned on the hemiorthodome (101). The gypsum crust ends abruptly where it meets a bed of highly plastic white clay (see fig. 5.2A).

The distortion of the gypsum beds and the isolation of the islet indicate that this is an old deposit. Nevertheless, the essential characteristics of evaporitic gypsum crusts are preserved. These characteristics are well exemplified in the case of crusts from the Chott el Rahim, an eastern arm of Chott el Rharsa,

north of Tozeur. The perimeter of this small chott is remarkable for the dramatic changes in relief. Bluffs up to 5.0 m in height fringe the chott floor. These have a stepped appearance, a higher bluff, 3.50 m high, is cut in to silty sand partially indurated with gypsum. These sediments lie upon bedded gypsum deposits, 1.50 m of which are exposed in the lower bluff, (see plate 5.5). The bench which separates the two cliffs is the surface of the evaporite deposits mantled with pebbles.

The upper 0.85 m of the lower exposure is composed of gypsum beds 50 mm to 100 mm thick. These exhibit the same variation in crystal size as was noted at Sabkha el Melah. In this case, however, microcrystalline gypsum at the top of each stratum grades downwards into lenticular crystals up to about 1.0 mm in diameter above the sharp discontinuity with next microcrystalline surface (see plate 5.6). It should be noted that this variation in crystal size through individual beds is the opposite to the inverse graded bedding described by Carozzi (1960, pp. 407 - 409). The examples he cited show decreasing crystal sizes downwards through the laminae. Several explanations may be proposed but since the origins of his examples are not discussed in detail, lengthy discussion is not meritted. Suffice it to point out here that a standing body of shallow water evaporating completely will tend to precipitate smaller crystals late in the cycle as the potential for prolonged ionic migration decreases. The strata are horizontal but have tepee structures at 0.50 m to 1.50 m intervals. The tepee points are formed by upturning and breaching of strata at discrete points. They have been described in bedded dolomites where the folds were 10 m in height (Termier and Termier, 1963). In this case the disruption was held to be the result of diapiric movement of gypsum beneath the dolomite. The tepee structures in the Chott el Rahim are probably desiccation cracks formed when the lake dried up completed resulting in contraction of clays (Butzer and Hansen, 1968) or the gypsum itself (Christiansen, 1963). Krinsley (1970) held that upward movement of brine may cause the upturned

edges of polygons on salt lakes as Bellair (1957) had suggested.

Below this zone there is 0.45 m of interbedded gypsum and clay. The strata are the same dimensions as those in the upper zone but the lower part of each varve is clay not lenticular gypsum. The top of each stratum is again microcrystalline gypsum. This horizon rests upon clay which extends to the base of the exposure (see fig. 5.2B).

Coque and Jauzein (1967) suggested that the gypsum bluffs found around Chott el Rharsa were structural terraces of tectonic origin. It is true that the chott is a product of subsidence (see chapter four) but it is unreasonable to propose rifting of these Quaternary sediments (they rest upon Mio-Pliocene deposits) on such a local scale. The topography is probably the result of deflation of the Pleistocene sediments in the chott. Besler (1977) described small yardangs composed of evaporitic gypsum and cemented gypsum dune sand on the southern shore of Chott el Fedjedj. Wind fluting is not well developed at Chott el Rahim but differential erosion dependent on the degree of gypsum cementation suggests aeolian excavation of small oued courses.

Of the two locations of evaporitic gypsum crusts examined in the Central Namib Desert, one is very similar to that at Sabkha el Melah and the other differs from both of the Tunisian examples. In the White Lady Salt Pan, south of Cape Cross (CC) on the sea-ward side of the Lagunenberg (see plate 5.7), there are a number of gypsum and basalt islands rising from the polygonal salt crust (see plate 5.8). The gypsum (CC 2.2) is found in distinct beds 50 mm to 100 mm thick, similar to those at Chott el Rahim in Tunisia. Though the islet that was examined was nearly 3.0 m in height, there were no vertical faces so the overall structure of the deposit was difficult to ascertain. It would appear that it is an isolated remnant of a previous lagoon floor or shore. Some gypsum banks are currently forming around the shores of the smaller lagoons

Plate 5.1 (overleaf, above)

Aïoun (salt springs) in the north-west part of Sabkha el Melah (CM), see table 6.1 for chemical analyses of water samples.

Plate 5.2 (overleaf, below)

Blistered evaporitic gypsum crust forming "Quellungshohle" on the south-eastern shore of Sabkha el Melah.





Plate 5.3 (overleaf, above)

Large gypsum crystals from the evaporitic crust at Sabkha el Melah (see fig. 5.2A).

Plate 5.4 (overleaf, below)

Arrowhead twinned gypsum crystals from the base of the evaporitic crust at Sabkha el Melah (see fig. 5.2A).


CM 5.2

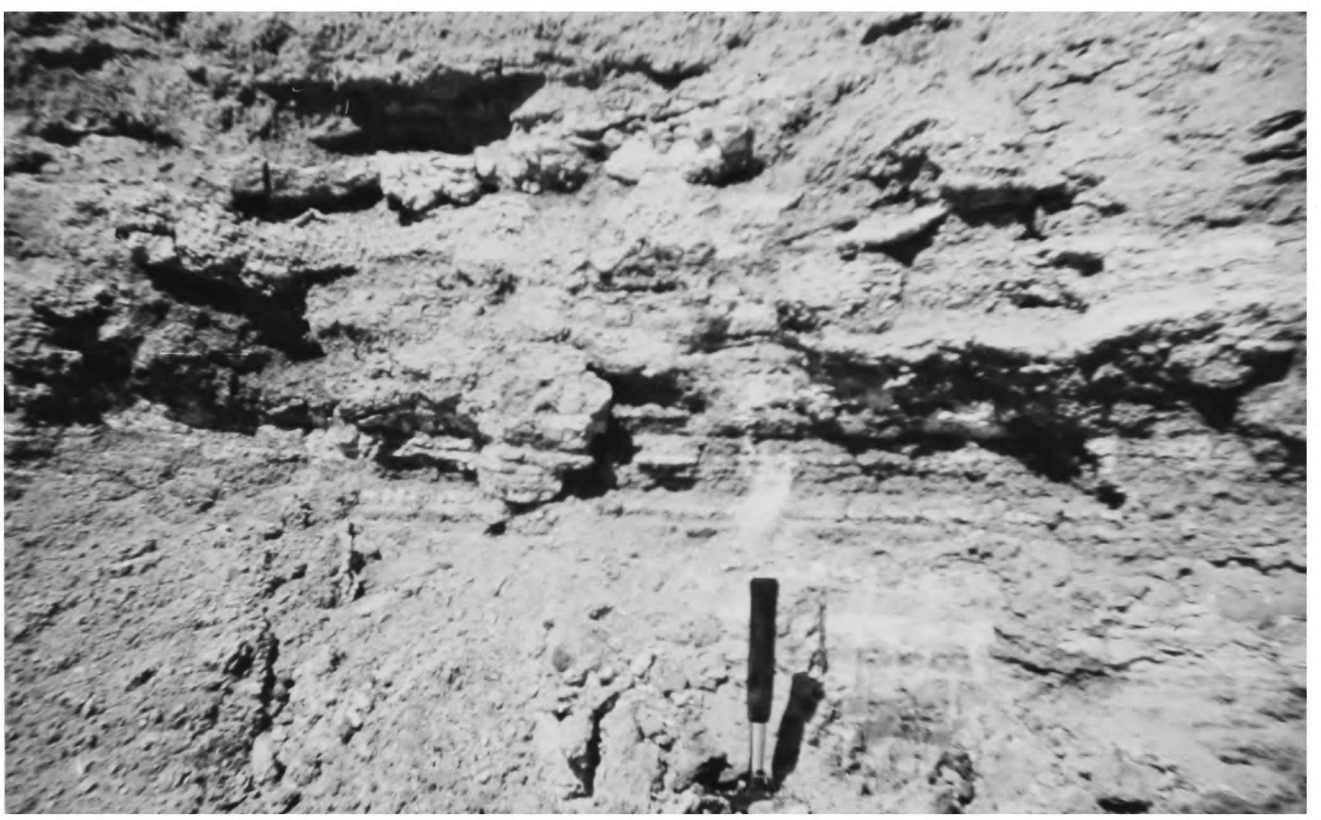

CM 54

Plate 5.5 (overleaf, above)

Horizontally bedded evaporitic gypsum at Chott el Rahim (CR) in southern Tunisia (see fig. 5.2 B and table 5.7 for chemical analyses).

Plate 5.6 (overleaf, below)

Bedded evaporitic gypsum from Chott el Rahim. The uppermost portion of the sample is on the left where the crystals are largest. They overlie a micro-crystalline band which grades downwards (left to right) into larger crystals (see plate 5.39 and table 5.7 for chemical analyses).





CR 1.2

Plate 5.7 (overleaf, above)

The White Lady Salt Pan (CC) in the Central Namib Desert seen from the Lagunenberg.

Plate 5.8 (overleaf, below)

Polygonal salt crust of the White Lady Pan.





in the area (Gevers, 1936).

North of Narabeb (NA) in the Central Namib dune field, an area of about 500 m^2 of the interdune valley floor is composed of gypsum (NA 1.1, 1.2) (see plate 5.9). The microcrystalline nature of the gypsum is in keeping with subaerial evaporites rather than subsurface croûte de nappe types. The deposit does not exhibit bedding, suggesting that it may represent a single evaporative phase. The material is less indurated than the other evaporitic gypsum deposits described above but it has undergone a small amount of doming. These 'blisters' are about 0.10 m to 0.20 m high and 0.50 m to 1.0 m in diameter and are often ruptured at the crest.

ii. Croûtes de nappe (subsurface crusts).

The term croûte de nappe refers specifically to a genetic type of gypsum crust formed at the surface of just above the water table. There are, however, two forms of subsurface crust which, for the purposes of this examination, will be classified as croûte de nappe. The first of these, the desert rose form, is characterized by large lenticular crystals up to 0.20 m in diameter which twin and interlock to form a continuous subsurface crust. The second form may be differentiated on several grounds. It does not appear to be associated with a permanent groundwater table, it is invariably composed of lenticular qypsum less than 1.0 mm in diameter and often is closely associated with various forms of surface crusts, including those on steep hill sides and hill crests. Hence the distinction between the forms is probably justified on genetic as well as structural grounds.

Desert rose croûtes de nappe have been described from a large number of areas (see chapter three). The most extensive examples are recorded in the coastal lagoons and sabkhas of Texas (Masson, 1955) and the Trucial Coast (Butler, 1969),

in Central Asia (Tolchel'nikov, 1962) and Namibia (Kaiser and Neumaier, 1932; Knetsch, 1937) around small pans and around the chotts of Algeria and Tunisia (Le Houérou, 1960; Bureau and Roederer, 1961; Coque, 1962, pp. 91 - 93; Pouget, 1968; Kulke, 1974; Vieillefon, 1976) (see tables 3.1 to 3.6). Though generally best developed in sandy, non-gypsiferous sediments, the large discoid crystals sometimes grow in silts and clays. Good examples are found in a small oued 50 km west of Gabès on the road to Kebili (GK) (see plate 5.10). Lenticular crystals up to 0.15 m in diameter are found in green and red Wealdian clays. The crystals have the same coloration as the host sediment owing to poikilitic inclusion of particles as the crystals grow displacively (see chapter one). The gypsum content may vary considerably, from 20% to 80% (Pouget, 1968), but it is usually between 40% and 60% (Brunsden, Doornkamp and Jones, 1976; Vieillefon, 1976), the remainder of the crust being composed of quartz sand in these cases. The crusts reach thickness of 5.0 m in the Algerian Souf (Kulke, 1974) and 3.50 m in Namibia (Knetsch, 1937) but usually they are 1.0 m to 2.0 m thick extending from about 1.0 m to 2.0 m or 3.0 m beneath the surface (Durand, 1949; Bellair, 1954). Pouget (1968) suggested that the zone of accumulation was restricted to aerated horizons of the host sediment. The origin of these crusts will be discussed below.

The desert rose croûte de nappe form was called deb-deb by Bureau and Roederer (1961) to distinguish it from a second form of hydromorphic crust which they gave the name terch. These Arabic names will not be used here since some confusion exists over their application. This second form differs from the desert rose type most strikingly in the size of the crystals, generally less than 1.0 mm in diameter compared with up to 200 mm in the case of the desert rose form. Besides this, however, there is less poikilitic inclusion of host material within the crystals resulting in more effective displacement of the host and thereby a greater gypsum content of the crust. Petrographic and chemical data presented below support this. Non-gypsiferous material makes up between 5% and

60% of the crust, the average gypsum content being about 60% to 65% (Vieillefon, 1976). Thicknesses of these crusts are less than those of many of the desert rose forms of subsurface crust being usually less than 1.0 m, though Dalloni (1953) reported examples up to 2.50 m thick in Algeria. The accumulations are generally beneath 0.25 m to 1.0 m of surface material which frequently has a markedly lower gypsum content than the underlying crust. This discontinuity is described by Jessup (1960 a.b., 1961 a.) and Ollier (1966) in Australia, Smith and Robertson (1962) in Iraq, Pias (1972) in Afghanistan and Gaddas (1971) and Fathi et al. (1972) in Tunisia and Egypt. Opinions differ greatly as to whether it is a result of different phases of sedimentation or preferential accumulation of gypsum in the host sediment. It is possible that the subsurface transition from 'flour' to 'seed' gypsum noted in the deposits of Australia (see chapter three) may be a result of processes similar to those producing this discontinuity. Bureau and Roederer (1961) pointed out that terch in Tunisia is generally overlain by 0.40 m to 0.50 m of unconsolidated gypsum. On the other hand break-down of 'seed' gypsum to 'flour' by partial dehydration at high temperatures may account for the transition.

The genetic distinction between the two forms of subsurface crust will be examined in chapter six. Suffice it to point out here that most workers have supported the groundwater model involving either direct precipitation near the surface of the water table, or from a fluctuating water table, or capillary rise and evaporation of water bringing about gypsum crystallization. Kulke (1974) felt that the two forms of subsurface crust could be accounted for by postulating different groundwater chemistries, gypsum saturated water evaporates to form desert roses and the less macrocrystalline form develops when the water is undersaturated with gypsum. Pouget (1968), however, suggested that gypsum roses crystallized from waters which were not highly saturated, the main condition for their development being long periods of chemical stability of the groundwater. A significant trait of the mesocrystalline form is the columnar

structures it occasionally exhibits. The fact that these are more akin to the structures developed in the surface crusts described below than to those of desert rose croûte de nappe suggests that there may be a genetic distinction between the forms of subsurface crust.

Nodular structures, ras-kelb, are sometimes associated with subsurface gypsum crusts. Pouget (1968) felt that nodular accretion was an important mechanism in croûte de nappe development, especially the desert rose form. Indeed, a true desert rose or gypsum rosette consists of a roughly spherical mass of interlocking lens shaped crystals (see plate 5.11) up to about 0.50 m in diameter (Masson, 1955). There is also evidence that nodular development occurs during the early stage of mesocrystalline croûte de nappe formation (Bureau and Roederer, 1961; Kulke, 1974)

Examples of subsurface gypsum crusts and accumulations composed of lenticular crystals visible with the naked eye are widespread in southern Tunisia. In the Central Namib, they are only rarely encountered perhaps owing to the scarcity of fluvial activity which produces the natural sections in which the Tunisian examples are exposed. It should be noted that in some cases the distinction between the desert rose and mesocrystalline forms is not clear cut. Plates 5.10, 5.11 and 5.12 show well developed lenticular gypsum crystals; 10 are from Wealdian clays, 11 from sands and 12 shows part of a subsurface crust composed of interlocking crystals. Many crusts, however, are made up of less well formed crystals and have frequently undergone modification through desiccation or solution after exposure.

The best examples of desert rose croûte de nappe in Tunisia come from low lying areas in the vicinity of the Chotts Djerid and el Fedjedj. In the area to the north, the Chotts are bordered by a chain of hills and there is only a narrow peripheral plain. To the south of Chott el Fedjedj there is a wider

plain but the most extensive low lying area lies to the east of Chott Djerid, south of the western end of Djebel Tebaga. This is the Nefzaoua. Examples of croûte de nappe come from sites at Djemma (DJ), Zaafrane (ZAA), El Faouar (EF), and Oued Tarfa (OT) (see fig. 4.1). At these locations gypsum crusts are found underlying sand dunes. 10 km west-north-west of Djemma the desert rose croûte de nappe has been exposed by the incision of a small oued into a partially fossilized dune, the underlying croûte de nappe and clays. While the surface of the dune is cemented by a microcrystalline gypsum crust, the underlying quartz sand has intermittent calcareous horizons which highlight aeolian cross-bedding. The desert rose croûte de nappe, 0.25 m to 0.30 m thick, lies beneath 2.70 m of these partially cemented dune sands. Beneath the croûte de nappe there is a clay horizon (see fig. 5.2C).

At Zaafrane, 30 km south-west of the site near Djemma, a small dune field extends south of the track from Douz to El Faouar and Sabria. The dunes are essentially barchanoid but tend to coalesce forming transverse ridges with crests trending north-west to south-east. The dunes are up to 5.0 m in height, the slipfaces reflecting dominantly north-easterly winds (see plate 5.13). In the depressions between the dune ridges, gypsum crusts are exposed at the surface. They are rarely more than 0.20 m thick and are composed of quartz sand cemented by lenticular gypsum crystals 2.0 mm to 5.0 mm in diameter, though altering to predominantly microcrystalline gypsum at the surface. Beneath the crust the damp sand is unconsolidated. The gypsum crust is superficially similar to those found at El Faouar, 20 km west-south-west of Zaafrane, and at Bir Zefazfa, in the Oued Tarfa (tarfa is the Arabic name for Tamarix spp., in this case Tamarix gallica L.) 30 km east of Douz. At closer examination, however, the examples from El Faouar and Bir Zefazfa are composed of microcrystalline material rather than larger lenticular crystals. An important difference in the location of the crusts is that, unlike Zaafrane and El Faouar which are on the shores of Chott Djerid, Oued Tarfa is a broad

shallow valley. The oued is up to 5.0 km wide and though the floor is no more than about 10 m below the surrounding land surface it is covered by a field of dunes 2.0 m to 3.5 m high (see fig. 5.3A). The chemical and grain-size characteristics of this sand are different from that making up the larger dunes at Djemma, Zaafrane and El Faouar (see tables 5.8 and 5.9 and chapter six).

Elsewhere in Tunisia, beyond the Nefzaoua, desert rose croûte de nappe where examined near Sebkhret Mekrecherma (SM), about 5.0 km west of Oudref, and near El M'Dou (EM), 10 km south-south-west of Gabès on the road to Matmata. At the first of these sites, a desert rose crust up to 0.25 m thick is found at the surface on a low terrace (2.0 m) of a oued draining the sebkhret. The crust lies upon 2.0 m of green clays (see fig. 5.2D), the lenticular crystals also being green in colour. About 200 m down the oued, the bed cuts through red Wealdian clays (sample SM 3, table 5.9) which underlie the green. The clays have a chicken-wire pattern of polygons from 0.20 m to 0.30 m in diameter. The polygon edges are composed of transparent crystalline gypsum which does not have well developed individual crystals. In alluvial sands overlying the clays, however, there are abundant selenite crystals up to 100 mm in length (see plate 5.14). These have well developed clinopinacoids (010), hemipyramid (111) and prism (110) and small hemiorthodomes (101). Twinning on the (100) plane is common, forming crucifixes. Though these crystals exhibit poikilitic inclusion of clay and sand, they contain less than the lenticular crystals developed on the green clays. The chemistry of the clays, described below may account for the different habits and the amount of sediment inclusion.

At El M'Dou the gypsum crusts are unusual because four different levels are represented (see fig. 5.3B). The uppermost crust is found on the main planation surface of the northern Djeffara. It differs from other surface crusts in the Djeffara, such as those at Kettana, being made up of lenticular crystals from 1.0 mm to 10 mm in diameter. Two more crusts composed of interlocking lenti-

Figure 5.2 (overleaf)

Location of samples taken from sections through evaporitic gypsum crusts and croûtes de nappe in southern Tunisia. Site locations are shown in fig. 4.1. Chemical analyses of the samples are presented as follows:

A. Sabkha el Melah (CM) table 5.7.

B. Chott el Rahim (CR) table 5.7.

C. Djemma (DJ) table 5.8.

D. Sebkhret Mekracherma (SM) table 5.9.

E. El M'Dou (EM) table 5.10.

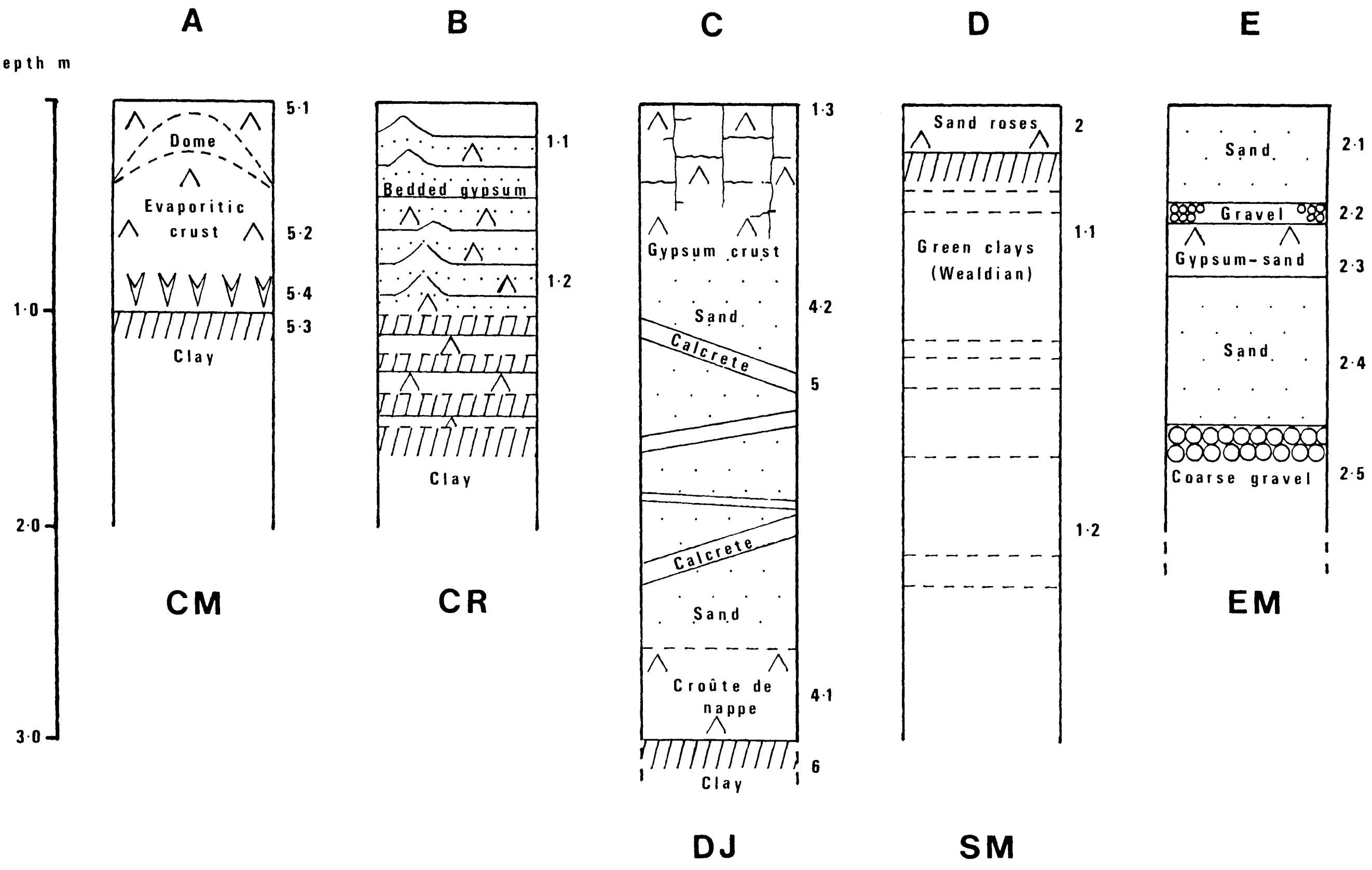
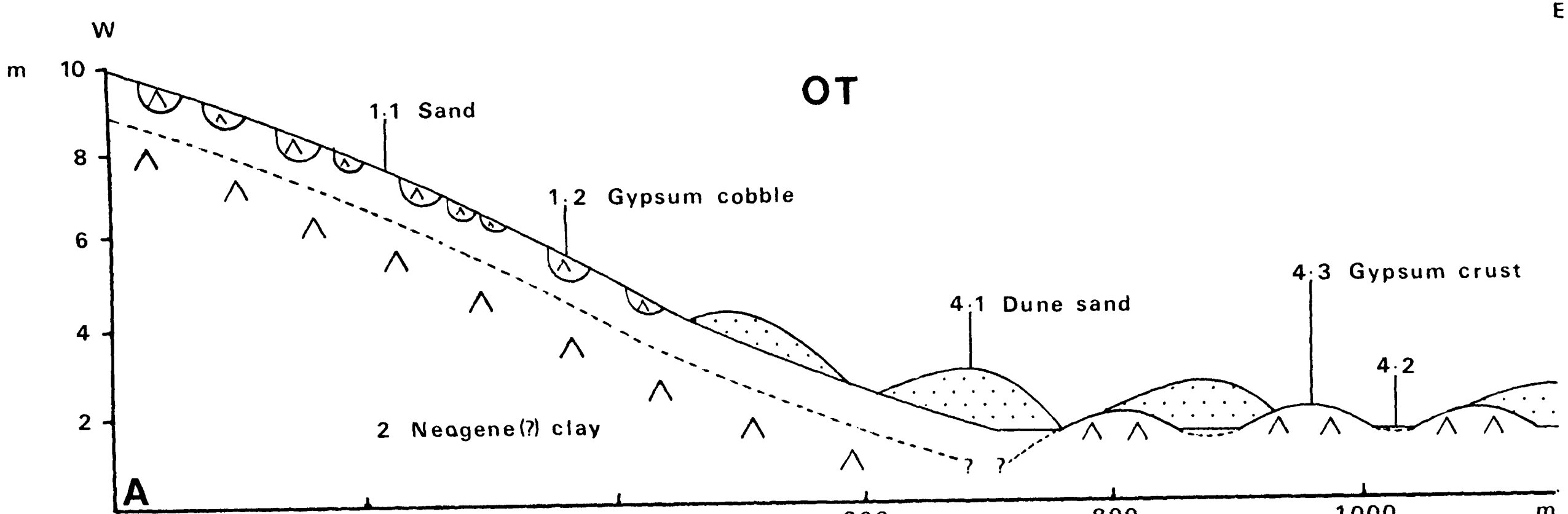

A
B
C
D
E
Depth m
1·0
2·0
3·0
Dome
Evaporitic crust
Clay
5·1
5·2
5·4
5·3
CM
Bedded gypsum
Clay
1·1
1·2
CR
Gypsum crust
Sand
Calcrete
Sand
Calcrete
Sand
Croûte de nappe
Clay
1·3
4·2
5
4·1
6
DJ
Sand roses
Green clays (Wealdian)
2
1·1
1·2
SM
Sand
Gravel
Gypsum-sand
Sand
Coarse gravel
2·1
2·2
2·3
2·4
2·5
EM

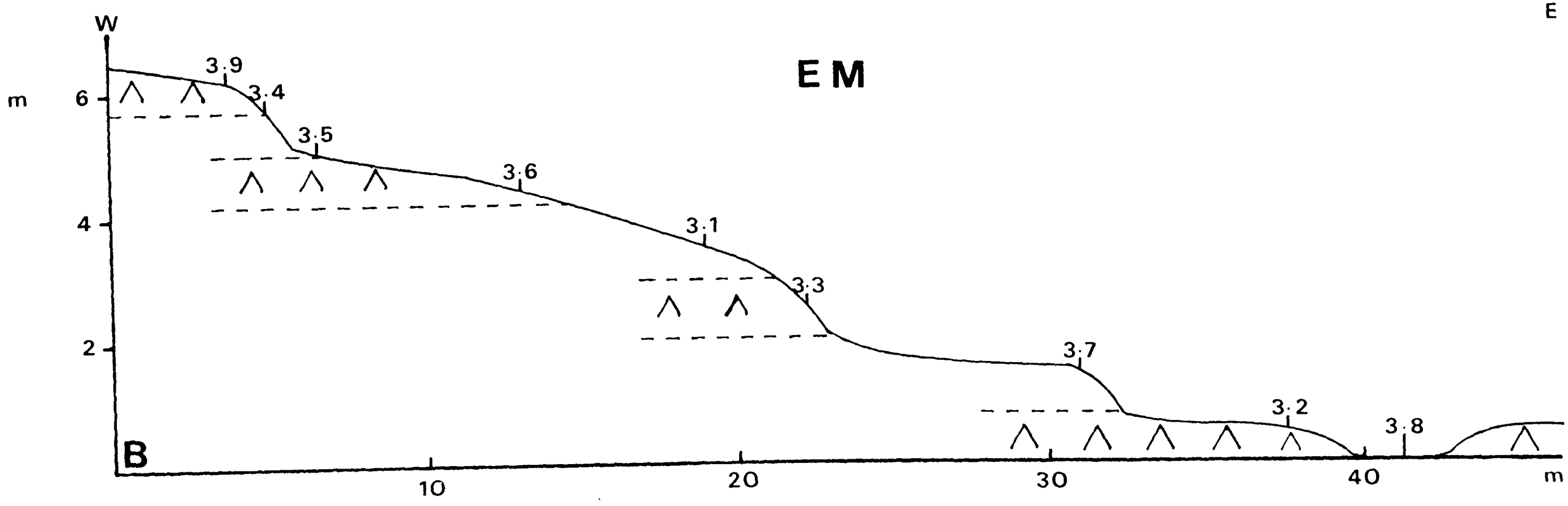
Figure 5.3 (overleaf)

Location of samples taken on two slope profiles in southern Tunisia.
Site locations are shown in fig. 4.1.
Chemical analyses of the samples are presented as follows:

A. Bir Zefazfa (OT) table 5.9.

B. El M'Dou (EM) table 5.11.


W
E
m
10
8
6
4
2
OT
1.1 Sand
1.2 Gypsum cobble
2 Neogene(?) clay
4.1 Dune sand
4.3 Gypsum crust
4.2
A
200
400
600
800
1000
m


W
E
m
6
4
2
EM
3.9
3.4
3.5
3.6
3.1
3.3
3.7
3.2
3.8
B
10
20
30
40
m

cular crystals are exposed in an oued but it is uncertain whether these represent different indurated horizons or parts of one thick crust exposed in notches in the sides of the oued. In the bed of the oued the crust is similar, in that it is of the desert rose form, but the crystals have more poikilitic inclusions of sand. Though Pouget (1968) regarded the gypsum crusts at El M'Dou as croûte de nappe, they are enigmatic owing to their thickness or multiplicity and their occurrence on a high land surface, up to 6.0 m above the river bed. Some oueds in the vicinity of El M'Dou have well developed calcretes cementing alluvial terrace deposits (samples EM 1.1 and 1.2) and some alluvial sands are cemented by microcrystalline gypsum (see fig. 5.2 E and table 5.10) While it would be logical to attribute the gypsum crusts on the main planation surface a greater age than the valley calcretes, it is unlikely that gypsum crusts would survive the wetter climate usually associated with calcrete formation.

In the Central Namib Desert, desert rose croûtes de nappe are rarely encountered. This may be because few streams have incised the land surface to provide natural sections. Crusts composed of lenticular gypsum crystals more than 1.0 mm in diameter have been identified at four locations. Between the Hamilton and Gungochoab Ranges (HA 3) there are several small surface exposures of lenticular gypsum in shallow stream channels. These exposures sometimes exhibit dome structures akin to those described north to Narabeb. Though doming has been described in evaporitic crusts, it is not a common feature of croûtes de nappe. Coque (1962, pp. 80 - 81) held that structurally similar features south of Chott Djerid represented nebkhas, small sand mounds associated with plant growth, fossilized by gypsum cementation. Rupturing of the domes lends support to the hypothesis that they may be the products of post-depositional deformation owing to volumetric changes or that the crusts are displacive accumulations. The Hamilton Range and Narabeb deposits are similar though the larger crystals and location in a river valley suggest a subsurface rather than subaerial origin for the Hamilton Range crusts. 5.0 km east of Rooibank, on the border of th

Namib Desert Park with the Walvis Bay enclave (PF 2.5, see fig. 5.4A), a gypsum croûte de nappe is exposed in a 200 m wide, 12.7 m deep stream channel. The crust is over 1.0 m thick and is composed of lenticular crystals 1.0 mm to 10 mm in diameter which enclose sand and clay particles poikilitically (see plate 5.15). This is the same type of crust as those in the Tumas River area which may reach 4.0 m in thickness (Goudie, 1972; Carlisle, 1978). Further inland a similar stream bed location provides another example of a croûte de nappe. About 13 km north-west of Gobabeb, near the confluence of the Aussinanis River and the Kuiseb (IMET) a crust is exposed beneath 0.10 m to 0.30 m of sands on the floor of the broad shallow stream channel. Several unusual features are exhibited. Firstly, although the surface of the crust is composed of lenticular gypsum crystals over 1.0 mm in diameter, at a depth of about 0.30 m it becomes micro-crystalline. Furthermore, columnar structures are in evidence forming polygonal patterns at the surface from 0.50 m to 2.0 m across. The polygon edges are composed of microcrystalline gypsum. Finally, in the lowest lying areas, the gypsum crust, about 0.30 m thick, overlies mica-schist bedrock which is brecciated and cemented by sodium chloride (see figs. 5.4C and D).

At Wlotzkas Baken (WB) about 32 km north of Swakopmund, a well developed gypsum crust is exposed at the surface over a large area less than 1.0 km inland, on the 17 m beach described by Rust and Wieneke (1972) and Wieneke and Rust (1973) near Swakopmund. The surface of the crust exhibits small domes of resilient, coarsely crystalline gypsum (WB 1.2) which protrude through a surface layer of sand and powdery gypsum usually less than 0.10 m thick (WB 1.1). The crust is over 1.0 m thick (the base could not be ascertained) and is composed of lenticular crystals through the 1.0 m excavated section (WB 1.3 from 1.50 m). It is possible that this crust represents a fossil sabkha deposit similar to the subsurface accumulations of lenticular gypsum that have been described along the Trucial Coast (Butler et al., 1965; Kinsman, 1969) If this were the case, it would argue a considerable age for the crust (see chapter four).

Figure 5.4 (overleaf)

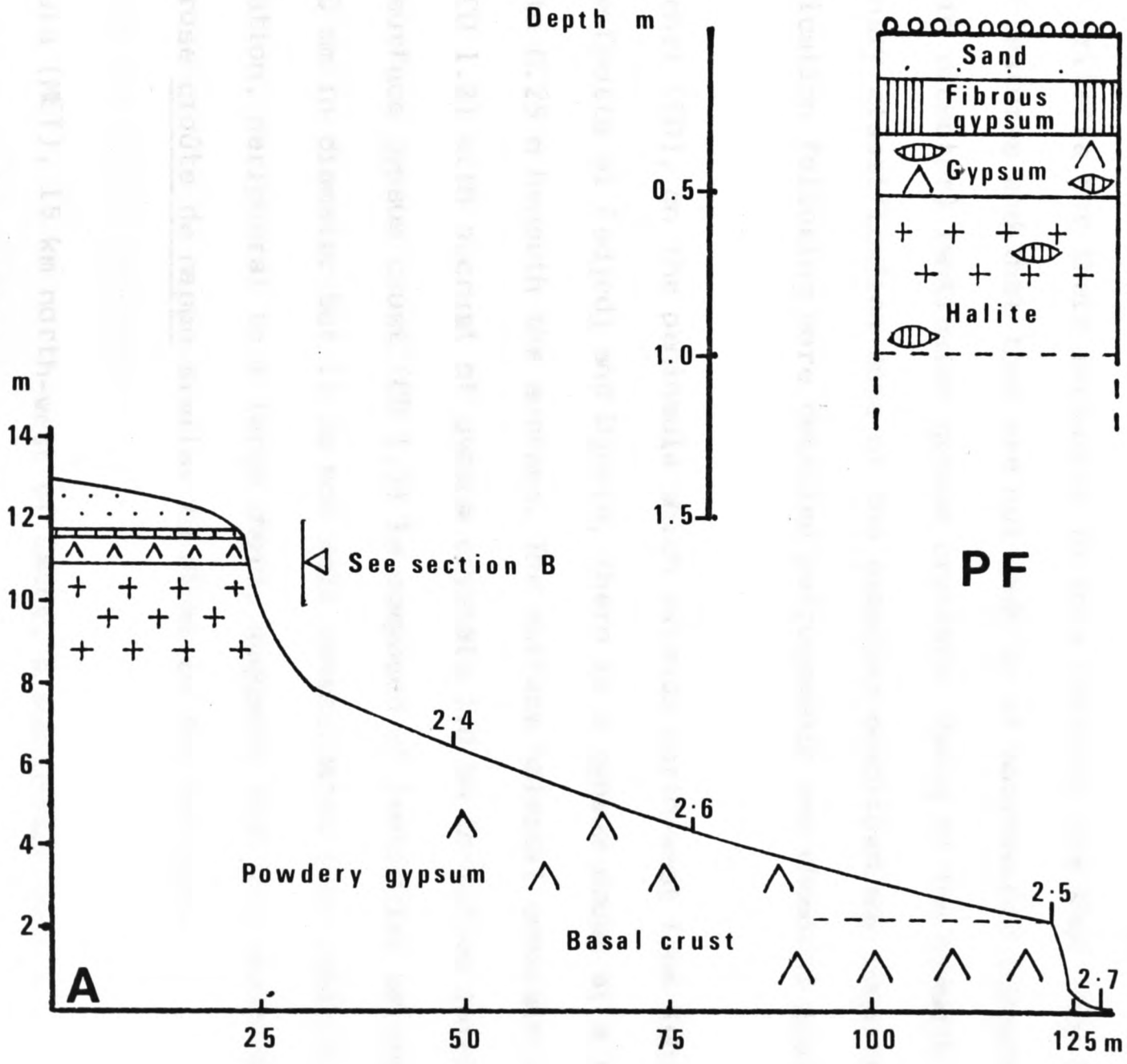
Location of samples taken from sections through gypsum and halite crusts in the Central Namib Desert. Site locations are shown in fig. 4.4.
Chemical analyses are presented as follows:

A. Near Rooibank (PF) tables 5.12 and 5.30.

B. Near Rooibank (PF) table 5.30.

C. Aussinanis (1MET) table 5.12.

D. Aussinanis (1MET) table 5.12.


m
14
12
10
8
6
4
2
See section B
2·4
2·6
2·5
2·7
Powdery gypsum
Basal crust
A
25
50
75
100
125 m

Depth m
0.5
1.0
1.5
B
Sand
Fibrous gypsum
2·1
2·2
Gypsum
3
2·3
Halite
C
Sand and gravel
Gypsum
3·3
3·1
3·2
Halite
brecciating
mica schist
D
2
1·1
1·2
1·3

PF

1MET

It should be noted here that small, well developed lenticular crystal aggregates are not uncommon in the Central Namib. Small desert roses are present in the Homeb silt deposits and are often found in association with surface gypsum crusts.

The second form of the croûte de nappe type of gypsum crust is less distinct from the surface crusts than the desert rose form. This is largely because of the less striking crystalline structure of the crusts as well as their close association with surface crusts. Here eight sites in Tunisia and three in the Central Namib will be described, their locations are shown on Figs. 4.1 and 4.4. Some crusts consist solely of mesocrystalline gypsum beneath 0.50 m to 1.0 m of surface sediment with a low gypsum content. Others, though structurally similar, are found beneath surface gypsum crusts or powdery gypsum horizons. The two criteria for their inclusion in this category are that they are sub-surface features and that they are not made up of macroscopic (greater than 1.0 mm in diameter) lenticular gypsum crystals. Owing to the breadth of this preliminary classification some of the examples described may require re-classification following more detailed petrographic and chemical analyses.

Near Bechri (CD), on the peninsula which extends north-west from Kebili separating the Chotts el Fedjedj and Djerid, there is a gypsum crust at a depth of 0.20 m to 0.25 m beneath the surface. The surface material consists of quartz sands (CD 1.2) with a crust of gypsum crystals 1.0 mm to 2.0 mm thick (CD 1.1). The subsurface gypsum crust (CD 1.3) is composed of lenticular crystals less than 1.0 mm in diameter but it is not well consolidated (see table 5.13). The location, peripheral to a large chott, suggests that the crust may be a desert rose croûte de nappe similar to other in the Nefzaoua.

At Metouia (MET), 15 km north-west of Gabès, a subsurface gypsum crust has been excavated to provide building material. The excavated pit reaches nearly 2.0 m

in depth and covers about 2,500 m^2. In some places the crust is covered by up to 0.50 m of silty sand. In others, a surface gypsum crust about 0.25 m thick overlies 0.25 m of silty sand which in turn is underlain by the subsurface gypsum crust which is 1.0 m to 2.0 m thick (see fig. 5.5A and B). The subsurface crust is composed of lenticular gypsum crystals, generally less than 1.0 mm in diameter, while the surface crust is noticeably more microcrystalline. The most striking structural characteristic of the subsurface crust is a columnar structure, similar to that described by Meckelein (1959) from the Central Sahara and by Tucker (1978) in Iraq. The columns are roughly hexagonal, 0.25 m to 0.75 m in diameter, and extend down through the whole of the crust (see plate 5.16). Columns, and the polygonal patterns they form at the surface, are often associated with surface crusts and mesocrystalline subsurface crusts, they are not developed on desert rose croûtes de nappe. The columns are characterized by hard microcrystalline outer walls, 10 mm to 30 mm thick, which frequently show lamination parallel to the column sides (see plate 5.17). The gypsum within the column is made up of mesocrystalline (less than 1.0 mm) discoids which are not well cemented when first exposed. This core material is slightly damp and consolidates upon drying. Small piles of excavated gypsum may be seen throughout the field area since when dried it is utilized as building stone.

During 1976 and 1977 a 1.50 m to 2.0 m deep trench was excavated between Medenine and Zarzis (MZ) for a water pipeline. White gypsum crusts 1.50 m to 2.0 m thick were exposed in the trench. The crusts lie beneath 0.10 m to 0.50 m of gypsiferous, orange silty sand and in section have an undulating surface profile with a wave length of 5.0 m to 10 m and an amplitude of 0.25 m to 0.50 m. The pattern is consistant over hills and within depressions. The crust is composed of mesocrystalline discoids, less than 1.0 mm in diameter. Columnar structures are not in evidence (see fig. 5.5C).

In the Oued el Melah (OMG), 25 km south-west of Gafsa on the road to Metlaoui,

thick subsurface gypsum crusts are exposed in sections forming the oued walls at a number of sites. Two of these have been examined. At both the crusts are 1.50 m thick. The first, in a cutting where the road fords the oued, lies beneath 1.50 m of orange sand and gravel and the second below 0.50 m of similar surface material. At the first location the hardness of the base of the gypsum crusts was determined using a Schmidt hammer (P type). Ten readings at each of three points gave the mean values of 10.0; 14.1; and 15.7, indicating a tendency toward hard and soft vertical zonation. Over a horizontal distance of about six metres the widths, in metres, of hard and soft gypsum zones were as follows.

soft	hard	soft	hard	soft	hard
0.457	0.381	0.584	0.457	0.127	0.660
soft	hard	soft	hard	soft	hard
1.016	0.457	0.610	0.330	0.305	0.711

Means; soft, 0.517; hard 0.499.

The pattern results from the alternate exposure of column walls and cores, hard and soft gypsum, in the man-made road cutting. At the second location, the section is a natural one forming the oued wall and columns have been removed completely leaving a geometric columnar face. The Schmidt hammer hardness of this exposure varies from 10 to 30, averaging 18.9.

Subsurface crusts at the next three sites differ from the aforementioned examples in that they are beneath surface crusts. Several other examples of subsurface crusts beneath surface crusts will be mentioned below but those described here provide the best subsurface sequences. About 5.0 km south of Limaguess (L) on the northern side of the western end of the Djebel Tebaga, gypsum crusts mantle the glacis and outlier's of Turonian sandstone and limestone (see plate 5.18). Surface crusts lying directly upon bedrock are found

on the crests of these ridges. Where the glacis are covered with alluvial fan deposits, oued sections frequently reveal a subsurface gypsum crust up to 1.0 m thick beneath 0.50 m to 1.0 m of superficial material (see fig. 5.5D). The samples were collected from a section in the middle of the three glacis represented at Limaguess. It should be noted that the crusts on the upper glacis were identical structurally and, as Page (1972) pointed out in other areas around the Chotts, the marked topographic discontinuity between adjacent glacis is not parallelled by a break in the gypsum crust. Indeed, a crust often mantles or underlies the break in slope. The crust is composed of lenticular crystals less than 1.0 mm in diameter and exhibits a columnar structure with hard, microcrystalline column walls. Both surface and subsurface crusts are white in colour. A very similar sequence is found at Oasis de Lesseps (OL) about 5.0 km north of Oudref. This is the point at which Ferdinand de Lesseps began excavating the canal which would allow Mediterranean water into those Chotts below sea level. A surface gypsum crust 0.50 m thick, with a weak columnar structure, overlies 0.50 m of powdery gypsum which also exhibits columns, their edges being defined by plant roots.

At a depth of 1.0 m there is a sharp discontinuity at the top of a hard white gypsum crust with marked columniation (see plate 5.19). The subsurface crust is about 1.0 m to 1.25 m thick and lies upon about 2.50 m of sands and silts. An auger hole revealed unpenetrable material at 2.56 m below the bed of the 2.0 m deep oued (see fig. 5.5E). Groundwater was not encountered. The third site with essentially the same sequence is also near Oudref (OU), about 5.0 km to the south-west, at the eastern end of Chott el Fedjedj. A columnar crust, 0.10 m to 0.20 m thick, covers the surface or lies beneath a discontinuous layer of sand of similar thickness. At a depth of about 1.0 m there is a second crust. This is separated from the surface crust by an horizon of powdery gypsiferous material exhibiting weak columniation (see plate 5.20). The subsurface crust differs from that at Oasis de Lesseps in not having a columnar structure

Figure 5.5 (overleaf)

Location of samples taken from sections through surface and subsurface gypsum crusts in southern Tunisia. Site locations are shown in fig. 4.1.
Chemical analyses are presented as follows:

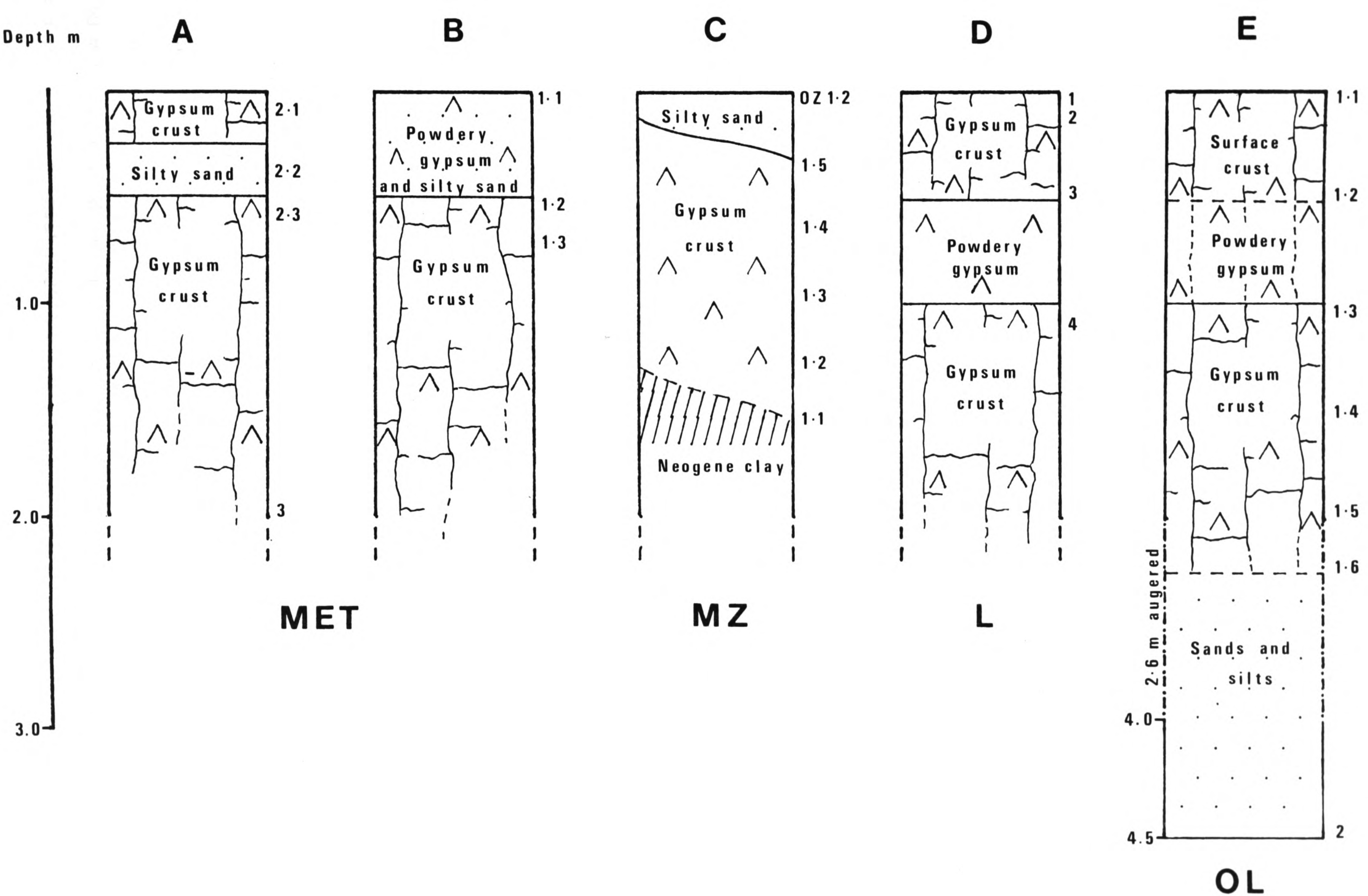
A. Metouia (MET) table 5.13.

B. Metouia (MET) table 5.13.

C. Medenine to Zarzis (MZ) table 5.14.

D. Limaguess (L) table 5.14.

E. Oasis de Lesseps (OL) table 5.15.


Depth m
A
B
C
D
E
Gypsum crust
2·1
Silty sand
2·2
2·3
Gypsum crust
1.0
2.0
3
3.0
MET
Powdery gypsum and silty sand
1·1
1·2
1·3
Gypsum crust
MZ
Silty sand
OZ 1·2
1·5
Gypsum crust
1·4
1·3
1·2
1·1
Neogene clay
L
Gypsum crust
1
2
3
Powdery gypsum
4
Gypsum crust
Surface crust
1·1
1·2
Powdery gypsum
1·3
Gypsum crust
1·4
1·5
1·6
2·6 m augered
Sands and silts
4.0
4.5
2
OL

and being composed of lenticular crystals at depth though it is largely microcrystalline in its upper portion. Small nodular structures with a diameter of 5.0 mm to 10 mm, composed of white lenticular gypsum crystals up to 1.0 mm in diameter are found within the crust. It is about 0.50 m thick and lies upon uncemented sands (see fig. 5.6A).

At La Skhirra, on the Golfe de Gabès about 40 km north of Gabès, a subsurface gypsum crust is found beneath a surface sand deposit. The crust is developed on Wealdian (?) gypsiferous clays. It is columnar, 0.50 m to 0.75 m thick but, unlike examples from other locations, it is composed of microcrystalline material though toward its base lenticular crystals are abundant. The red clays which underlie the crust have horizons of translucent selenite 10 mm to 20 mm thick. At a depth of 2.50 m the clay is heavily indurated forming an orange subsurface crust at least 1.0 m thick. This possesses neither well developed gypsum crystals nor columniation (see fig. 5.6B). Page (1972) reported examples of two subsurface crusts in this area. The gypsum crust described here represents the lower of these.

The final Tunisian example of subsurface mesocrystalline gypsum crust comes from the northern Djeffara, 5.0 km south-east of Kettana (K). The site is about 2.0 km from the coast, on the north bank of a small oued which has incised the Mio-Pliocene sands and clays which now form a series of low plateaux which late Pleistocene and Holocene coastal sediments abut. The land surface above the oued incision is scattered with cobbles of calcrete and gypsum which lie upon and within the surface of an accumulation of orange powdery gypsum up to 0.50 m thick. This material exhibits columniation. The column walls form white upstanding ridges, 20 mm to 50 mm high and the same width, creating a polygonal pattern at the surface (see plate 5.21). The polygons are roughly hexagonal, 0.50 m to 1.0 m in diameter. Beneath the unconsolidated surface horizon there is a white gypsum crust 0.25 m to 0.50 m thick. This is composed of lenticular

Figure 5.6 (overleaf)

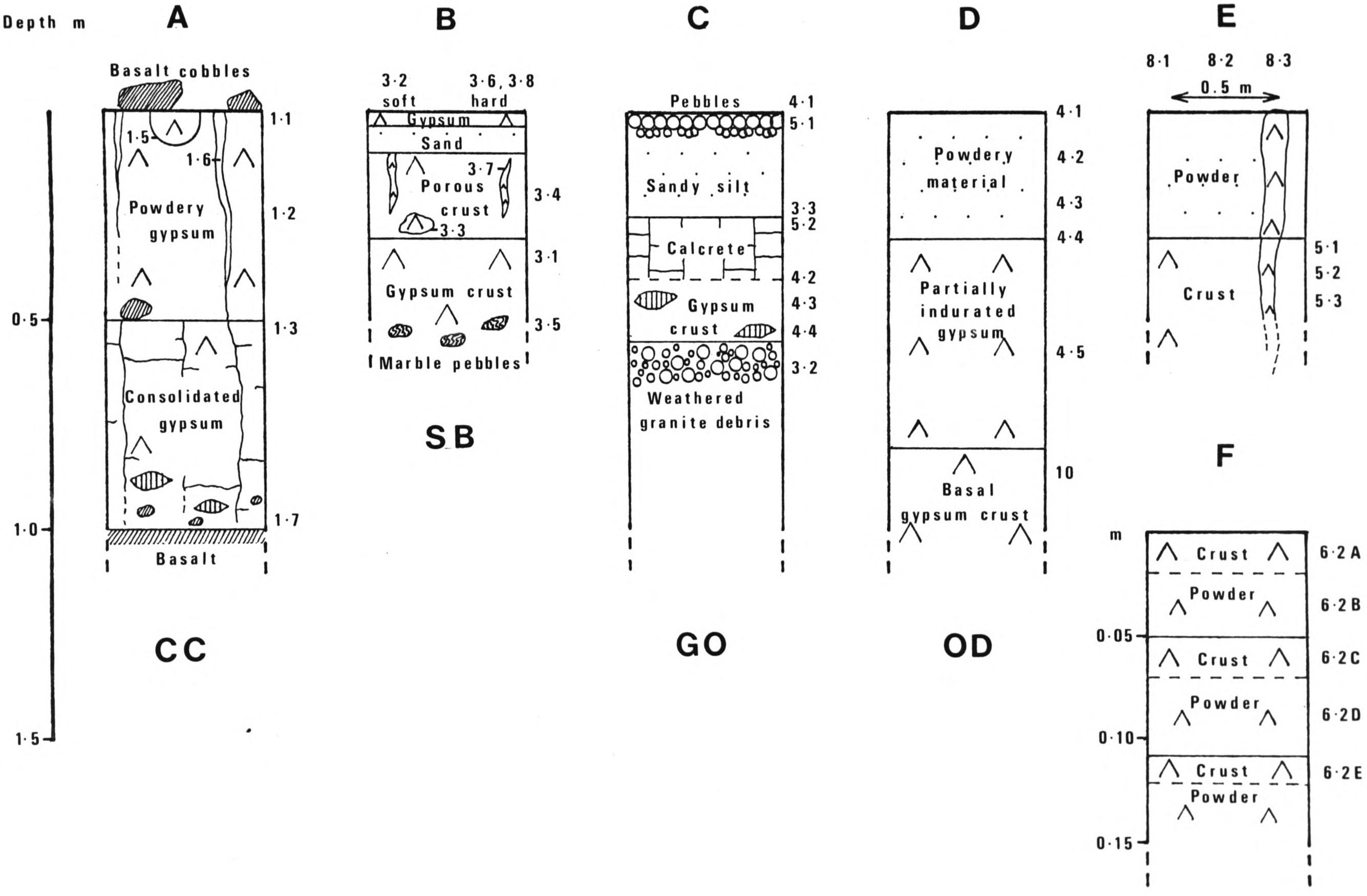
Location of samples taken from sections through gypsum crusts in southern Tunisia. Site locations are shown in fig. 4.1.
Chemical analyses are presented as follows:

A. Oudref (OU) table 5.16.

B. La Skhirra (LS) table 5.17.

C. Kettana (K) table 5.18.

D. Kettana (K) table 5.18.


Depth m
A
Basalt cobbles
1·1
1·5
1·6
Powdery gypsum
1·2
0·5
1·3
Consolidated gypsum
1·7
1·0
Basalt
1·5
CC
B
3·2 soft
3·6, 3·8 hard
Gypsum
Sand
3·7
Porous crust
3·4
3·3
3·1
Gypsum crust
3·5
Marble pebbles
SB
C
Pebbles
4·1
5·1
Sandy silt
3·3
5·2
Calcrete
4·2
Gypsum crust
4·3
4·4
3·2
Weathered granite debris
GO
D
4·1
Powdery material
4·2
4·3
4·4
Partially indurated gypsum
4·5
10
Basal gypsum crust
OD
E
8·1
8·2
8·3
0·5 m
Powder
Crust
5·1
5·2
5·3
F
m
Crust
6·2A
Powder
6·2B
0·05
Crust
6·2C
Powder
6·2D
0·10
Crust
6·2E
Powder
0·15

gypsum crystals usually less than 1.0 mm in diameter and exhibits weak columniation emphasized by plant roots. The crust contains calcrete pebbles towards its base (see fig. 5.6C).

A profile from the upper surface to the bed of the oued (see fig. 5.6D) reveals an exposure of the subsurface crust down-slope. The exposed crust has a better developed columnar structure than the buried example. The oued floor consists of a hard, porous calcrete.

In the Central Namib Desert, three sites provide good examples of mesocrystalline subsurface gypsum crusts. The first is located in the north of the study area near Cape Cross (CC) on the western side of the Lagunen Range where it rises from the White Lady Salt Pan, just above sea level, up to 173 m at its crest (see plate 5.22). The white gypsum crusts cover most of the land surface, being found on slopes of up to 30°. Though they are overlain by basalt boulders and wind blown quartz sand (see plate 5.23), gypsum cobbles and polygonal patterns produced by microcrystalline column edges are exposed locally. Beneath the non-gypsiferous surface materials there are 0.50 m of poorly consolidated gypsum containing hard column wall structures. This material differs from the corresponding powdery deposit commonly found in Tunisia, in that it consists of lenticular crystals up to 1.0 mm in diameter admixed with microcrystalline gypsum. Underlying this is a hard white gypsum crust up to 0.50 m thick. It rests upon basalt bedrock, fragments of which are included in the basal crust. Columnar structures are present. A significant feature of this subsurface crust is the presence of fibrous gypsum textures especially near the base. The fibres are orientated vertically forming lens-shaped structures up to 30 mm thick and 50 mm to 100 mm in diameter (see fig. 5.7A).

The second of the sites in the Central Namib Desert where subsurface mesocrystalline crusts are represented, is on the western slopes of the marble

ridge west of Swartbankberg (SB) on the north bank of the Kuiseb halfway between Walvis Bay and Gobabeb. The ridge forms the southern end of an outcrop of inclined Palaeozoic marble beds with later doleritic intrusions. The western slopes are markedly convex with an average inclination of 3.7°. The lower portions are covered with fan deposits which are gypsiferous. A surface gypsum crust composed of loosely consolidated, powdery material or a hard microcrystalline crust extends from the surface to a depth of 0.05 m. This may be overlain with marble fragments even on the ridge crest. Beneath the crust, a discontinuous sandy horizon reaches about 0.10 m where a loosely cemented accumulation of admixed powdery and crystalline gypsum is found. This contains cobbles of hard gypsum composed of lenticular crystals less than 1.0 mm in diameter and gypsified rootlets made up of similar material. At a depth of 0.30 m to 0.40 m a continuous, consolidated gypsum crust is encountered. The crust, consisting of discoids of gypsum less than 1.0 mm in diameter, has inclusions of marble pebbles. Columnar structures are not apparent (see fig. 5.7B).

North of the Desert Research Station at Gobabeb (GO), the landscape consists of undulating gravel-covered plains dissected by shallow stream channels and dotted with granite inselbergs and less pronounced feldspathic outcrops. An excavation near Gobabeb revealed 0.25 m of sandy silt beneath a surface lag of granite and calcrete pebbles. At a depth of about 0.25 m there is a 0.10 m to 0.20 m thick calcrete horizon which overlies a similar thickness of gypsum which cements fragments of granite towards its base at 0.50 m to 0.70 m. The gypsum crust consists of lenticular crystals, usually less than 1.0 mm in diameter, and lenses of fibrous gypsum. These lenses are 0.10 m to 0.20 m in diameter and up to 50 mm thick with the fibres orientated vertically. Though the calcrete component of the crust shows vertical cracking and a blocky structure, evidence of columniation of the gypsum crust is absent (see fig. 5.7C). Polygonal patterns 2.0 m to 5.0 m in diameter defined by vegetation growth at the boundaries are present in some areas (Watson, 1981). Their origin will be discussed

Figure 5.7 (overleaf)

Location of samples taken from sections through surface and subsurface gypsum crusts in southern Tunisia and the Central Namib Desert. Site locations for A, B and C are shown in fig. 4.4 and for D, E and F in fig. 4.1. Chemical analyses are presented as follows:

A. Lagunenberg (CC) table 5.19.

B. Swartbankberg (SB) table 5.20.

C. Gobabeb (GO) table 5.21.

D. Oued Demna (OD) table 5.22.

E. Oued Demna (OD) table 5.23.

F. Oued Demna (OD) table 5.23.

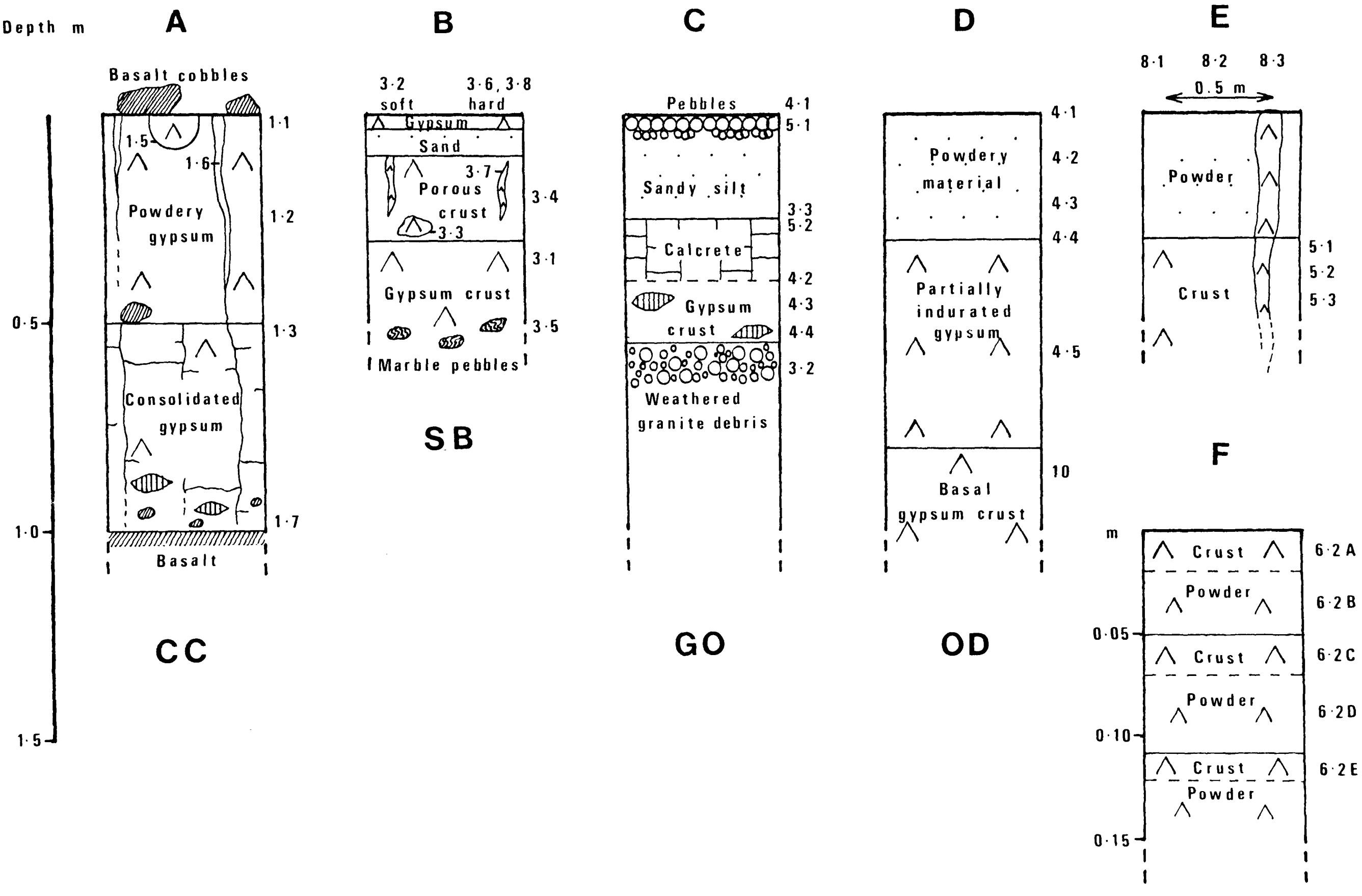

Depth m
A
Basalt cobbles
1·5
1·6
Powdery gypsum
Consolidated gypsum
Basalt
1·1
1·2
1·3
1·7
0·5
1·0
1·5
CC
B
3·2 soft
3·6, 3·8 hard
Gypsum
Sand
3·7
Porous crust
3·3
Gypsum crust
Marble pebbles
3·4
3·1
3·5
SB
C
Pebbles
4·1
5·1
Sandy silt
3·3
5·2
Calcrete
4·2
Gypsum crust
4·3
4·4
3·2
Weathered granite debris
GO
D
4·1
Powdery material
4·2
4·3
4·4
Partially indurated gypsum
4·5
Basal gypsum crust
10
OD
E
8·1
8·2
8·3
0·5 m
Powder
Crust
5·1
5·2
5·3
F
m
Crust
6·2A
Powder
6·2B
0·05
Crust
6·2C
Powder
6·2D
0·10
Crust
6·2E
Powder
0·15

Plate 5.9 (overleaf, above)

Evaporitic gypsum deposit in an interdune valley near Narabeb (NA) in the Dune Namib (see plate 5.40 and table 5.7 for chemical analysis).

Plate 5.10 (overleaf, below)

Lenticular gypsum crystals from red Wealdian clays south of Chott el Fedjedj (GK) in southern Tunisia.




GK 1.1

Plate 5.11 (overleaf, above)

Small desert roses from the northern shore of Chott Djerid near Tozeur in southern Tunisia.

Plate 5.12 (overleaf, below)

Mass of lenticular gypsum crystals from a desert rose croûte de nappe to the east of Gafsa in southern Tunisia.


CHOTT DJERID
(TOZEUR)


G 1

Plate 5.13 (overleaf, above)

Gypsum dunes near Zaafrane (ZAA) in southern Tunisia (see table 5.8 for chemical analyses).

Plate 5.14 (overleaf, below)

Prismatic selenite crystals from Wealdian(?) clays near Sebkhret Mekracherma (SM) in southern Tunisia.




SM 3.2

Plate 5.15 (overleaf, above)

Croûte de nappe exposed in a dry stream channel near Rooibank (PF) in the Central Namib Desert (see fig. 5.4A and table 5.12 for chemical analyses).

Plate 5.16 (overleaf, below)

Columnar structures developed in a subsurface mesocrystalline gypsum crust at Metouia (MET) in southern Tunisia (see plate 5.44, figs. 5.5A and 5.5B and table 5.13 for chemical analyses).





Plate 5.17 (overleaf, above)

Column edge from Limaguess, (L) south of Chott el Fedjedj, showing characteristic banding resulting from solution and reprecipitation of gypsum under the influence of water percolating down the fissures between the columns. The top of the specimen is to the left (see plates 5.63 and 5.64, fig. 5.5D and table 5.14 for chemical analyses).

Plate 5.18 (overleaf, below)

Outliers of southward dipping Turonian sandstone south of Chott el Fedjedj near Limaguess (L). Note the large cobbles of sandstone overlying a surface gypsum crust in the foreground.


L 4



Plate 5.19 (overleaf, above)

Columnar subsurface gypsum crust exposed in a oued at Oasis de Lesseps (OL) in southern Tunisia (see plate 5.46, fig. 5.5E and table 5.15 for chemical analyses).

Plate 5.20 (overleaf, below)

Columnar structures in a powdery gypsum crust at Oudref (OU) in southern Tunisia (see fig. 5.6A and table 5.16 for chemical analyses).



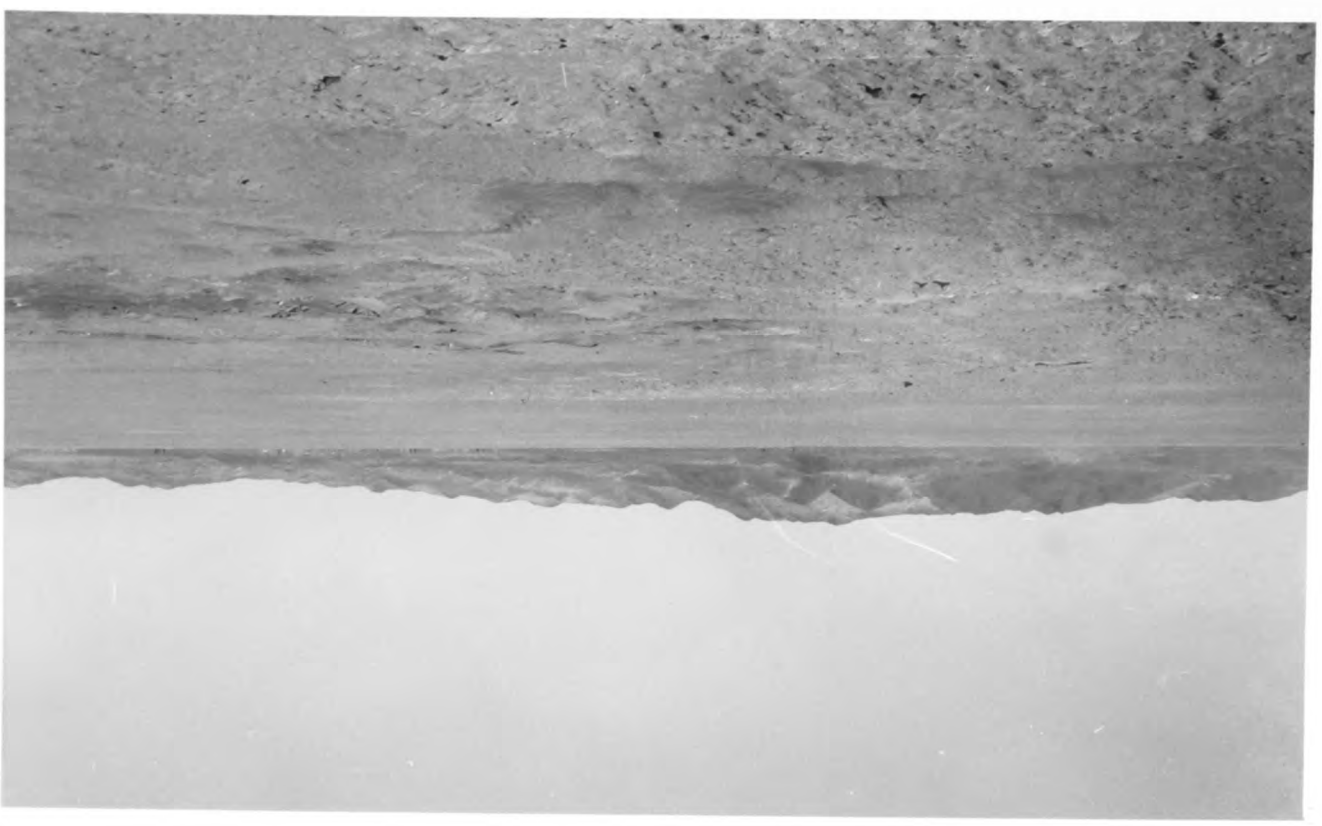


Plate 5.21 (overleaf, above)

Hexagonal polygons formed by exposure of consolidated gypsum at the boundaries of columns in a powdery gypsum crust near Kettana (K) in southern Tunisia (see figs. 5.6C and 5.6D and table 5.18 for chemical analyses).

Plate 5.22 (overleaf, below)

The Lagunen Range (CC) in the Central Namib Desert seen from the White Lady Salt Pan.





below. 6.0 km east of Gobabeb, a gypsum crust is present at a depth of about 1.0 m (GO 6).

Several structural features justify the subdivision of subsurface crusts into the desert rose form and the mesocrystalline form. The topographic location of the two forms is often distinct since the desert rose croûte de nappe are almost invariably found close to the top of the water table in low-lying areas while the mesocrystalline crusts are found on hill sides and crests as well as in areas of low relief. Crystal size is an important structural distinction with desert rose forms usually being built up of interlocking discoids over 1.0 mm and up to 0.20 m in diameter. The mesocrystalline form only rarely exhibits crystals over 1.0 mm in diameter though they are made up of lenticular crystals visible with the naked eye, generally larger than about 0.20 mm. Also of great significance is the frequent presence of columnar structures within the meso-crystalline form of subsurface crust. Their importance to the understanding of the origin of the crusts will be explained below. Despite these differences, however, the two crust forms are justifiably classified as one type on the basis of their subsurface location and genesis.

iii. Surface gypsum crusts and accumulations

The third main type of gypsum crusts comprises those which are found at the surface. The term encroûtement has been used by a number of French workers (Le Houérou, 1960; Bureau and Roederer, 1961; Coque, 1962, p. 82; Vieillefon, 1976) but some confusion exists as to whether it may be applied to all sur-face accumulation, regardless of the degree of induration, or solely to powdery surface accumulations. For the purposes of this examination three forms of surface gypsum crust will be differentiated.

Firstly, powdery accumulations which are often orange coloured and have gypsum

contents in the region of 40% to 70% and up to 90% (Durand, 1963), the remainder consisting of calcium carbonate and quartz grains (Gaddas, 1971). The material may be found in a wide variety of topographic locations but frequently it overlies subsurface mesocrystalline gypsum crusts as at Oudref and Metouia, described above. The deposits are usually 0.50 m to 1.5 m thick (Page, 1972).

The second form of surface accumulation constitutes the most widely reported examples of gypsum crusts. They are usually 0.50 m to 0.20 m thick and are characterized by their columnar structure visible as a polygonal pattern on the surface, an hexagonal net being defined by the column edges. This columniation is identical to that reported in subsurface mesocrystalline crusts. Columns extend through the crust and have diameters of 0.25 m to 1.0 m. The white, or greyish crust is frequently composed solely of microcrystalline gypsum though, as in the case of the subsurface crusts, the core of the column is less indurated than the walls and retains some moisture. The gypsum content reaches 75% to 95%, the remainder being made up of quartz grains and small amounts of calcium carbonate (Coque, 1962). Page (1972) reported examples with between 20% and 98% gypsum. As in the case of the powdery accumulations, surface crusts are often found to overlie a subsurface mesocrystalline crust at a depth of 0.50 m to 1.0 m.

This is also true of the third form of surface gypsum crust. This consists of cobbles of white, or greyish, microcrystalline gypsum from 30 mm to 0.50 m in diameter, studding a powdery accumulation. The cobbles are flat on their upper surface, which is flush with the land surface, roughly circular when viewed from above and domed, conical or flat on their undersides (see plates 5.24 and 5.25). Often only the outer 10 mm to 30 mm of the larger cobbles are indurated microcrystalline gypsum, the cores being softer and damp though usually also microcrystalline. They are probably an intermediate stage between the powdery form and the indurated columnar crust.

Surface crusts have been reported from all the main areas of gypsum crusts (see tables 3.1 to 3.6). Examples of the columnar form, sometimes called prismatic or polygonal, are found in north Africa (Coque, 1955 a.b., 1962; Le Houérou, 1960; Bureau and Roederer, 1961; Page, 1972; Vieillefon, 1976, from Tunisia and Kulke, 1974; Horta, 1979, from Algeria) the Central Sahara (Meckelein, 1959), Egypt (Butzer and Hansen, 1968, pp. 105 - 107) and the Middle East (Gabriel, 1964, from Iran; Tucker, 1978, from Iraq and Brunsden, Doornkamp and Jones, 1976, from Bahrain). They have also been mentioned in Central Asia (Fersmann and Wlodawetz, 1926; Dregne, 1976), the United States (Hunt et al., 1966) and the Central Namib Desert (Goudie, 1972). In contrast with these numerous reports of columnar surface crusts, there are relatively few descriptions of powdery accumulations. Le Houérou (1960), Bureau and Roederer (1961) and Vieillefon (1976) refered to encroûtement pulvérulents in Tunisia and Durand (1963) to them in Algeria. Examples are also reported from California (Hess, 1909) and South and Western Australia (Jack, 1921; De la Hunty and Low, 1958). The intermediate form of surface crust, that consisting of indurated cobbles set in a powdery surface horizon, has received little attention in the literature. It is widespread in southern Tunisia and not uncommon in the Central Namib Desert, yet the only features described elsewhere which bear any resemblance to the aforementioned cobbles are 'gypsum beards' on the undersides of pebbles (Rozanov, 1961; Pias, 1972; Petrov, 1976).

Though the various forms of surface crust are markedly different structurally, it will be shown that a genetic relationship exists. The classification of the forms as one type is justifiable at the outset of the analysis on the grounds that the intermediate form of surface crust suggests an interrelationship. Furthermore, the similarities in the distribution of the forms supports this approach. Not only are the various forms found in close association but they may be differentiated from the other types of gypsum crust on the basis of their distributional characteristics. While some mesocrystalline subsurface crusts

underlie terrain regardless of the topography, just as surface crusts can mantle the landscape, the nature of the underlying sediments show a greater variability in the case of the surface accumulations. Subsurface crusts develop only within permeable sediments, though occasionally they may overlie clays or other types of bedrock as at La Skhirra in Tunisia and near Cape Cross in the Namib Desert (see above). In contrast, columnar surface crusts are frequently developed directly upon bedrock without any evidence of a permeable host sediment (Coque, 1962, p. 82; Page, 1972) The porosity of the surface crust itself shows considerable variability owing to the alternation from dense microcrystalline column walls to soft porous column and cobble cores. Page (1972) reported porosities of about 25% by volume for indurated surface crusts and Kulke (1974) 31% to 45% in the case of Algerian croûtes de nappe. The porosity was determined of five cubes, each of about 10 g, cut from a gypsum cobble from Kettana, Tunisia (see sample K 9.1), composed of about 90% gypsum Each cube was dried at 35°C and weighed before being soaked in water under vacuum for 6 hours. The saturated cube was then reweighed, in water, and the volume determined by displacement.

$$\text{Porosity (\%)} = \frac{100}{\text{Cube Vol. (Dry)}} \times \text{Vol. of water in saturated cube}$$

It should be pointed out that there will be over-determination owing to gypsum solution during the saturation process, though several test runs were undertaken to ensure saturation of the water with gypsum. Loss of water owing to evaporation during the determination of the volume of water in the saturated cube will result in slight under-determination of the porosity. Since volume was determined by weight, another small source of error is the assumption that the density of the water is 1.0. Owing to the presence of dissolved gypsum it will be slightly greater. The average volumetric porosity was 37.90% with a range from 35.58% to 40.43%.

As in the case of the mesocrystalline subsurface crusts described above, a number of the sites and sections that will be referred to and discussed below have a variety of forms and types of crust. Rather than subdivide individual sites according to the various crust types and forms, they will be examined under the heading of the most characteristic form that is present. Surface forms from nine sites in southern Tunisia and four in the Central Namib will be examined. The locations are shown on fig. 4.1 and 4.4. These fall into the following categories, two sites in each field area representing powdery crusts and two in each representing the intermediate cobble form; columnar surface crusts will be described from five locations in southern Tunisia and from one in the Central Namib Desert.

Near Oued Demna (OD), 10 km north-west of Gabès, a powdery gypsum crust, usually light grey (7.5 YR 8/1 - Japanese Colour Chart) to white in colour, is partially exposed beneath orange (7.5 YR 6/6) sands and silty material. Topographically the area is a plain, about 10 m to 20 m above sea level, dissected by steep-sided oueds. Mio-Pliocene silts and clays, overlying gypsiferous Wealdian clays (sample OD 9) are exposed in the oued sections. Vegetated dunes, up to 3.0 m in height, form isolated hillocks on the otherwise featureless plain. Sample OD 7 (see table 5.22) is taken from a depth of 0.50 m beneath the crest of one of these. 0.20 m to 0.30 m of orange powdery material overlies a partially consolidated, powdery gypsum horizon about 0.50 m thick (see fig 5.7D). This exhibits horizontal stratification into alternating beds of weakly cemented, microcrystalline gypsum and loose, greyish powder. This structure is best developed in areas where the surface material has been removed by deflation or stripping prior to excavation of basal gypsum for building stone. When excavated the white subsurface gypsum indurates (see samples OD 6.1 and 6.3, tables 5.4 and 5.22). These are also encountered beneath small nebkha dunes (sample OD 6.4, table 5.22). At depths of between 0.80 m and 1.5 m beneath the surface, an indurated gypsum crust composed of

lenticular gypsum crystals less than 1.0 mm in diameter underlies the partially consolidated powdery gypsum.

The most striking structural trait of the powdery crusts around Oued Demna is their tendancy to exhibit large-scale polygonation. Irregular polygons, from 1.0 m to 10 m across, are defined by light grey to white lines 0.10 m wide (see fig. 5.7E) which sometimes form ridges 20 mm to 30 mm high. The nets are developed on both the powdery orange surface material and exposures of white powdery gypsum. They are found in many areas of southern Tunisia where the surface is mantled by powdery gypsum deposits and have also been identified in the Central Namib Desert (Goudie, 1972). Watson (1981) discussed their origin but additional data will be examined below.

The second southern Tunisian site exemplifying the powdery gypsum crust form is located near the Oued Zigzaou (ZIG) between Gabès and Medenine in the Djeffara. The oued has incised a 1.50 m thick calcrete conglomerate (ZIG 1.1, 1.2, 1.3) overlying red clays of Mio-Pliocene age (ZIG 1.4). On the northern bank, alluvial sands and silts form a 5.0 m high terrace which abuts the incised calcrete and clays. A powdery gypsum crust is developed upon this alluvium. Its surface is indurated to a depth of less than 10 mm. Beneath this white, powdery gypsum admixed with quartz sand extends to a depth of 1.50 m to 2.0 m. 1.0 m below the surface, gypsified rootlets from a distinct dikaka horizon. Beneath the gypsum horizons there are 5.0 m of partially cemented sands and silts. Induration is greatest below 4.0 m where calcium carbonate has precipitated from groundwater. Neither columns nor polygonal nets are developed at the surface (see fig. 5.8A).

In the Central Namib Desert, two sites with well developed powdery gypsum crusts were examined though many other areas in the vicinity of Swartbankberg have powdery gypsum deposits associated with subsurface mesocrystalline crusts

and cobble forms. 10 km north-west of Swartbankberg (WG, tables 5.3 and 5.25) the landscape is flat with small marble and dolerite outcrops and shallow stream channels. The surface consists of 0.05 m to 0.10 m of quartz sand and lag pebbles 10 mm to 20 mm in diameter. Beneath, a loosely cemented gypsum crust (WG 2) made up of powdery microcrystalline gypsum extends to about 1.0 m. Small lumps, up to about 0.10 m in diameter, of mesocrystalline, lenticular gypsum crystals are found within the powdery material. At the surface, polygonal nets, 5.0 m to 10 m in diameter, defined by lines of the grass Stipagrostis gonatostachys, are well developed.

The structure of these polygonal features has been examined at a site 1.5 km north-west of the summit of Swartbankberg (SB) on the plain between that inselberg and a narrow marble ridge trending north-south which lies 3.0 km to the east. Microcrystalline surface crusts are found over much of the land surface but the lowest lying areas are composed of a layer of orange (7.5 YR 6/6) quartz sand and gravel about 0.05 m thick which overlies powdery gypsum generally bright brown (7.5 YR 5/8) in colour. This is admixed with lenticular crystals up to 1.0 mm in diameter and powdery calcium carbonate which may amount to over 60.0% by weight. A polygonal pattern is defined by Stipagrostis gonatostachys growing along the polygon edges. The polygons are irregular and from 2.0 m to 6.0 m in diameter (see plate 5.26). Excavation revealed that the polygon boundaries are quartz sand infilling 0.05 m to 0.10 m wide fissures between plinths of gypsum which form the polygon faces. The edges of the plinths are composed of dull orange (7.5 YR 7/3) material with white mottling and are more indurated than the rest of the powdery gypsum (see fig. 5.8B). Chemical analyses of the material reveal that the mottled zone has a considerably higher gypsum content (65.84%) than the adjacent powdery material (40% to 50%), though this may be higher in areas with indurated surface crusts (59.17%). Gypsum enrichment adjacent to the fissure is probably the result of precipitation of gypsum from water moving through the sands or enhanced evaporation at the

interface between the moisture-retentive powdery gypsum and more porous sands. Watson (1981) showed that the polygonal patterns, defined by vegetation growth either along the polygon edges or on the faces, are found on three different types of soil in the Central Namib. Besides those on powdery gypsum soils, there are similar features on the mesocrystalline, subsurface gypsum crusts overlain by calcrete north of Gobabeb and large examples on calcium carbonate cemented sands in the interdune valleys south of the Kuiseb near Gobabeb (Ollier and Seely, 1977) (see plates 5.27 and 5.28). It was concluded that 'the main factor in the production of patterned ground in the Central Namib Desert is desiccation of gypsum-rich sediments', the interdune polygons probably being fossil features which developed before the bulk of the gypsum within the sands was leached out. Residual gypsum horizons are present locally within these sands (1W 4.2 and 5.1, table 5.25).

Polygonal structures on the powdery gypsum crusts of southern Tunisia are essentially the same as those in the Central Namib Desert. Gruet (1954) proposed seven mechanisms which might account for their development. He rejected percussion, tectonic and gravitational fracturing on the grounds that the distribution of the polygons was too restricted. Gypsum dissolution would produce a different form of polygonation. Chemical dehydration causing volume reduction was held to be unlikely within the temperature ranges experienced in the area and desiccation would not produce the necessary contraction. Gruet argued in favour of ice wedging and later infilling of the fissures with gypsum. This, however, would seem unlikely in the absence of any supporting evidence for periglacial conditions in southern Tunisia in the late Pleistocene. Evidence which supports the desiccation model comes from salt lakes in the western United States (Lang, 1943; Willden and Mabey, 1961; Christiansen, 1963; Neal and Motts, 1967). The depth of the fissures in the salt crust produced by tensional stresses is usually one tenth of the surface diameter of the polygons. This is in keeping with the dimensions of the features associated with gypsum

crusts. Volume changes resulting from desiccation of clays (Jessup, 1960 a.; Ollier, 1966; Butzer and Hansen, 1968, pp. 105 - 107) may play a role but many of the materials exhibiting polygonation have a negligible clay content. Gypsum dehydration resulting in volume reduction (Chatterji and Jeffery, 1963; Hunt et al., 1966) cannot be dismissed since hemihydrate occurs naturally in association with surface gypsum deposits. The importance of this process will be discussed in chapter seven in relation to the development of columnar structures in gypsum crusts.

Surface crusts composed of microcrystalline gypsum are widely distributed in the Djeffara and around the Chotts of southern Tunisia (see fig. 5.1). They frequently exhibit columnar structures of similar dimensions to those associated with mesocrystalline subsurface crusts described above. Sections exposed in the Oued Magroum (EH), 15 km west of El Hamma de Gabès south of Chott el Fedjedj, provide some of the most striking examples of these crusts (see plate 5.29 and fig. 5.8C) A layer of quartz sand 20 mm to 30 mm thick overlies a hard light grey to white gypsum crust. When exposed, the crust surface has a marked polygonal pattern formed by roughly hexagonal faces 0.15 m to 0.30 m in diameter. The edges of the faces are either slightly raised owing to greater induration of the walls of the columns which extend down from the surface or the edges are marked by 5.0 mm to 10 mm wide gaps filled with sand. The surface is hard microcrystalline gypsum 10 mm to 30 mm thick. Beneath this rind, the white gypsum is softer and often damp. The column walls and the surface are harder because evaporation has precipitated gypsum from the moisture within the crust. Once this has occurred the columns become sealed and further evaporation of moisture in the interior is slow or negligible. Raised column walls have a laminated structure which testifies to precipitation of gypsum from moisture moving through the joints. Small gaps between them at the surface probably indicate gypsum solution by downward percolating water. The columnar crust is 1.0 m to 2.0 m thick grading downwards into a zone of gypsum-rich

clays and sandstone. The gypsum in this zone is microcrystalline and columniation is developed weakly. A large lump of iron-rich material from within this zone was examined (see table 5.26). It appears to be associated with the bedrock rather than the gypsum crust since similar structures are present in the undisturbed Wealdian strata below. If this interpretation is correct, the gypsum crust has developed directly upon the bedrock and to some extent within the clay and sandstone beds, causing their disruption. At a depth of 3.0 m to 4.0 m sub-horizontally bedded red and green Wealdian clays appear. Isolated beds of selenite and satinspar 10 mm to 30 mm thick are found within these clays.

The section on the eastern side of the Oued Magroum (see fig. 5.8D) differs from that on the west, described above, in that alluvial sands and gravels form a 6.0 m to 7.0 m high terrace abutting the Wealdian deposits. The sediments are similar to those described from Oued Zigzaou. A columnar microcrystalline gypsum crust underlies about 1.0 m of orange silty sands. The alluvial sands and gravels beneath the crust are cemented by microcrystalline gypsum, though small aggregates of lenticular crystals up to 5.0 mm in diameter are found toward the base of the section. These may indicate a transition from a surface crust to a croûte de nappe down the section.

10 km west of Gabès, the Djebel ed Dissa (DI) forms a north-south trending ridge of Turonian limestone (samples DI 3.3 B and 3.4, table 5.6) and clays (sample DI 4.2, table 5.28) separating the Chott el Fedjedj from the Golfe de Gabès. Columnar surface gypsum crusts cover most of the Djebel, often lying directly upon limestone bedrock (see fig. 5.8E and plate 5.30). A surface scatter of sand and lag gravel 0.10 m to 0.20 m thick overlies silty sand which in turn rests upon the columnar gypsum crust. The material is microcrystalline, though some small lenticular crystals are evident near the base of the 1.0 m to 1.50 m thick crust. The columnar structure is the same as that

Figure 5.8 (overleaf)

Location of samples taken from sections through surface gypsum crusts in southern Tunisia and the Central Namib Desert. Site locations for A, C, D and E are shown in fig. 4.1 and for B in fig. 4.4.
Chemical analyses are presented as follows:

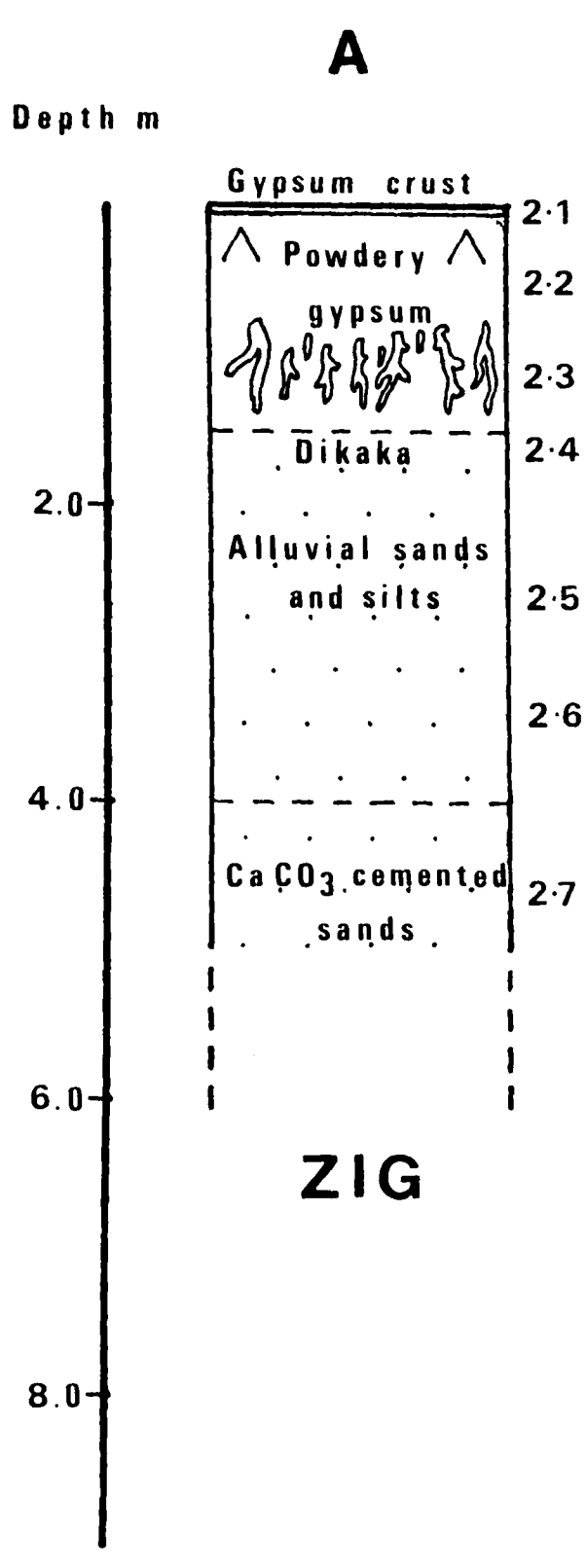
A. Oued Zigzaou (ZIG) table 5.24.

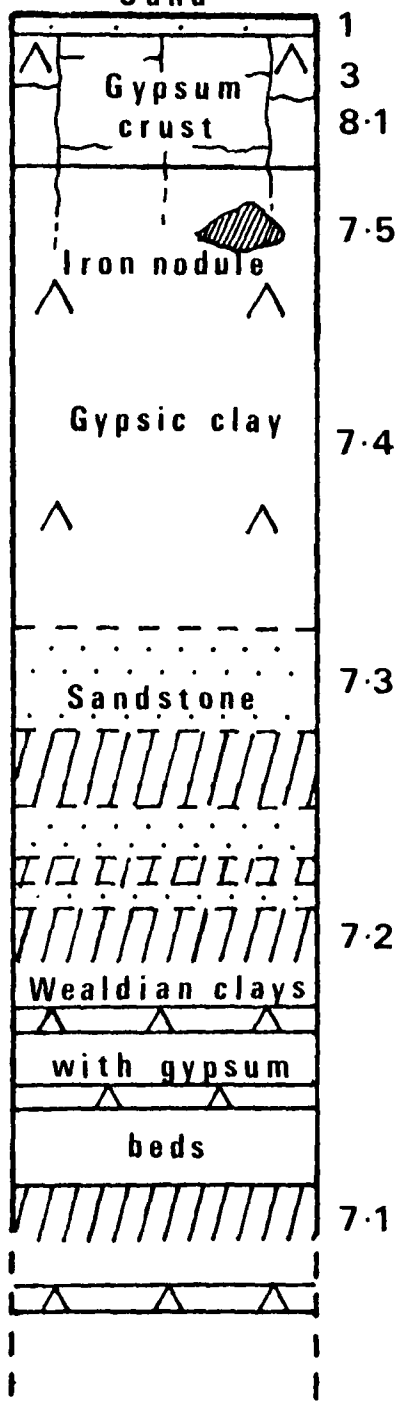
B. Swartbankberg (SB) table 5.25.

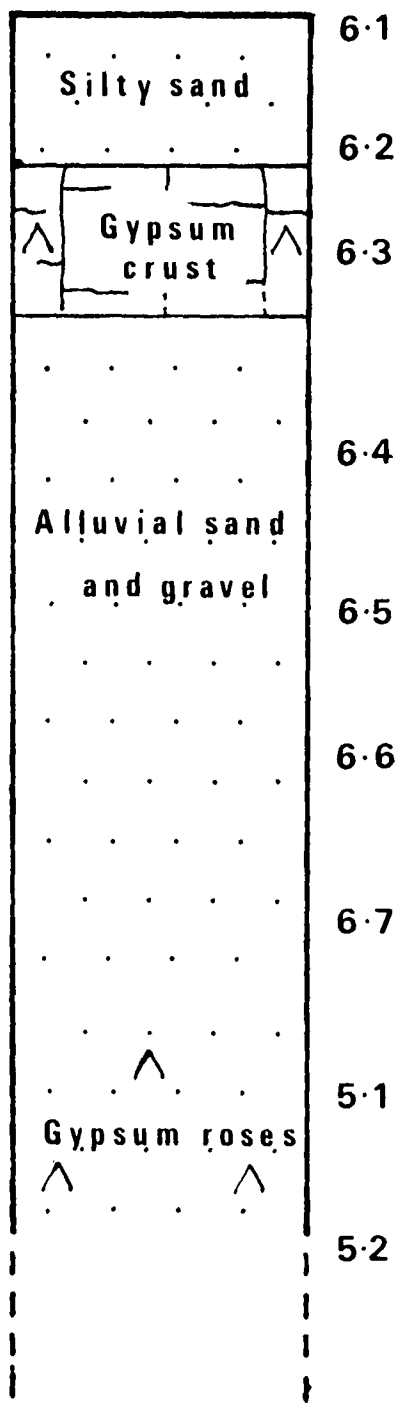
C. Oued Magroum (EH) table 5.26.

D. Oued Magroum (EH) table 5.27.

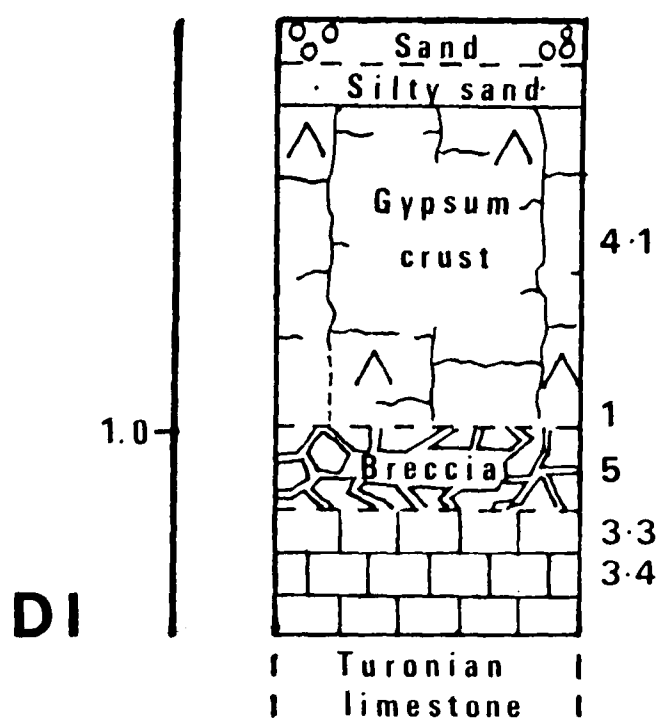
E. Djebel ed Dissa (DI) table 5.28.

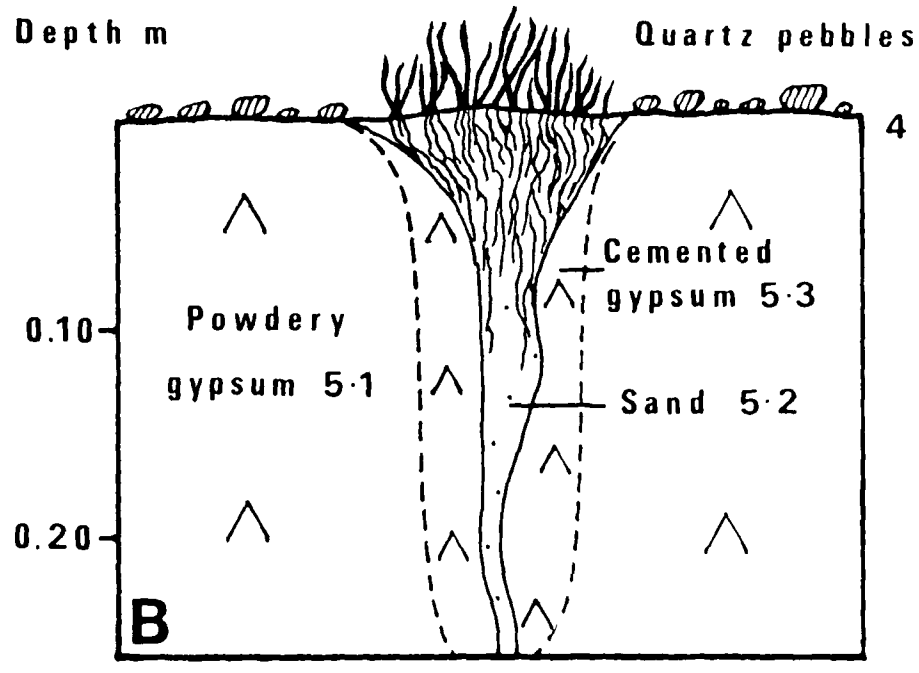

A
Depth m
Gypsum crust
2·1
Powdery
gypsum
2·2
2·3
Dikaka
2·4
2.0
Alluvial sands
and silts
2·5
2·6
4.0
CaCO3 cemented
sands
2·7
6.0
8.0
ZIG


C
Sand
1
Gypsum
crust
3
8·1
Iron nodule
7·5
Gypsic clay
7·4
Sandstone
7·3
7·2
Wealdian clays
with gypsum
beds
7·1


D
6·1
Silty sand
6·2
Gypsum
crust
6·3
6·4
Alluvial sand
and gravel
6·5
6·6
Gypsum roses
6·7
5·1
5·2

EH


E
Depth m
1.0
Sand
Silty sand
Gypsum
crust
4·1
1
Breccia
5
3·3B
3·4
DI
Turonian
limestone


Depth m
Quartz pebbles
4
Cemented
gypsum 5·3
0.10
Powdery
gypsum 5·1
Sand 5·2
0.20
B

SB

Plate 5.23 (overleaf, above)

A subsurface gypsum crust beneath basalt boulders on the western side of the Lagunenberg (CC) in the Central Namib Desert (see fig. 5.7A and table 5.19 for chemical analyses).

Plate 5.24 (overleaf, centre)

The surface of an excavated gypsum cobble from Swartbankberg (SB) in the Central Namib Desert.

Plate 5.25 (overleaf, below)

The underside of the cobble shown above.







Plate 5.26 (overleaf, above)

Polygonal patterns defined by growth of Stipagrostis gonatostachys near Swartbankberg (SB) in the Central Namib Desert. Note the white band of gypsum in the trench (see fig. 5.8B and table 5.25 for chemical analyses).

Plate 5.27 (overleaf, below)

Polygonal patterns developed on calcium carbonate sands in an interdune valley south of Gobabeb (GO) in the Central Namib Desert.



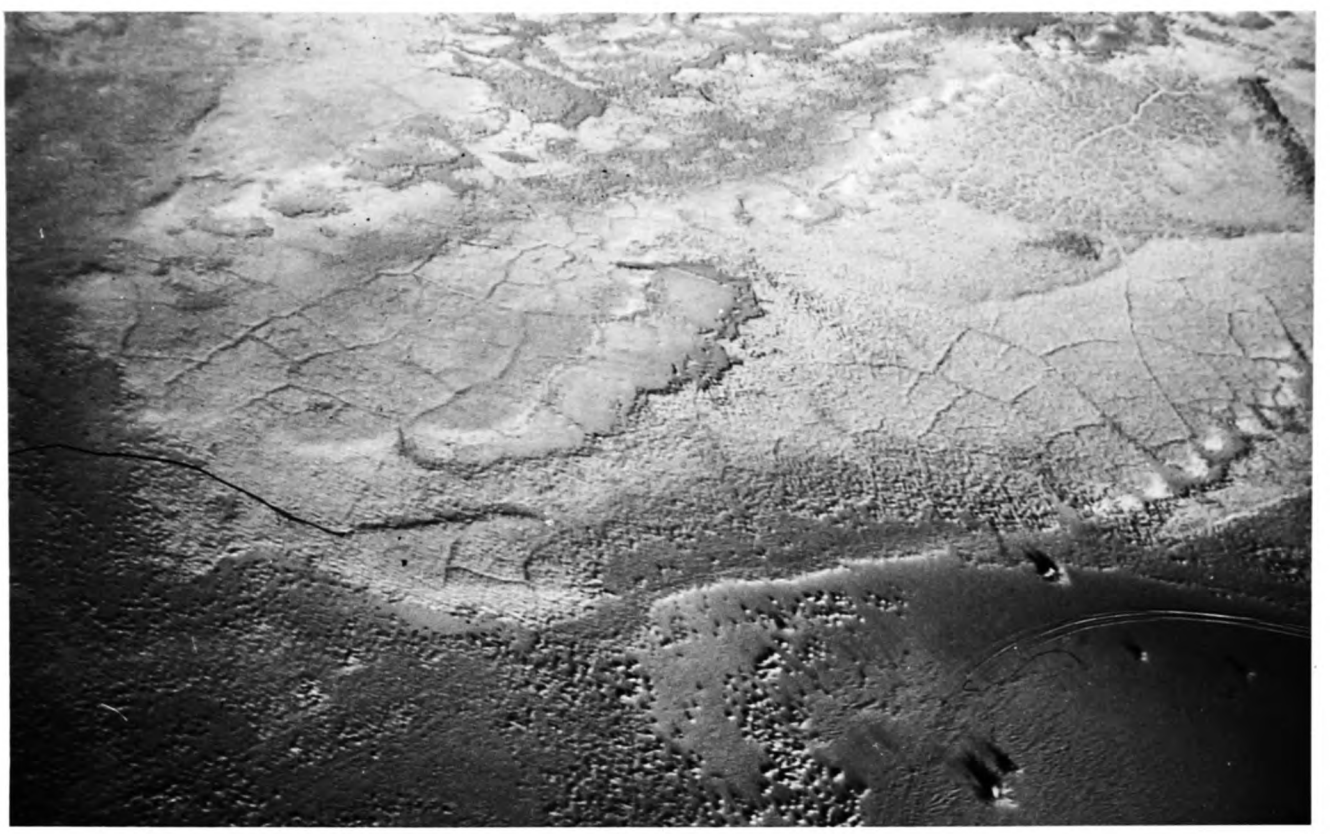


Plate 5.28 (overleaf, above)

A variety of vegetation patterning south of Gobabeb (GO) in the Central Namib Desert. Landrover tracks bottom right provide a scale.

Plate 5.29 (overleaf, below)

Columnar surface gypsum crust overlying Wealdian clays in the Oued Magroum (EH) south of Chott el Fedjedj in southern Tunisia (see plate 5.53, fig. 5.8C and table 5.26 for chemical analyses).




RST
11855

Plate 5.30 (overleaf, above)

Gypsum crust immediately overlying Turonian limestone at Djebel ed Dissa (DI) in southern Tunisia (see plate 5.51 and fig. 5.8E).

Plate 5.31 (overleaf, below)

Large microcrystalline gypsum cobble near Limaguess (L) in southern Tunisia. Note the profusion of sandstone fragments on top of the crust.





at Oued Magroum described above. The western flank of the Djebel ed Dissa is covered by columnar surface gypsum crusts which, on the lower slopes, are covered by silty sands and scattered vegetated dunes. To the south-east, a similar sand dune has several large limestone blocks of Roman age (J. Vieillefon pers. comm.) on its crest. This dune, 2.0 m to 3.0 m in height, rests upon a powdery gypsum crust exhibiting polygonal patterns 4.0 m to 6.0 m in diameter. These are identical to those described above at Oued Demma.

In the Oued Zerkine (ZER), 30 km south-east of Gabès, a surface gypsum crust is exposed in section where the oued incises the Djeffara plain where it meets the coastal shelf. In this area, Mio-Pliocene sediments have been capped by a calcrete horizon, 1.0 m to 2.0 m thick, which may be contemporaneous with that described above at Oued Zigzaou 10 km further south. The 0.75 m to 1.0 m thick gypsum crust is composed of microcrystalline gypsum and exhibits columniation the same as that described at other sites. It lies directly upon the calcrete, incorporating some calcrete pebbles in the basal zone (see fig. 5.9A).

Halfway between Tozeur and Nefta (TN, table 5.28) on the northern shore of Chott Djerid, surface microcrystalline gypsum crusts are found overlying fluvial sands (sample TN 2) and beds of Cardium edule. The 0.50 m to 0.75 m thick columnar crust (sample TN 1.1) is covered by patches of coarse quartz sand (sample TN 1.2) which have large but poorly developed selenite crystals on their surface (sample TN 1.3). These crystals are up to 100 mm in length and exhibit clinopinacoid, hemipyramid and prism, being similar to those described near Sebkhret Mekracherma. While it is unlikely that the gypsum crust is a croûte de nappe type since it is microcrystalline and columnar, it is possible that the crystals represent surface or near-surface deposits from a high water table. About 100 m down a gentle 1° to 2° slope toward the Chott, the surface is encrusted with sodium chloride efflorescences hence the gypsum crystals and indeed the crust may have a similar origin. Bedded selenite is

also present in the underlying Upper Cretaceous (?) bedrock (sample TN 3.2, table 5.6).

The final site in southern Tunisia where a surface microcrystalline gypsum crust was examined was near Djebel Tebaga Fatnassa (DTF, tables 5.6 and 5.29), a 275 m high ridge forming the eastern end of the chain of Djebel to the north of the Chotts Djerid and el Fedjedj. The site is 30 km north-west of Gabès. The gypsu crusts (DTF 1.2), on the lower slopes to the east of the Djebel, lie beneath a surface of coarse sand and pebbles derived from upslope. The crust contains more quartz grains than most described above and columnar structures are poorly developed. It is 0.80 m to 1.20 m thick and rests upon Upper Cretaceous, Turonian (?), clays (sample DTF 1.3) and conglomeritic limestones (sample DTF 1.4). Bedded gypsum is present within these strata (sample DTF 1.1).

Besides the shortage of good natural sections in the Central Namib Desert, there are few well developed surface crusts which have not undergone modification or deterioration. Only one site with near-surface crusts composed of microcrystalline gypsum will be described here though two different sections will be examined. They are located in the same stream channel 5.0 km east of Rooibank, as the croûte de nappe described above. The first section (see fig. 5.9B) is found on the side of a small tributary channel to the south-east of the site exhibiting a croûte de nappe. An upper horizon of powdery microcrystalline gypsum 0.10 m thick may be subdivided into an upper zone of compact material and a lower which is loosely packed. Beneath, there is a microcrystalline gypsum crust 0.20 m to 0.25 m thick exhibiting weak columniation. This in turn overlies 0.05 m to 0.10 m of mesocrystalline gypsum and a 0.10 m to 0.20 m thick clay horizon with lenticular gypsum crystals up to 5.0 mm in diameter. This lies upon a hard pinkish crust, up to 1.50 m thick, composed of calcrete pebbles and quartz fragments cemented by sodium chloride which sometimes exhibits fibrous textures. The second section (see figs. 5.4A and B), located on the wall of the

aforementioned dry river valley, has a similar sequence to the first. The surface consists of a lag gravel overlying 0.10 m to 0.20 m of sand resting upon a 0.30 m to 0.40 m thick gypsum crust. This crust, however, has two distinct zones. The upper 0.10 m to 0.20 m is made up of fibrous crystals sometimes aggregated into rhombic structures aligned vertically as are the fibres. Exposed faces show topotactic alteration to white hemihydrate but the bulk of the material is transparent gypsum with lens-shaped inclusions of sand and gravel oriented perpendicular to the fibres (see plate 5.32). Beneath this horizon the crust is composed of microcrystalline gypsum with lenses of fibrous gypsum 20 mm to 30 mm thick and 0.05 m to 0.10 m in diameter. Columnar structures are well developed. At a depth of 0.40 m to 0.60 m there is a sudden transition to a sodium chloride crust which incorporates fragments of calcrete and quartz. Fibrous lenses of similar dimensions to those in the gypsum horizon are composed of sodium chloride. This crust is at least 2.0 m thick but the absence of a vertical section precluded determination of the full depth. Samples from the slope down to the stream bed show a decrease in sodium chloride but the surface is mantled with gypsiferous material. Sand in the stream bed has a high sodium chloride content (sample PF 2.7, table 5.30) presumably derived from evaporating groundwater or surface water.

In the vicinity of this site, there are exposures of bedded selenite (sample PF 5), structurally akin to those in the Upper Cretaceous rocks of southern Tunisia, and also slightly gypsiferous clays (sample PF 4.2, tables 5.6 and 5.30). The Palaeozoic bedrock, however, does not contain evaporites and, since the area is about 170 m above sea level, recent transgressions cannot have resulted in sabkha facies this far in land as Gevers and Westhuyzen (1931) suggested. It is possible that this material represents a lacustrine or groundwater precipitate since it is invariably located in lowlying depressions in contrast with the fibrous and microcrystalline crusts which are frequently found on topographic highs.

The final form of surface gypsum crust is here termed the intermediate. This is justified on structural as well as genetic grounds since the crusts consist of cobbles of microcrystalline gypsum set in the surface of a powder gypsum horizon (see plate 5.31). Though often very scattered, the cobbles are occasionally densely packed revealing their origin as either embryonic or deteriorating columnar structures. Their importance lies in the fact that by ascertaining the relationship between them and microcrystalline columns it is possible to show the genetic relationship between not only powdery and microcrystalline crusts but also between surface crusts and the mesocrystalline subsurface form. The first of two sites that will be described in southern Tunisia is on the northern slopes of the most easterly part of the Djebel Tebaga, the Djebel el Aziza (DT) south of Chott el Fedjedj, about 10 km west-south-west of El Hamma de Gabès. The bedrock consists of Upper Cretaceous, Turonian, dolomites and limestones with intermittant clay and gypsum strata and is similar to that at Djebel ed Dissa and Djebel Tebaga Fatnassa. The strata dip south at about 10° and have eroded to produce a striking stepped appearance to the northern face of the cuesta. Most of the surface, excluding the near vertical faces, is covered by a thin powdery gypsum crust with cobbles set in the surface (see fig. 5.9D). These cobbles are 0.10 m to 0.30 m in diameter on the flat upper surface and up to 0.10 m thick, being domed or conical on their undersides. They are composed of microcrystalline gypsum and show the aforementioned hardening on all surfaces which protects softer gypsum at the core. Where the powdery gypsum or sandy horizon is more than about 0.20 m to 0.30 m thick and not resting directly upon bedrock, a subsurface crust is present. This is usually composed of lenticular gypsum crystals less than 1.0 mm in diameter but may contain much microcrystalline gypsum. The section shown in figure 5.9C shows surface cobbles set in sand which exhibits large scale polygonal patterns 2.0 m to 5.0 m in diameter, the polygon edges being defined by ridges of white gypsum 20 mm to 50 mm wide. At a depth of about 0.25 m large, fragmented blocks of Turonian dolomite up to 0.25 m thick overlie an

unconsolidated subsurface gypsum crust exhibiting weak columniation. This horizon is up to 0.50 m thick and overlies Turonian dolomite bedrock.

The location of the dolomite blocks above the subsurface crust is unusual. The local relief is insufficient for them to have fallen onto the crust therefore the gypsum has either displaced them upwards or replaced a pre-existing sediment. The possibilities that the subsurface gypsum crust may be of Turonian age is discounted on the grounds that the gypsum found within bedrock in the area invariably consists of either satinspar veins (see plate 5.32) or lenses or massive selenite (chemical analyses and thin sections DT 2.1, 2.2, 3.1, 3.2 and 7.2, tables 5.6 and 5.32). Plate 5.34 shows a lump of satinspar found within Turonian clays which is weathering to form lenticular gypsum crystals up to 10 mm in diameter (chemical analysis DT 6.2, table 5.32). It is possible, however, that weathering processes have resulted in alteration of gypsum from within bedrock to a gypsum crust. This will be examined in chapter six.

The second site at which the intermediate form of surface gypsum crust is found is that near Kettara, from where subsurface mesocrystalline crusts were described above (see figs. 5.6C and D). Cobbles are found on the hill crests and upper slopes where they are set in a powdery accumulation which is weakly columnar, the walls being defined by white microcrystalline gypsum as described above. The cobbles are from 0.05 m to 0.25 m in diameter at the surface and 0.03 m to 0.10 m thick being domed or conical on the underside. They are identical to those from Djebel el Aziza though some are markedly calcretized.

The intermediate form of surface gypsum crust has been examined at two locations in the Central Namib Desert. Both sites are on the marble ridge trending roughly north-south from the Hamilton Range west of Vogelfederberg in the north, to the Kuiseb west of Swartbankberg in the south, a distance of about 35 km. Near Swartbankberg (SB), the summit of the ridge has a well developed cobbled

Plate 5.32 (overleaf, above)

Fibrous gypsum overlying a columnar gypsum crust near Rooibank (PF) in the Central Namib Desert (see plate 5.55, figs. 5.4A and 5.4B and table 5.30 for chemical analyses).

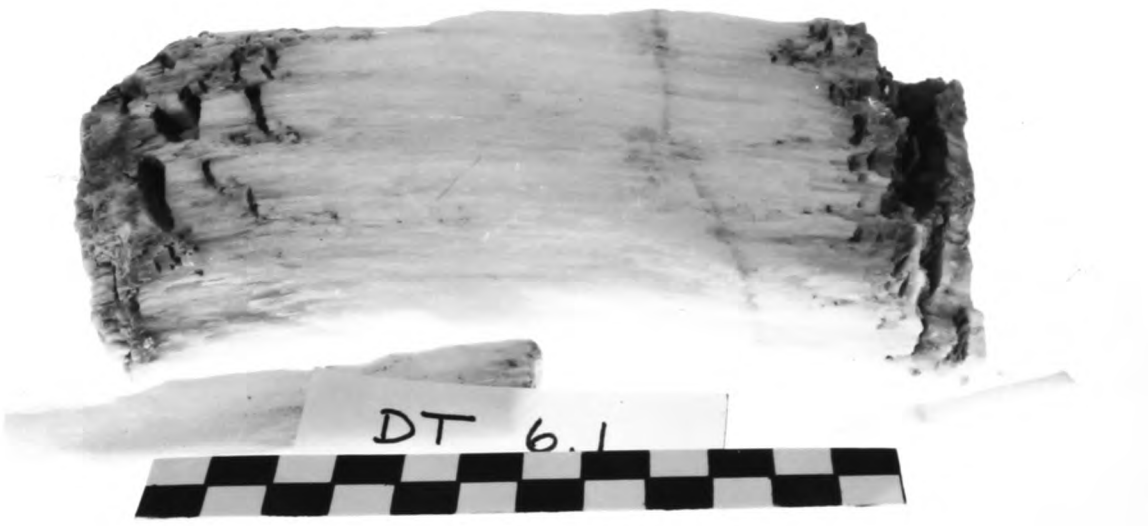
Plate 5.33 (overleaf, centre)

Satinspar from Djebel el Aziza (DT) in southern Tunisia (see plates 5.71 and 5.72).

Plate 5.34 (overleaf, below)

An exposed piece of fibrous gypsum from Djebel el Aziza (DT) showing surface alteration to lenticular gypsum crystals.




DT 6.1


DT 6.2

surface, the cobbles being 0.05 m to 0.25 m in diameter and identical in shape to those in southern Tunisia (see plates 5.24 and 5.25). Unlike the cobbles from Tunisia, however, the external hardening is less pronounced, in that the core is also composed of hard microcrystalline gypsum. Furthermore, the exterior of the cobble, except on the upper face exposed to the atmosphere, has areas of lenticular gypsum crystals up to 1.0 mm in diameter and zones of poorly developed fibrous crystals up to 1.0 mm long. The upper surfaces of the cobbles are frequently coloured bright yellow by a thin layer of algae. The powdery material, in which the cobbles are set, is made up of microcrystalline gypsum mixed with lenticular crystals up to 1.0 mm in diameter. This 0.20 m thick horizon overlies a weakly cemented, columnar mesocrystalline crust of about the same thickness which in turn rests upon marble bedrock (see fig. 5.9E)

15 km further north along the marble ridge, at the southern end of the Hamilton Range (HA), there are similar well developed columnar crusts and surface cobbles. On the eastern side of the ridge, below the marble outcrops which dip eastwards at over 42°, columnar gypsum crusts are found on slopes of up to 25°. These appear to be basal mesocrystalline crusts exposed by erosion of less consolidated surface materials. On the crest of the ridge, where erosion is less active, a columnar subsurface gypsum crust is overlain by partially consolidated powdery gypsum which also exhibits weak columniation. The surface is densely covered with gypsum cobbles of similar dimensions and structure to those near Swartbankberg. The 0.25 m to 0.30 m thick gypsum sequence overlies marble bedrock (see fig. 5.9F).

In order to ascertain the precise relationships among the various types of gypsum crusts and among their subforms, it is necessary to examine their nature in greater detail. These preliminary descriptions of their structural characteristics and field associations show that evaporitic crusts are quite distinct from the other types but that the mesocrystalline subsurface form and the

Figure 5.9 (overleaf)

Location of samples taken from sections through surface gypsum crusts in southern Tunisia and the Central Namib Desert. Site locations for A, C and D are shown in fig. 4.1 and for B, E and F in fig. 4.4.
Chemical analyses are presented as follows:

A. Oued Zerkine (ZER) table 5.28.

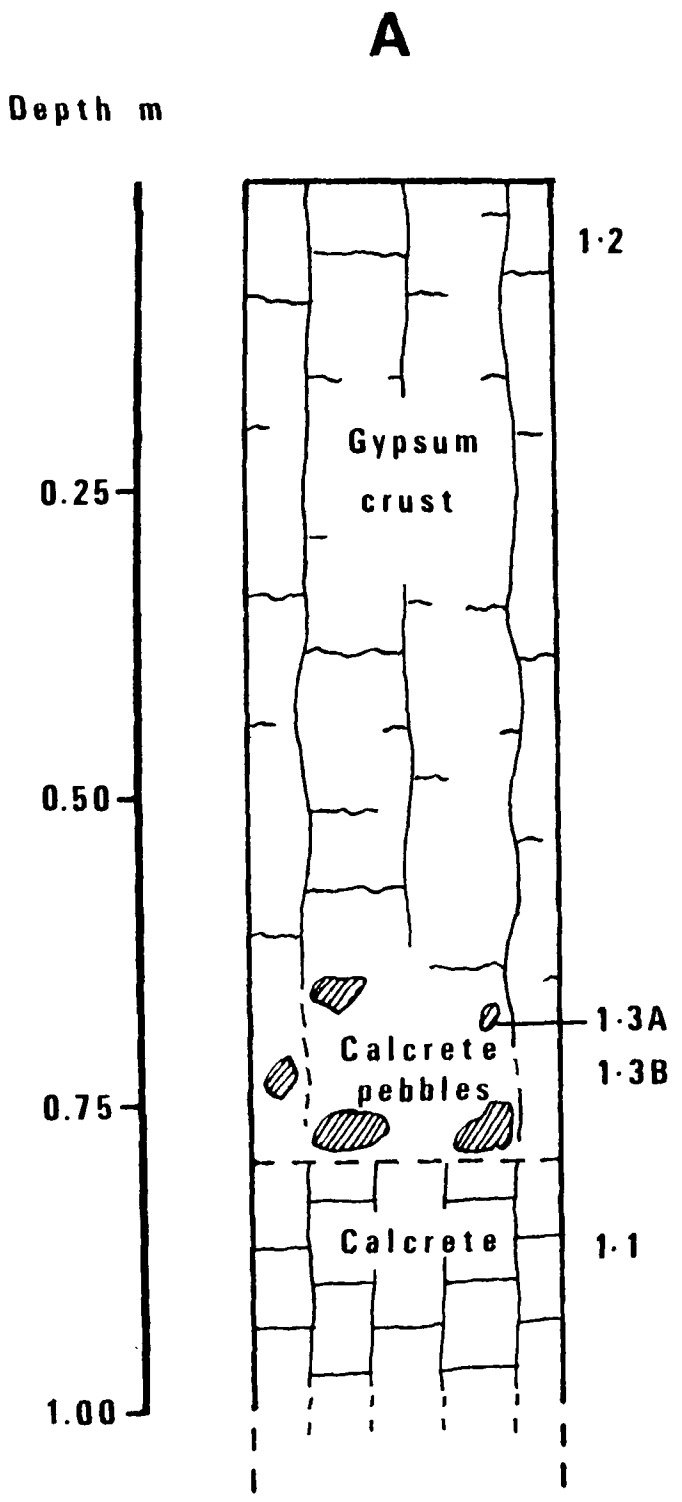
B. Near Rooibank (PF) table 5.29.

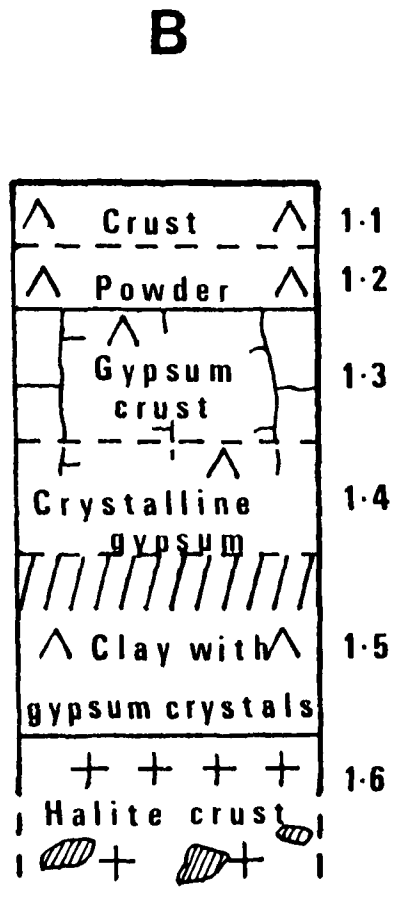
C. Djebel el Aziza (DT) table 5.31.

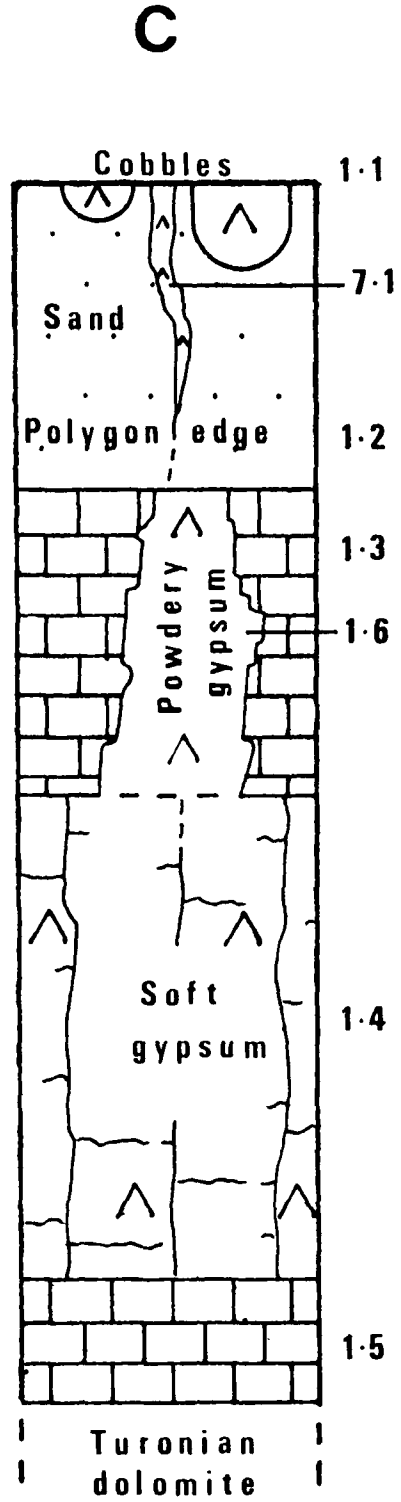
D. Djebel el Aziza (DT) table 5.32.

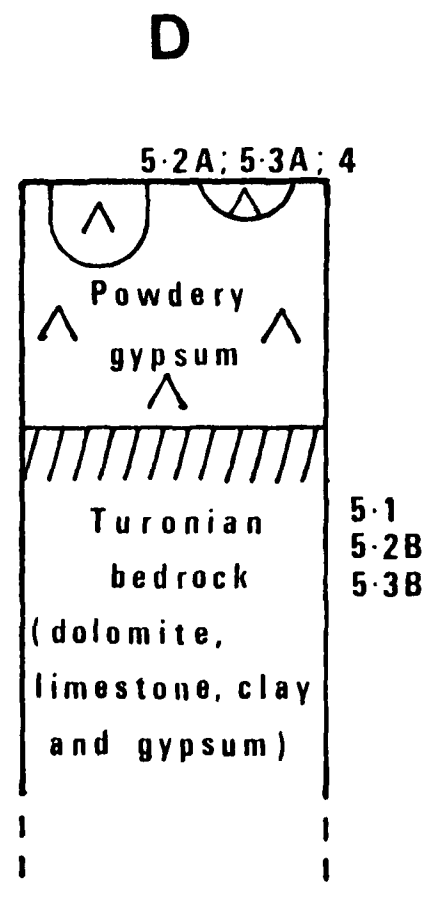
E. Swartbankberg (SB) table 5.33.

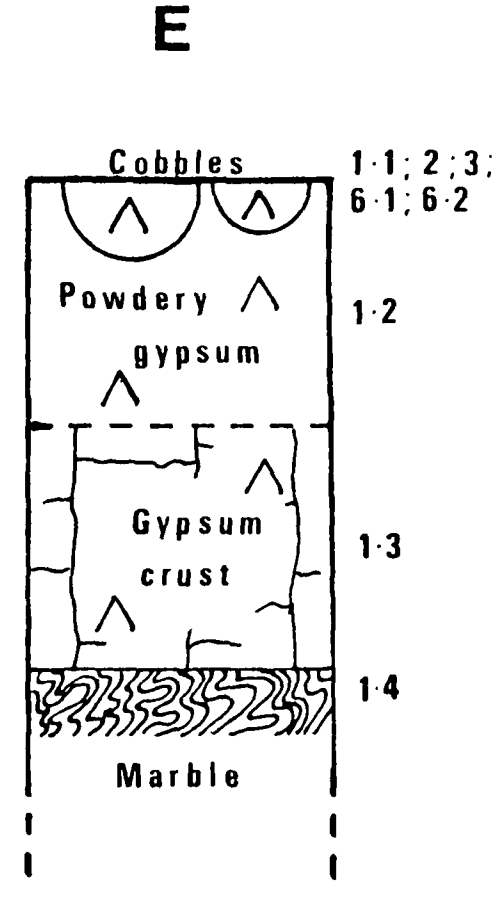
F. Hamilton Range (HA) table 5.34.

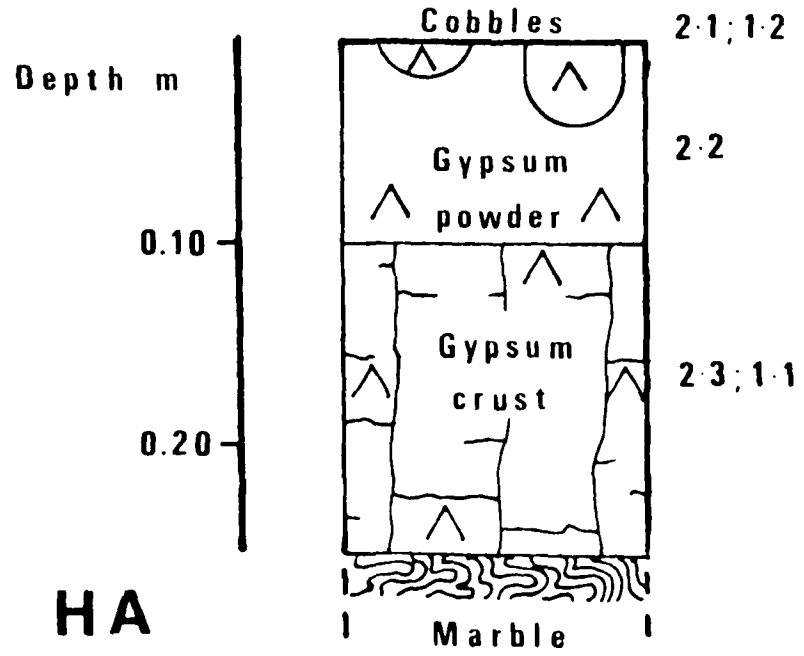

A
Depth m
0.25
0.50
0.75
1.00
Gypsum crust
1·2
Calcrete pebbles
1·3A
1·3B
Calcrete
1·1
ZER


B
Crust
1·1
Powder
1·2
Gypsum crust
1·3
Crystalline gypsum
1·4
Clay with gypsum crystals
1·5
Halite crust
1·6
PF


C
Cobbles
1·1
Sand
7·1
Polygon edge
1·2
Powdery gypsum
1·6
Soft gypsum
1·4
1·5
Turonian dolomite
DT


D
5·2A; 5·3A; 4
Powdery gypsum
Turonian bedrock (dolomite, limestone, clay and gypsum)
5·1
5·2B
5·3B
DT


E
Cobbles
1·1; 2; 3; 6·1; 6·2
Powdery gypsum
1·2
Gypsum crust
1·3
1·4
Marble
SB


Depth m
0.10
0.20
Cobbles
2·1; 1·2
Gypsum powder
2·2
Gypsum crust
2·3; 1·1
Marble
F, HA

various forms of surface crust are not only often found in close association but also possess structural similarities. The most striking of these are columnar structures and, to some extent, large scale surface polygonation. Beyond these, there are clear distinctions between the mesocrystalline and microcrystalline forms and between the desert rose subsurface forms and various evaporitic types in terms of their topographic locations. The next stage in the examination is to conduct detailed analyses of the microstructural characteristics of the gypsum crust.