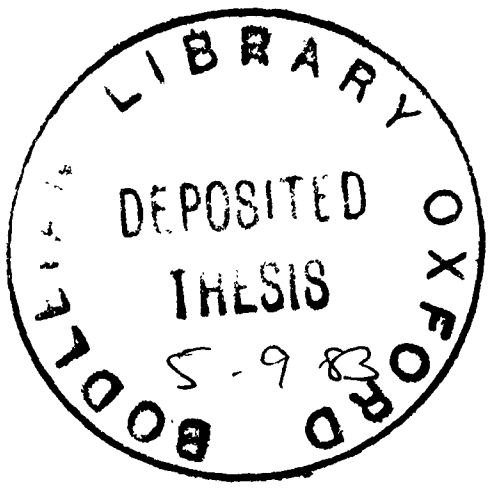
THE ORIGIN, NATURE AND DISTRIBUTION OF

GYPSUM CRUSTS IN DESERTS

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THE NATURE OF GYPSUM CRUSTS

c. Micropetrography of gypsum crusts.

Thin sections were prepared of 140 samples. All of the rock and crust samples which could be cut without disintegrating were thin sectioned. Impregnation of unconsolidated and poorly cemented samples was not attempted owing to the difficulties outlined below. Of these 4 were evaporitic gypsum crusts, 13 desert rose croûtes de nappe, 27 mesocrystalline subsurface crusts, 22 surface crusts, predominantly columnar, and 12 cobbles. The remaining 65 thin sections comprised, 14 of calcretes associated with gypsum crusts, 9 of gypsum bedrock from southern Tunisia, 8 of non-gypsiferous bedrock, 32 of calcrete, mostly from the Central Namib, and the one aeolianite sample from Djerba described in chapter four.

i. Methods of thin section preparation.

Several difficulties are encountered in the preparation of thin sections of gypsum-bearing rock and crust samples (Milton, 1942). These may all be attributed to two properties of gypsum, firstly, its solubility and, secondly, its tendancy to dehydrate to hemihydrate at temperatures above about 35^{o}C in the absence of water. In order to avoid excessive dissolution of gypsum and any halite in the thin section, sawing and grinding of the sample must be undertaken in the absence of water. When these procedures are carried out dry, however, frictional heating is sufficient to cause dehydration which markedly alters the internal structure of individual gypsum crystals. Plates 5.35 and 5.36 show two different aggregates of granular gypsum crystals. In plate 5.36, however, the crystals are neither translucent nor do they have uniform extinction. The cleavage planes are visible as small laths or fibres of hemihydrate. The large gypsum crystals have been replaced by large numbers of hemihydrate crystals and, although gypsum crystal outlines are preserved

as pseudomorphs, the internal structures and textures are masked. This is particularly confusing in the case of mesocrystalline gypsum which dehydrates to produce textures akin to microcrystalline or alabastrine material when viewed at magnification of less than about x 20.

To overcome dehydration and dissolution, sawing and grinding must be undertaken using a coolant such as acetone or methanol in which gypsum is only slightly soluble. Nevertheless, care must be taken to keep the rock surfaces cool at all times and this is best achieved by slow grinding and polishing. The mounting medium must also be cool when applied to avoid dehydration of the gypsum. Walker (1969) advocated the use of liquid casting plastic in the preparation of halite thin sections to avoid cleaving during grinding of the sample. Though similar problems are encountered in the case of gypsum crust samples, especially those containing resistant quartz grains, the difficulties in avoiding dehydration during casting and during grinding of the hard plastic, offset the benefits. For this reason slow grinding and polishing, even by hand, were preferred to impregnation. A problem which could not be surmounted was that of variable thickness among or even within thin sections. This produces a wide range of interference colours for the large gypsum crystals under polarized light but mineral identification was not severely hindered.

Staining of thin sections was not undertaken since in nearly all cases mineral identification was straightforward. Gypsum, hemihydrite and anhydrite may be easily differentiated in thin sections by cleavage characteristics, even if pseudomorphic structures are present, anhydrite being distinctively rectangular pseudocubic. Gypsum may be readily distinguished from calcite and quartz, the other main constituents of the samples examined. Identification of dolomite and aragonite and their differentiation from calcite proved problematic in two thin sections but chemical analysis and X-ray diffraction provided

additional information on the mineral constituents of the samples (see below).

ii. Terminology.

The terminology of the crystalline textures in gypsum rocks has been developed largely in the context of bedrock deposits which have experienced hydration from anhydrite (Hammerschmidt, 1883; Holliday, 1970; Mossop and Shearman, 1973). The terminology will be used in this examination of gypsum crusts, though the genetic interpretation of the textures, in terms of the rates of hydration and groundwater chemistry, do not apply.

The most common texture in the non-evaporitic types of gypsum crust is of rhombic crystals usually tending towards a lozenge shape. This represents aggregates of lenticular crystals which will appear as lozenges when sectioned through any plane except those parallel to the rim of the disc. For the purposes of this examination the lozenge textures are subdivided into those with crystals with a maximum diameter greater than 1.0 mm and those with diameters less than 1.0 mm but greater than 0.05 mm. This lower limit marks the boundary between the small lozenge-shaped form and the alabastrine (Hammerschmidt, 1883; Holliday, 1970). Alabastrine textures range from small (0.05 mm) lozenge-shaped crystals or subcrystals to cryptocrystalline masses. Granular textures consist of irregularly-shaped interlocking gypsum crystals which may grade into alabastrine textures when the crystal diameters fall below 0.05 mm. Holliday(1970) felt that the granoblastic form was the product of slow hydration from anhydrite, while alabastrine textures indicated rapid transition. Gypsum crusts rarely exhibit granular textures though they may be localized in evaporitic crusts and along column walls in the other types. Lozenge-shaped gypsum crystal porphyroblasts from 0.50 mm to 5.0 mm in diameter set within a groundmass of alabastrine gypsum are found

in all the crust forms except the desert rose croûtes de nappe. The term porphyroblast is applied to large, isolated euhedral crystals which have developed in the solid state within a finer groundmass, usually in metamorphic rocks. In the case of secondary gypsum rocks, porphyroblasts may indicate slow hydration (Mossop and Shearman, 1973) or may be remnants of earlier gypsification (Holliday, 1970). In gypsum crusts, the porphyroblasts may be relic or possibly replacive or displacive crystal growths or detrital inclusions, as Misch (1971) concluded in relation to metamorphic rocks. The final gypsum texture is the fibrous form. Though a rare feature of gypsum crusts, these textures are of great importance to the understanding of the origin of several different forms of crust. They will be discussed in chapter seven, suffice it to point out here that when the fibres are aligned parallel to the plane of the thin section the elongated crystals each with uniform, or rarely undulose extinction, and well developed cleavage-cracks are unmistakable.

Though a wide variety of calcite textures are encountered in the gypsum crust and calcrete samples examined in thin section, for the purposes of this analysis, they will be greatly simplified. The distinction between sparry and micritic calcite is made on the basis of crystal size, the micritic texture being that made up of crystals less than 0.01 mm in diameter. Fibrous calcite (beef) is present in some calcretes from the Central Namib Desert but the occurrences are rare. A calcite texture important in relation to gypsum crusts is that composed of crystals pseudomorphic after gypsum. These take the form of lozenge-shaped ghosts composed either of individual calcite crystals or, more commonly, micritic calcite. They are present in both gypsum crusts and calcretes associated with gypsum crusts. Though Coque (1962, p. 88) alluded to chemical replacement of gypsum by calcite, such pseudomorphic structures have never been described from near-surface crusts and only occasionally from bedrock (Lacroix, 1897).

Detailed microscopic determination of the quartz grain content of the samples has not been undertaken because, although there are a number of potential sources of error, chemical analyses provide this information. The location of quartz grains within the matrix is classified as either poikilitic, within gypsum crystals, or as a floating texture with the grains isolated by areas of mesocrystalline or alabastrine gypsum.

In many gypsum crust and calcrete thin sections, especially from southern Tunisia, structures are encountered which are probably biogenic. They are confined to the near-surface soil zone where burrowing organisms produce tunnels which are later filled with gypsum and calcite precipitated from waters moving through them. Botanic influences may also be important, enhancing crystallization around roots and in the voids they form. The structures are here grouped under the term pedoforms. Pedoforms have not been subdivided according to different structures but according to their mineralogy. In most cases they consist of elongated or circular zones, 0.50 mm to 1.0 mm in diameter, composed of micritic calcite or gypsum crystals more densely packed than the matrix.

In tables 5.1 to 5.6 the results of micro-petrographic analyses are summarized. Simple quantitative parameters are provided by the notations (I) for the texture or feature being present, (II) for abundant and (III) for dominant. In the case of isolated crystal textures or structures abundance is defined as over one porphyroblast or pseudomorph for every 25 mm^2 of thin section. The same applies to pedoforms. In the case of quartz grain textures more than about 5 grains/mm^2 (0.10 mm to 0.20 mm in diameter) constitutes abundance and corresponds to over 10% by weight of insoluble residue in the chemical analyses. A dominance of quartz grains, whether poikilitic or floating, will usually indicate a grain supported texture with over 50% of the area of the slide

TABLES 5.1 to 5.6

The following tables describe the micropetrographic characteristics of thin sectioned gypsum crusts, calcretes and bedrock from the two field areas. The notations are as follows; (1) the feature or texture is present, (11) it is abundant, (111) it is dominant. A detailed explanation is provided in the text.

TABLE 5.1

SAMPLE	TYPE of CRUST	STRUC-TURE	GYPSUM CRYSTAL TEXTURES						CALCITE TEXTURES			QUARTZ GRAINS		PEDOFORMS		PLATE NUMBER
			LOZENGE >1mm	LOZENGE <1mm	ALABAS-TRINE	GRANU-LAR	PORPHY-ROBLASTS	FIBROUS	SPARRY	MICRITIC	PSEUDO-MORPHS	POIKIL-ITIC	FLOATING	CALCITE	GYPSUM	
CR 1.1	Evaporitic	varved	111	1	11	11	1			1						5.37 and 5.38
CR 1.2	"	massive	111	11		1										5.39
CC 2.2	"	varved		1	111	1	1			1			11			
NA 1.2	"		1	111						11		1	1			5.40
ZAA 1	desert rose croûte de		1	111		11				1	1	1	11			
ZAA 3.1	nappe "		1	111	1	1				1		1	11			
EF 1.2	"				111								11			5.41
OT 4.3	"			1	111					1			1	1	11	
EM 3.9	"		1	111	11							1	1			5.42
EM 3.8	"		111	1								11				5.43
IMET 1.1	"		111	1	1	1				1		1	11			
IMET 1.2	"		111	1	11							1	1			
IMET 1.3	"		11	11	111							1	1			
IMET 2	"	column edge	1	1	111				1	1	1		1			
IMET 3.3	"				111	11			1	1			11			
WB 1.2	"		11	111	11					1		1	11			
WB 1.3	"		11	111	1	11				11		1	1			

TABLE 5.2

SAMPLE	TYPE of CRUST	STRUCTURE	GYPSUM CRYSTAL TEXTURES						CALCITE TEXTURES			QUARTZ GRAINS		PEDOFORMS		PLATE NUMBER
			LOZENGE >1mm	LOZENGE <1mm	ALABASTRINE	GRANULAR	PORPHYROBLASTS	FIBROUS	SPARRY	MICRITIC	PSEUDOMORPHS	POIKILITIC	FLOATING	CALCITE	GYPSUM	
MFI 1.3	mesocrystalline subsurface croûte de nappe	columnar		III	II					I			I	I	II	5.44
MZ 1.4	"			III	II					I			I	I	II	5.45
MZ 1.3	"			III	I					II			II	I	I	
MZ 1.2	"			III	II					II			II	I	I	
MZ 1.1	"			III	I					II			II	I	I	
L 4	"	column edge		III	II	II	I		I	I			I	I	I	5.63 and 5.64
DL 1.3	"	columnar	I	III	I		I			I	I		I	I	II	
DL 1.4	"	columnar	I	III	I		II		I	I	II		I		I	5.46
DL 1.6	"	(basal zone)	III	II		I			I			II	I			
DU 2.1	"	(surface zone)		I	III					I			II		I	5.47
LS 1.2	"	(surface zone) columnar		I	III					I			I	I	I	5.48
LS 1.3	"	columnar	I	III	II		I		I	I			I	I	I	5.49
K 5 *	"		I	III	I					I			I	I	I	
K 8.1	"	(exposed)		III	II					I			I	I	I	
K 8.2	"	(exposed)		II	III					I			I	I	II	
CC 1.3	"	columnar	II	III	I	I	I	I	I			I	I			

* N.B. The chemical analysis is of a single calcrete pebble

TABLE 5.2 (cont.)

SAMPLE	TYPE of CRUST	STRUCTURE	GYPSUM CRYSTAL TEXTURES						CALCITE TEXTURES			QUARTZ GRAINS		PEDOFORMS		PLATE NUMBER
			LOZENGE >1mm	LOZENGE <1mm	ALABASTRINE	GRANULAR	PORPHYROBLASTS	FIBROUS	SPARRY	MICRITIC	PSEUDOMORPHS	POIKILITIC	FLOATING	CALCITE	GYPSUM	
CC 1.7	mesocrystalline subsurface croûte de nappe	(basal zone)	1	111		1	1						1			
SB 3.4	"	nodule	1	111		1						1	1			
SB 3.7	"	root?		111								1			?	
GO 4.4	"	fibrous	1	111	1			11	1	1	1		11			
OD 1	"	exposed		111	11	1	1			11			1	1	11	5.50
OD 2	"	exposed		111	11					1			1	1	1	
OD 3	"	exposed		111	1	1	1		1	1	1		1	1	11	
OD 5.2	"	column edge		111	11	1	1			1			1	1	1	
DI 5	Base of overlying Turonian limestone	breccia				11				111						5.51
DT 1.4		columnar		11	111				1	1	1		1	11	11	5.52
HA 2.3	mesocrystalline subsurface croûte de nappe	columnar		111	1	1				1			1		1	

TABLE 5.3

SAMPLE	TYPE of CRUST	STRUC-TURE	GYPSUM CRYSTAL TEXTURES					CALCITE TEXTURES			QUARTZ GRAINS		PEDOFORMS		PLATE NUMBER
			LOZENGE >1mm	LOZENGE <1mm	ALABAS-TRINE	GRANU-LAR	PORPHY-ROBLASTS	FIBROUS	SPARRY	MICRITIC	PSEUDO-MORPHS	POIKIL-ITIC	FLOATING	CALCITE	GYPSUM	
EH 1	columnar surface crust				111								11		1	5.53
EH 3	"				111								1		1	
EH 8.1	"				111								1		1	
DI 4.1	"			11	111					11			1	11	11	
DI 1	"			11	111					1			1	1	11	5.54
IN 1.1				1	111	1	1			1			1	1	1	
PE 2.2		fibrous						11	1	111						5.55
MEI 2.1				11	111					1			1	1	11	
L 1				111	11	1	1		1	1	1		1	1	11	5.56
L 2				111	11		1			1	1		1	1	1	5.57
L 3				111	11	1	1		1	1	1		1	1	11	
OL 1.1				111		1	1		1	1	1		1	1	11	
OL 1.2				11	111		1		1	11	1		1	1	11	
DU 1.2				1	111					1			1	1	1	
K 6	polygonal powdery surface crust	polygon edge		11	111								1		1	5.58

TABLE 5.3 (cont.)

SAMPLE	TYPE of CRUST	STRUC-TURE	GYPSUM CRYSTAL TEXTURES						CALCITE TEXTURES			QUARTZ GRAINS		PEDOFORMS		PLATE NUMBER
			LOZENGE >1mm	LOZENGE <1mm	ALABAS-TRINE	GRANU-LAR	PORPHY-ROBLASTS	FIBROUS	SPARRY	MICRITIC	PSEUDO-MORPHS	POIKIL-ITIC	FLOATING	CALCITE	GYPSUM	
CC 1.6	columnar surface crust	column edge		111	11	1	1						1			
SB 3.8	surface crust			111	11				1	1	1					5.59
SB 4	surface crust			111	1								11			
OD 6.2A	"			1	111					1			1	1	1	5.60
WG 1	powdery surface crust			111	1					1			1			
DI 7.1	"	polygon edge														
HA 2.2	"	nodule		11	111				1	1	1		1		1	5.61 and 5.62

TABLE 5.4

SAMPLE	TYPE of CRUST	STRUCTURE	GYPSUM CRYSTAL TEXTURES						CALCITE TEXTURES			QUARTZ GRAINS		PEDOFORMS		PLATE NUMBER
			LOZENGE >1mm	LOZENGE <1mm	ALABASTRINE	GRANULAR	PORPHYROBLASTS	FIBROUS	SPARRY	MICRITIC	PSEUDOMORPHS	POIKILITIC	FLOATING	CALCITE	GYPSUM	
DT 1.1	microcrystalline surface cobble			1	111								1		11	
DT 4	"			11	111								1		11	
DT 5.2A	"			111		11							1		11	5.65
DT 5.3A	"			11	111					1			1	1	11	
K 2	"			11	111					1			1	1	1	
K 9.2	"			11	111								1		11	5.66
HA 2.1	"			111		1	1		1	1	1		1			5.67
OT 1.2	"			1	111	1			1	1			1	1	11	
CC 1.5	"		1	1		111						1				
OD 6.1	"			1	111		1			1			1	1	1	
OD 6.3	"			111	1	1				1			1	1	11	
OD 6.4	"	beneath nebkha		11	111		1			1			1	1	1	

TABLE 5.5

SAMPLE	TYPE of CRUST	STRUC-TURE	GYPSUM CRYSTAL TEXTURES						CALCITE TEXTURES			QUARTZ GRAINS		PEDOFORMS		PLATE NUMBER
			LOZENGE >1mm	LOZENGE <1mm	ALABAS-TRINE	GRANU-LAR	PORPHY-ROBLASTS	FIBROUS	SPARRY	MICRITIC	PSEUDO-MORPHS	POIKIL-ITIC	FLOATING	CALCITE	GYPSUM	
DJ 5	calcrete	dune							11	111			11	1		
EM 1.1	"									111			1			
EM 1.2	"									111			11			
GO 4.1	"	pebble							111	11			11			5.68
GO 5.1	"	"							111				11			
GO 5.2	"								11	111			11			
ZIG 1.1	"								1	111			11	1		
ZIG 1.2	"								1	111			11	1		
ZIG 1.3	"	conglom-erate							11	111			11			
ZER 1.3A	"	pebble							111	1			1	1		
ZER 1.1	"			1					11	111	11		1	1	21	5.69 and 5.70
PF 1.6 *	"	pebble							11	111			11			
KF	"	pebble							111	1				1		
K 3	"	basal							111							

* differs from chemical analysis

TABLE 5.6

SAMPLE	TYPE of CRUST	STRUCTURE	GYPSUM CRYSTAL TEXTURES						CALCITE TEXTURES			QUARTZ GRAINS		PEDOFORMS		PLATE NUMBER
			LOZENGE >1mm	LOZENGE <1mm	ALABASTRINE	GRANULAR	PORPHYROBLASTS	FIBROUS	SPARRY	MICRITIC	PSEUDOMORPHS	POIKILITIC	FLOATING	CALCITE	GYPSUM	
OM (G) 1	gypsum bedrock					III			I	I		I	I			
IN 3.1	gypsified root	massive				II				III						
IN 3.2	gypsum bedrock		I	I	II	III	II		I	I	I		I			
DIF 1.1	"	bedded				III				I						
DIF 1.4	"	conglomerate				II			III	II						
DI 2.1	"					III										
DI 2.2	"					III				II						
DI 3.1	"	satin-spar				II		III								5.71
DI 3.2	"	"						III								5.72
SB 3.5	marble bedrock	pebble							II	III			I			
DI 3.3B	limestone (turonian)	cobble							III							
DI 3.4	"								I	III						
PF 4.2	clay/calcrete sodium chloride								I	III			I			
DT 1.5	dolomite (Turonian)								dolomite III	II						
DT 5.1	limestone (Turonian)	nodular overgrowths							III	II						
DT 5.2B	"								II	III			I			
DT 5.3B	"								III							

being sectioned grains. Since the thin sections represent only a small sample of the rock, whereas the chemical analyses were conducted on aliquots of larger samples, some disagreement should be anticipated. Furthermore, the chemically determined insoluble residue will in some cases include material in addition to quartz grains. An abundance of sparry or micritic calcite is defined as more than about 20% of the area of the slide; once again comparison with the chemical analyses is not always justified. Large and small lozenge, alabastrine and granular textures are similarly classified as abundant when they cover more than 20% of the slide area. Fibrous textures, being less common overall, are described as abundant if they amount to over 10% of the area. A dominance of any of these textures generally implies more than 50% of the area though in some cases, where several are abundant, the most common texture is designated dominant.

iii. Results.

The results of the micropetrographic examination are subdivided into the types and forms of gypsum crust they represent according to the preliminary assessment and listed in the same order as the foregoing site descriptions.
Since several forms are represented at certain sites, thin section data from crusts different from the characteristic form under which the site has been described above are presented after the samples from the main sites.

Evaporitic gypsum crusts show variable gypsum crystal sizes and textures within single thin sections (see table 5.1) owing to the frequently varved structure. Lozenge-shaped crystals varying in size from 3.0 mm to 0.05 mm are interbedded with alabastrine material (see plate 5.37) which may contain lozenge-shaped porphyroblasts. These may be detrital or displacive growths (see plate 5.38). Zones of granular crystals from 1.0 mm to 0.05 mm in diame-

ter are often present within the crystalline bands (see plate 5.39). Non-gypsiferous inclusions which may be present, consist of micritic calcite, often forming disrupted varves near the base of crystalline gypsum bands, and isolated floating quartz grains which do not show a preferential location. In the sample from the gypsum pan deposit near Narabeb in the Central Namib, some quartz grains are included within gypsum crystals poikilitically. It is likely that calcite was deposited from the evaporating water early in the desiccation cycle and the beds later disrupted by lenticular gypsum crystal growth (see plate 5.40). Quartz grains are probably introduced by aeolian deposition. Fibrous gypsum textures, calcite pseudomorphs and pedoforms are absent.

Gypsum croûtes de nappe composed of large lenticular crystals show considerable variability among samples from single sites (see table 5.1). While the crusts have been classified as consisting of lenticular crystals over 1.0 mm in diameter, there is a tendency for the upper portions to be mesocrystalline or sometimes alabastrine. This is evident in thin sections of crusts from El Faouar and Oued Tarfa in the Tunisian Nefzaoua. The absence of any large lozenge-shaped crystals in these thin sections is unusual. At Zaafrane and Wlotzkas Baken lozenge-shaped crystals are dominantly mesocrystalline (1.0 mm to 0.05 mm) and show signs of peripheral solution. This suggests that exposed croûtes de nappe are prone to destruction when they come in contact with meteoric waters which are not saturated with gypsum. This is well exemplified at El M'Dou. The high level crust (EM 3.9) (see plate 5.42) is composed of corroded mesocrystalline gypsum and some large lozenges with abundant alabastrine gypsum. The crust on the oued floor, however, is dominantly macrocrystalline and does not contain alabastrine material (see plate 5.43). An unusual pattern is found at Aussinanis (1MET) where large lozenges become less frequent down the section. The polygon edge (1MET 2) at this site is dominantly alabastrine, as is the surface of the buried crust nearby (1MET 3.3). This

may be the result of enhanced solution beneath the surface where the presence of sodium chloride increases gypsum solubility. Granular gypsum crystal textures are found in some thin sections but they are less frequent than in evaporitic crusts. Porphyroblasts and fibrous textures are absent.

Micritic calcite is present in many of the thin sections and, though rare, pseudomorphs are found in one sample from Zaafrane and the polygon edge at Aussinanis. This latter thin section and that of the subsurface crust (1MET 3.3) also contains small amounts of sparry calcite. The location of quartz grains within these samples signals them out as anomalies. In most cases, the desert rose croûtes de nappe contain grains poikilitically enclosed within lozenges of gypsum but in these two examples and those from El Faouar and Oued Tarfa they are absent. This is because poikilitic inclusion is evident only when crystals over 1.0 mm in diameter are present (see plate 5.43) and of these four samples only the column edge has any such crystals. It should be pointed out that while other forms of gypsum crust, such as the evaporitic and the mesocrystalline subsurface form, frequently contain large gypsum lozenges, poikilitic inclusions are very rare. Floating quartz grain textures are present in all thin sections except that of the basal crust at El M'Dou in which the whole matrix is composed of twinned lozenge-shaped crystals. Pedoforms are absent from all thin sections except that of the sample from Oued Tarfa. This fact, combined with the alabastrine texture, may justify its reclassification.

There is a striking consistency in the structure and textures of subsurface mesocrystalline crusts (see table 5.2). Of 27 samples examined in thin section, all but six have a dominantly small (up to 1.0 mm) lozenge crystal texture The exceptions are the basal crust sample from Oasis de Lesseps, which constitutes the only sample made up of large lozenges of gypsum, three surface

or exposed samples from Oudref (see plate 5.47), La Skhirra (see plate 5.48) and Kettana, which are alabastrine, a basal breccia from Djebel ed Dissa (DI 5, see plate 5.51) and a sample from Djebel el Aziza (DT 1.4, see plate 5.52). As in the case of desert rose croûtes de nappe, crystal size appears to be reduced under the influence of meteoric waters and wetting and drying cycles near the surface. The lozenges show signs of peripheral solution (see plate 5.49) and are surrounded by alabastrine material, suggesting that water undersaturated with gypsum has been introduced but has not percolated through the crust before evaporating. Rapid evaporation of thin films of water will result in precipitation of alabastrine gypsum because ionic migration to preferential growth sites and crystal surfaces will be limited. Exposed samples which are mesocrystalline rather than alabastrine, such as those from Oued Demna (see plate 5.50), probably represent recent exposures, that at Oued Demna being man-made. Large lozenges (greater than 1.0 mm) are relatively rare features of this form of crust and are restricted to either basal horizons or to subsurface crusts beneath surface forms, or calcrete in the case of GO 4.4. Their development probably required long periods of moist conditions under which ions may migrate to the growing crystal faces.

Sample OL 1.6 may constitute a desert rose form, though the absence of groundwater within 2.5 m argues against this unless water flowing in the oued is saturated with gypsum. Granular gypsum textures are also rare, being restricted to column walls and basal crust samples, suggesting that their development or preservation is dependent on moist conditions. The basal sample from Djebel ed Dissa (see plate 5.51) is made up of granular gypsum brecciating micritic Turonian limestone bedrock and is therefore anomalous. Porphyroblasts are restricted to those samples containing large, lozenge-shaped crystals, since they often represent the same feature. In the thin sections of samples from Oued Demna, small porphyroblasts (1.0 mm in diameter) are found within zones of alabastrine gypsum. Solution features indicate that they are residual

lenticular crystals. Fibrous textures are found in two thin sections from the Central Namib. The slightly curved fibres are oriented vertically and exhibit uniform extinction indicating that curvature occurred during formation (Shearman, et al. 1972). The fibrous zones are located within areas of meso-crystalline gypsum.

Calcite textures are common and predominantly micritic, though sparry crystals may be present, often in basal samples which also contain large porphyroblastic gypsum lozenges. Pseudomorphs after gypsum are found in only five thin sections. Most of the lozenge-shaped ghosts are composed of sparry calcite (see plate 5.52) and are again more common in basal crusts. This provides a clue to their origin. In some cases, lozenge-shaped gypsum crystals show partial alteration to calcite along cleavage planes, usually in the (010) plane (see plate 5.46). Sparry calcite crystal faces abut the gypsum without evidence of interstitial voids. The calcite appears to be growing into the gypsum by replacing it therefore the pseudomorphs do not represent voids created by gypsum solution later filled with calcite. It is difficult to envisage an inorganic chemical process which could account for this direct replacement; bacterial sulphate reduction is the most probable process. Ashirov and Sazonova (1962) reported alteration of gypsum and anhydrite to calcite by bacteria in the presence of hydrocarbons (see chapter two) but reduction may be effected by Desulphovibrio desulphuricans under aqueous anaerobic conditions (Butlin, 1953) liberating hydrogen sulphide. Hence, the processes involved in the formation of the pseudomorphs are the same as those Blanckenhorn (1921, pp. 203-204) postulated for Egyptian calcrete formation and Lattman and Lauffenburger (1974) invoked to explain the development of calcretes on gypsum bedrock in Nevada, though pseudomorphs were not described. The pseudomorphs are located in the lower horizons of the Tunisian and Namibian gypsum crusts because there anaerobic conditions prevail.

Floating quartz grain textures occur in all the thin sections except the granular gypsum breccia from Djebel ed Dissa, described above, and SB 3.7 which is a gypsum filled root void. This latter section contains poikilitically included quartz grains, one of only four to do so. The others are two coarsely crystalline samples from Swartbankberg (SB 3.4) and Lagunenberg (CC 1.3) and the anomalous basal crust at Oasis de Lesseps (OL 1.6). This final thin section is also one of only two of 21 from southern Tunisia which do not exhibit pedoforms (plates 5.44 and 5.45 show good examples), the other being the breccia from Djebel ed Dissa. In contrast, only one example from the Central Namib has pedoforms (HA 2.3) though SB 3.7 is itself a root fill. This difference in the crusts from the two field areas is enigmatic; it may reflect the greater abundance of rooting woody plant species in the less arid field area or, if the features are filled termite or ant tunnels, absence of certain insect species in the more arid Central Namib. The dimensions and structure of the features support a floral rather than faunal origin.

Surface crusts (see table 5.3), which are usually columnar, have marked micropetrographic similarities. Of 22 thin sections examined, all but 9 are composed predominantly of alabastrine gypsum, the exceptions comprise five of the six samples from the Central Namib which are made up of small lozenges and some alabastrine material. This suggests that if reduced crystal sizes are a result of solution of larger crystals and reprecipitation, then this process is less active in the coastal Namib than southern Tunisia. The anomalous thin sections from Tunisia are also mesocrystalline. They come from Limaguess (L) and one from Oasis de Lesseps (OL), both are sites with well developed mesocrystalline subsurface crusts. None of the thin sections contain lozenges greater than 1.0 mm in diameter. Porphyroblasts are restricted to the samples from Limaguess (see plate 5.56), Oasis de Lesseps and Lagunenberg (CC 1.6), again the sites with subsurface mesocrystalline crusts. An exception is the

crust TN 1.1. Porphyroblasts show evidence of solution, which suggests they are residual lozenges, being extant only in the least weathered crusts. Granular textures are rare, being restricted to polygon edges or column walls (see plates 5.58 and 5.63 and 5.64) and the examples of mesocrystalline surface crust, suggesting that they develop under moist conditions associated with lines of water movement or moisture retention. Only one thin section contains fibrous gypsum textures, PF 2.2 (see plate 5.55). The crystals, which are within a micritic calcite matrix, exhibit uniform extinction even in cases where the crystal is not continuous in the section. This characteristic results from very irregular surfaces on the fibres rather than calcite brecciation. The crystals are in fact continuous.

Most thin sections contain calcite textures. Many are micritic but the sections which are dominantly mesocrystalline gypsum and above subsurface crusts often exhibit sparry textures as well. Pseudomorphs after gypsum are similarly restricted to the mesocrystalline surface crusts containing sparry calcite textures (see plates 5.56, 5.59, 5.61 and 5.62). The pseudomorphic calcite is usually sparry. Quartz grains, if present, are invariably floating (see plates 5.53 and 5.57). Poikilitic grain inclusions are absent, though it is impossible to ascertain whether this represents a genetic difference from other forms of crust or simply reflects the smaller crystal size of the surface crusts. Gypsum pedoforms are present in all Tunisian surface crust sections (see plate 5.54), calcite forms are also common. The structures are noticeably more abundant in the surface crusts than in the subsurface mesocrystalline form in which they are also invariably found. Pedoforms are again absent from all but one of the sectioned samples from the Central Namib. The exception is from the Hamilton Range, the site which provided the solitary subsurface crust pedoform from the Central Namib.

Surface cobbles (see table 5.4) are commonly more alabastrine than the surface crusts, though all of the twelve samples examined in thin section contained some mesocrystalline lozenges. Three sections are dominantly mesocrystalline and one is granular. The granular sample (CC 1.5) is the only one containing any lozenges over 1.0 mm in diamater. Five samples exhibit granular textures (see plate 5.56), all except one (OT 1.2) being the more mesocrystalline examples. Porphyroblasts are rare and often show signs of considerable surface solution (see plate 5.67). Fibrous textures are absent.

Micritic calcite is found in over half of the thin sectioned cobbles. Sparry calcite is rare but one of the two samples with it present is the only cobble to contain pseudomorphs after gypsum (HA 2.1). Floating quartz grain textures are found in all the thin sections except the granular sample from Lagunenberg (CC 1.5). This is the only cobble exhibiting quartz grains poikilitically included in gypsum crystals. Most Tunisian cobbles contain abundant gypsum pedoforms (see plate 5.66) and many have calcitic forms as well. The two samples from the Central Namib do not contain pedoforms; one of these is from the Hamilton Range where the underlying crusts do have gypsum pedoforms.

Several valuable conclusions may be drawn from the micropetrographic analyses. These do not only further distinguish the different types and forms but aid the understanding of their origins and interrelationships. While evaporitic crusts show considerable variation in crystal size, owing to their varved structure, they often possess granular gypsum textures and do not contain pedoforms. Desert rose croûtes de nappe are composed of large to medium-sized gypsum lozenges, representing lenticular crystals. These frequently enclose quartz grains poikilitically. Porphyroblasts, calcite pseudomorphs after gypsum and pedoforms are all rare or absent. In contrast, subsurface mesocrystalline crusts consisting of lozenges 1.0 mm to 0.05 mm in diameter, do

not usually exhibit poikilitic quartz grain inclusion but rather floating grain textures. Gypsum pedoforms are found in all examples from southern Tunisia but are very rare in those from the Central Namib. These subsurface crusts often show a close interrelationship between a more coarsely crystalline texture and the presence of porphyroblasts, sparry calcite and pseudomorphs, suggesting that the features development depends upon similar conditions. Most probably these are prolonged moist conditions, which prevail at depth. Surface crusts differ from the mesocrystalline subsurface variety only in that they are more alabastrine and pedoforms are more abundant in Tunisia, though absent in the Namib. Porphyroblasts, sparry calcite and pseudomorphs are less common in cobbles than surface crusts and solutional features are more prevalent. This suggests a genetic interrelationship between these forms and types.

Several micropetrographic analyses of calcretes associated with gypsum crusts are presented in table 5.5. Most of these calcretes are composed of micritic calcite, though four sections through pebbles are dominantly sparry as is the basal sample from Kettana (K 3) which may represent a calcrete croûte de nappe. Most sections exhibit floating quartz grain textures and several have calcite pedoforms. The most interesting thin section, in terms of its relevance to an understanding of gypsum crusts, is ZER 1.1 (Oued Zerkine) which comes from a calcrete horizon which is overlain by a gypsum crust. This calcrete contains some small, lozenge-shaped gypsum crystals within sparry calcite walled voids (see plate 5.69) but there are also abundant calcite pseudomorphs after gypsum in similar relic voids. These represent the best developed aggregates of pseudomorphs in any of the thin sections (see plate 5.70).

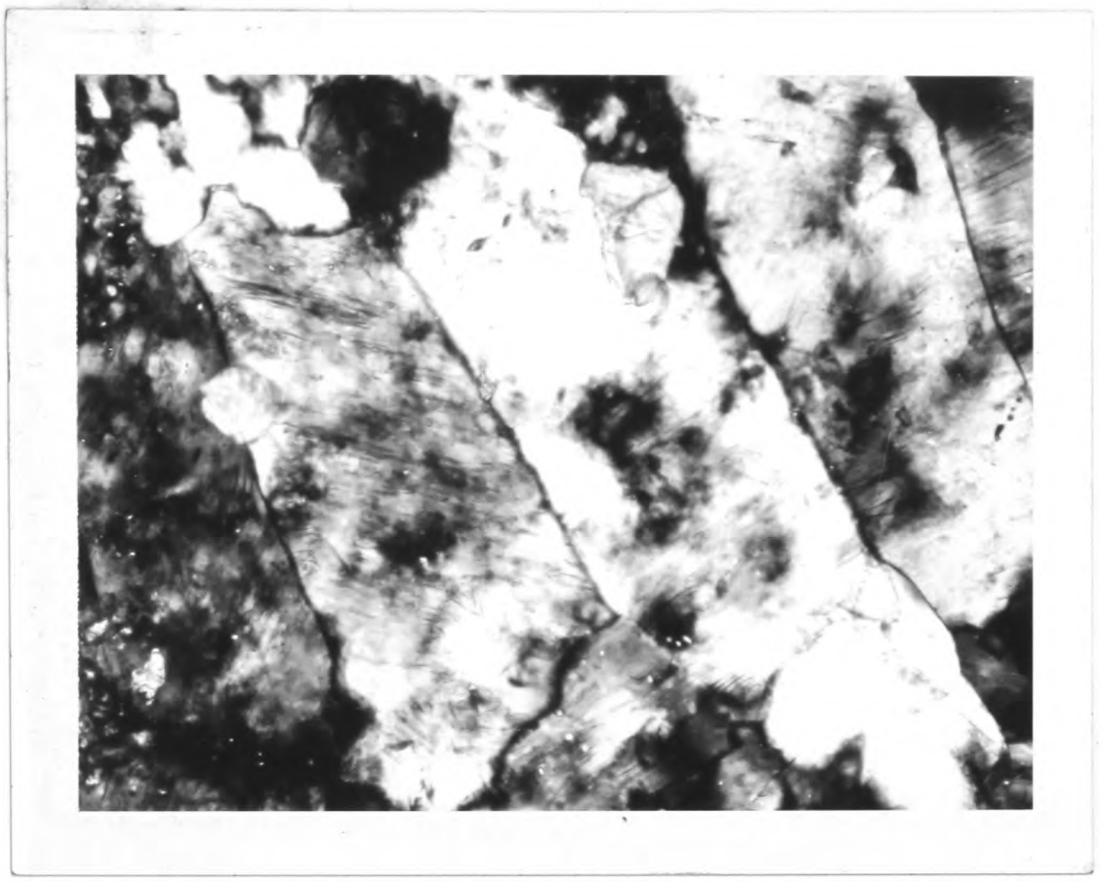
Table 5.6 gives micropetrographic information on some of the bedrock overlain by gypsum crusts. Most of the gypsiferous thin sections, all of which origi-

Plates 5.35 to 5.72

The following series of plates are photomicrographs of thin sections of gypsum crusts, calcrete and gypsum bedrock from southern Tunisia and the Central Namib Desert.
The micropetrographic characteristics of the thin sections are presented in tables 5.1 to 5.6. Unless stated otherwise the photomicrographs are correctly oriented. Unless stated otherwise they are taken under crossed-nicols.

Plates 5.35 and 5.36 (overleaf)

Two thin sections of fibrous gypsum. In plate 5.35 (above) the internal crystal structure of cleavage planes and some dark inclusions are readily identified. In plate 5.36 (below), however, the fibrous gypsum crystals have dehydrated to hemihydrate during thin section preparation. Though the minute, lath-like hemihydrate crystals are oriented in the direction of the former cleavage planes of the gypsum crystals, most of the original textural characteristics are masked. The photomicrographs are not oriented, magnification is x40.



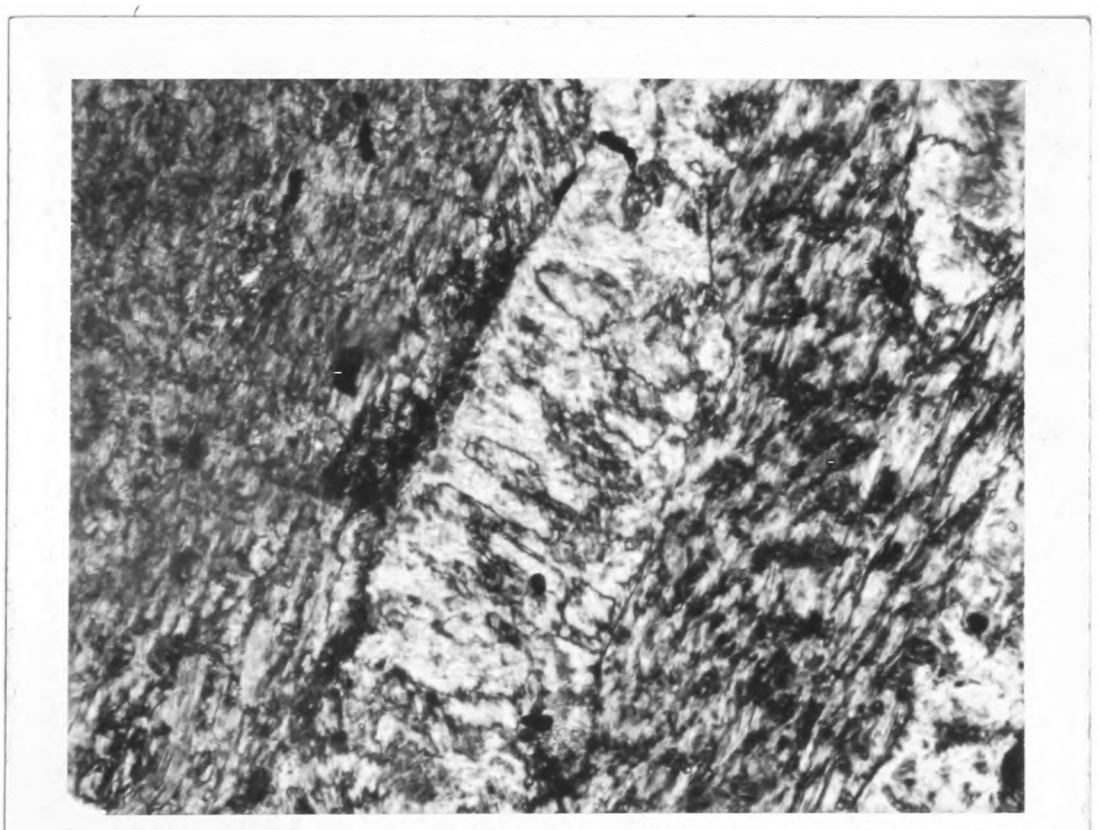


Plate 5.37 (overleaf, above)

Sample CR 1.1, x40, see table 5.7 for chemical analyses. Large, lozenge-shaped gypsum crystals forming a discontinuous horizon within an alabastrine gypsum groundmass in an evaporitic gypsum crust.

Plate 5.38 (overleaf, below)

Sample CR 1.1, x50, see table 5.7 for chemical analyses. Gypsum porphyroblast within a groundmass of alabastrine gypsum in an evaporitic gypsum crust.



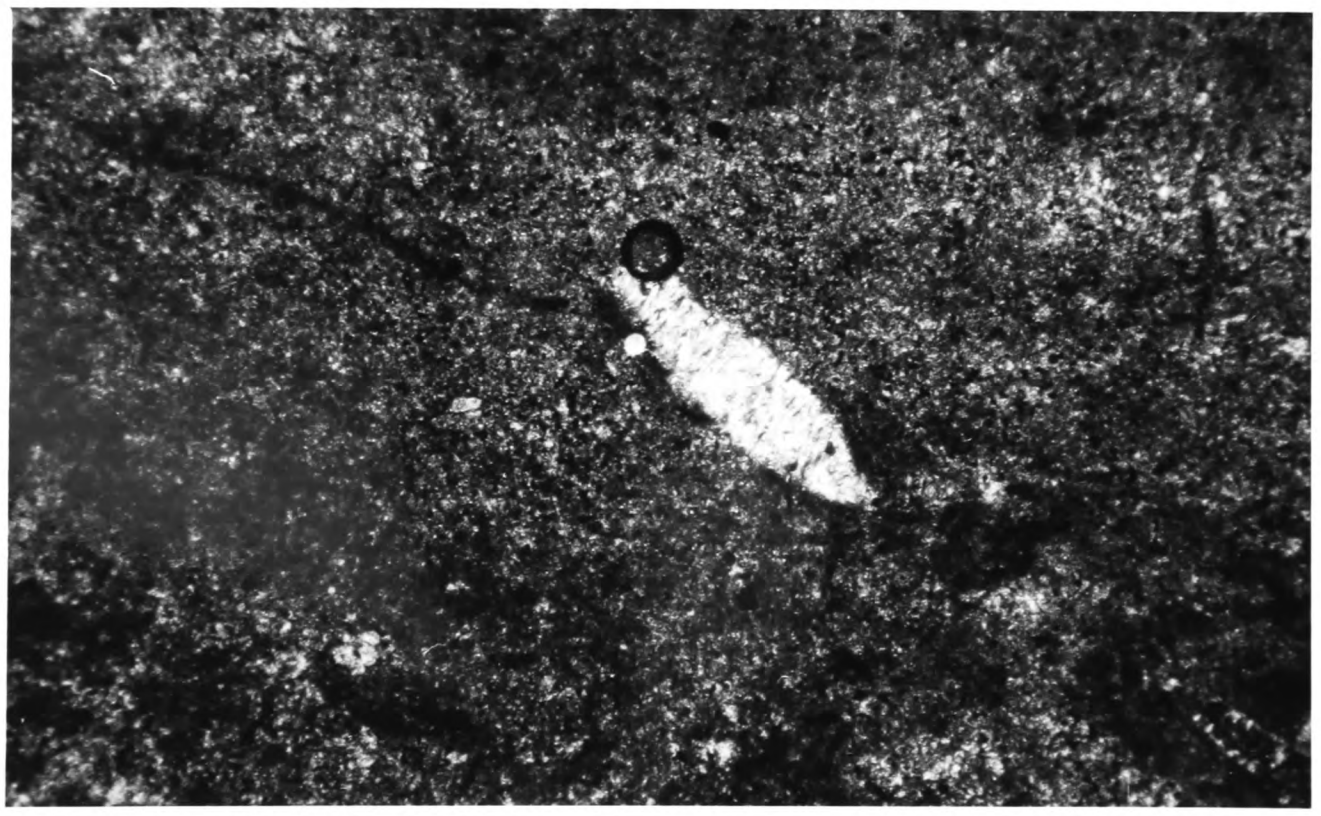
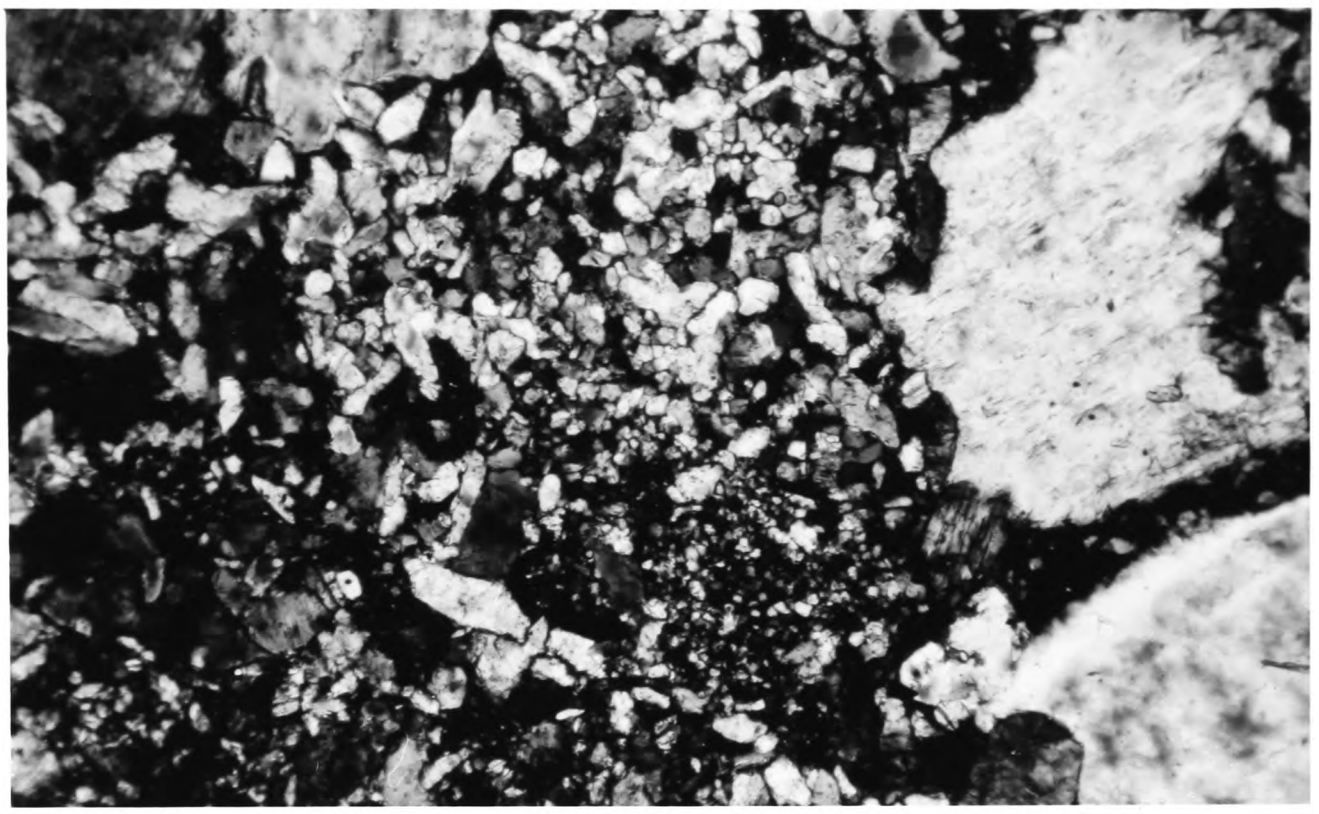


Plate 5.39 (overleaf, above)

Sample CR 1.2, x50, see table 5.7 for chemical analyses. Mass of granular gypsum crystals of markedly different sizes from an evaporitic gypsum crust.

Plate 5.40 (overleaf, below)

Sample NA 1.2, x40, see table 5.7 for chemical analyses. Quartz grains (large, irregularly-shaped white areas), micritic calcite particles (dark masses) and diffuse calcite crystals (grey areas) within a groundmass of small, white, lozenge-shaped gypsum crystals and alabastrine gypsum. The sample is from an evaporitic crust near Narabeb in the Central Namib.



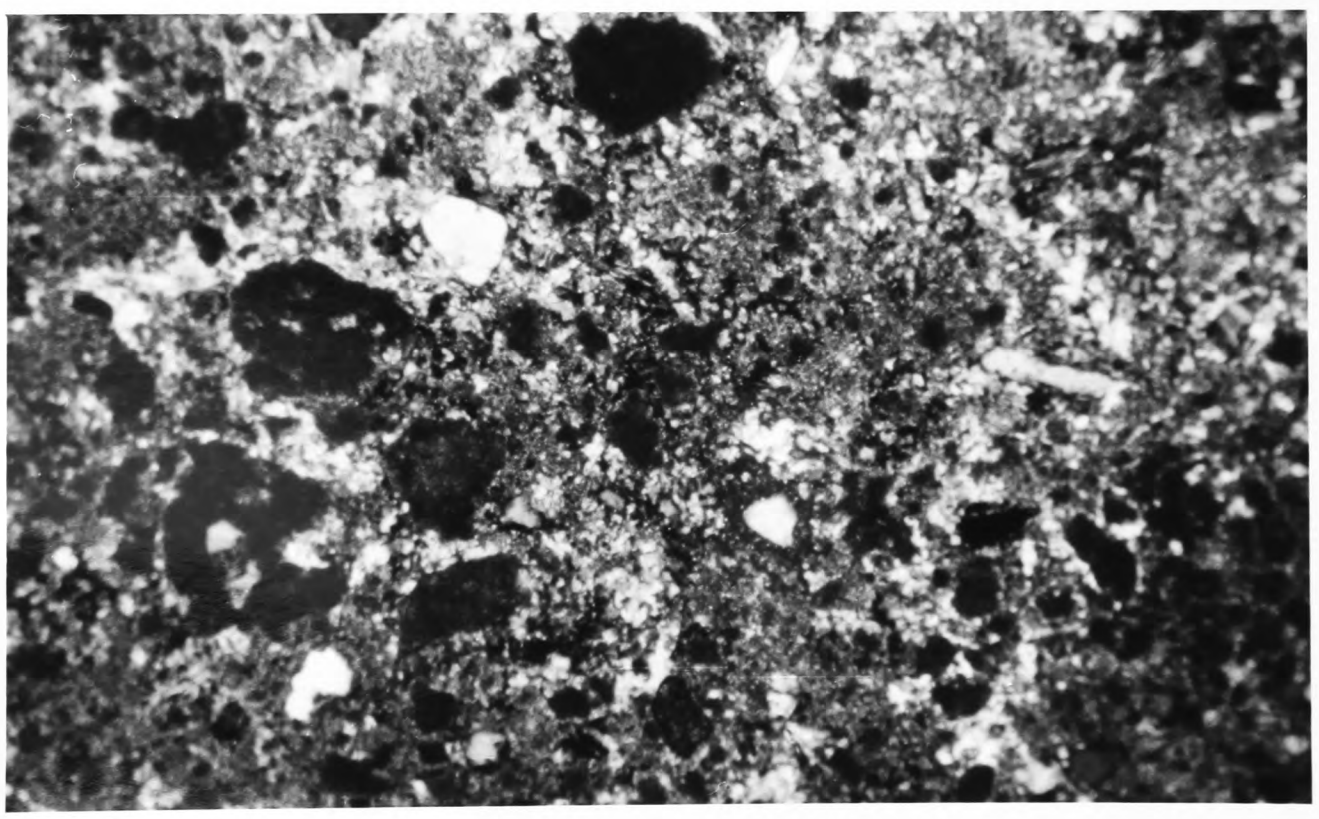
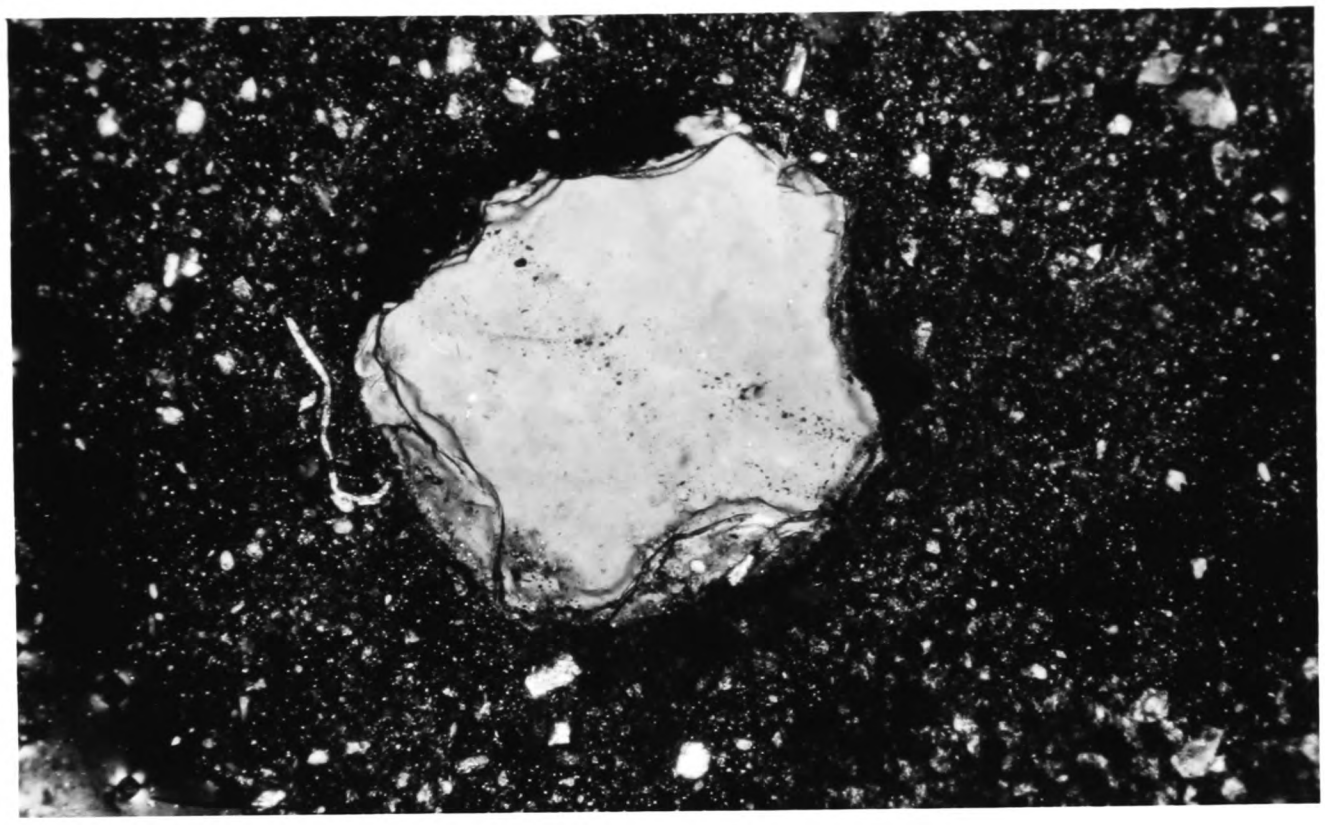


Plate 5.41 (overleaf, above)

Sample EF 1.2, not oriented, x50, see table 5.9 for chemical analyses. Large quartz grain from a gypsum croûte de nappe. The periphery of the grain shows probable solution features (left) and some optically continuous overgrowths of quartz (top). On the lower side of the grain alabastrine gypsum is infilling the gaps between the grain and detached fragments of quartz which have a slightly different extinction than the grain.

Plate 5.42 (overleaf, below)

Sample EM 3.9, not oriented, x40, see table 5.11 for chemical analyses. Masses of lozenge-shaped gypsum crystals in a groundmass of alabastrine gypsum. The mesocrystalline lozenges show evidence of solution and partial breakdown to an alabastrine gypsum texture. The sample is from the exposed croûte de nappe at El M'Dou in southern Tunisia.



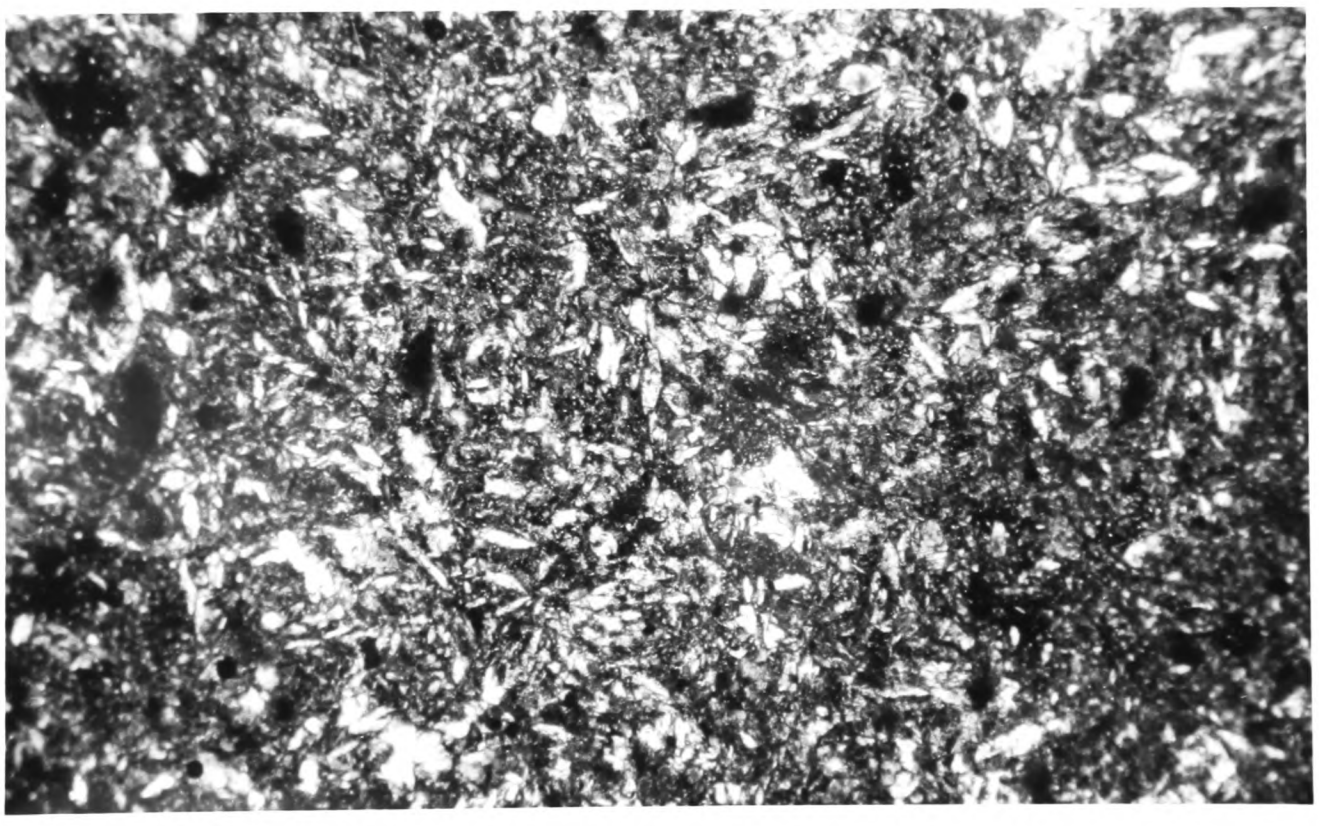
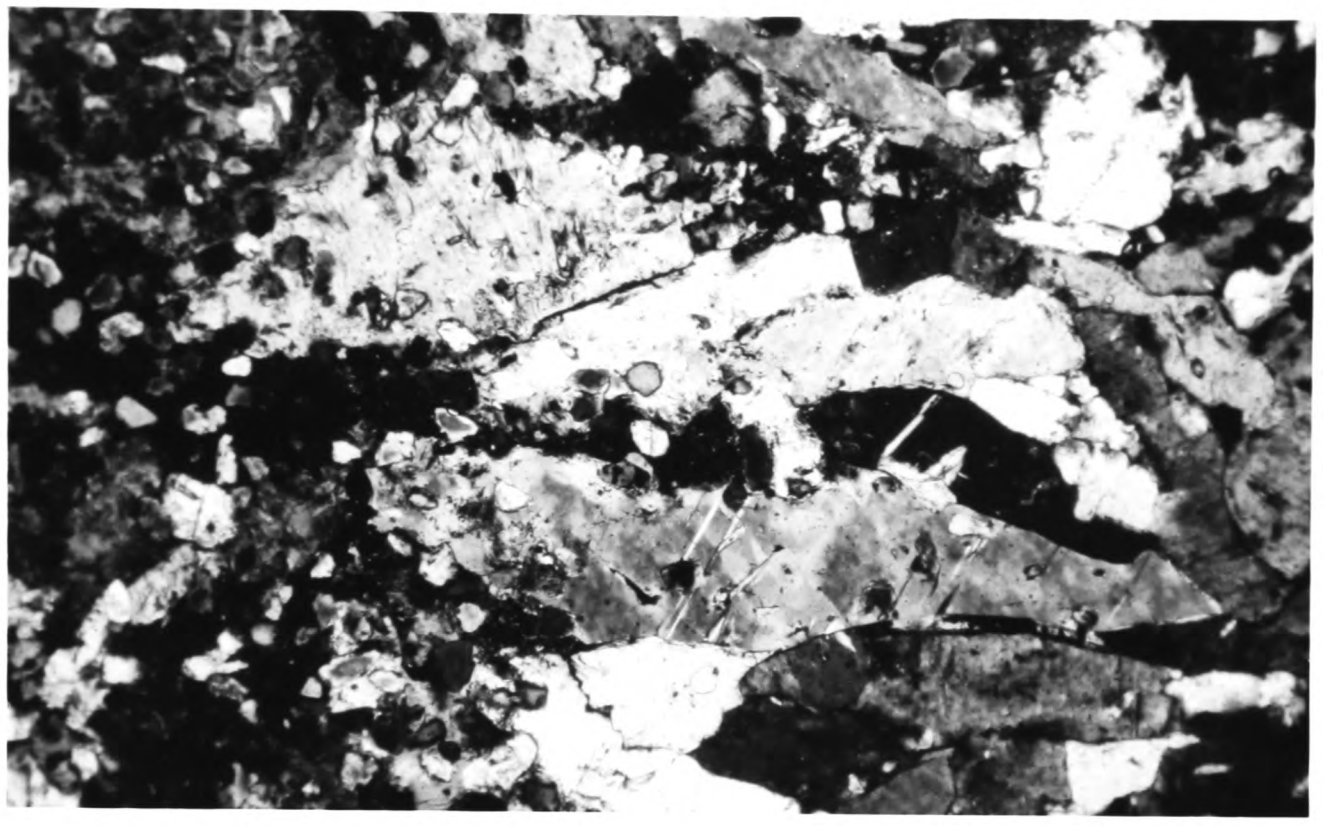


Plate 5.43 (overleaf, above)

Sample EM 3.8, not oriented, x40, see table 5.11 for chemical analyses. Very large, lozenge-shaped and irregularly-shaped gypsum crystals incorporating small quartz sand grains poikilitically.

Plate 5.44 (overleaf, below)

Sample MET 1.3, x40, see table 5.13 for chemical analyses. Gypsum pedoforms from a subsurface mesocrystalline gypsum crust. Two pedoforms consist of rings of lozenge-shaped gypsum crystals with a core of alabastrine gypsum. The bulk of the plate consists of small, lozenge-shaped gypsum crystals and quartz grains in a groundmass of alabastrine gypsum.



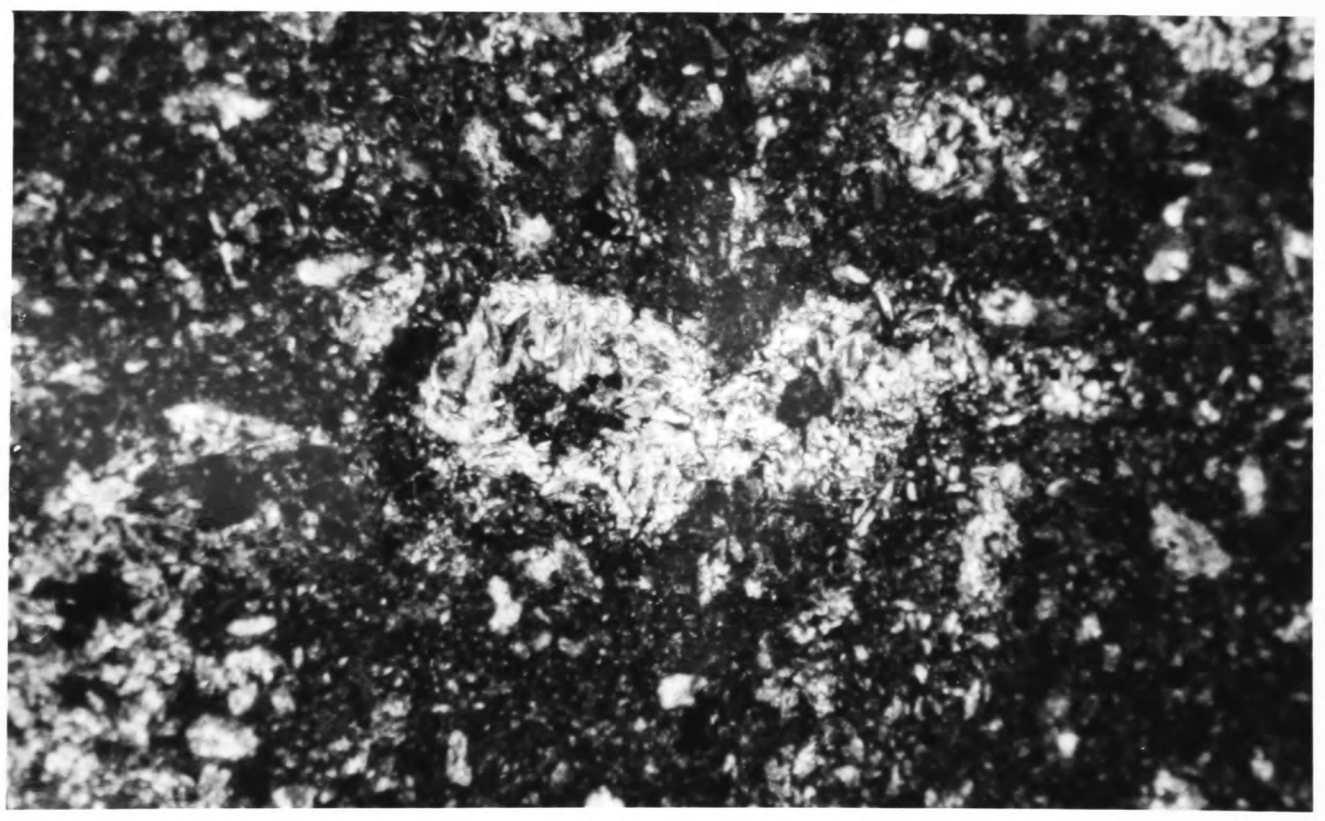


Plate 5.45 (overleaf, above)

Sample MZ 1.4, x40, see table 5.14 for chemical analyses. Pedoforms in a meso-crystalline subsurface crust from southern Tunisia. The pedoforms, within a groundmass of admixed lozenge-shaped gypsum crystals and alabastrine gypsum, comprise well-defined, lozenge-shaped gypsum crystals partly surrounded by dark bands of micritic calcite.

Plate 5.46 (overleaf, below)

Sample OL 1.4, not oriented, x50, see table 5.15 for chemical analyses. Large, porphyroblastic, lozenge-shaped gypsum crystal within a groundmass of small gypsum lozenges and alabastrine gypsum. The dark areas within the large crystals are micritic to sparry calcite which is replacing the gypsum along the cleavage plane (011) (see fig. 1.11).

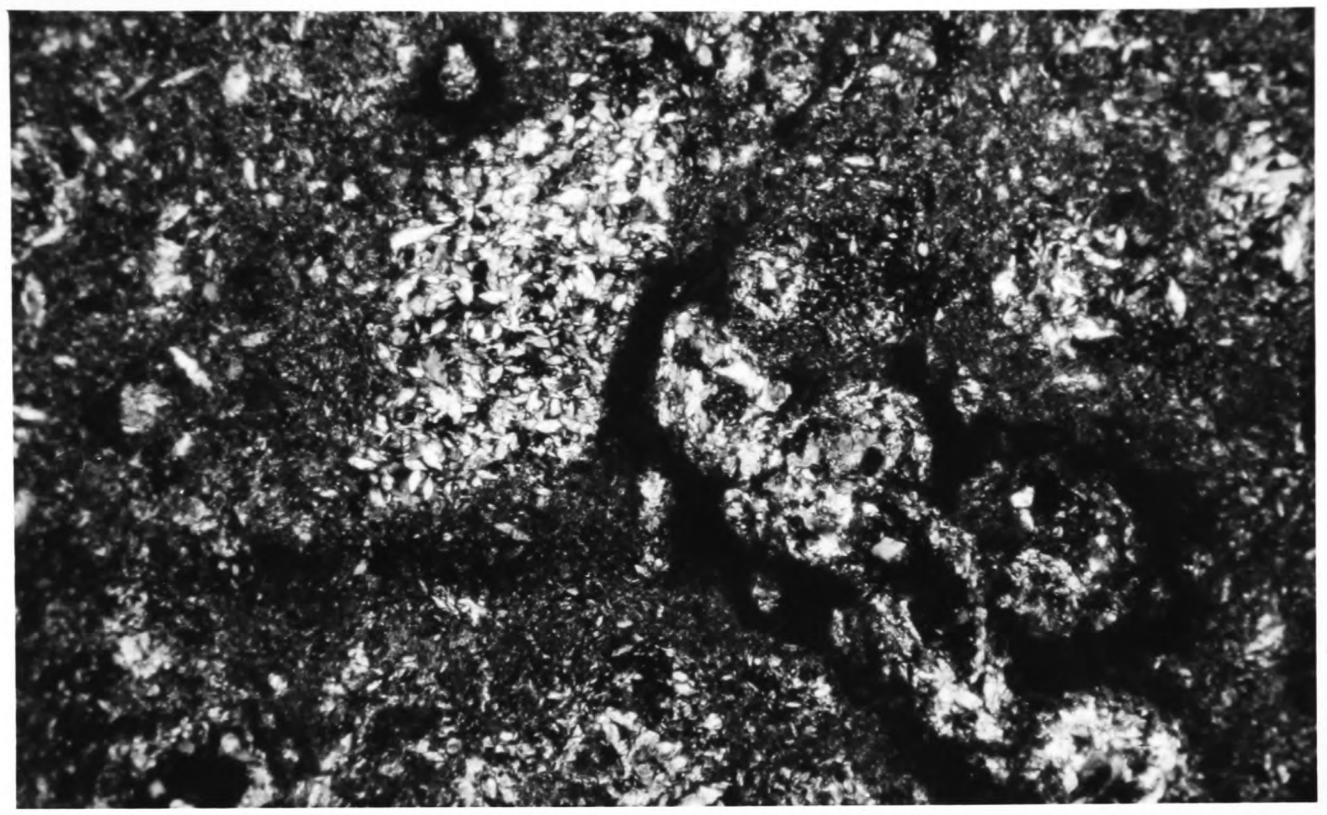


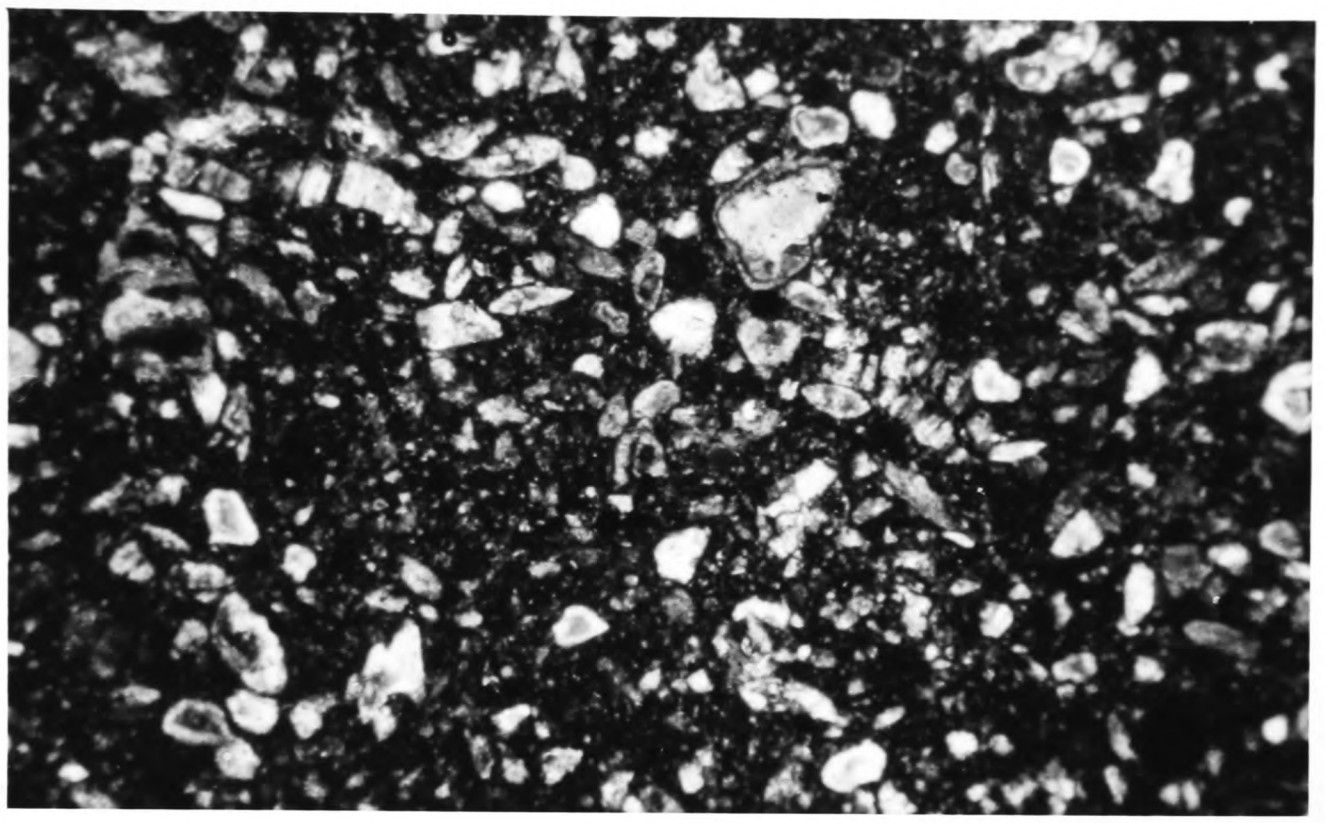


Plate 5.47 (overleaf, above)

Sample OU 2.1, x40, see table 5.16 for chemical analyses. Quartz grains floating within a groundmass of alabastrine gypsum in a sample from the top of a meso-crystalline subsurface gypsum crust.

Plate 5.48 (overleaf, below)

Sample LS 1.2, x40, see table 5.17 for chemical analyses. Alabastrine gypsum with small, lozenge-shaped gypsum crystals along fissures in the subsurface crust sample. Note two irregularly-shaped quartz grains left of centre.



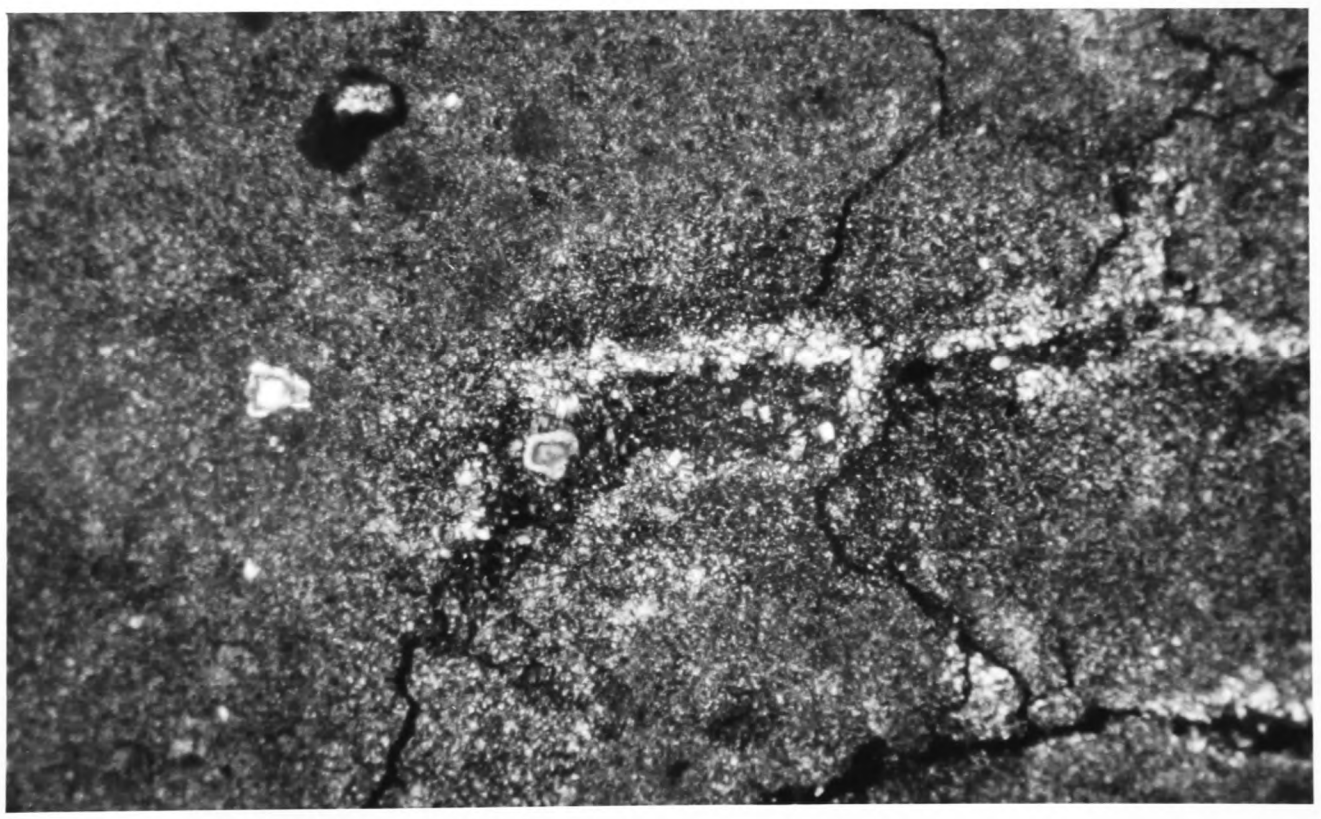
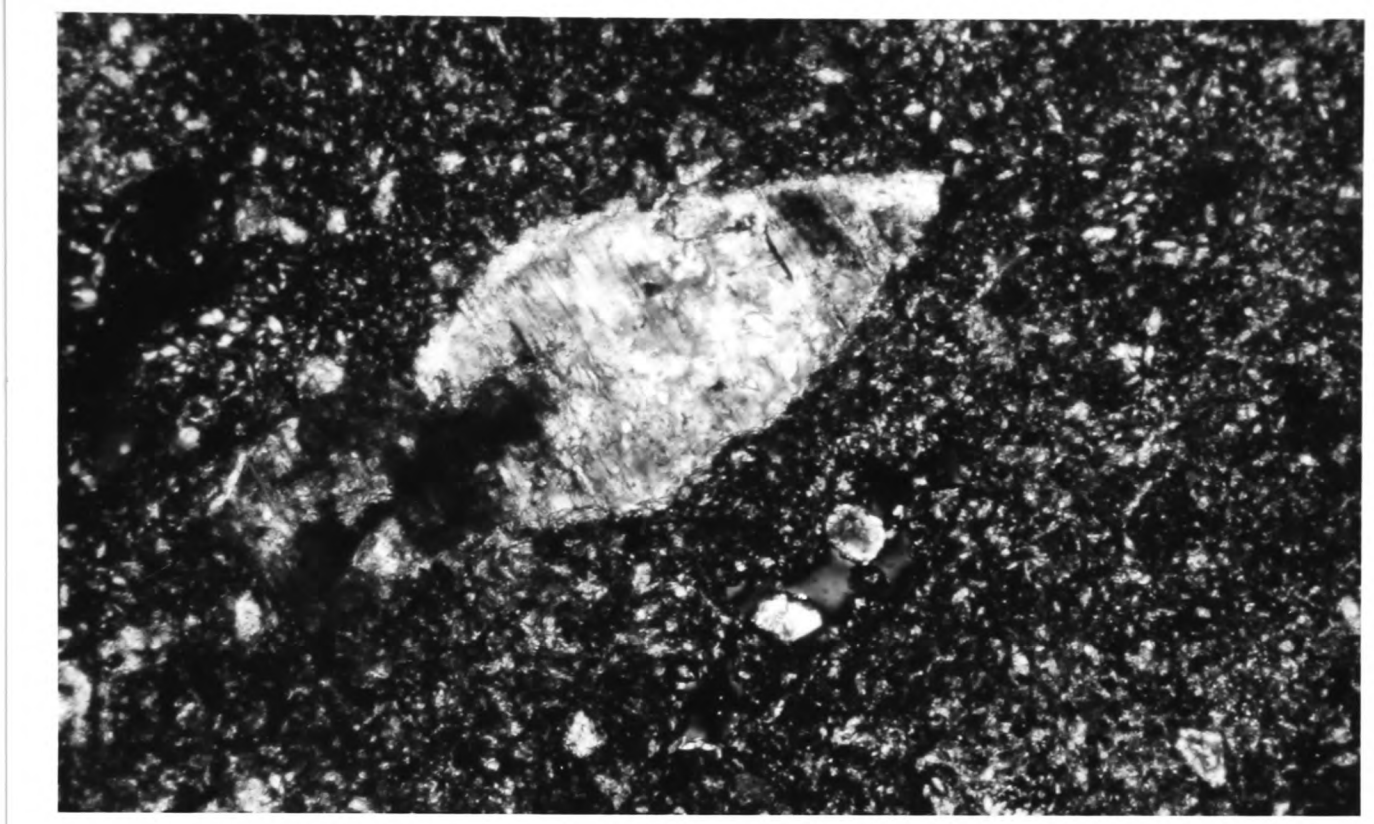


Plate 5.49 (overleaf, above)

Sample LS 1.3, x64, see table 5.17 for chemical analyses. Large, lozenge-shaped gypsum porphyroblast in a groundmass of small, lozenge-shaped gypsum crystals and alabastrine gypsum. Note the poorly defined periphery to the porphyroblast, a result of gypsum dissolution. The sample is from a mesocrystalline subsurface crust in southern Tunisia.

Plate 5.50 (overleaf, below)

Sample OD 1, x64, see table 5.22 for chemical analyses. Pedoform defined by aggregates of lozenge-shaped gypsum crystals around a core of alabastrine gypsum. The groundmass consists of small, lozenge-shaped crystals and alabastrine gypsum. The sample is from a Tunisian subsurface mesocrystalline gypsum crust.



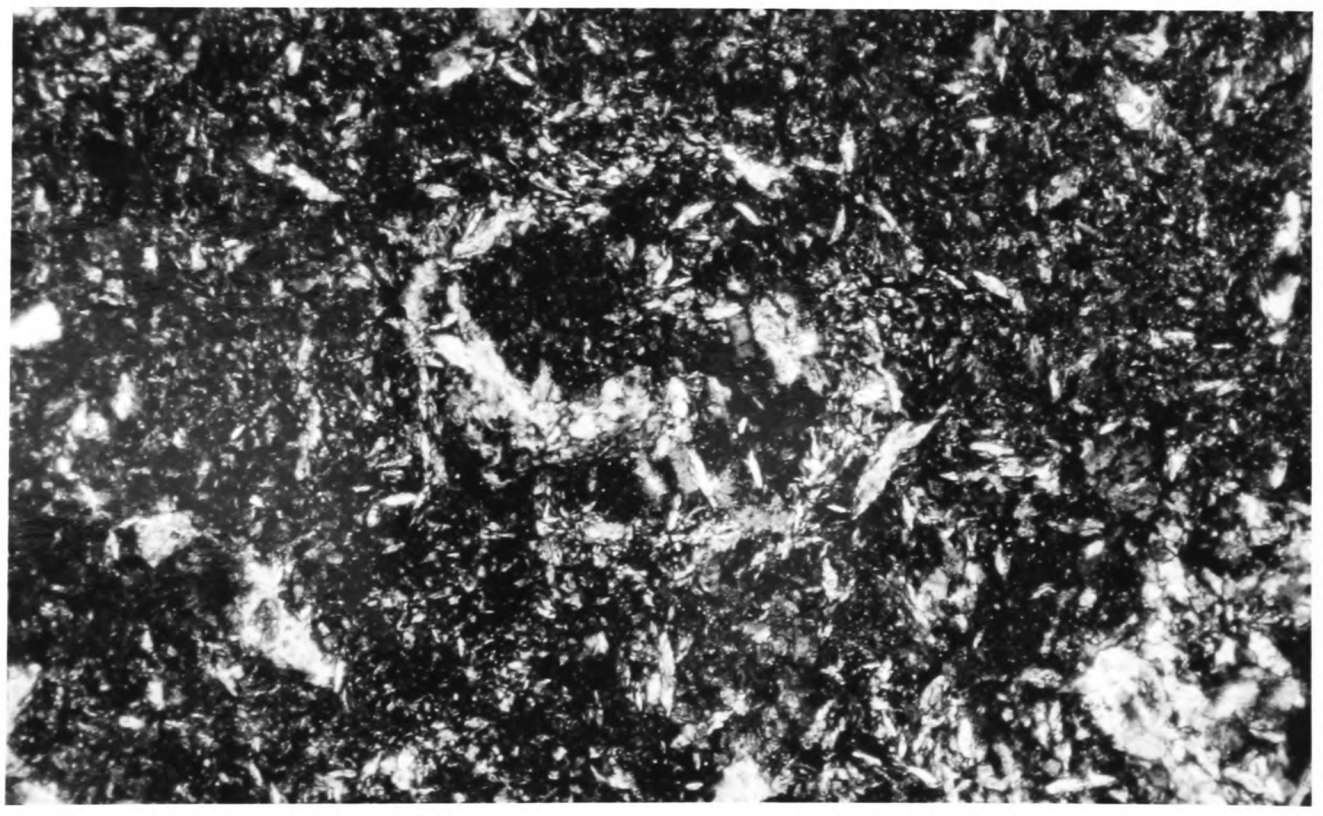
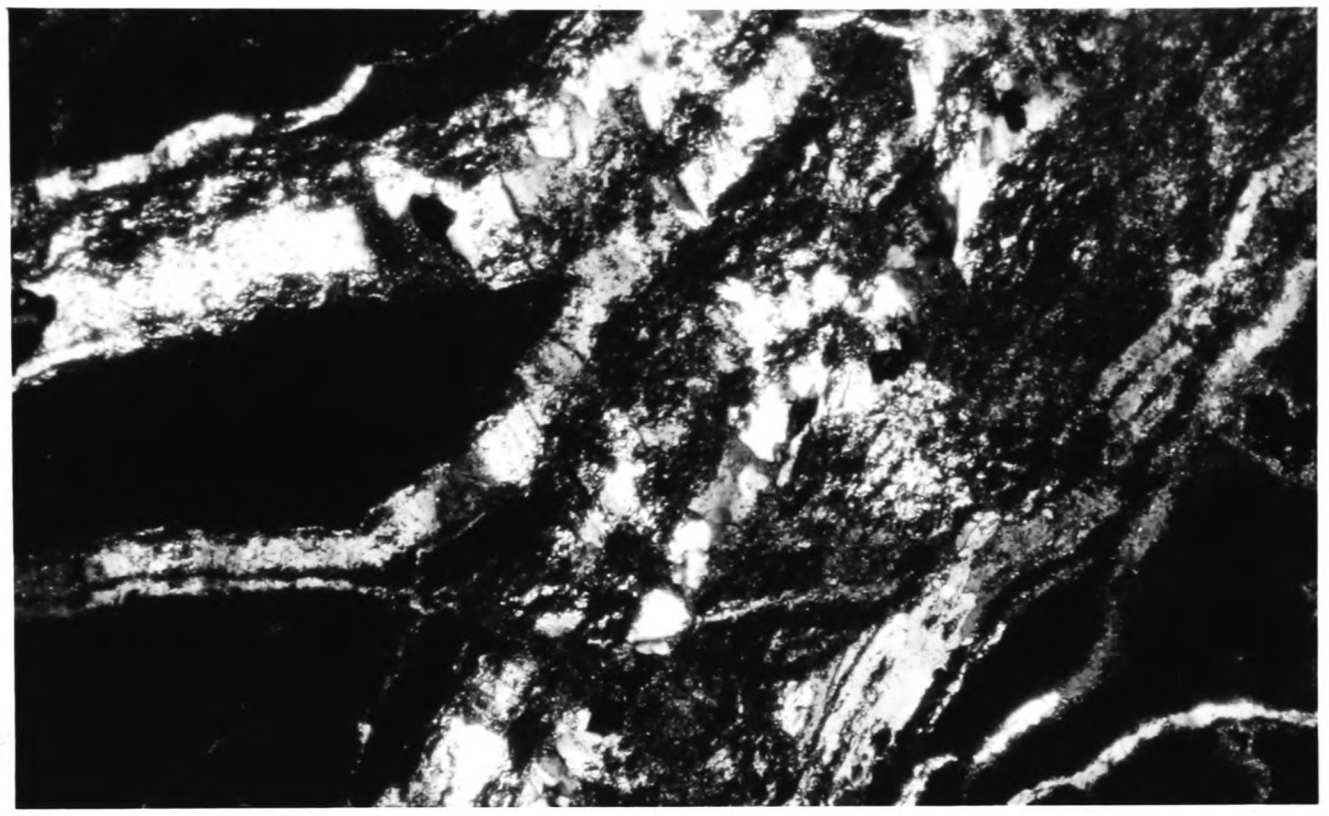


Plate 5.51 (overleaf, above)

Sample DI 5, x64. Bands of granular gypsum crystals brecciating dark areas of micritic calcite (Turonian limestone). Detached bands of micrite within the gypsum crack-fill may indicate displacive crystallization of gypsum within insipient fissures.

Plate 5.52 (overleaf, below)

Sample DT 1.4, x64, plain light, see table 5.31 for chemical analyses. Zone of aggregated lozenge-shaped gypsum crystals within a groundmass of dominantly alabastrine gypsum. Note particularly the complete replacement of some of the gypsum lozenges by sparry calcite forming the darker, lozenge-shaped calcite masses. These are pseudomorphs after gypsum. The sample comes from a subsurface crust in southern Tunisia.



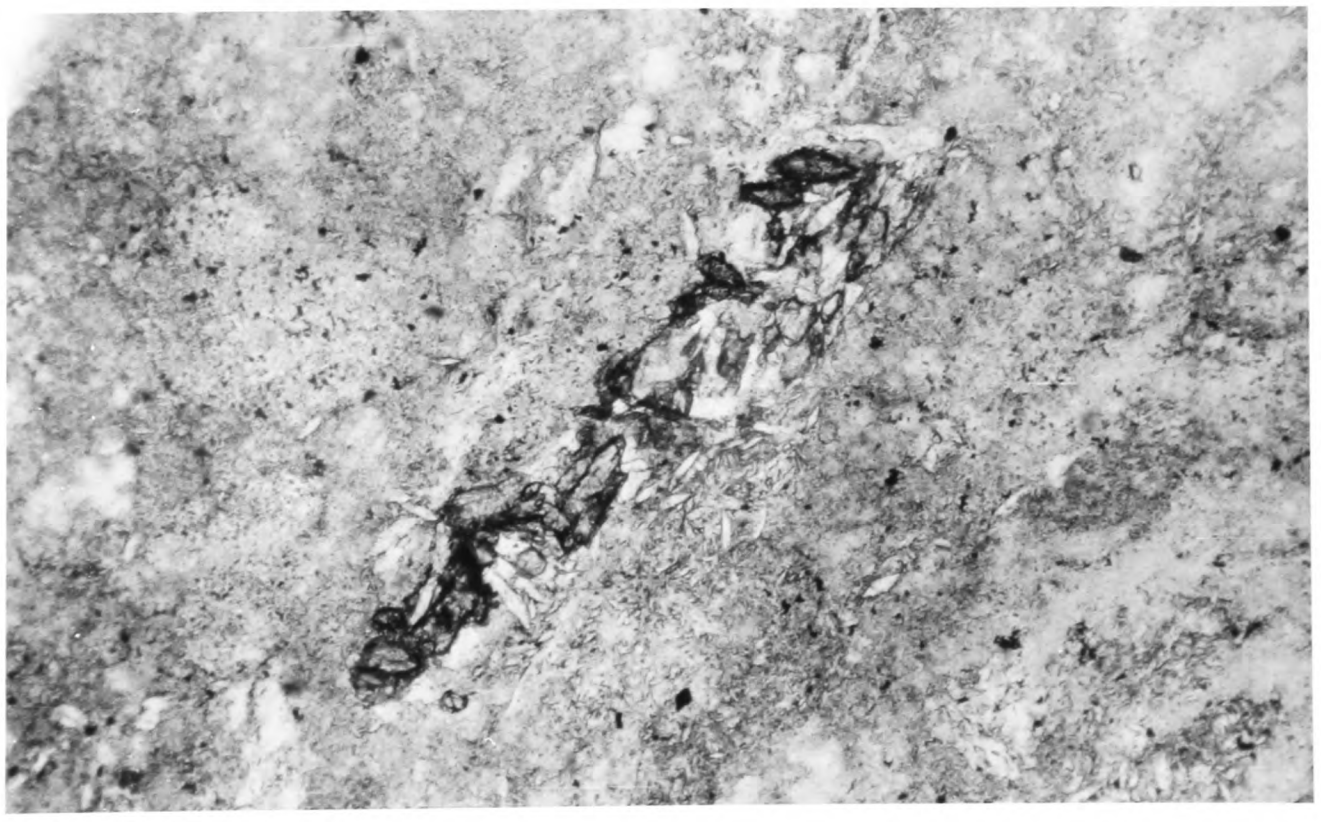
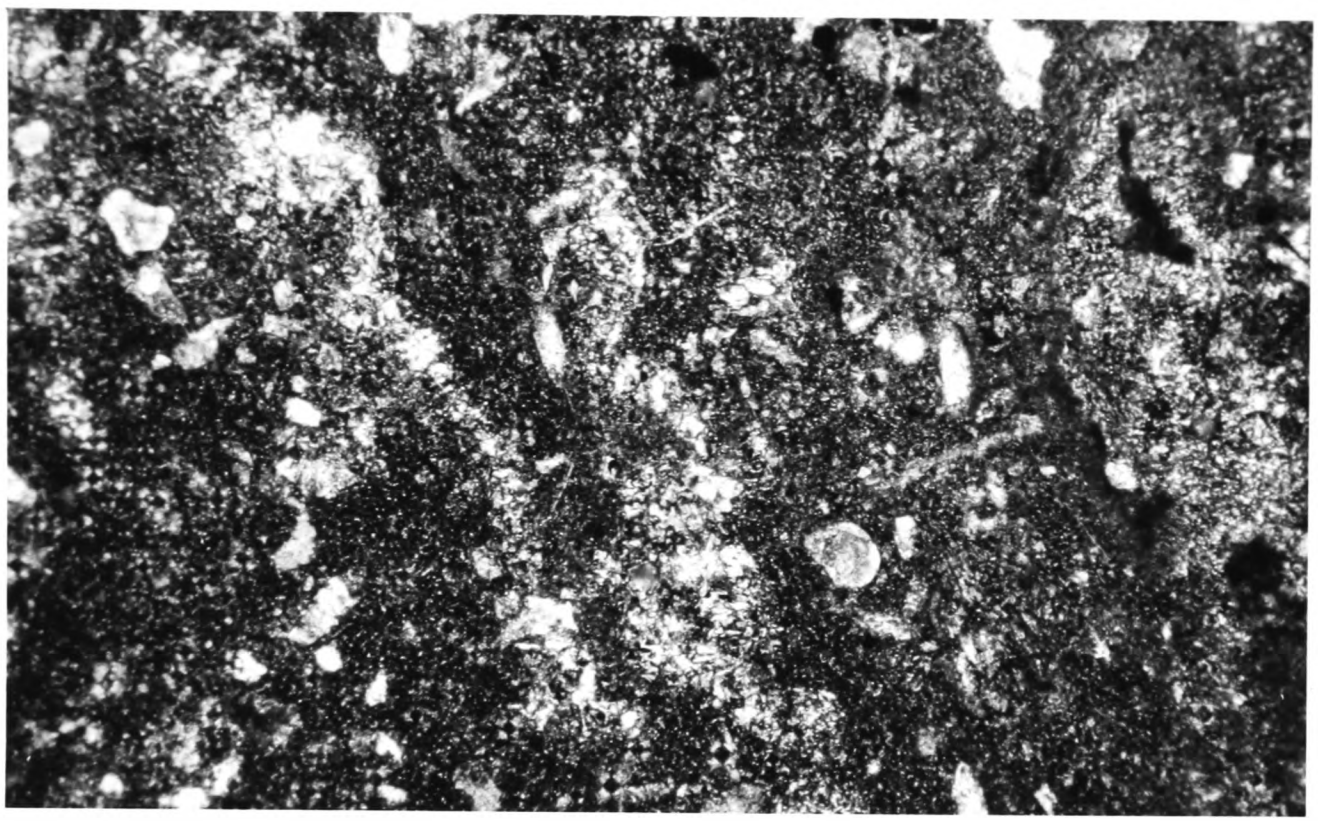


Plate 5.53 (overleaf, above)

Sample EH 1, x40, see table 5.26 for chemical analyses. Gypsum pedoforms composed of small, lozenge-shaped crystals (white bands) in a groundmass of alabastrine gypsum with sub-angular to sub-rounded quartz grains. The sample is from a columnar surface crust.

Plate 5.54 (overleaf, below)

Sample DI 1, x40, see table 5.28 for chemical analyses. Pedoform composed of small lozenge-shaped gypsum crystals. The groundmass consists of small, lozenge-shaped crystals and alabastrine gypsum. The sample is from a columnar surface crust.



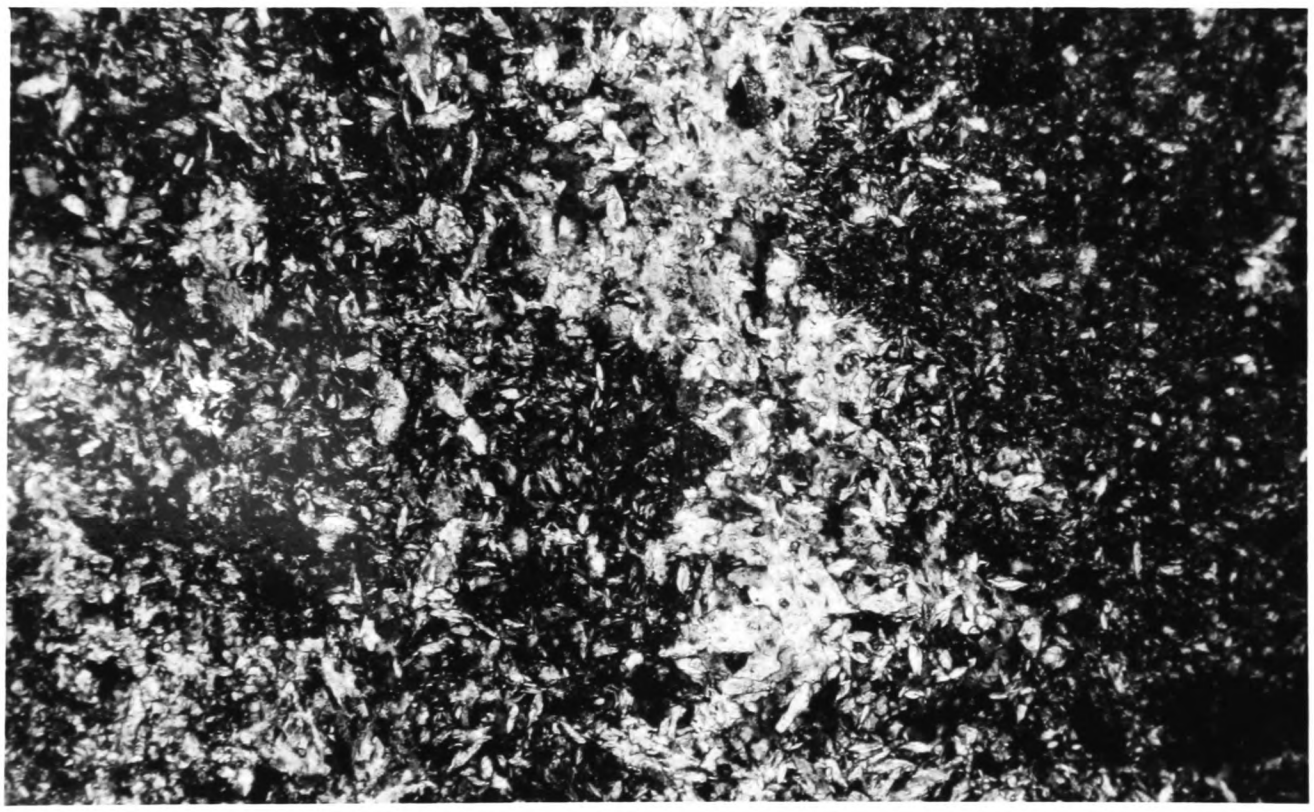


Plate 5.55 (overleaf, above)

Sample PF 2.2, x40. Fibrous gypsum crystals in a dark groundmass of clay and calcite. Though the crystals are not conjoined from top to bottom in the photograph, they are optically continuous. The sample is from the Central Namib Desert (see plate 5.32).

Plate 5.56 (overleaf, below)

Sample L 1, not oriented, x64, see table 5.14 for chemical analyses. Large, lozenge-shaped porphyroblast in a groundmass of small, lozenge-shaped gypsum crystals and darker masses of micritic calcite. The gypsum porphyroblast is partially altered to calcite along the 011 cleavage plane and is an insipient pseudomorph. The sample comes from a columnar surface crust.



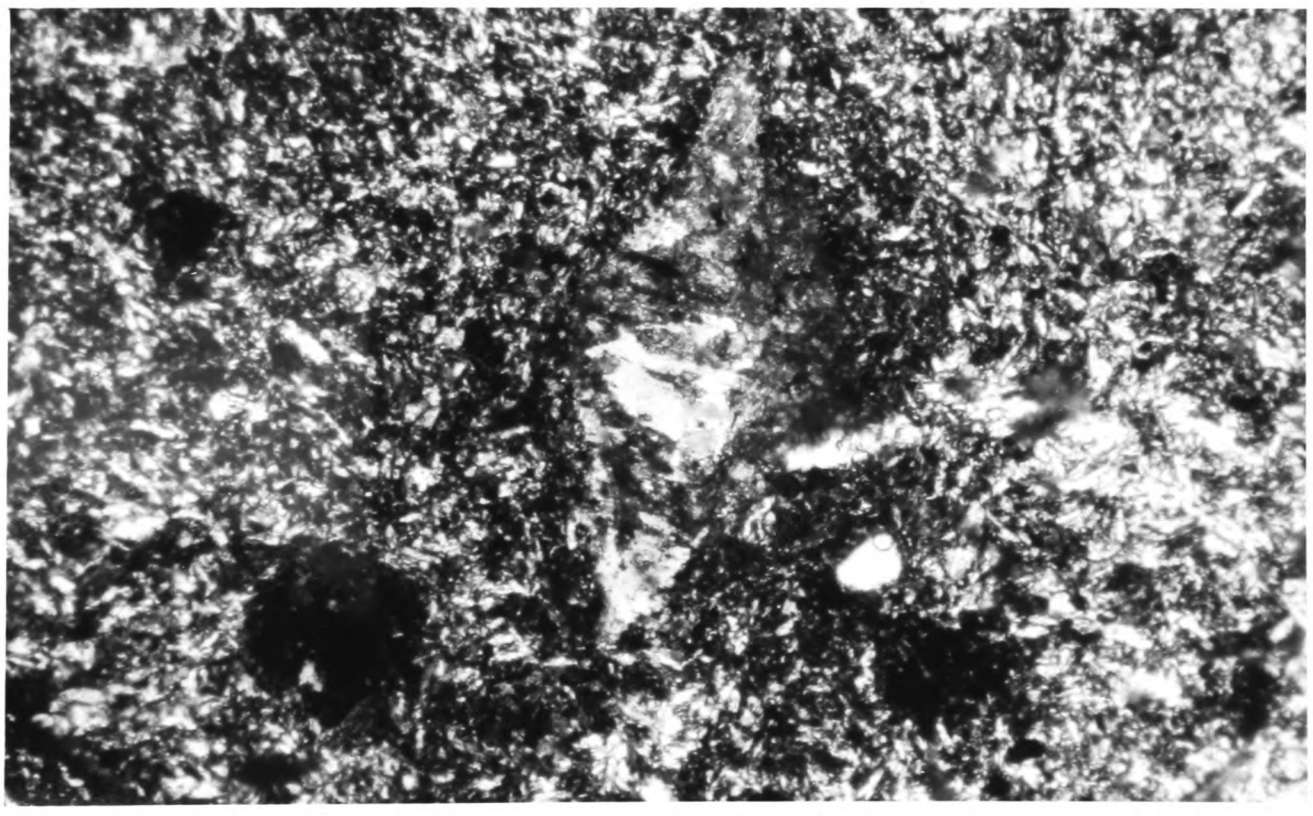
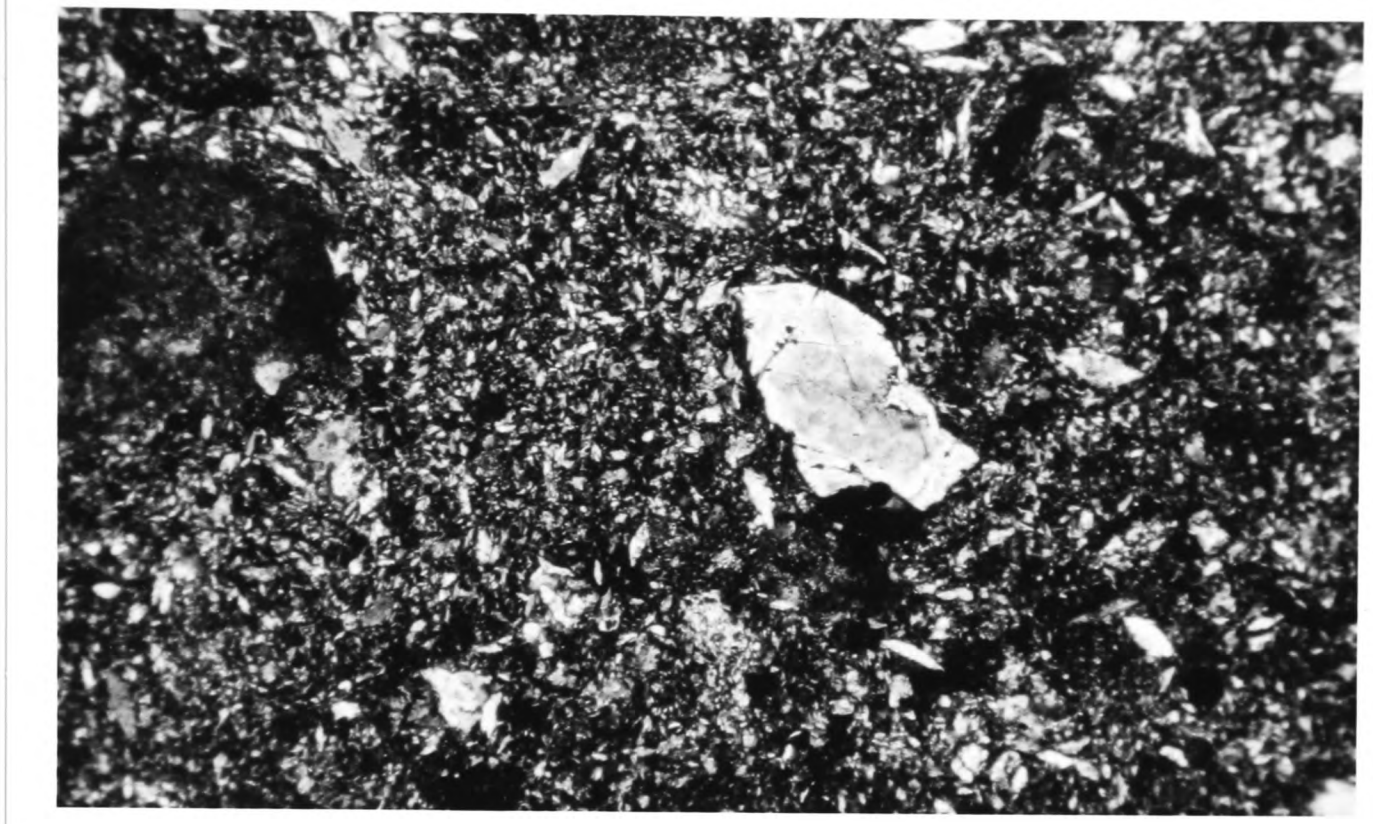


Plate 5.57 (overleaf, above)

Sample L 2, x50, see table 5.14 for chemical analyses. Large quartz grain in a groundmass of alabastrine gypsum and small, lozenge-shaped gypsum crystals forming a columnar surface crust. Note the solutional pitting on the quartz grain and also the large, dark calcite pedoform to the left.

Plate 5.58 (overleaf, below)

Sample K 6, x40, see table 5.18 for chemical analyses. The edge of a column from a surface gypsum crust showing alternating vertical bands of alabastrine gypsum and small, lozenge-shaped gypsum crystals. Some bands consist of gypsum with a granular texture (top centre).



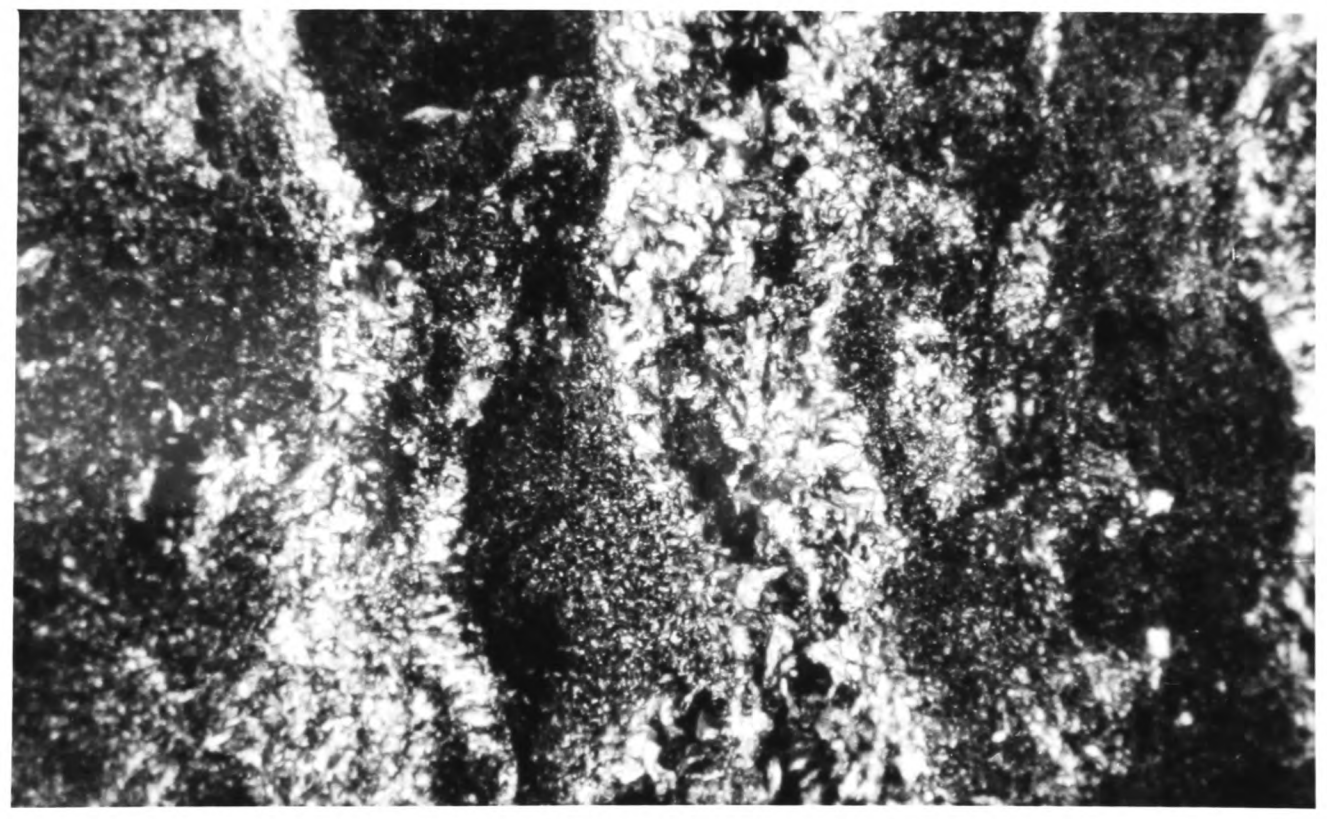
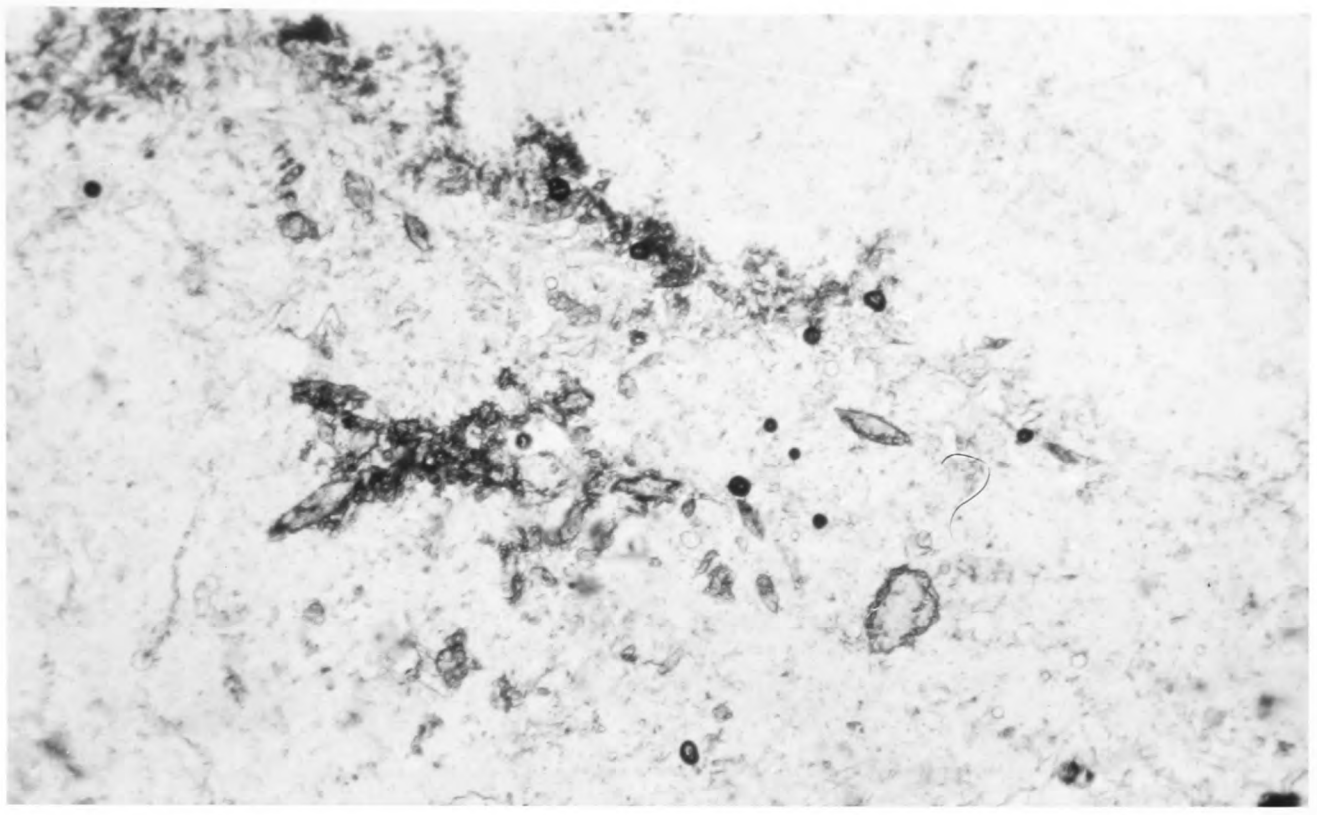


Plate 5.59 (overleaf, above)

Sample SB.3.8, x64, plain light, see table 5.20 for chemical analyses. Surface gypsum crust sample composed of alabastrine gypsum and small, lozenge-shaped gypsum crystals showing partial or surficial alteration to sparry calcite (darker areas). Several small pseudomorphs after gypsum are in evidence.

Plate 5.60 (overleaf, below)

Sample OD 6.2A, x40, see table 5.23 for chemical analyses. Horizontal banding at the surface of an indurated gypsum crust. The upper alabastrine gypsum zone has a discontinuous band of micritic calcite which further hardens the surface of the crust. Beneath the alabastrine rind, the crust is composed of small, lozenge-shaped gypsum crystals.



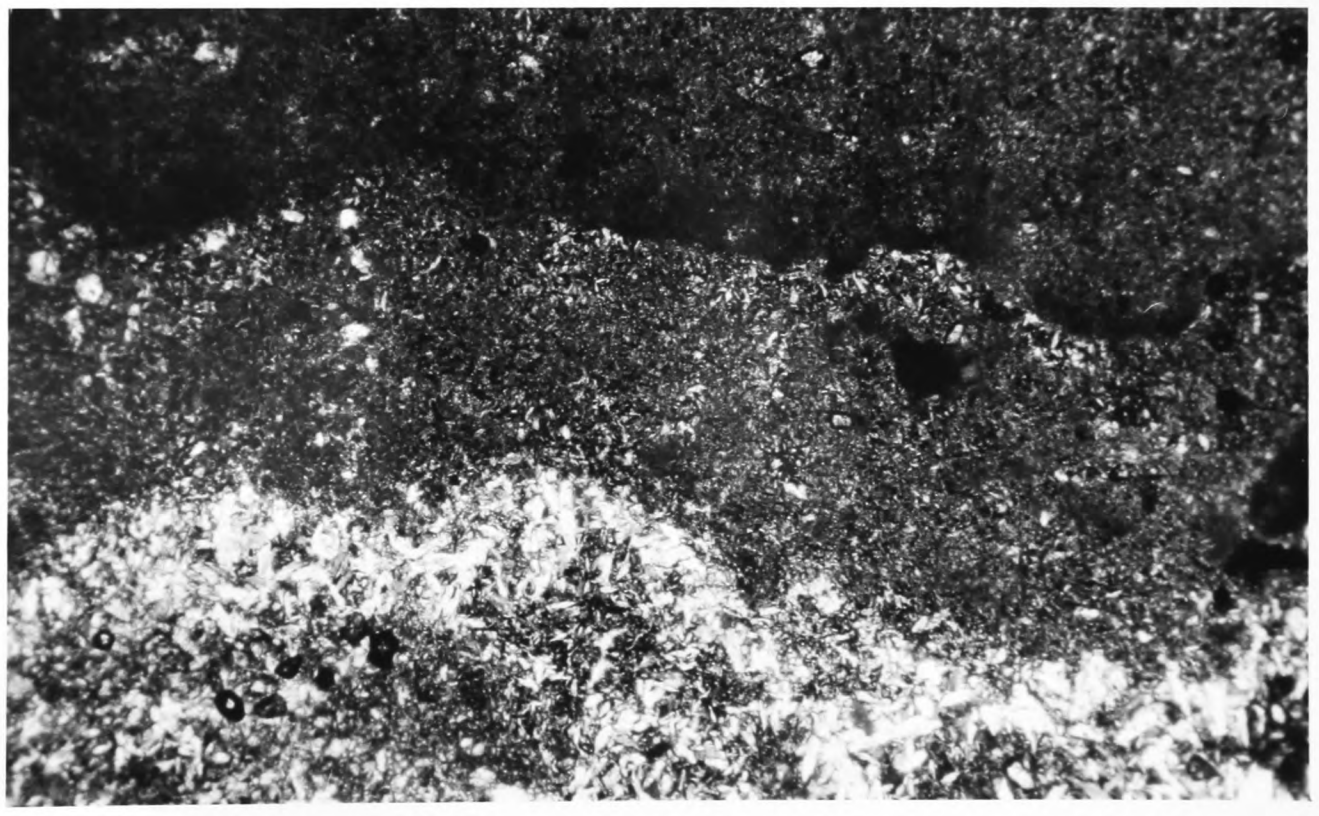
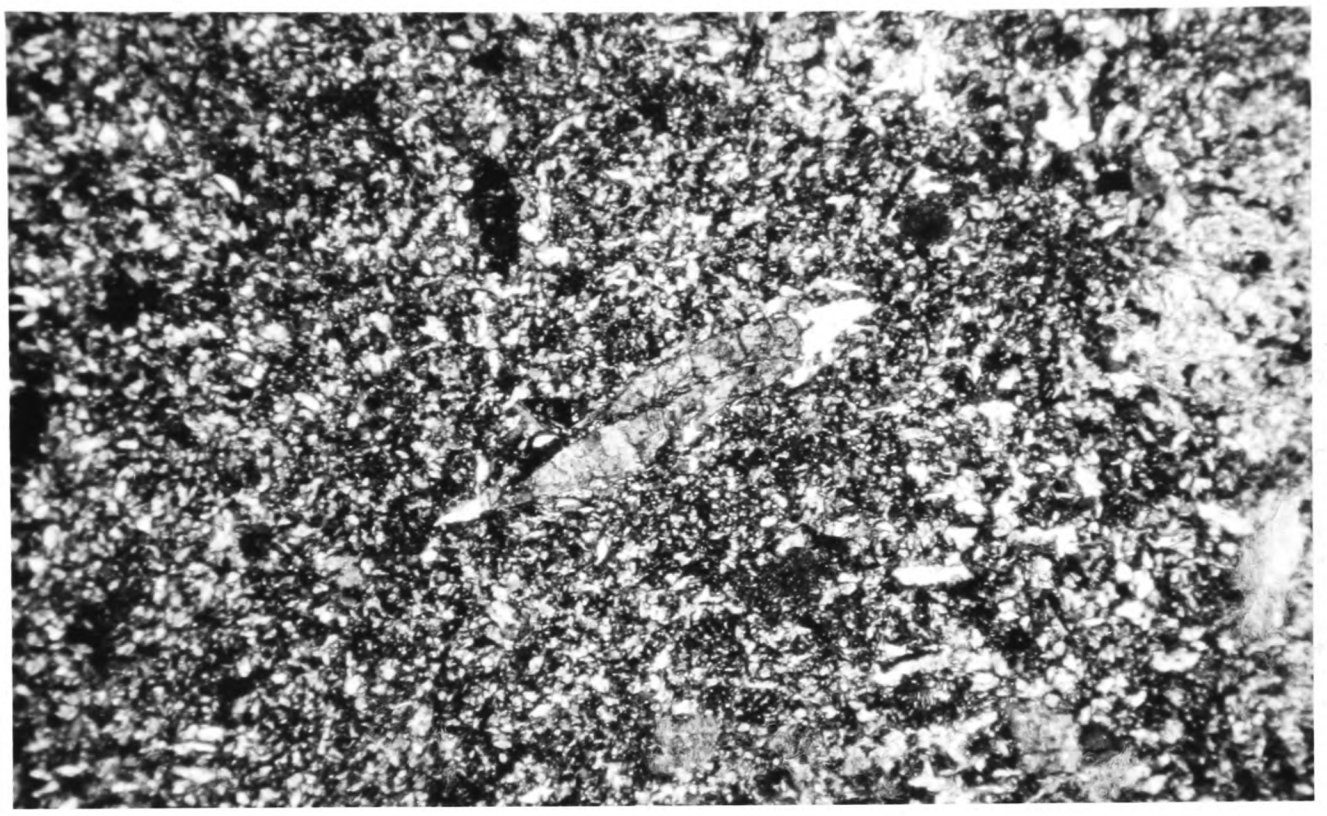


Plate 5.61 (overleaf, above)

Sample HA 2.2, oriented 45° clockwise of vertical, x50, see table 5.34 for chemical analyses. Large, lozenge-shaped porphyroblast in a groundmass of alabastrine gypsum and lozenge-shaped gypsum crystals. The porphyroblast is almost completely altered to micritic calcite, only the two ends of the lozenge are still composed of the original gypsum. The sample is a gypsum nodule from within a powdery surface crust in the Central Namib Desert.

Plate 5.62 (overleaf, below)

As above but under plain light. Note additional calcite aggregates (dark areas).



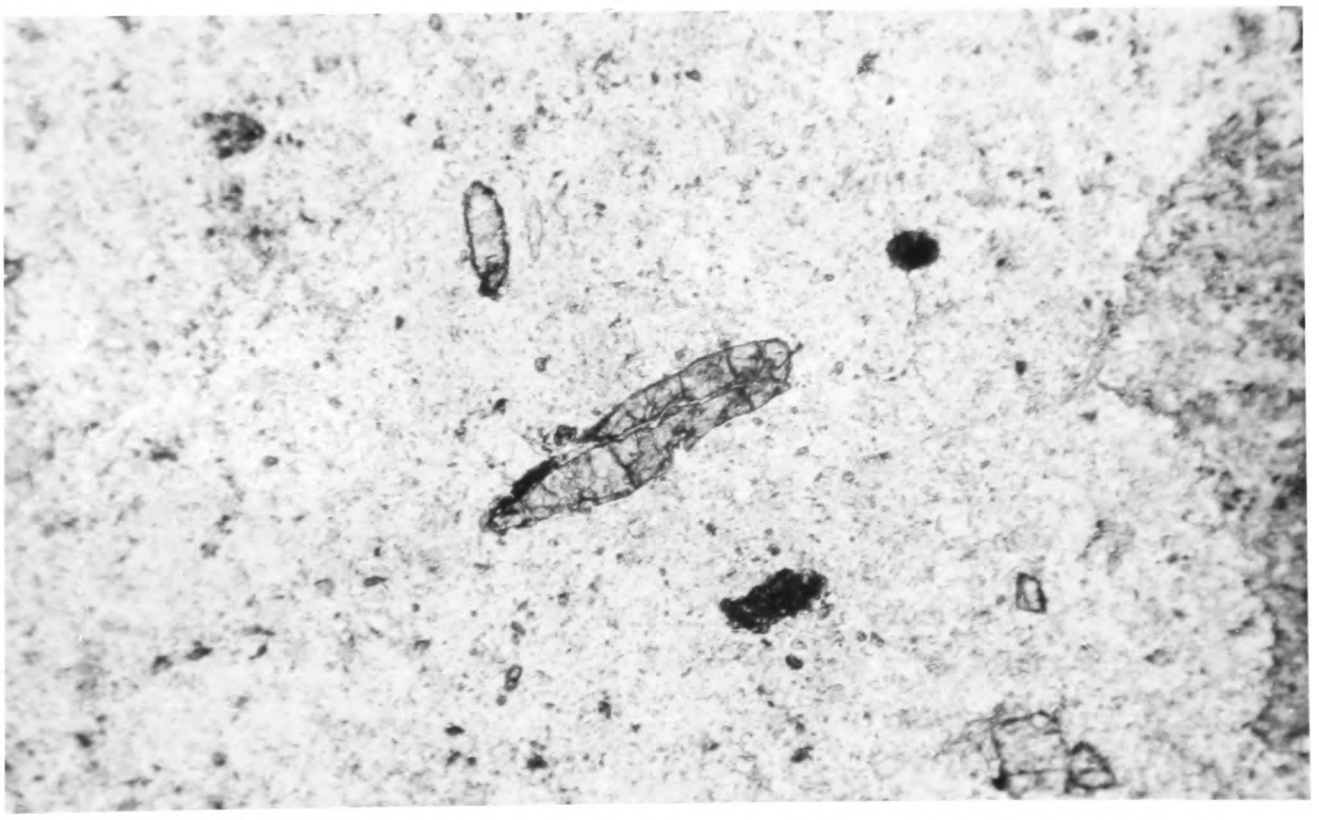


Plate 5.63 (overleaf, above)

Sample L 4, x40, see table 5.14 for chemical analyses. Column wall from a mesocrystalline subsurface crust. Note the dark zones of clayey material on either side of the band of granular gypsum crystals. Bands of small, lozenge-shaped gypsum crystals separate the granular and opaque zones.

Plate 5.64 (overleaf, below)

Sample L 4, x40, see table 5.14 for chemical analyses. As above but note also the displacive nature of the band of lozenge-shaped gypsum crystals between the dark zone of clay and micrite to the left and the zone of large granular gypsum crystals to the right. The small lozenges are well-formed and extend into the granular band.

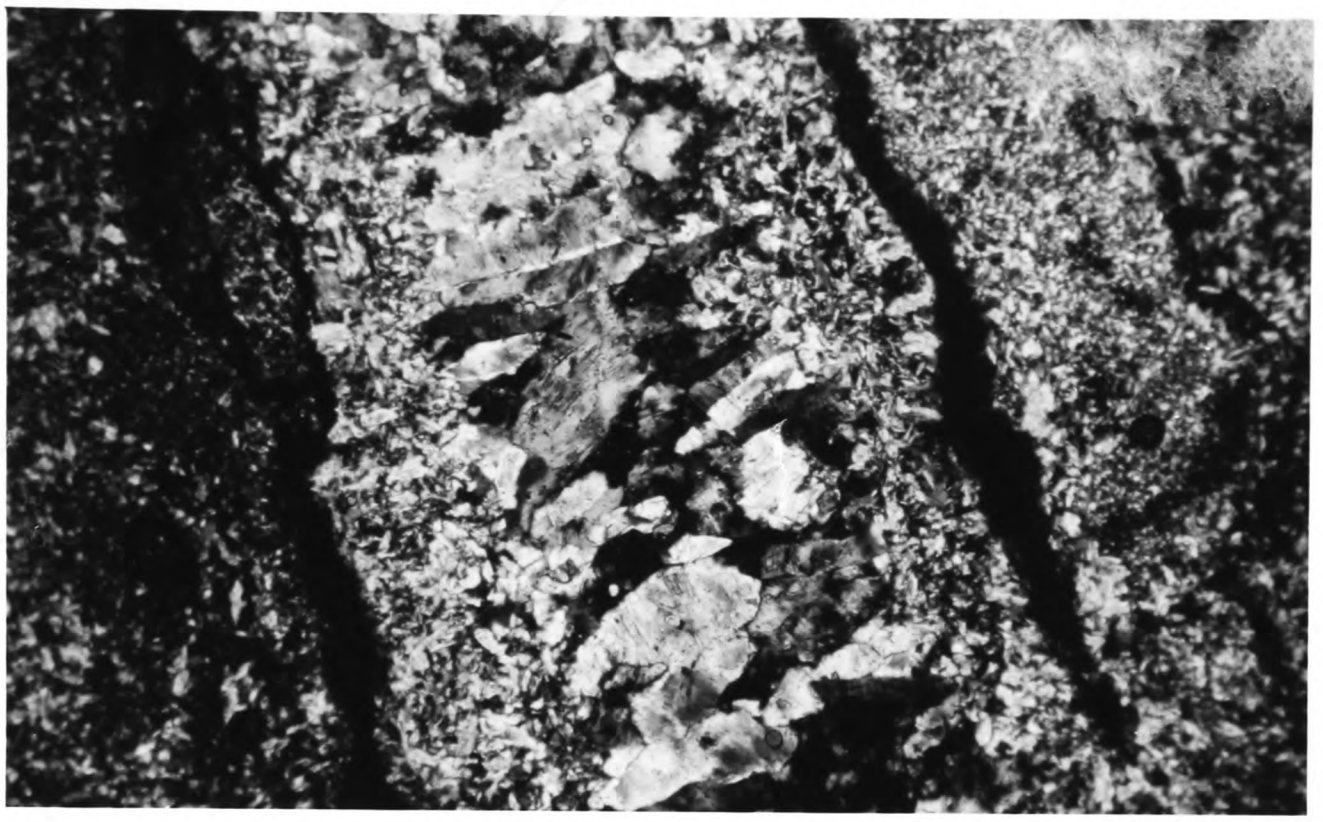




Plate 5.65 (overleaf, above)

Sample DT 5.2A, x40, see table 5.32 for chemical analyses. Unusual occurrence of granular gypsum crystal textures in a gypsum cobble from southern Tunisia. Moreover, the groundmass is of small, lozenge-shaped crystals rather than alabastrine gypsum.

Plate 5.66 (overleaf, below)

Sample K 9.2, x40, see table 5.18 for chemical analyses. Gypsum cobble composed of alabastrine material with pedoforms and crack-fill consisting of small, lozenge-shaped gypsum crystals.



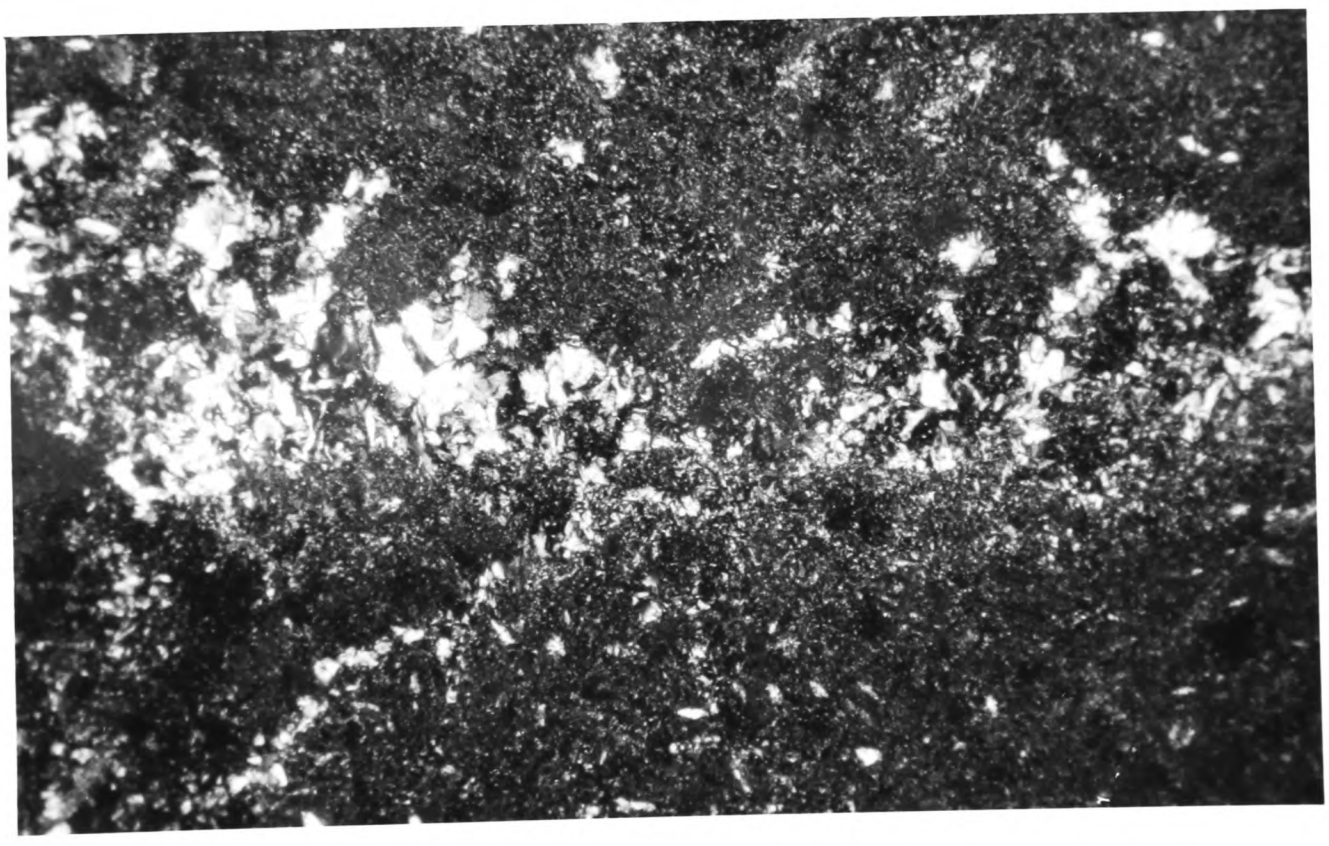
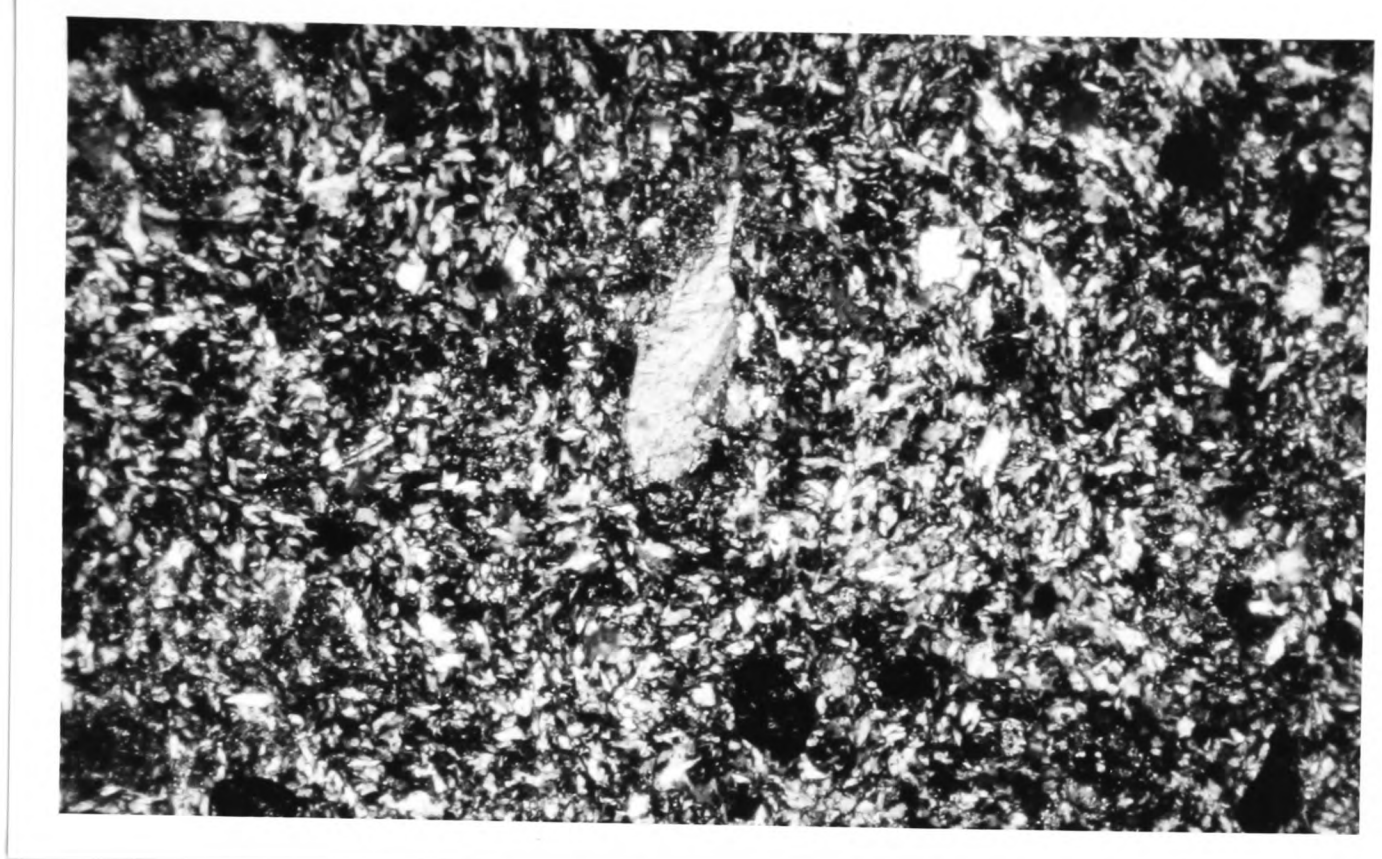


Plate 5.67 (overleaf, above)

Sample HA 2.1, oriented 45^{o} anticlockwise of vertical, x64, see table 5.34 for chemical analyses. Large gypsum porphyroblast in a groundmass of small, lozenge-shaped gypsum crystals in a cobble sample from the Central Namib Desert. Note the marked surficial solution features around the periphery of the porphyroblast and the alabastrine textures immediately above the feature.

Plate 5.68 (overleaf, below)

Sample GO 4.1, not oriented, x40, see table 5.21 for chemical analyses. Large lath-like fragments of biotite (dark) and angular quartz fragments (grey and white) in a groundmass of calcite (speckled). Note the exploded textures of the book-leaf micas and brecciation of quartz particles resulting from displacive calcite crystallization. The sample is of a calcrete from the Central Namib Desert.



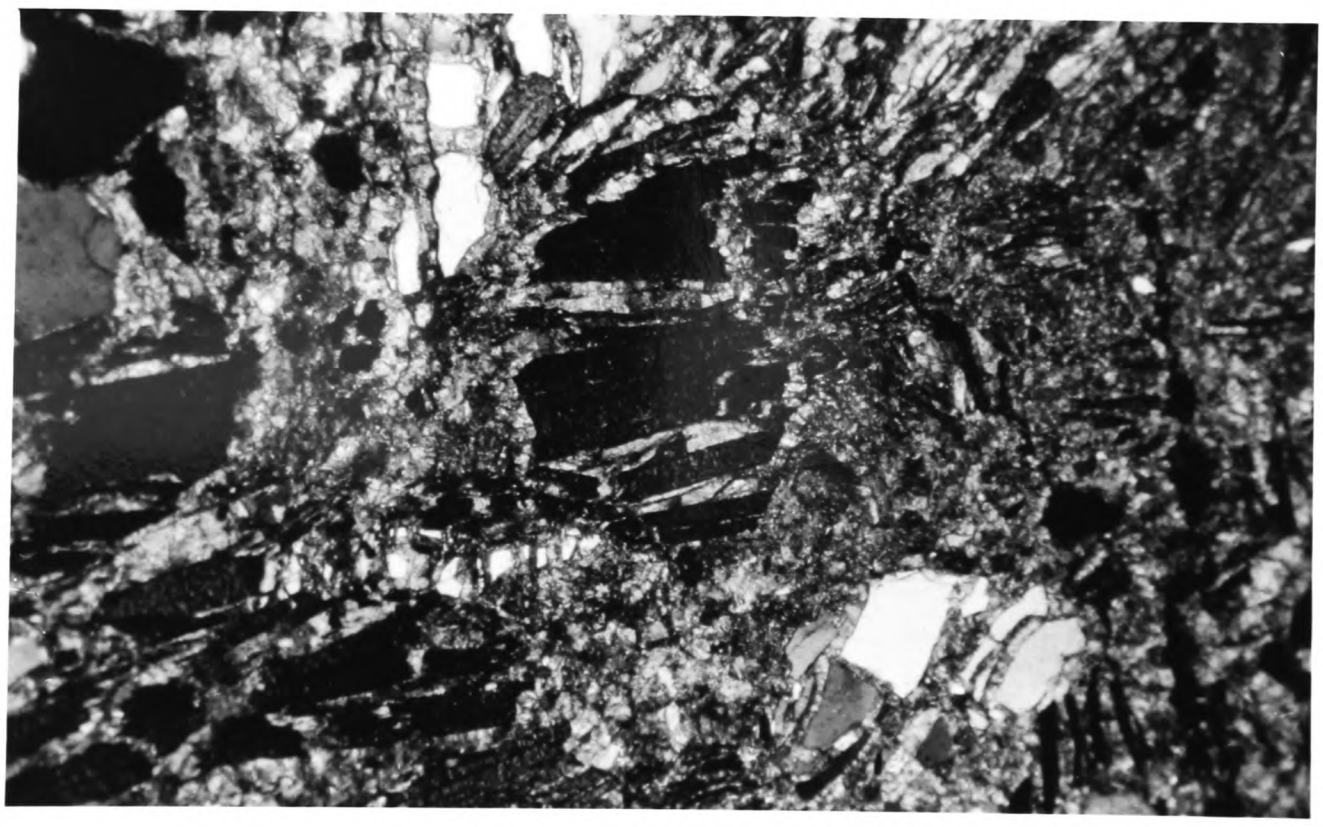
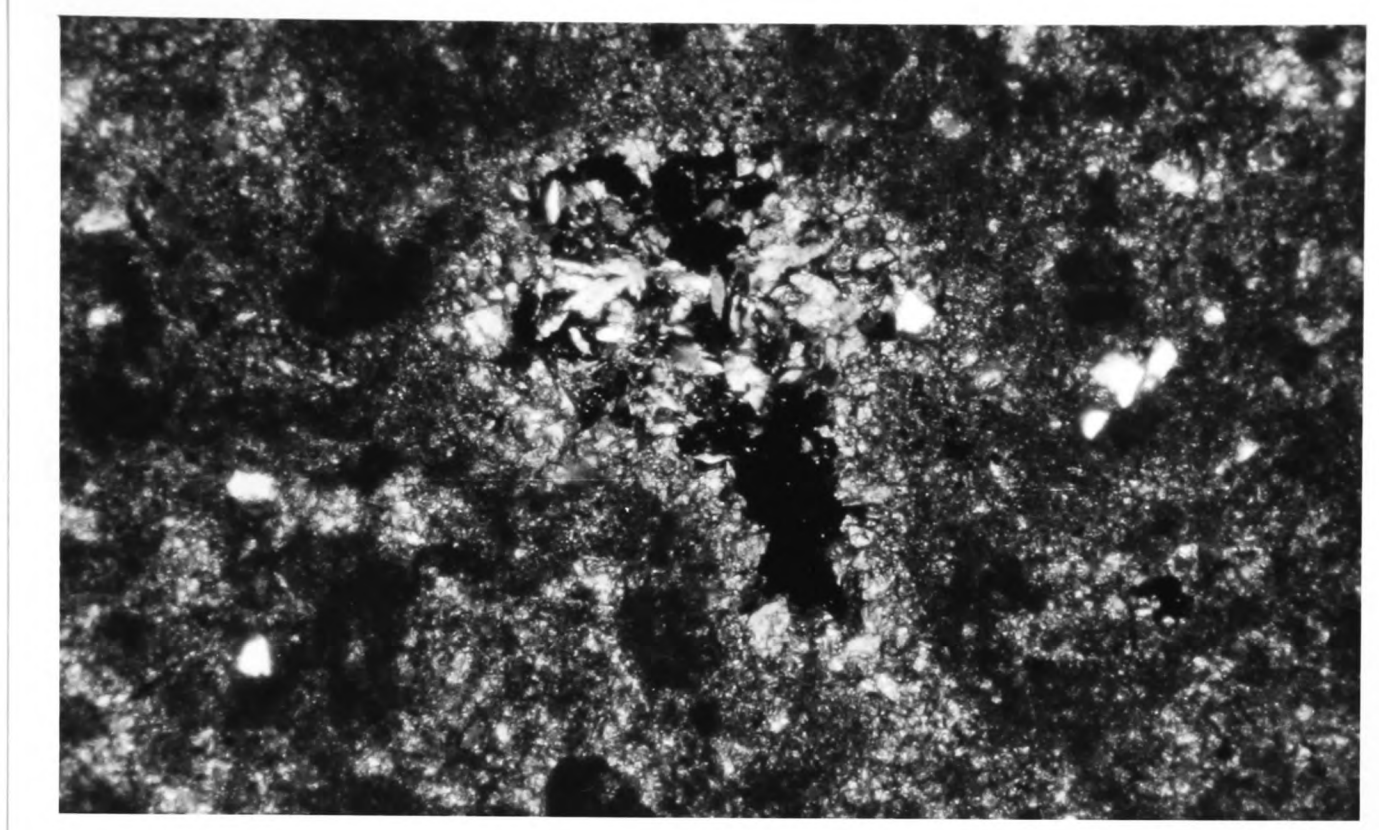


Plate 5.69 (overleaf, above)

Sample ZER 1.1, x64, see table 5.28 for chemical analyses. Calcrete from southern Tunisia showing dominantly micritic calcite with sub-angular quartz grain inclusions (white). The central void is lines with sparry calcite and has a number of small, lozenge-shaped gypsum crystals within it. Some poorly developed calcite pseudomorphs after gypsum are evident at the base of the void.

Plate 5.70 (overleaf, below)

Sample ZER 1.1, x64, plain light, see table 5.28 for chemical analyses. Gypsum crystal ghosts in the micritic calcrete described above. The lozenge-shaped features were formerly lenticular gypsum crystals but have been replaced by calcite. White, sub-angular inclusions are quartz grains and the bright area (bottom right) is a void lined with sparry calcite.



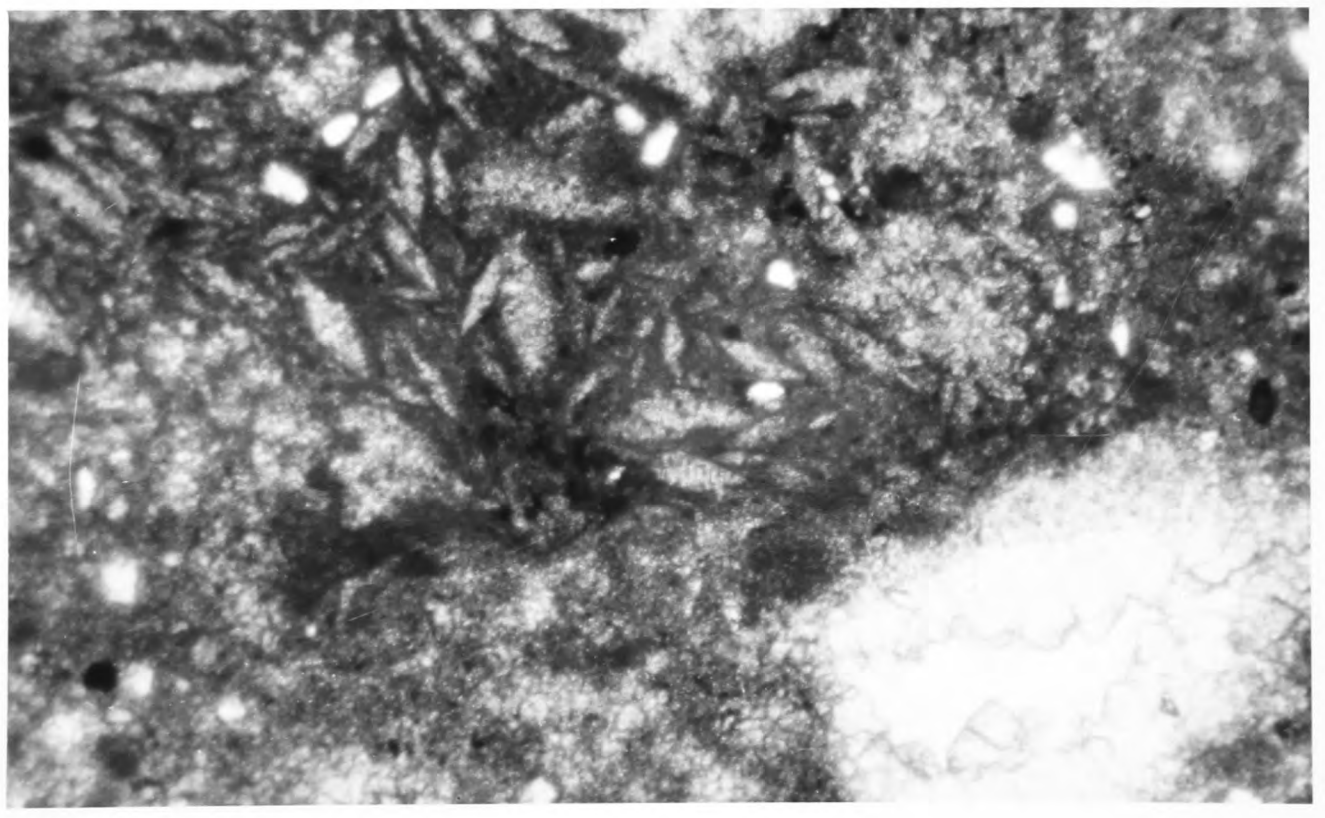
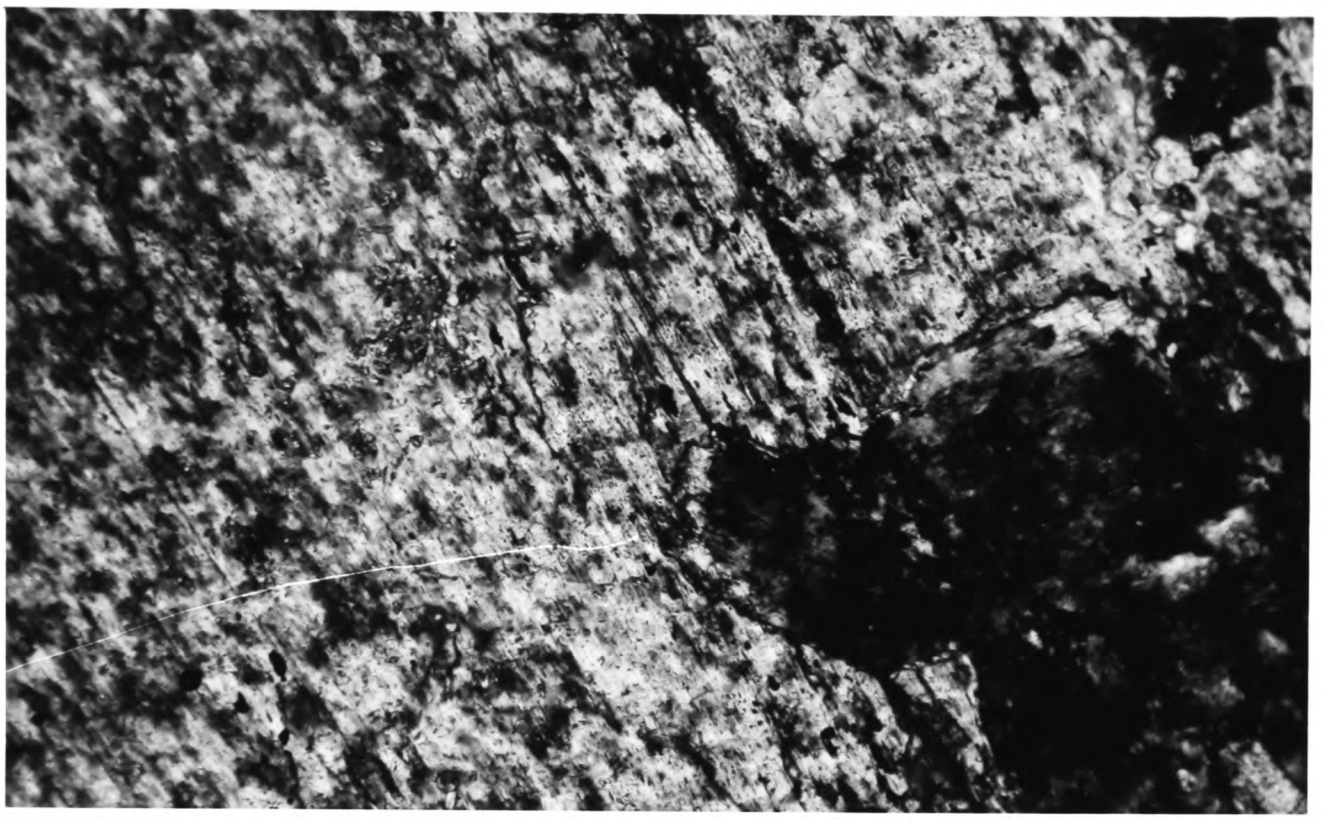


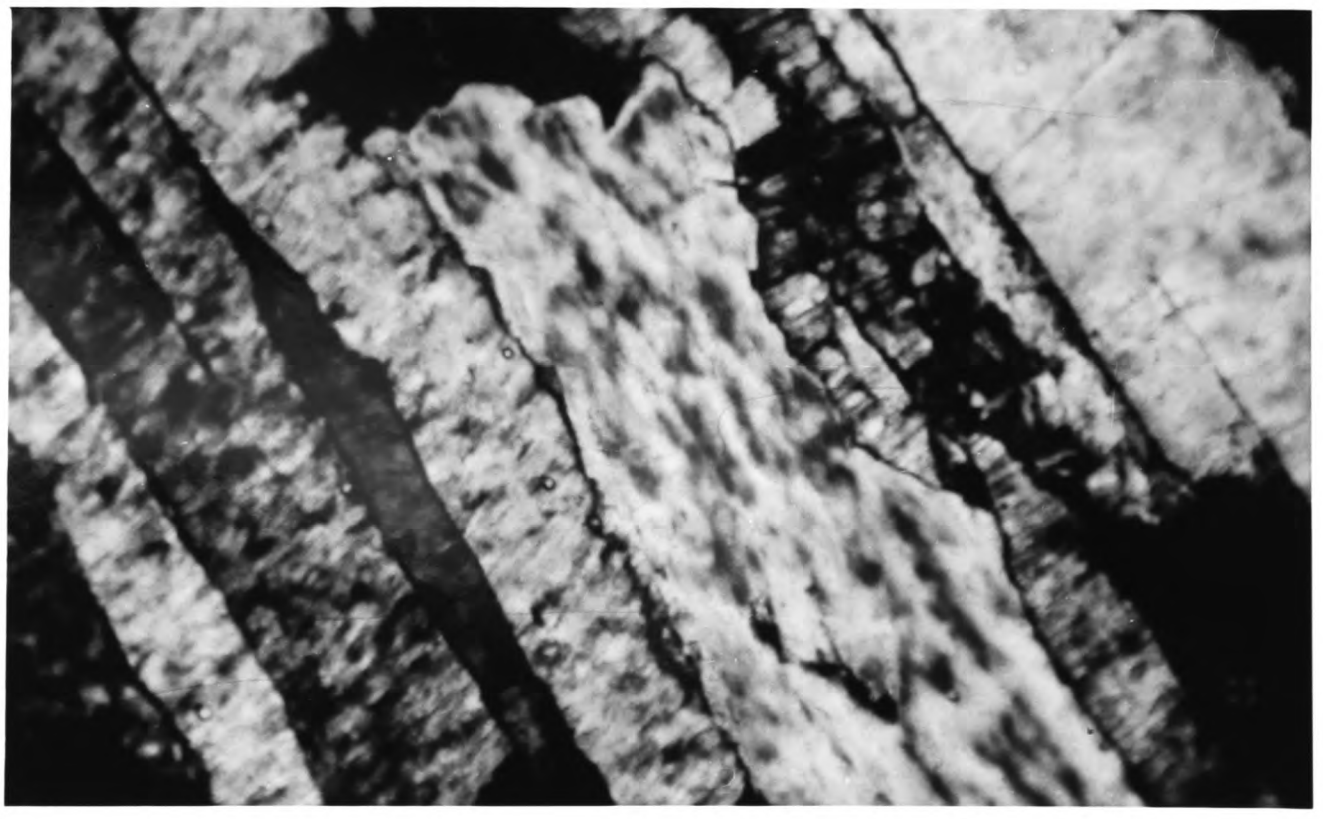
Plate 5.71 (overleaf, above)

Sample DT 3.1, x40, see table 5.32 for chemical analyses. Fibrous gypsum from Turonian bedrock in southern Tunisia showing the internal structure of a large fibrous crystal.

Plate 5.72 (overleaf, below)

Sample DT 3.2, oriented 20^{o} anticlockwise from vertical, x40, see table 5.32 for chemical analyses. Bent fibrous crystals showing undulose extinction along the fibre are indicative of tensile strain crystallization. Internal cleavage planes are also visible.





nate from southern Tunisia, are of granular texture. The bedded gypsum (DTF 1.4) is remarkable in having strata of selenite, up to 5.0 mm thick and of indeterminate lateral extent, with uniform extinction. The beds are separated by thin horizons of micritic calcite. All the examples have undergone considerable metamorphism but appear to be of evaporitic origin. Two thin sections of satinspar (DT 3.1 and 3.2) (see plate 5.71 and 5.72) from Djebel el Aziza show undulose extinction along some fibres and bending and differential extinction along others. These traits are indicative of stress or tensile strain crystallization and post-depositional disruption (Shearman et al., 1972); Phillips, 1974) and are the main distinguishing features between these bedrock fibrous forms and those associated with gypsum crusts. The two samples from between Tozeur and Nefta are interesting in that one is a micritic calcite root cast with a core of granular gypsum and the other contains not only large and small lozenges and alabastrine gypsum features but also porphyroblasts and calcite pseudomorphs after gypsum. Unfortunately the age of the bedrock horizon is undeterminate, it certainly predates the surface gypsum crust above it but it may be of any age from late Pleistocene to Mio-Pliocene or even Upper Cretaceous. The thin sections of non-gypsiferous bedrock include Turonian limestones and dolomites and Palaeozoic marbles.

d. The chemical composition of gypsum crusts.

Few detailed chemical analyses of gypsum crusts have been undertaken and only rarely are the methods determining of gypsum content presented. The results of these studies are presented in tables 3.1 to 3.6. Suffice it to point out here that few workers have provided information on calcium carbonate and insoluble residue content let alone "minor" cations. Some exceptions are the work of Coque (1962, p. 92), Pouget (1968), Gaddas (1971) and Vieillefon (1976) in Tunisia, Kulke (1974) in Algeria and Brunsden, Doornkamp and Jones

(1976) in Bahrain. Though the results of these analyses are very variable, the general pattern is of an increase in gypsum content and decrease in insoluble residue from desert rose croûtes de nappe through mesocrystalline subsurface crusts, to powdery accumulations and columnar surface crusts. In terms of evaporitic crusts, supratidal gypsum accumulations, which are essentially croûtes de nappe, have lower gypsum content than shallow water and pan deposits.

In all 344 samples of gypsum crusts and associated materials were analyzed; 213 from southern Tunisia and 131 from the Central Namib Desert. A quantitative analytical approach was adopted despite the complexity of some of the procedures and the need to perfect the techniques in some cases. An alternative approach could have been semi-quantitative X-ray diffraction analysis. This was undertaken on 21 samples of sabkha salt crusts, gypsum crusts, calcretes and several samples of unknown composition, for example the aragonitic miliolite from Djerba (D 1) and Turonian dolomite from Djebel el Aziza (DT 1.3). A powder camera technique was employed producing results with very poor resolution of minor constituents. Rejection of the X-ray diffraction approach was justified on the grounds that atomic absorption spectrophotometric analyses provided a resolution of 100 ppm (0.01%) in some cases which has proved critical for certain cations. The greatest drawback of atomic absorption spectrophotometry is the problem of assessing the anionic constituents. This was overcome by determining two principal anions (SO_4^{--} and CO_3^{--}) chemically. In effect, a combination of analytical approaches provided a good quantitative assessment of both the major mineral constituents of the samples and the scarce but significant cations.

See appendix one for a detailed description of the methods employed in the chemical analyses.

i. Results.

The tabulated results of the chemical analyses are presented by site rather than gypsum crust type or form to facilitate comparison with the section drawings and site descriptions.

The seven evaporitic gypsum crusts analysed (see table 5.7) have gypsum contents of between 45% and 80% with usually less than 10% calcium carbonate and less than 5% insoluble residue. The exceptions are one of the samples from the pan near Narabeb, which has over 30% calcium carbonate, and the bedded selenite from Sabkha el Melah and samples from the White Lady Salt Pan near Cape Cross, which have over 10% insoluble residue. This last of these samples is also anomalous in that it has unusually large potassium, magnesium, iron and aluminium contents. These result from the presence of basalt bedrock fragments in the sample which probably contain leucite or nepheline, increasing the potassium value. In the other samples, potassium is variable, though generally below 1000 ppm. Sodium content has been converted to sodium chloride, which varies from 0.38% to nearly 4.0% as would be expected in the presence of halite beds. Strontium values are significant in that they are higher in the samples from southern Tunisia than in those from the Central Namib, 1000 to 5000 ppm compared with 200 to 500 ppm. An explanation for this is presented below. Magnesium shows considerable variability, even at individual sites, being between 500 and 6000 ppm. The anomalous values from the clay sample beneath the crust in the Sabkha el Melah and the crust from the White Lady Pan reflect its presence in clay minerals, and possibly dolomite, in the former and basaltic augite or olivine in the latter. Total iron and aluminium contents are usually less than 1000 ppm and, as mentioned above, increase with insoluble residue content, indicating that they are present in clay minerals or oxide coatings on quartz grains.

The 22 desert rose croûtes de nappe (tables 5.8, 5.9, 5.11, 5.12 and 5.13) have gypsum contents of between 50% and 70% with less than 5.0% calcium carbonate and 20% to 30% insoluble residue, in southern Tunisia and 10% to 20%, in the Central Namib. Of the samples analysed, the major exceptions were that from Bir Zefazfa (OT 4.3) (see table 5.9), which contained more gypsum and less calcium carbonate and insoluble residue, as did the sample from the uppermost crust at El M'Dou (EM 3.9, see table 5.11). This adds support to the suggestion, put forward on the basis of micropetrographic evidence, that the crust at Bir Zefazfa may not be a croûte de nappe. The other chemical analyses support this, the sample having less sodium, magnesium, iron and aluminium and more potassium than most of the other desert rose croûte de nappe samples. Chemically the sample is very similar to the surface cobbles found at the same site. EM 3.9 may also represent a surface crust form. Of the samples from the Central Namib, PF 2.5 and HA 3 (see table 5.12) contain more than 70% gypsum and the latter has less than 10% insoluble residue. The polygon edge sample from Aussinanis (1MET 2) contains considerably more calcium carbonate (18.75%) than most samples and two examples from Wlotzkas Baken (WB 1.1 and 1.3) contain slightly more than 5.0% (see table 5.13).

Sodium chloride content is between 0.50% and 1.5%, with fewer high values than in evaporitic crusts. All the samples from El M'Dou have below 0.50%, as does the surface sample from Wlotzkas Baken. This probably reflects leaching of the soluble salts after exposure, as the solutional features on gypsum lozenges suggest. Basal samples from Aussinanis (1MET 3.1 and 3.2) are dominantly sodium chloride while the polygon edge has 2.29%. This figure and the high calcium carbonate value suggest salt precipitation from water moving along the fissure. Potassium content is generally below 1000 ppm, though the sodium chloride and mica breccia at Aussinanis has 2000 to 3700 ppm. This anomalously high figure and also those for magnesium, iron and aluminium result from

the biotite mica content. Strontium content is between 700 and 3500 ppm in the Tunisian samples, broadly the same as that for evaporitic crusts, while in the Central Namib it is between 200 and 2000 ppm. The samples from Aussinanis have high concentrations, except for the sodium chloride breccia in which strontium is absent. It is interesting to note that the strontium content of the samples from El M'Dou does not show a consistent pattern, varying from 1600 to 11400 ppm, though the gypsum crusts have higher concentrations than other surface materials. Magnesium varies in concentration from 1500 to 5500 ppm, though the desert rose crust from Sebkhret Mekracherma has 7000 ppm, owing to the influence of the underlying clays (sample SM 1.1, see table 5.9), and two samples from Wlotzkas Baken (WB 1.1 and 1.3) have 8100 and 10100 ppm. High iron and aluminium values at both sites also reflect the influence of underlying clays and bedrock. Low magnesium values are less common in desert rose croûtes de nappe than evaporitic crusts, only sample HA 3 (400 ppm) being below 1000 ppm. As in the case of evaporitic crusts, iron and aluminium levels are similar to each other. Since insoluble residue is more abundant in the desert rose croûtes de nappe, these elements are also more concentrated, generally 1000 to 2500 ppm are present but up to 3500 ppm is not unusual. The anomalously high values may be attributed to the influence of underlying materials. Low levels in the samples from Bir Zefazfa, the uppermost level at El M'Dou and HA 3, confirm the impression gained from the foregoing analyses that either they are not true croûtes de nappe or have undergone alteration. The possibility that HA 3 is an evaporitic form must be considered since, besides some chemical affinities, it also bears structural similarities to the evaporite gypsum deposits near Narabeb.

Chemical analyses of 40 mesocrystalline subsurface crusts were undertaken (see especially tables 5.13 to 5.21). The gypsum contents are within the range 65% to 90%, though lower values down to 50% are recorded in basal

samples such as those at Oasis de Lesseps (table 5.15), Oudref (table 5.16) and Oued Demna (table 5.22). In these cases, the non-gypsic component is made up of insoluble residue and calcium carbonate. Some surface samples, such as those from the Lagunenberg (CC 1.3) and Swartbankberg (SB 3.1 and 1.3, tables 5.19, 5.20 and 5.33), also have less than 65% gypsum owing to dissolution. The sample from La Skhirra (LS 1.4, table 5.17) is anomalous in having 95% gypsum. Calcium carbonate content is below 5.0% in most cases and almost invariably below 10%. Areas such as Kettana (table 5.18) and between Medenine and Zarzis, where there are well developed calcretes, are the only exceptions in Tunisia. PF 1.4, from near Rooibank (table 5.29), is the only exception in the Central Namib; underlying calcretes are the cause. Insoluble residue amounts to less than 15%, though the low gypsum levels in basal samples result in high percentages of insoluble residue, up to 30% in some cases. Exposed mesocrystalline crusts and the upper zones of shallow examples also have relatively high insoluble residue contents owing to gypsum dissolution by meteoric waters. Compared with desert rose croûtes de nappe, the mesocrystalline crusts have, in most cases, less insoluble residue, the deficit being made up of additional gypsum.

Sodium chloride levels are also lower in the mesocrystalline subsurface crusts, generally less than 0.5% (2000 ppm sodium). A few higher levels are recorded from gypsum crusts associated with underlying sodium chloride crusts in the Central Namib and some surface and basal samples from areas where salts are precipitated from ephemeral perched water tables. Potassium content is also lower than in desert rose croûtes de nappe, less than 500 ppm. Exceptional levels, up to 5500 ppm (see table 5.21), are recorded in those samples containing fragments of basalt or mica bedrock or associated with halite horizons, in the Central Namib. A sample from Oudref (OU 3), in Tunisia, has an unusually high level (1000 ppm) which may be derived from clay minerals since

iron and aluminium are also unusually concentrated.

Strontium content again shows a marked difference between southern Tunisia and the Central Namib Desert. In the former area, the values lie between 1500 and 3000 ppm, a narrower range than in the desert rose form. Some surface and basal samples, however, have lower levels reflecting, at least in part, lower gypsum content. Exceptionally high strontium concentrations, reaching 6800 ppm, are recorded from crusts between Medenine and Zarzis (MZ, table 5.14) but there is considerable variation within the section. In the Central Namib, strontium levels range from 200 to 5500 ppm but are usually below 1500 ppm. Since in most cases strontium levels are closely related to gypsum content, it is valid to assume that the strontium ions are in some way bound to the gypsum crystal lattice. They may be captured between the weakly bonded layers of water molecules during crystallization or possibly bonded with sulphate ions in place of calcium ions. Kushnir (1980) pointed out that differences in the ionic radii of calcium and strontium may preclude this. A more likely explanation is that gypsum crystals contain solid inclusions of celestite ($SrSO_4$) or liquid inclusions rich in strontium ions. Kulke (1974) found that Algerian gypsum croûtes de nappe had strontium contents increasing with depth, 1050 to 1600 $\pm$50 ppm near the surface, 1320 to 1800 $\pm$50 ppm in the middle and 1800 to 2200 $\pm$50 ppm at the base. She suggested that at the base, groundwater evaporation promoted higher levels. It is possible that weak ionic bonding attracts strontium ions to foci of gypsum precipitation thereby concentrating the cations which, with the gypsum, have been dissolved from diffuse deposits within bedrock. At this stage it is impossible to ascertain whether precipitation is directly from groundwater, as Kulke (1974) suggested, or from surface Chott waters followed by deflation of the strontium-rich gypsum and its remobilization on the surrounding landsurface. This bedrock source of strontium would explain the differences between

southern Tunisia and the Central Namib. Strontium is found widely in bedrock and some coastal sabkha deposits in Tunisia (see gypsum bedrock analyses from various sites, TN, DTF, DT, tables 5.28, 5.29 and 5.32), while in the Central Namib it is scarce, being derived from sea salts and from the Palaeozoic marbles in small quantities. Some may also be present as strontianite ($SrCO_3$) in igneous bedrock.

Magnesium levels are very variable in the mesocrystalline subsurface gypsum crusts. While concentrations in many crusts are in the range 1000 to 3000 ppm, some are as low as 100 ppm, as at La Skhirra, and as high as 7800 ppm, at Oued Demna (OD 10). Exposed crusts and surface horizons frequently have low magnesium levels suggesting that the element may be present in the form of soluble salts which have been leached out. Alternatively, the high concentrations in the basal zones may represent derivation of the element from underlying clays or bedrock, such as the marble of Swartbankberg, basalts of Lagunenberg, micas near Gobabeb and dolomites of southern Tunisia. Iron and aluminium contents are again similar to each other and rise and fall with insoluble residue levels, resulting in basal crusts having the highest concentrations. Once again bedrock may contribute, particularly in the Namib, but clays are usually the dominant source.

The 40 powdery gypsum accumulations examined (see especially tables 5.22 to 5.25) show considerable variations even from the same sections. Gypsum contents range from 20% to over 90%, the rest of the sample being composed of usually less than 5.0% calcium carbonate and up to 60% insoluble residue. At Oued Demna (OD 6.2 A to E) the harder horizons in the sequence contain more calcium carbonate and less insoluble residue than the soft. This may result from aeolian bedding of the gypsum-rich material or the host sediment into which it has been introduced (see fig. 5.7F and table 5.23). Gypsum con-

tent often increases down section until a consolidated subsurface crust appears but in some cases the surface is also rich in gypsum relative to the underlying powdery material, as at Oued Magroum (EH 6.1, table 5.27). This may result from introduction of gypsum at the surface and leaching during wet seasons. The edges of polygons have higher gypsum contents than the surrounding powdery material and are low in calcium carbonate and insoluble residue. Similarly, gypsified roots from Swartbankberg (SB 3.7, table 5.20) and Oued Zigzaou (ZIG 2.3, table 5.24) are gypsum-rich and, though having insoluble material, have very little calcium carbonate. Both types of feature probably represent voids filled with gypsum precipitated from percolating water.

Sodium concentrations are all below 2000 ppm, or 0.50% sodium chloride, except for several examples from the coastal fog belt of the Central Namib, a number from Oued Demna (see tables 5.22 and 5.23), which is also a coastal location, and two samples from the shores of Chott Djerid (see table 5.13), in which the maximum levels of 2.49% and 3.05% sodium chloride are recorded. Potassium is below 1500 ppm except in two samples (CC 1.2 and WG 2, tables 5.19 and 5.25) which may be influenced by bedrock or clay minerals. Strontium is very variable owing to the range of gypsum contents. It is generally between 200 and 2000 ppm in Tunisia, though in some samples with a high gypsum content it is greater. In the Central Namib, the strontium content is less than 1000 ppm except in two samples from near Swartbankberg (SB 1.2 and 3.7). The second of these is the above-mentioned root cast, the example from Tunisia (ZIG 2.3) also has an unusually high strontium content. Magnesium concentrations vary from less than 100 to 6000 ppm but show the same trends as in the subsurface mesocrystalline crusts. Two unusually high values were obtained from the samples from Bechri near Chott Djerid (CD 1.2, 15700 ppm) and Metouia (MET 1.1, 12800 ppm) (see table 5.13). These may result from the presence of dolomite.

Iron and aluminium are very variable since insoluble residue is as high as 60%. Iron levels are usually below 3000 ppm unless insoluble residue exceeds 45%. Aluminium levels closely parallel those of iron.

25 columnar surface gypsum crusts were analysed (see especially table 5.26 to 5.30). They show a striking consistency in their gypsum content, which is from 70% to 90%, and calcium carbonate content, which is below 10%. Insoluble residue is above 10% in only four samples with gypsum contents below 70% with the exception of the crust on dune sand at Djemma (DJ 1.3, table 5.8). Column walls and polygon edges are chemically similar to the crusts, though that from Djebel el Aziza (DT 7.1) has a high calcium carbonate content perhaps resulting from the presence of limestone bedrock.

Sodium levels rarely exceed 4000 ppm, 1.0% sodium chloride, the exceptions being in the Namib fog belt (see tables 5.29 and 5.30). Potassium content is below 1000 ppm in all but two samples, L 1 and EH 6.3 (see tables 5.14 and 5.27). The latter anomaly may be explained in terms of clay mineral content since the sample has a large amount of insoluble residue and high iron and aluminium values. The sample from Limaguess is enigmatic since the high potassium value is parallelled by neither high iron, aluminium nor sodium levels. Strontium is found to between 500 and 2500 ppm in Tunisia and 200 to 400 ppm in the Central Namib. Very high concentrations are found at Oasis de Lesseps (OL 1.1 and 1.2, table 5.15), which accord with high levels in the underlying mesocrystalline crust. The source may be local clays or the atmosphere, the site being near the coast. High strontium levels in one sample from Oued Zerkine (ZER 1.3B) are enigmatic since the gypsum crust above and the calcrete beneath have comparatively little (see table 5.28). Magnesium concentrations are variable, as in the case of powdery gypsum accumulations. Generally they range from 500 to 4000 ppm, though the crust on dune sand at

Djemma (DJ 1.3, table 5.8), that on limestone at Djebel ed Dissa (DI 4.1, table 5.28), and a surface crust from near Swartbankberg (SB 4, table 5.25), all have over 5000 ppm. Bedrock or atmospheric inputs may explain these anomalies. Iron and aluminium values exceed 1000 ppm only in those samples with over 20% insoluble residue and all have over 600 ppm potassium, indicating that their source is clay minerals. An exception is ZER 1.3B (see table 5.28) which may contain elements derived from the calcrete beneath or a former soil horizon upon it.

The intermediate form of surface gypsum crust is characterized by cobbles with gypsum contents of the order 75% to 90%, less than 5.0% calcium carbonate and less than 10% insoluble residue. Of the 19 samples analyzed (see especially tables 5.31 to 5.34) the only exceptions were from the Central Namib Desert, where four examples have less than 75% gypsum, though in some cases large deficits in the total determination suggest the gypsum figures are too low. In three of these cases, however, there is more than 10% insoluble residue and in two of the three also more than 5.0% calcium carbonate.

Sodium concentrations are below 2000 ppm in all but two samples from near Swartbankberg (SB 1.1 and 2, table 5.33) and one from Oued Demna (OD 6.1, table 5.22) and one from the Hamilton Range (HA 1.2, table 5.34), which also have high magnesium, iron and aluminium levels. This suggests that the source is basalt bedrock in the former and perhaps clays associated with insoluble residue in the latter. The high level (760 ppm) from Djebel el Aziza (DT 1.1, table 5.31) is enigmatic since it is not accompanied by any high values for other elements. Strontium lies between 500 and 2000 ppm in the Tunisian samples and 200 and 500 ppm in those from the Central Namib. High levels are recorded in DT 5.2A (see table 5.32), from Djebel el Aziza

(4000 ppm) and SB 1.1 (see table 5.33) from near Swartbankberg (1100 ppm), though why individual samples should differ so greatly is uncertain. Magnesium concentrations exceed 2000 ppm only in three samples which have high iron and aluminium levels, two of which have high potassium levels, suggesting bedrock or clay minerals are the source. An exception is DT 5.3A (see table 5.32) which comes from above dolomite bedrock at Djebel el Aziza. Once again iron and aluminium levels are very similar, though aluminium is slightly higher where calcretes are present as at Kettana (K 3, 5 and 7, table 5.18). Iron is below 500 ppm and aluminium below 1000 ppm in all samples except those with more than 10% insoluble residue or influenced by bedrock (CC 1.5, table 5.19).

The results of the chemical analyses of the different types and forms of gypsum crusts are summarized in tables 5.35, 5.36 and 5.37. The most striking differences are the high levels of insoluble residue in the desert rose form compared with all others except powdery accumulations, which are much more variable. The most interesting trait in terms of insoluble residue and gypsum content is the close similarity between subsurface mesocrystalline crusts, columnar surface crusts and cobbles. This similarity is also present in the other constituents though strontium and magnesium concentrations decrease from mesocrystalline to cobbles through the columnar form. They are highest in those crusts associated with abundant moisture, whether in the form of chott and sabkha waters, groundwater or soil moisture. Sodium levels are low, less than 1.0% sodium chloride except in the evaporitic and desert rose crusts. Potassium, iron and aluminium show little variation other than that resulting from the influences of bedrock and clay mineralogy which are evident from the insoluble residue content. There is some suggestion that iron and aluminium levels in surface crusts and cobbles are disproportionately lower than those in subsurface crusts, if the amounts of insoluble

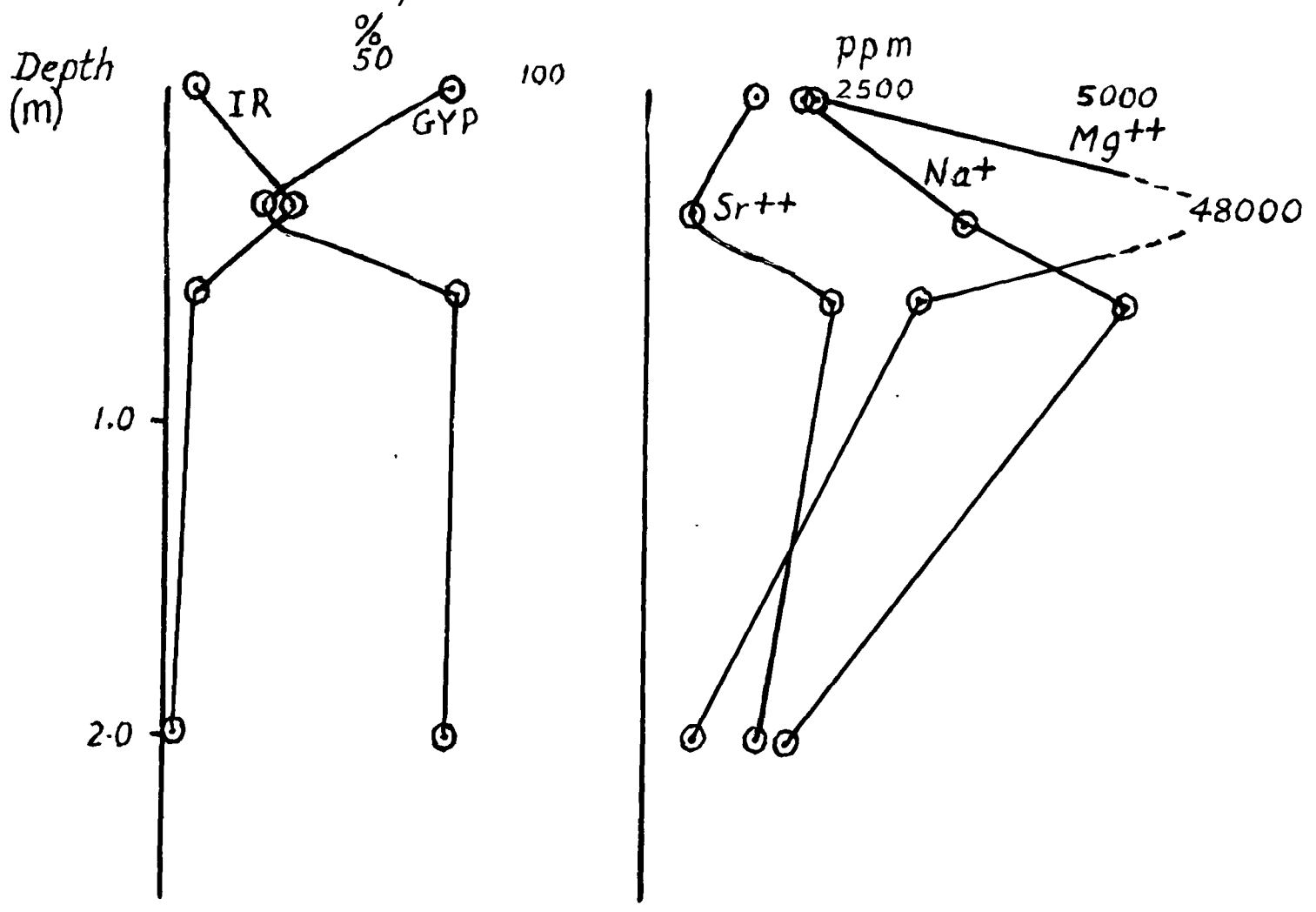
residue are compared. This may indicate leaching of some of the minerals containing these elements. For example, oxide coatings on quartz grains may be destroyed under the influence of meteoric waters.

Examination of the chemical characteristics of individual sections which exhibit more than one type of gypsum crust, such as those with columnar, powdery or cobble surface forms overlying mesocrystalline subsurface crusts, highlights some of the similarities and differences between them. Figures 5.10 and 5.11 show the variation in gypsum, insoluble residue and, where significant, calcium carbonate percentgage by weight through the sections at Metouia (MET), Oasis de Lesseps (OL), Oudref (OU) and the Lagunenberg (CC). Variations in the amounts of sodium, strontium and magnesium in parts per million are also shown. Gypsum levels usually mirror insoluble residue content since it never exceeds about 14%. It is striking that in all cases gypsum levels in the subsurface crust are as high if not higher than as at the surface, even where this consists of an indurated columnar crust. Since samples were not collected at close intervals, any conclusions regarding vertical variations are tentative. It is apparent from all the sections, that gypsum levels peak in two horizons, one at or near the surface and a second at depth. Furthermore, the sections at Oasis de Lesseps and Oudref show a gradual decrease in gypsum content in the lower horizons of the subsurface mesocrystalline crust. In the four profiles, sodium levels do not show any consistent pattern, though in some cases there is a slight peak at the top of the subsurface crust horizon. Strontium levels at the sites in southern Tunisia closely parallel gypsum content. While strontium concentrations are higher in the subsurface than surface crusts at Metouia and Oudref, the converse is true at Oasis de Lesseps where the surface has an extremely high strontium level (8900 ppm). Magnesium levels are very variable. While the mean values presented in

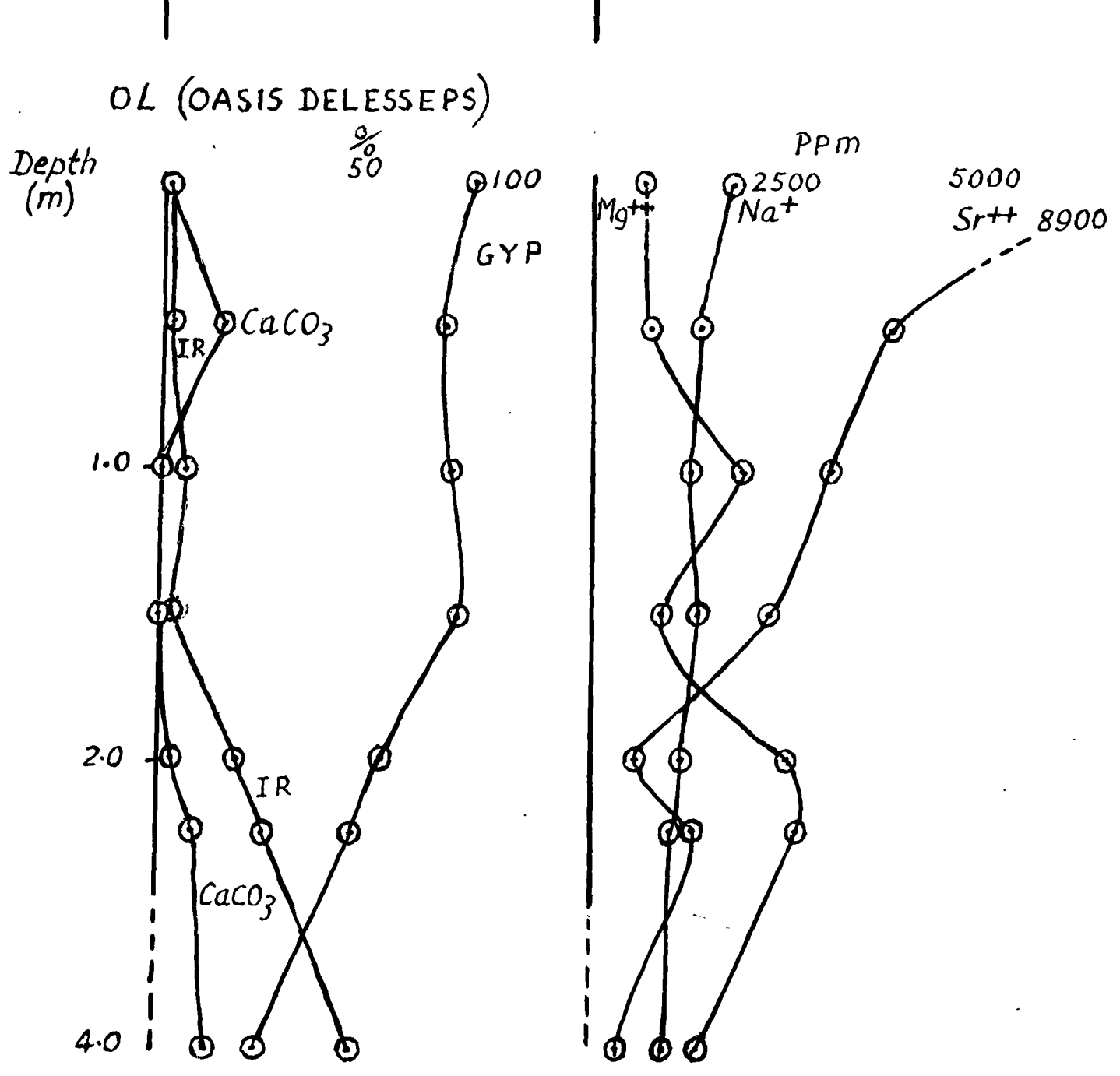
Figure 5.10 (overleaf)

Chemical and mineralogical variations with depth through the gypsum crusts at Metouia, above (see figure 5.5A) and Oasis de Lesseps, below (see fig. 5.5E).

A MET (METOUIA)



B OL (OASIS DELESSEPS)



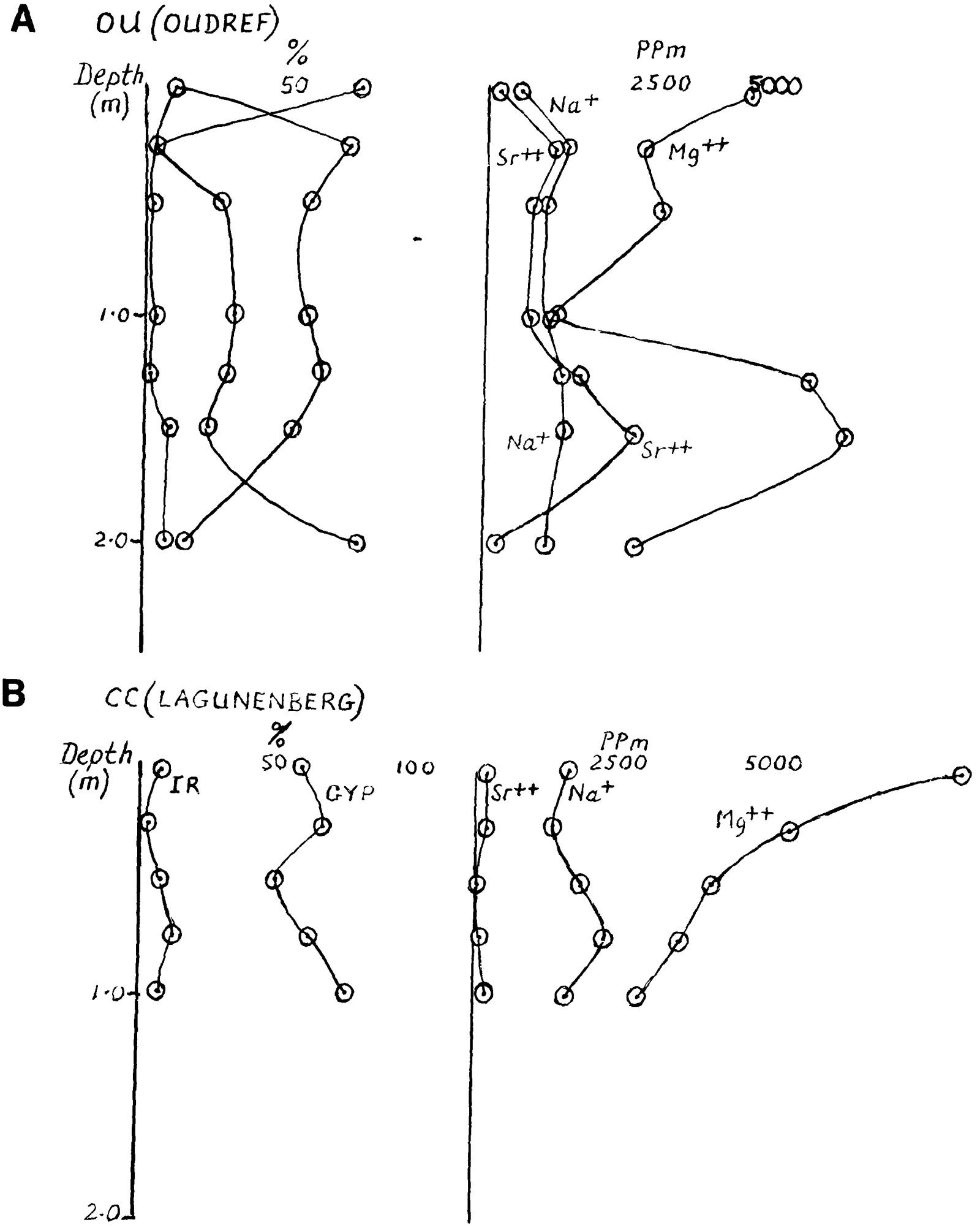


Figure 5.11

Chemical and mineralogical variations with depth through the gypsum crusts at Oudref, above (see figure 5.6A) and the Lagunenberg, below (see figure 5.7A).

tables 5.36 and 5.37 indicate that subsurface mesocrystalline crusts generally have higher magnesium concentrations than surface crust forms, this is not evident at Metouia or Lagunenberg. At the former site it is interesting to note that the intercrustal zone, which has a low gypsum content, is particularly rich in magnesium. This may suggest the presence of dolomite in the sediment. There is a suggestion in the diagrams of the chemical variations through the Tunisian profiles that, although there is little comparability between strontium and magnesium in the upper horizons, there is a tendency for both to decrease with depth, and gypsum content, in the subsurface mesocrystalline zone. The implications of this will be expanded upon in chapter seven; suffice it to point out here that this trait is also apparent in several other sections, notably that at Limaguess (L, table 5.14). In general, the evidence from these four profiles confirms that the broad characteristics provided by the mean chemical values are valid in individual cases.

In addition to the chemical analyses of the gypsum crust forms, those of associated sands, calcretes and bedrock are also provided in the tables. The information from these will not be summarized since the most important points have been alluded to in the foregoing discussion. Some additional analyses of sands and calcretes are presented in chapter six where their relevance to the understanding of the origin of different crust types and forms is examined.

TABLES 5.7 to 5.37

The following tables describe the mineralogical and chemical characteristics of all of the samples collected from the two field areas. In those cases where the samples come from recorded sections the figure numbers are provided. A detailed explanation of the notations used and the analytical methods is provided in appendix one.

In tables 5.35, 5.36 and 5.37 the mineralogical and chemical characteristics of the different types and forms of gypsum crust are summarized.

TABLE 5.7

	Bedded selenite CM 2	See fig. 5.2A 5.1	5.3	See fig. 5.2B CR 1.1	1.2
IR%	24.18	2.48	0.38	3.52	1.90
Gyp%	46.46	80.39	24.61	63.80	78.33
$CaCO_3$%	8.05	0.92	*42.76	6.21	2.30
Na^+ppm	1400	1500	16000	3040	11200
(NaCl%)	0.36	0.38	4.07	0.77	2.85
K^+ppm	2600	60	1200	260	1000
Sr^{++}"	5400	1000	2800	5000	1600
Mg^{++}"	3680	500	*89400	10800	5300
Fe^{tot}"	960	360	220	760	400
Al^{+++}"	1000	-	200	600	1000
$xsCa^{++}$"	47600	29100	-	52600	18400

*30.68% dolomite with 49000 ppm $xsMg^{++}$

	CC 2.2	NA 1.1	NA 1.2
IR%	11.02	1.00	7.62
Gyp%	49.29	65.66	48.88
$CaCO_3$%	-	7.14	31.19
Na^+ppm	15000	2200	2000
(NaCl%)	3.81	0.56	0.51
K^+ppm	*25000	100	300
Sr^{++}"	200	500	200
Mg^{++}"	*15900	1400	5700
Fe^{tot}"	9800	600	1700
Al^{+++}"	4600	500	1200
$xsCa^{++}$"	51200	60600	30300

*Basalt - Calcic plagioclase... ... $CaAl_2Si_2O_8$ (anorthite)
pyroxene. ... ... $(Ca,Mg,Fe,Al)_2(Al,Si)_2O_6$ (augite)
olivine.. ... ... $(Mg,Fe)_2SiO_4$ ($xsMg^{++}$)
leusite and nepheline will increase K^+

TABLE 5.8

See fig. 5.2C

	DJ 1.3	4.2	5	4.1	6
IR%	14.76	65.50	65.44	23.82	53.94
Gyp%	73.87	13.99	0.46	58.83	0.69
$CaCO_3$%	7.59	4.57	31.05	5.71	*39.04
Na^+ppm	2400	1100	1740	4800	5200
(NaCl%)	0.61	0.28	0.44	1.22	1.32
K^+ppm	340	480	700	600	1200
Sr^{++}"	1800	400	200	2600	200
Mg^{++}"	5400	3000	3800	5200	*47400
Fe^{tot}"	400	1000	3800	800	2400
Al^{+++}"	tr.	600	1200	400	1400
$xsCa^{++}$"	-	33100	-	18900	-

*35.14% dol. with 1100ppm $xsMg^{++}$

	Crust ZAA 1	Basal sand 2.1	Dune sand 2.2	Crust 3.1	Basal sand 3.2
IR%	26.02	42.82	41.90	22.32	50.76
Gyp%	60.77	48.72	41.39	64.46	36.86
$CaCO_3$%	1.40	-	6.05	2.33	2.33
Na^+ppm	5800	1950	1950	2700	1630
(NaCl%)	1.47	0.50	0.50	0.69	0.41
K^+ppm	900	600	700	650	700
Sr^+ "	800	1000	700	700	600
Mg^{++}"	4600	3200	2700	3700	4700
Fe^{tot}"	1000	1400	1100	1000	1800
Al^{+++}"	1400	1300	1000	1000	1800
$xsCa^{++}$"	22900	16600	20200	22600	14800

TABLE 5.9

See fig. 5.3A

	OT 1.1	1.2	2	4.1	4.2	4.3
IR%	92.72	6.36	71.66	93.26	82.10	1.22
Gyp%	0.21	85.55	1.77	0.16	0.19	82.27
$CaCO_3$%	1.19	1.16	17.91	4.76	7.14	2.33
Na^+ppm	150	1600	1300	200	380	1500
(NaCl%)	0.04	0.41	0.33	0.05	0.10	0.38
K^+ppm	1050	320	920	900	1500	170
Sr^{++}"	-	500	100	-	-	800
Mg^{++}"	1900	300	2600	1500	3600	200
Fe^{tot}"	3300	300	4500	2700	5700	100
Al^{+++}"	3900	500	4700	3400	6500	100
$xsCa^{++}$"	3000	10200	6100	-	5000	41100

See fig. 5.2D

	Dune sand EF 1.1	Crust 1.2	SM 1.1	1.2	2	Chicken-wire Clays 3
IR%	53.52	29.54	44.40	27.24	20.52	9.12
Gyp%	37.48	56.19	27.11	43.17	59.67	68.39
$CaCO_3$%	2.98	3.67	15.12	12.09	0.23	3.49
Na^+ppm	1600	8900	6500	15500	2600	6700
(NaCl%)	0.41	2.26	1.65	3.94	0.66	1.70
K^+ppm	800	1200	1600	2320	1050	1350
Sr^{++}"	500	900	1600	900	400	200
Mg^{++}"	2500	3800	10200	7800	7000	2900
Fe^{tot}"	1800	1000	4800	4400	5000	3900
Al^{+++}"	1700	1040	4000	4700	5300	4500
$xsCa^{++}$"	6800	12400	12300	4100	71200	32800

TABLE 5.10

	Calcrete EM 1.1	Calcrete 1.2	See fig. 5.2E 2.1	2.2	2.3	2.4	2.5
IR%	29.14	32.00	67.10	45.48	64.30	76.84	46.98
Gyp%	1.39	1.07	13.34	22.77	22.10	6.27	3.87
$CaCO_3$%	64.73	59.16	11.86	26.22	7.42	9.74	38.98
Na^+ppm	2450	3950	1100	1800	1000	1550	2300
(NaCl%)	0.62	1.00	0.28	0.46	0.25	0.39	0.58
K^+ppm	900	1250	1750	950	1450	1400	1400
Sr^{++}"	200	100	400	1400	600	1200	100
Mg^{++}"	2800	3300	3400	2400	2300	2800	3100
Fe^{tot}"	4100	5200	6000	4400	4300	6700	5100
Al^{+++}"	2900	3780	5540	3200	4200	5400	4200
$xsCa^{++}$"	-	-	1400	-	4900	400	12900

TABLE 5.11

	See fig. 5.3B EM 3.9	3.4	3.5	3.6	3.1	3.3	3.7	3.2	3.8
IR%	3.38	9.12	29.02	19.84	42.26	24.12	60.34	28.92	28.78
Gyp%	74.79	82.59	55.98	69.06	48.27	59.28	29.68	59.36	49.15
$CaCO_3$%	11.83	1.39	6.26	2.55	2.78	6.96	3.48	3.94	4.18
Na^+ppm	1700	1550	1150	1200	1000	1400	700	1250	1100
(NaCl%)	0.42	0.39	0.29	0.31	0.25	0.36	0.18	0.32	0.28
K^+ppm	250	350	950	600	800	650	1250	700	950
Sr^{++}"	3300	1600	3100	2100	900	11400	800	1600	1600
Mg^{++}"	3500	1200	3500	1600	1500	1400	2200	1700	2300
Fe^{tot}"	600	1200	2700	1900	2400	1700	3700	2200	2500
Al^{++}"	1100	800	2900	1600	2800	1820	3600	2300	2700
$xsCa^{++}$"	15300	6200	12600	9100	8600	16200	2900	8100	32800

TABLE 5.12

	HA 3	See fig. 5.4A PF 2.5	IMET 1.1	See figs. 5.4C and 5.4D 1.2	1.3	2	3.1	3.2	3.3
IR%	6.82	12.94	14.40	14.54	10.80	6.18	8.54	44.86	29.26
Gyp%	83.19	77.19	67.41	70.04	71.28	60.89	2.54	2.06	56.61
$CaCO_3$%	trace	0.69	*2.76	*2.53	-	18.75	-	0.93	0.23
Na^+ppm	2000	2850	2400	2650	3450	9000	290000	140000	6000
(NaCl%)	0.51	0.72	0.61	0.67	0.88	2.29	73.72	35.59	1.53
K^+ppm	150	880	700	1200	950	1050	3700	2150	950
Sr^{++}"	200	200	1900	1700	1200	1800	-	-	600
Mg^{++}"	400	1500	*23100	*13200	*15600	10600	3500	1900	5100
Fe^{tot}"	600	2000	1000	2400	2000	2500	7100	2300	2400
Al^{+++}"	500	2300	5000	5000	3900	2700	4700	3800	1700
xsCa^{++}"	6300	15500	*14000	*16800	*34000	39100	100	1500	35500

*11.50% dol. with 7900 ppm xsMg^{++}

*10.02% dol. with 5100ppm xsMg^{++}

*11.84% dol. with 8300ppm xsMg^{++}

*Mg^{++} probably from mica

TABLE 5.13

	Crust surface WB 1.1	Nebkha 1.2	0.50m 1.3	Surface CD 1.1	Sand 1.2	Crust 1.3
IR%	16.90	18.38	13.80	19.50	22.08	18.88
Gyp%	68.46	66.91	61.34	49.34	58.14	68.47
$CaCO_3$%	8.28	1.38	7.13	*5.52	*1.61	0.23
Na^+ppm	1450	1900	4000	12000	9800	2900
(NaCl%)	0.37	0.48	1.02	3.05	2.49	0.74
K^+ppm	700	850	1050	1200	1500	400
Sr^{++}"	700	400	600	1800	1200	3200
Mg^{++}"	8100	3500	10100	*15500	*15700	8000
Fe^{tot}"	3500	3200	2300	1500	2900	980
Al^{+++}"	2500	3300	6300	1000	2400	800
$xsCa^{++}$"	-	24700	22600	*17000	6200	9700

*11.7% dol. with 20500 ppm $xsCa^{++}$

*5.8% dol. with 8100 ppm $xsMg^{++}$

See figs. 5.5A and 5.5B

	MET 1.1	1.2	1.3	2.1	2.2	2.3	3
IR%	43.06	4.10	3.98	7.68	29.66	7.52	1.96
Gyp%	33.10	92.55	84.13	69.94	24.13	72.31	69.81
$CaCO_3$%	7.66	1.39	3.94	6.03	*7.52	1.86	3.48
Na^+ppm	1100	1500	1600	1750	3950	5000	1750
(NaCl%)	0.28	0.38	0.41	0.44	1.00	1.27	0.44
K^+ppm	600	150	150	300	1120	800	250
SR^{++}"	600	1600	2100	1300	500	2300	1450
Mg^{++}"	12800	1500	600	1800	*48000	3400	600
Fe^{tot}"	2600	500	200	500	2500	700	300
Al^{+++}"	2900	400	500	600	2100	1100	600
$xsCa^{++}$"	32300	-	16300	29100	-	25300	53500

*36.42% dol. with 4600 ppm xs Ca^{++}

TABLE 5.14

	See fig. 5.5C MZ 1.5	1.4	1.3	1.2	1.1	Powdery surface OZ 1.2	See fig. 5.5D L 1	2	3	4
IR%	2.98	1.24	14.42	9.14	11.34	19.18	1.44	1.12	7.50	0.98
Gyp%	86.21	80.78	50.01	67.22	62.32	56.18	88.85	87.11	82.34	88.99
$CaCO_3$%	4.64	1.16	29.70	14.39	13.23	15.35	1.16	0.46	2.55	1.62
Na^+ppm	1400	1500	1850	1650	1750	1500	2800	1600	2100	1400
(NaCl%)	0.36	0.38	0.47	0.42	0.44	0.38	0.71	0.41	0.53	0.36
K^+ppm	150	200	450	300	350	500	1500	200	250	50
Sr^{++}"	2900	6700	6200	700	6800	800	900	1300	800	1600
Mg^{++}"	300	700	1800	1100	6100	1600	3100	2000	1200	1600
Fe^{tot}"	300	200	1300	900	1500	1500	100	200	700	-
Al^{+++}"	700	200	700	1300	2400	1800	500	200	900	200
$xsCa^{++}$"	2700	27300	-	9800	17200	17700	14500	15400	5100	10300

TABLE 5.15

See fig. 5.5E

	OL 1.1	1.2	1.3	1.4	1.5	1.6	2
IR%	2.28	2.92	5.70	3.72	20.88	28.82	54.60
Gyp%	86.10	77.70	79.75	82.13	60.07	52.08	26.02
$CaCO_3$%	1.16	8.12	-	0.70	2.56	-	13.26
Na^+ppm	1900	1500	1350	1450	1200	1150	930
(NaCl%)	0.48	0.38	0.31	0.34	0.37	0.29	0.24
K^+ppm	200	300	400	220	550	730	630
Sr^{++}"	8900	4200	3400	2500	600	1400	400
Mg^{++}"	700	800	2100	1000	2800	2900	1500
Fe^{tot}"	200	400	600	400	1800	2800	3600
Al^{+++}"	200	400	1000	700	2600	3900	3000
$xsCa^{++}$"	40900	18600	28300	34000	35900	36800	4300

TABLE 5.16

See fig. 5.6A

	OU 1.1	1.2	1.3	DI 2.1	DI 2.2	OU 2.1	4	3	2.2
IR%	74.80	3.72	26.60	1.78	52.86	30.34	28.90	22.66	76.66
Gyp%	7.75	72.15	57.70	88.09	39.08	56.51	61.86	53.56	14.21
$CaCO_3$%	9.77	4.42	2.09	1.38	6.67	3.02	0.93	8.84	6.28
Na^+ppm	600	1500	1150	1700	1400	1150	1400	1400	1100
(NaCl%)	0.15	0.38	0.29	0.43	0.36	0.29	0.36	0.36	0.28
K^+ppm	1350	250	700	400	700	450	450	1000	700
Sr^{++}"	200	1300	900	2000	800	800	1700	2700	200
Mg^{++}"	4800	2900	3200	60	900	1200	5900	6500	2700
Fe^{tot}"	4800	300	1600	60	860	1100	1500	3500	2500
Al^{+++}"	4500	300	1700	800	1200	1000	1500	3300	2500
$xsCa^{+++}$"	-	38300	35300	13400	16300	24400	16300	22000	-

TABLE 5.17

See fig. 5.6B

	LS 1.1	1.2	1.3	1.4	1.5
IR%	75.06	9.52	1.38	15.78	48.26
Gyp%	1.08	82.43	95.02	68.73	27.93
$CaCO_3$%	12.99	2.55	-	4.87	7.42
Na^+ppm	650	1400	2250	1600	6000
(NaCl%)	0.17	0.36	0.57	0.41	1.53
K^+ppm	1650	210	120	1050	1800
Sr^{++}"	-	500	2300	100	100
Mg^{++}"	5000	100	100	2300	5500
Fe^{tot}"	7800	700	200	4200	8400
Al^{+++}"	7300	900	700	4700	8700
$xsCa^{++}$"	13400	1900	2800	4500	9300

TABLE 5.18

See figs. 5.6C and 5.6D

	K 2	6	4	5	9.1	9.2	7	8.1	8.2	3
IR%	1.20	1.54	8.34	18.24	1.12	1.70	3.28	1.30	6.82	24.24
Gyp%	88.13	85.58	66.33	5.96	91.21	89.61	2.33	84.81	70.41	1.06
$CaCO_3$%	4.64	0.70	17.40	67.05	1.62	3.02	91.65	1.86	6.03	65.19
Na^+ppm	2600	1500	1550	2300	1400	1500	3550	1500	1300	1850
(NaCl%)	0.66	0.38	0.40	0.58	0.36	0.38	0.90	0.38	0.33	0.47
K^+ppm	100	100	650	1600	120	150	400	500	250	1650
Sr^{++}"	1200	1600	1300	-	1100	900	100	1600	1300	100
Mg^{++}"	200	500	1700	3700	100	-	300	200	400	5800
Fe^{tot}"	200	400	2400	5600	200	200	1300	200	700	7800
Al^{+++}"	200	500	2000	6000	700	800	2000	150	700	10400
$xsCa^{++}$"	8200	25900	7900	-	3200	1300	-	25100	30000	-

TABLE 5.19

See fig. 5.7A

	CC 1.5	1.6	1.1	1.2	1.3	1.7
IR%	7.42	5.14	2.92	6.60	11.64	7.18
Gyp%	57.62	78.50	64.89	48.64	61.38	74.35
$CaCO_3$%	-	-	trace	1.71	0.73	-
Na^+ppm	1650	1450	1400	1920	2250	1650
(NaCl%)	0.42	0.37	0.36	0.49	0.57	0.42
K^+ppm	620	300	250	1670	880	470
Sr^{++}"	200	200	200	-	100	200
Mg^{++}"	8800	1000	5600	4200	3700	2900
Fe^{tot}"	8900	1500	1700	9000	6500	3300
Al^{+++}"	3500	600	600	4200	1900	2000
$xsCa^{++}$"	67800	37200	66900	79900	52100	37000

TABLE 5.20

	See fig. 5.7B SB 3.2	3.6	3.8	3.3	3.4	3.7	3.1	3.5
IR%	10.38	4.00	2.18	9.82	15.82	3.10	14.12	6.80
Gyp%	77.73	83.85	83.06	79.36	55.93	85.61	59.18	23.35
$CaCO_3$	5.06	4.06	5.52	3.68	14.48	0.69	9.43	63.91
Na^+ppm	1450	6400	1600	1500	2150	6000	4000	2950
(NaCl%)	0.37	1.63	0.41	0.38	0.55	1.53	1.02	0.75
K^+ppm	270	400	50	150	550	400	1450	300
Sr^{++}"	200	400	200	1000	500	1800	500	300
Mg^{++}"	4800	3600	900	3800	4300	2500	8900	4100
Fe^{tot}"	1800	600	200	800	1400	400	3000	800
Al^{+++}"	1600	800	100	800	1000	500	3200	1000
$xsCa^{++}$"	6800	40400	30500	8500	29800	61900	30500	33700

TABLE 5.21

See fig. 5.7C

	GO 4.1	5.1	3.3	5.2	4.2	4.3	4.4	3.2	6
IR%	23.96	28.74	26.90	27.98	22.54	19.68	15.72	13.08	24.82
Gyp%	1.12	0.58	3.70	0.17	50.41	63.21	54.35	66.06	59.08
$CaCO_3$%	67.14	*61.46	61.46	*70.49	5.12	-	1707	-	0.95
Na^+ppm	3000	2250	8000	1650	2600	2750	1950	1700	1100
(NaCl%)	0.76	0.57	2.03	0.42	0.66	0.70	0.50	0.43	0.28
K^+ppm	2200	145	3450	1400	3400	1200	750	5500	580
Sr^{++}"	100	200	100	300	400	1200	200	1300	1600
Mg^{++}"	4000	*25000	9300	*72000	13200	6900	6700	10300	9400
Fe^{tot}"	6500	4300	7700	3600	8600	3400	3000	10000	2200
Al^{+++}"	4000	3100	5700	3400	5600	2400	2000	9900	1800
$xsCa^{++}$"	500	* -	-	* -	32200	26900	21100	34200	22700

*18.97% dol.
46.80% $CaCO_3$

*54.63% dol.
12.39% $CaCO_3$

*Magnesium probably derived from micas

TABLE 5.22

	Gypsum from clays OD 9	Dune sand 7	Surface 1	Surface 2	Surface 3	Cobble 6.1	Cobble 6.3	Beneath dune 6.4
IR%	6.26	86.96	0.90	1.38	0.50	0.90	4.86	1.52
Gyp%	63.89	0.77	76.72	86.56	87.24	75.96	82.42	82.83
$CaCO_3$%	-	6.50	9.28	0.70	1.39	5.34	4.64	2.32
Na^+ppm	1750	300	1700	1500	2200	2300	1450	1500
(NaCl%)	0.44	0.08	0.43	0.38	0.56	0.58	0.37	0.38
K^+ppm	500	900	100	150	150	150	200	100
Sr^{++}"	200	100	2500	900	2100	2000	1600	2000
Mg^{++}"	2200	2500	100	-	100	400	200	200
Fe^{tot}"	2300	3600	100	200	100	100	400	100
Al^{+++}"	2400	3400	900	200	100	200	400	100
$xsCa^{++}$"	63300	6300	18200	25700	25400	35800	11500	29800

TABLE 5.22 (cont.)

See fig. 5.7D

	OD 4.1	4.2	4.3	4.4	4.5	10
IR%	82.92	84.66	77.00	60.06	3.28	15.88
Gyp%	3.89	0.87	8.57	23.96	85.39	63.93
$CaCO_3$%	6.96	6.48	7.19	8.59	-	3.48
Na^+ppm	450	400	650	1000	2100	1500
(NaCl%)	0.11	0.10	0.17	0.25	0.53	0.38
K^+ppm	800	900	950	850	220	550
Sr^{++}"	200	100	200	400	2200	1600
Mg^{++}"	3000	3300	2600	2300	200	7800
Fe^{tot}"	3600	4000	3900	3200	200	1800
Al^{+++}"	2900	3500	3500	3000	200	1900
xsCa^{++}"	6900	12100	9300	7800	17200	35200

TABLE 5.23

See fig. 5.7E

	OD 8.1	8.2	8.3	5.1	5.2	5.3
IR%	41.86	47.90	14.76	4.06	2.36	2.06
Gyp%	46.15	19.74	71.79	82.01	83.36	75.52
$CaCO_3$%	6.50	6.50	2.32	4.41	2.78	18.10
Na^+ppm	1050	1200	1400	2000	1850	2750
(NaCl%)	0.27	0.31	0.36	0.51	0.47	0.70
K^+ppm	700	800	370	200	200	250
Sr^{++}"	500	500	1200	1500	1200	1000
Mg^{++}"	1100	1300	600	300	100	300
Fe^{tot}"	2900	2900	1100	300	200	200
Al^{+++}"	2500	2800	1300	400	200	400
$xsCa^{++}$"	16500	60100	25500	31400	22800	-

TABLE 5.23 (cont.)

See fig. 5.7F

OD 6.2	A	B	C	D	E
IR%	1.98	8.18	0.54	2.88	0.98
Gyp%	78.44	78.70	83.09	81.08	79.35
$CaCo_3$%	7.42	1.86	3.48	0.93	3.02
Na ppm	2500	2950	3000	3050	2950
(NaCl%)	0.64	0.75	0.76	0.78	0.75
K^+ppm	180	200	150	250	350
Sr^{++}"	1500	1200	1600	1500	2100
Mg^{++}"	400	900	1100	1000	1100
Fe^{tot}"	100	400	100	200	100
Al^{+++}"	300	200	100	200	300
$xsCa^{++}$"	17700	21400	26600	35500	39200

TABLE 5.24

ZIG	Calcrete surface 1.1	Calcrete 0.25 m 1.2	Calcrete 1.5 m 1.3	Clay 2.5 m 1.4
IR%	16.52	18.80	8.32	45.14
Gyp%	0.68	0.50	0.39	1.72
$CaCO_3$%	73.49	76.05	*91.86	42.33
Na^+ppm	3400	3350	2850	2500
(NaCl%)	0.86	0.85	0.72	0.64
K^+ppm	950	850	450	1800
Sr^{++}"	500	600	100	-
Mg^{++}"	4400	4400	*48000	4400
Fe^{tot}"	2100	1900	2800	11200
Al^{+++}"	2000	2300	1100	7300
$xsCa^{++}$"	10200	-	* -	500

* 36.42% dol.
52.92% $CaCO_3$.

TABLE 5.24 (cont.)

See fig. 5.8A

ZIG	2.1	2.2	2.3	2.4	2.5	2.6	2.7
IR%	24.32	46.90	13.66	82.82	76.68	81.26	73.26
Gyp%	55.93	40.19	72.25	0.74	4.12	2.11	0.38
$CaCO_3$%	5.12	2.33	-	3.95	8.60	6.98	*16.51
Na^+ppm	1200	900	1550	820	1120	1250	810
(NaCl%)	0.31	0.23	0.39	0.21	0.28	0.32	0.21
K^+ppm	680	800	350	1150	1020	1070	1200
Sr^{++}"	700	500	2700	100	100	20	100
Mg^{++}"	6400	5500	1400	7800	9300	7300	*14900
Fe^{tot}"	3000	3700	800	9900	5900	4500	3900
Al^{+++}"	2700	2900	700	4200	3700	3700	4100
$xsCa^{++}$"	27400	17100	27800	10500	7800	5200	* -

* 11.31% dol.
8.62% $CaCO_3$.

TABLE 5.25

	Powdery crust WG 2	WG 1	See fig. 5.8B SB 4	SB 5.1	SB 5.2	SB 5.3
IR%	30.06	12.94	24.30	10.98	89.28	22.68
Gyp%	57.55	73.14	59.17	24.84	1.33	65.84
$CaCO_3$%	1.38	1.38	7.82	64.37	2.76	4.60
Na^+ppm	1950	1350	1900	3250	550	1400
(NaCl%)	0.50	0.34	0.48	0.83	0.14	0.36
K^+ppm	2100	400	700	250	1500	600
Sr^{++}"	500	400	200	500	-	100
Mg^{++}"	3700	3000	5200	5700	3100	2100
Fe^{tot}"	6500	1400	2500	1600	7000	2500
Al^{+++}"	4500	1100	2000	4500	4500	1800
$xsCa^{++}$"	26500	32200	13000	1900	1900	18800

	Powder lW 4.2	Powder lW 5.1
IR%	34.94	21.98
Gyp%	58.73	69.00
$CaCO_3$%	1.39	0.47
Na^+ppm	1100	1550
(NaCl%)	0.28	0.39
K^+ppm	110	400
Sr^{++}"	600	800
Mg^{++}"	2300	2900
Fe^{tot}"	1300	1600
Al^{+++}"	1000	1000
$xsCa^{++}$"	13800	21500

TABLE 5.26

See fig. 5.8C

	EH 1	3	8.1	7.5	7.4	7.3	7.2	7.1
IR%	22.74	6.12	10.44	55.30	92.30	94.94	38.62	47.62
Gyp%	64.20	77.70	76.14	16.69	1.16	1.54	1.56	1.46
$CaCO_3$%	4.59	4.36	3.48	-	-	4.82	44.72	29.82
Na^+ppm	1650	1800	1500	3500	750	350	2800	2600
(NaCl%)	0.42	0.46	0.38	0.89	0.19	0.09	0.71	0.66
K^+ppm	600	300	550	2500	2650	1150	4650	11550
Sr^{++}"	400	500	600	200	400	1400	-	-
Mg^{++}"	800	400	800	1000	1400	1200	52800	38400
Fe^{tot}"	1700	600	800	11600	5700	4200	26600	32400
Al^{+++}"	1300	600	700	2200	3600	2500	13200	20300
$xsCa^{++}$"	12100	29600	18800	3100	-	-	-	-

TABLE 5.27

See fig. 5.8D

EH	6.1	6.2	6.3	6.4	6.5	6.6	6.7	5.1	5.2
IR%	51.90	71.76	43.44	69.72	50.04	43.36	46.94	15.04	28.64
Gyp%	36.09	17.22	43.00	22.67	40.44	48.89	32.91	62.55	46.74
$CaCO_3$%	11.70	3.90	2.75	1.83	4.59	8.03	2.29	3.44	11.93
Na^+ppm	1350	700	1200	900	1100	1150	1500	4100	3800
(NaCl%)	0.34	0.18	0.31	0.23	0.28	0.29	0.38	1.04	0.97
K^+ppm	1500	2800	1200	1600	1000	1000	3150	3400	1000
Sr^{++}"	900	200	3100	500	700	900	800	1200	1000
Mg^{++}"	2300	2500	1800	2100	1400	1300	6100	5600	2100
Fe^{tot}"	3200	3700	2300	3300	2300	1800	7100	8000	2300
Al^{+++}"	2500	4300	2100	3200	2000	1900	6000	6200	1600
$xsCa^{++}$"	-	4200	18900	5900	1500	-	28200	8600	17400

TABLE 5.28

See fig. 5.8E

See fig. 5.9A

	Clay							Surface crystals	Surface sand	Surface crust	Under-lying sand	Bedrock
	DI 4.1	1	4.2	ZER 1.2	1.3A	1.3B	1.1	TN 1.3	1.2	1.1	2	3.2
IR%	1.48	0.90	14.16	4.24	6.32	8.92	5.54	1.96	91.94	10.56	91.12	2.34
Gyp%	78.76	81.39	55.09	80.72	21.03	72.49	6.23	82.66	0.57	74.97	2.58	92.49
$CaCo_3$%	6.67	9.43	4.14	10.70	69.07	2.33	84.88	0.70	8.14	4.65	2.33	-
Na^+ppm	2200	2200	2800	1500	3650	2250	3000	1850	1000	1400	2400	1300
(NaCl%)	0.56	0.56	0.71	0.38	0.93	0.57	0.76	0.47	0.25	0.36	0.61	0.33
K^+ppm	340	400	820	250	400	650	400	150	580	300	300	150
Sr^{++}"	2200	4200	400	1300	400	3300	300	300	100	700	100	1300
Mg^{++}"	5200	4400	13200	800	2500	2400	2600	100	1200	1100	1100	200
Fe^{tot}"	-	100	600	400	1200	1600	1300	400	1900	500	1500	200
Al^{+++}"	-	400	1400	400	1400	2100	1500	300	1400	800	1500	300
$xsCa^{++}$"	30000	12800	49200	-	8500	29900	5600	32800	-	18900	6700	14700

TABLE 5.29

	Surface crust DTF 1.2	Turonian(?) clay 1.3	Bedrock 1.4	Gypsum bedrock 1.1
IR%	19.86	40.00	15.24	3.26
Gyp%	65.00	3.91	1.3 1	82.95
$CaCO_3$%	7.11	37.16	77.93	0.46
Na^+ppm	1600	2250	4000	1950
(NaCl%)	0.41	0.57	1.02	0.50
K^+ppm	800	2500	550	400
Sr^{++}"	1000	400	400	300
Mg^{++}"	1000	6100	5800	300
Fe^{tot}"	2300	13400	5900	3500
Al^{+++}"	1700	7600	1200	1000
$xsCa^{++}$"	12200	-	-	27000

TABLE 5.29 (cont.)

See fig. 5.9B

PF	1.1	1.2	1.3	1.4	1.5	1.6
IR%	1.24	15.98	27.64	10.82	64.58	47.76
Gyp%	86.29	70.63	52.92	53.67	*16.22	2.32
$CaCO_3$%	-	1.84	1.15	22.99	3.68	2.53
Na^+ppm	2550	2150	3900	9000	72000	130000
(NaCl%)	0.65	0.55	0.96	2.29	*18.30	33.05
K^+ppm	-	600	620	1100	4650	1750
Sr^{++}"	300	200	200	200	600	100
Mg^{++}"	200	2800	3000	4400	7600	6300
Fe^{tot}"	100	2200	2700	1500	9700	3000
Al^{+++}"	200	2500	2100	2000	8500	3800
$xsCa^{++}$"	29100	14200	32200	15000	1500	400

*Possible effect of thenardite or mirabilite.

TABLE 5.30

See figs. 5.4A and 5.4B

	PF 2.1	3	2.3	2.4	2.6	2.7	Clay 4.2	Selenite 5
IR%	1.84	28.52	68.76	78.00	37.02	33.90	24.80	0.56
Gyp4	87.39	66.35	2.30	0.69	47.19	2.61	12.79	85.05
$CaCO_3$%	0.92	2.99	0.23	2.30	2.76	23.22	25.29	3.45
Na^+ppm	4700	2300	37000	17000	6200	110000	80000	2400
(NaCl%)	1.19	0.58	9.41	4.32	1.58	27.96	20.34	0.61
K^+ppm	300	300	4850	4700	2000	2100	2600	-
Sr^{++}"	400	200	-	-	200	100	100	200
Mg^{++}"	2800	1100	17700	8000	6300	6800	3900	-
Fe^{tot}"	400	1500	6100	7300	6400	3700	2700	100
Al^{+++}"	400	1000	7700	9200	5400	4300	3400	100
$xsCa^{++}$"	28800	-	1700	-	55100	4900	15900	16200

TABLE 5.31

See fig. 5.9C

DT	1.1	7.1	1.2	1.3	1.6	1.4
IR%	4.40	3.44	66.66	1.66	2.54	5.42
Gyp%	78.10	65.58	1.66	*N.A.	76.01	72.91
$CaCO_3$%	2.97	14.16	23.56	*(51.37)	2.51	6.62
Na^+ppm	1700	2050	1600	1700	1920	2820
(NaCl%)	0.43	0.52	0.41	0.43	0.49	0.72
K^+ppm	760	600	1160	200	200	420
Sr^{++}"	800	800	200	400	2200	1200
Mg^{++}"	1000	600	6600	*68000	800	2600
Fe^{tot}"	400	400	3800	4600	200	400
Al^{+++}"	300	400	3000	600	200	1400
$xsCa^{++}$"	42400	40600	1800	*109800	31000	23800

*estimating that all Mg^{++} is as dolomite gives 51.59% $MgCa(CO_3)_2$ and that the excess Ca^{++} makes up the deficit 45% of the sample gives 3.5% $CaCO_3$, 41.5% $CaSO_4.2H_2O$.

TABLE 5.32

See fig. 5.9D

	DT 5.2A	5.3A	4	Bedrock gypsum 2.1	Bedrock gypsum 2.2	Bedrock gypsum 3.1	Fibrous gypsum 3.2	Fibres 7.2	Rosettes 6.2
IR%	1.78	5.68	4.50	0.80	3.10	0.70	3.00	4.96	6.18
Gyp%	86.25	76.70	75.17	92.92	86.05	90.96	79.10	84.77	77.33
$CaCO_3$%	0.46	2.97	3.65	4.34	1.83	0.23	0.23	2.05	1.14
Na^+ppm	1440	1820	1740	1400	1840	1420	1440	1750	1400
(NaCl%)	0.37	0.46	0.44	0.36	0.48	0.36	0.37	0.44	0.36
K^+ppm	60	160	80	100	400	80	320	450	500
Sr^{++}"	4000	1000	1000	800	2800	1600	400	400	400
Mg^{++}"	1000	3200	600	200	2600	200	800	4300	2400
Fe^{tot}"	200	200	200	200	400	200	200	1300	800
Al^{+++}"	400	600	200	400	250	300	400	600	600
$xsCa^{++}$"	13300	13600	30400	-	18300	13300	29900	19400	12400

TABLE 5.33

See fig. 5.9E

SB	2	3	6.1	6.2	1.1	1.2	1.3	1.4
IR%	9.64	14.00	1.76	10.86	10.30	8.90	10.70	0.50
Gyp%	77.86	74.10	88.20	73.66	57.11	63.70	63.31	0.90
$CaCO_3$%	3.22	3.45	1.61	5.74	10.34	17.24	6.67	*88.74
Na^+ppm	2200	1650	1400	1350	2200	1800	2600	4650
(NaCl%)	0.56	0.42	0.36	0.34	0.56	0.46	0.66	1.18
K^+ppm	280	350	50	280	310	350	360	150
Sr^{++}"	400	300	500	300	1100	1600	2000	400
Mg^{++}"	1900	1900	600	1200	3700	3600	5700	*20100
Fe^{tot}"	1200	1600	400	1500	1700	2600	2200	300
Al^{+++}"	1000	1200	300	1200	1100	1100	1300	300
$xsCa^{++}$"	9800	15700	12200	25500	21000	14700	29600	*10700

*15.25% dol.
82.59 $CaCO_3$.

TABLE 5.34

See fig. 5.9F

HA	1.2	2.1	2.2	2.3	1.1
IR%	11.26	1.06	3.26	4.50	1.98
Gyp%	55.51	88.88	85.81	83.87	84.39
$CaCo_3$%	0.95	trace	trace	-	3.81
Na^+ppm	1450	1350	1320	2050	1500
(NaCl%)	0.37	0.34	0.34	0.52	0.38
K^+ppm	830	10	100	100	100
Sr^{++}"	200	300	500	500	200
Mg^{++}"	4100	300	900	500	1300
Fe^{tot}"	7200	300	500	500	500
Al^{+++}"	2400	100	400	400	600
$xsCa^{++}$"	75000	25100	24200	22800	22300

TABLE 5.35

Evaporitic 7 Samples	MEAN (Figures in brackets are with CM 2 excluded)	TOTAL RANGE	MODAL CHARACTERISTICS
IR%	7.39 (4.59)	1.90 to 24.18 (11.02)	<5.0
Gyp%	63.97 (66.89)	45.88 to 80.39	50.0 to 80.0
$CaCO_3$%	7.97 (7.96)	0 to 31.19	<10.0
Na^+ppm	5200 (5800)	1400 (15007 to 15000	1500 to 15000
NaCl%	1.32 (1.47)	0.36 (0.38) to 3.81	0.38 to 3.81
K^+ppm	4200 (4500)	60 to 25000	<1000
Sr^{++}" T(4) N(3)	3250 (2500) 300	1000 to 5000 200 to 500	1000 to 5000 200 to 500
Mg^{++}"	6200 (6600)	500 to 15900	500 to 6000
Fe^{tot}"	2100 (2300)	360 to 9800	<2000
Al^{+++}"	1300 (1300)	0 to 4600	<1500

TABLE 5.35 (cont.)

Desert rose
croûte de nappe

22 Samples (Figures in brackets are with OT 4.3 excluded)

IR%	23.71 (24.78)	1.22 (6.82) to 29.26	20.0 to 30.0
Gyp%	66.35 (65.59)	49.14 to 83.19	50.0 to 70.0
$CaCO_3$%	4.01 (4.09)	0 to 18.75	<5.0
Na^+ppm	3200 (3200)	1100 to 8900	2000 to 6000
NaCl%	0.81 (0.81)	0.28 to 2.26	0.50 to 1.50
K^+ppm	750 (750)	150 to 1200	<1000
Sr^{++}" T(12) N(10)	2300 (2400) 950	400 to 11400 200 to 1900	700 to 3500 200 to 2000
Mg^{++}"	5800 (6100)	200 (400) to 15600	1500 to 5500
Fe^{tot}"	1850 (1950)	100 (600) to 5000	1000 to 2500
Al^{+++}"	2400 (2500)	100 (400) to 6300	1000 to 2500

TABLE 5.36

Mesocrystalline subsurface crusts	MEAN	TOTAL RANGE	MODAL CHARACTERISTICS
40 Samples	(Figures in brackets are with OL 1.6, LS 1.2 and 1.3 excluded)		
IR%	10.51 (10.29)	0.98 to 30.34	<15.0
Gyp%	69.41 (68.83)	50.01 to 95.02 (88.99)	65.0 to 90.0
$CaCO_3$%	5.90 (6.31)	0 to 29.70	< 5.0
Na^+ppm	2450 (2500)	1150 to 9000	<2000
NaCl%	0.62 (0.64)	0.29 to 2.29	<0.50
K^+ppm	670 (690)	50 to 5500	< 500
Sr^{++}" T(28) N(12)	2200 (2300) 550	600 to 6700 100 to 2000	1500 to 3000 100 to 2000
Mg^{++}"	3250 (3400)	100 to 13200	100 to 6500
Fe^{tot}"	1800 (1840)	0 to 10000	<1500
Al^{+++}"	1650 (1650)	150 to 9900	<1500

TABLE 5.36 (cont.)

Powdery accumulations

40 Samples

IR%	19.70	0.54 to 60.34	up to 60.0
Gyp%	60.98	19.74 to 88.09	20.0 to 90.0
$CaCO_3$%	5.84	0 to 64.37	< 10.0
Na^+ppm	2350	700 to 12000	< 2000
NaCl%	0.60	0.18 to 305	< 0.50
K^+ppm	620	100 to 2100	< 1500
Sr^{++}" T(26)	1300	500 to 3100	500 to 2000
N(14)	550	0 to 1800	< 1000
Mg^{++}"	4300	60 to 15700	up to 6000
Fe^{tot}"	1960	60 to 9000	< 3000
Al^{+++}"	1550	100 to 4500	< 3000

TABLE 5.37

Columnar and other surface crusts 25 Samples	MEAN	TOTAL RANGE	MODAL CHARACTERISTICS
IR%	7.83	0.50 to 28.52	< 10.0
Gyp%	77.46	59.17 to 88.85	70.0 to 90.0
$CaCO_3$%	4.52	0 to 10.70	< 10.0
Na^+ppm	2000	1400 to 4700	< 4000
NaCl%	0.51	0.36 to 1.19	< 1.0
K^+ppm	380	0 to 1500	< 800
Sr^{++}" T(20) N(5)	2000 260	400 to 8900 200 to 400	400 to 2500 200 to 400
Mg^{++}"	2000	0 to 5400	500 to 4000
Fe^{tot}"	700	0 to 2500	< 1000
Al^{+++}"	700	0 to 2100	< 1000

TABLE 5.37 (cont.)

Cobbles

19 Samples

IR%	5.28	0.19 to 14.00	< 10.0
Gyp%	78.15	55.51 to 91.21	75.0 to 90.0
$CaCO_3$%	3.05	0 to 10.34	< 5.0
Na^+ppm	1700	1350 to 2600	< 2000
NaCl%	0.43	0.34 to 0.66	< 0.50
K^+ppm	260	10 to 760	< 400
Sr^{++}" T(11)	1450	500 to 4000	500 to 2000
N(8)	410	200 to 1100	200 to 500
Mg^{++}"	1560	0 to 8800	< 2000
Fe^{tot}"	1330	100 to 8900	< 500
Al^{+++}"	800	100 to 3500	< 1000

CHAPTER 6

THE ORIGINS OF GYPSUM CRUSTS

As was pointed out in chapter three, the various areas of the world which exhibit well developed gypsum crusts have provided a wide variety of hypotheses as to their origin. In Australia, most workers have advocated an aeolian model involving deflation of gypsum-rich lake sediments and their deposition on the surrounding landscape (Jack, 1921; De la Hunty and Low, 1958; Jessup, 1960 a., 1961 a.). Though Coque (1955 a.b., 1958, 1960, 1962), Coque and Jauzein (1966, 1967), Page (1972) and Vieillefon (1976) proposed similar models to account for the surface crusts of southern Tunisia, most earlier workers in north Africa supported hypotheses based on capillary rise or fluctuation of groundwater. In contrast, the gypsum forming the crusts of the Central Namib Desert was held to be authigenic (Martin, 1963; Scholz, 1963, 1968; Besler, 1972). It will be shown here that the three main types of gypsum crust, as outlined in chapter five, have distinct origins and that the processes involved are the same in all desert areas

In order to do this, the genetic models will be grouped into five categories and each will be examined to assess their applicability to the different types of gypsum crust. The first category is that of models involving lacustrine and lagoonal sedimentation, deep-water and shallow-water and the sabkha type. Secondly, the crust forming processes involving capillary rise of water saturated with soluble salts from the groundwater table or soil moisture will be examined under the heading of per ascensum models. Thirdly, lateral movements of gypsum-rich materials and waters will be assessed as a potential source of the crust forming deposits. Fourthly, authigenic gypsum crust development will be discussed and finally, in chapter seven, several models involving downward movement of gypsum from the surface into the soil zone will be described and developed into a model which explains the genesis of several of the forms of gypsum crust which have

been examined. These are termed the per descensum models.

a. Lacustrine and lagoonal models.

i. Deep-water evaporite sedimentation (see fig. 6.1A).

Though Durand (1963) advocated the deposition of powdery calcium carbonate and gypsum sediments in deep lakes and Stainier (1912) held that many of the gypsum crusts in the Algerian Souf were lacustrine, the evidence presented in chapter two argues against this. High-magnesium calcite is deposited on the floors of the Persian Gulf and Red Sea but, probably owing to the presence of sulphate reducing bacteria in the surface waters, gypsum is rarely deposited in waters more than 1.0 m in depth, since it is altered to calcite. Moreover, the locations of many gypsum crusts in the two field areas are not compatible with a lacustrine origin. It has been pointed out that the lowest of the glacis identified around the Chotts of southern Tunisia is not covered by a gypsum crust. Assuming a maximum lake depth of 10 m, many of the crusts are well above the most recent high stands, indeed many are at higher altitudes than the uppermost Cardium bed which may be of Mio-Pliocene age. In effect, the only crust type which is found exclusively within the feasible Chott transgression zone is the evaporitic. In the Sabkha el Melah, near Zarzis, Busson and Perthuisot (1977) suggested a maximum water depth of 10 m but the restriction of gypsum deposits to the periphery and to small islets indicates a shallow subaqueous or subaerial origin. In the Central Namib Desert, tectonic disturbances have complicated the correlation of raised beaches along the coast. Nevertheless, it is unlikely that anywhere in the field area land over 18 m above present sea level has been within the tidal range within the last 30,000 years. Most of the gypsum crusts are above this height and on the Lagunenberg they are found up to 100 m or more above sea level. Only the evaporitic crust sample from the White Lady Pan (CC 2.2) and the desert rose croûte de nappe samples from Wlotzkas Baken could be lagoonal deposits. Further

inland in the Central Namib, the topography precludes the previous existence of any deep lakes. The evaporite gypsum deposits from Narabeb and possibly the Hamilton Range (HA 3) could not have been deposited in water bodies more than 1.0 m to 2.0 m in depth.

ii. Shallow-water evaporites (see fig. 6.1B).

While an evaporitic origin for the bulk of the gypsum crusts described from the field areas may be discounted on the grounds of topography, several sabkha, chott and pan deposits may be the products of shallow-water sedimentation.

The two samples from Sabkha el Melah (CM 2 and 5.1, table 5.7) are different in their composition as well as structure. The bedded selenite (CM 2) has more insoluble residue, calcium carbonate and accessory elements, notably potassium, strontium and magnesium. Large amounts of included pebbles suggest a subterranean development possibly under the influence of terrestrial water enriched in minerals derived from bedrock. The thick gypsum deposit (CM 5) may represent an older evaporite laid down during the early phases of sedimentation in the sabkha. Busson and Perthuisot (1977) and Perthuisot (1980) proposed a date of 6,000 years B.P. for the onset of sedimentation. Limited enrichment of the accessory elements may result from a marine origin of the earliest brines. Present sabkha waters reach the surface in salt springs (aïoun). Their composition, determined by atomic absorption spectrophotometry, flame photometry and chloride by titration with silver nitrate using potassium chromate as an indicator, is shown in table 6.1. When these are compared with the chemical analyses of sediments from the surface of the sabkha it is apparent that in the peripheral zone (CM 3) the calcium, magnesium and strontium constituents are precipitated while in the hyper-saline areas (CM 4.1) only the sodium and some magnesium remains to be deposited. The halite deposits are pink in colour owing to the presence of dinoflagellates, possibly Dunaliella salina (Bonython, 1956). The bedded selenite

(CM 2) is a recent evaporite containing minerals of terrestrial origin while the thick bed forming the islet (CM 5.1) is not. The clay at the base of this deposit (CM 5.3) is within the present flood range and may contain some recent precipitates. Alternatively, it may be an early terrestrial sediment upon which the "marine" Gypsum was laid down. The excess magnesium in the clay indicates the presence of other minerals, possibly huntite, though none were evident in the X-ray diffraction analysis.

The high strontium and magnesium levels in the samples from Chott el Rahim (CR, table 5.7) indicate that these gypsum deposits are also composed of minerals of terrestrial origin. Water samples from the centre of Chott Djerid (see table 6.2) have a high strontium and magnesium content as have the gypsum crusts. The standards are samples of deionized water treated, in the same way as the natural water samples, with 3 drops of chloroform to prevent algal growth and enough hydrochloric acid (1.0 N) to lower pH to less than 2.0 in order to minimize precipitation of calcium salts. It is interesting to note that surface waters flowing into the Chott at Blidet in the Nefzaoua contain considerably less magnesium, sodium and potassium than that at the centre of the Chott while calcium, strontium and iron show a proportionally smaller increase toward the centre. This reflects the loss of these elements by precipitation as the brine evaporates becoming enriched in the most soluble minerals as it retreats to the lowest lying areas. Since some supratidal sabkha deposits may be varved owing to the structure of the host sediment or algal mat development at or beneath the surface, this mode of origin cannot be discounted. In the case of the bedded deposits from Chott el Rahim, however, the absence of clastic material in the alabastrine horizons, indeed throughout the deposit, argues against an origin similar to that of the sabkha deposits of the Persian Gulf. Post-depositional enrichment by displacement of the host sediment cannot be ruled out but the well defined crystalline varving argues in favour of a subaqueous or subaerial origin in an intermittantly flooded basin. The flood waters are supplied by inflowing

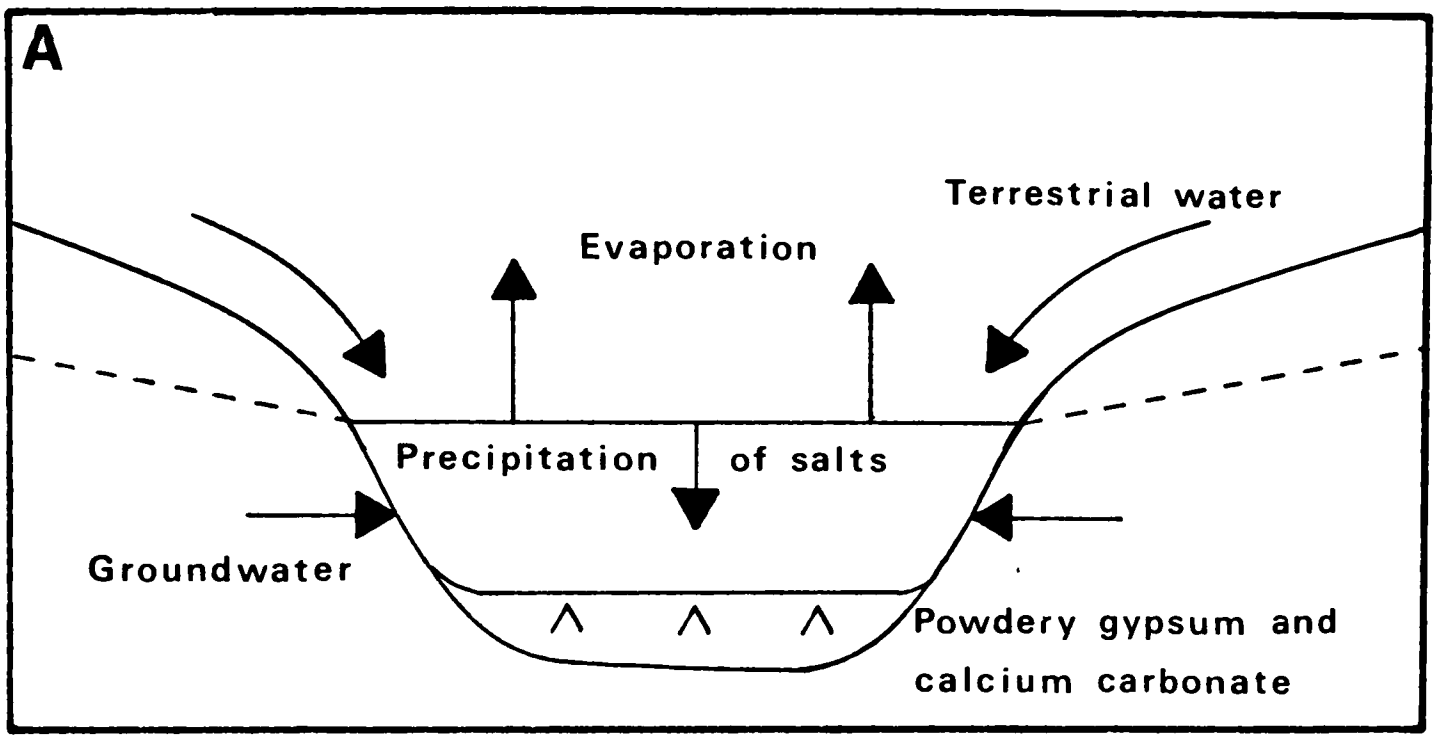
surface waters or rising groundwaters which contain soluble minerals and accessory elements derived from the bedrock within the catchment. The chemistry of the gypsum crusts is not in keeping with those derived from the evaporation of marine water.

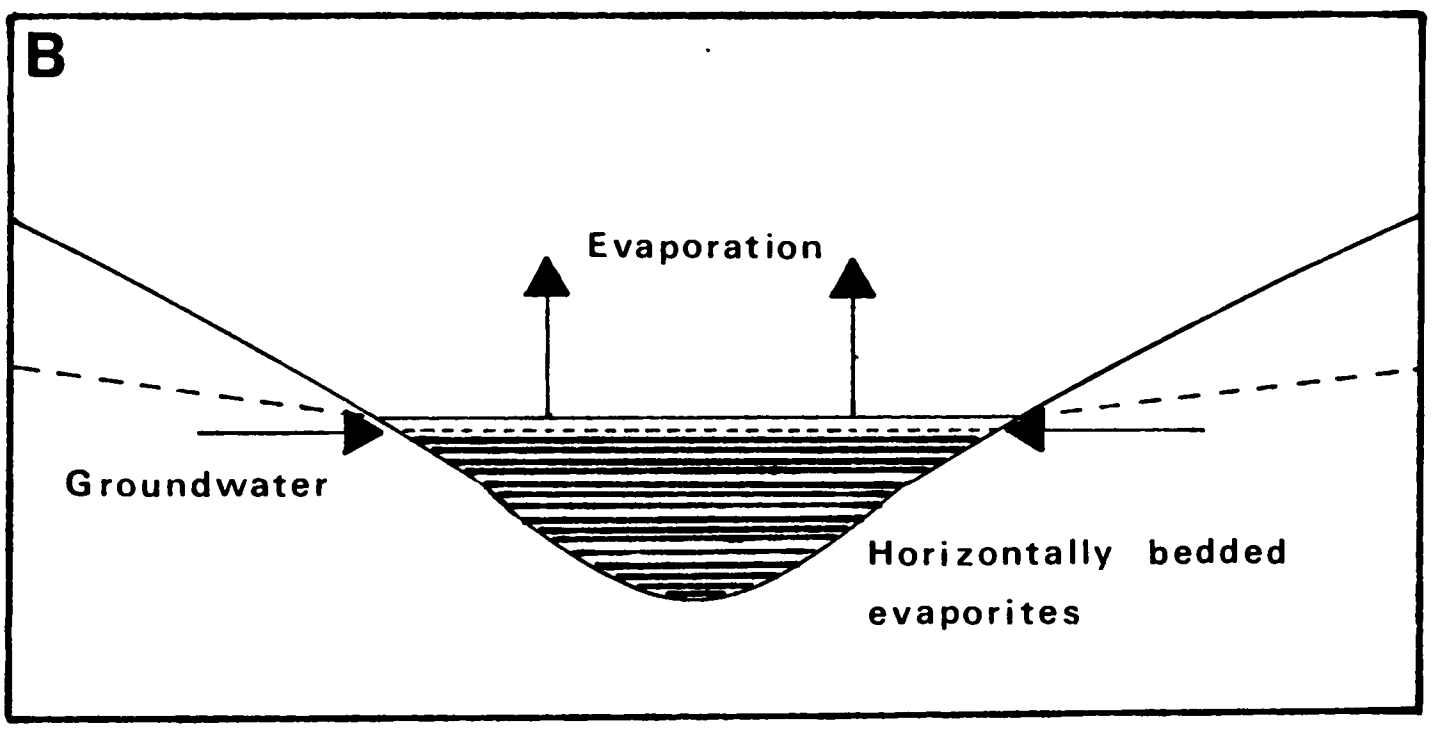
iii. Sabkha evaporite facies (see fig. 6.1C).

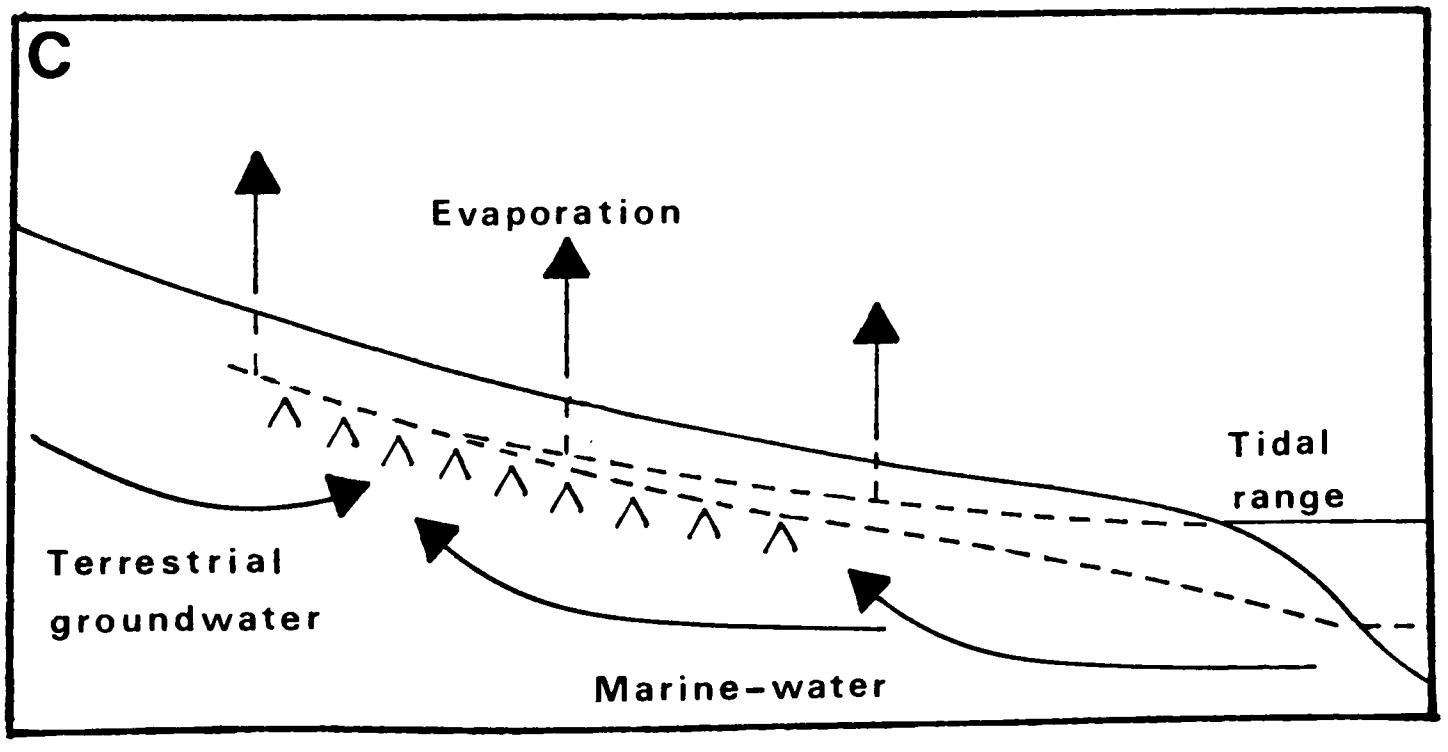
The samples of evaporite gypsum crusts from the Central Namib Desert contain more insoluble residue, in two cases, than those of southern Tunisia. That from the White Lady Salt Pan (CC 2.2, table 5.7) may represent a sabkha deposit since it comes from a gently sloping lagoonal area in many ways similar, though on a much smaller scale, to those around the Persian Gulf. The present pan surface does not represent a supratidal sabkha, in that it is occasionally inundated by storm tides when halite is deposited from the evaporating waters trapped behind the coastal bar. Samples of the halite (CC 2.1 and 2.3, table 6.1) from the lowest levels of the pan show a loss of calcium carbonate and gypsum relative to sea-water. These salts are deposited in peripheral zones as in the Sabkha el Melah. Besides certain wind blown deposits making up the insoluble residue and some accessory elements, the only major constituents are highly soluble salts. The gypsum crust has well developed varves formed by beds of alabastrine and lenticular crystals which suggest cyclic shallow subaqueous or subaerial precipitation Mineralogically, the sample is complex owing to the presence of fragments of basaltic bedrock. The absence of carbonates indicates that most of the accessory minerals are derived from these rather than from soluble minerals. Nevertheless, the low strontium content argues against the considerable brine enrichment characteristic of most sabkha facies. Significantly higher strontium levels in the desert rose croûte de nappe at Wlotzkas Baken than in other examples of this form of gypsum crust in the Central Namib, may be a result of mineral enrichment in a sabkha environment. The proximity of the crust to the 17 m raised beach lends support to this hypothesis. Anhydrite is not encountered in any of the

Figure 6.1 (overleaf)

Models of lacustrine and lagoonal evaporite sedimentation.
A. Deep-water sedimentation (after Durand, 1963).
B. Shallow-water sedimentation.
C. Sabkha sedimentation.


A
Terrestrial water
Evaporation
Precipitation
of salts
Groundwater
Powdery gypsum and
calcium carbonate


B
Evaporation
Groundwater
Horizontally bedded
evaporites


C
Evaporation
Tidal
range
Terrestrial
groundwater
Marine-water

Tunisian or Namibian samples that have been analyzed. While the low mean annual temperatures in both field areas may account for this, it is possible that its absence testifies to the lack of true supratidal sabkha environments.

The gypsum deposits near Narabeb and perhaps those from near the Hamilton Range are from small, shallow enclosed basins and differ in this respect, as well as their structure, from the aforementioned evaporitic crusts. The underlying sediments in these pans are partially consolidated sands and gravels within which the soluble salts crystallize. Since the bedrock does not contain evaporitic minerals, those in the pans must be derived from atmospheric salts deposited on the surface of the catchment. Seepage within the upper soil zone will carry the salts, dissolved in rainwater, to the depressions where evaporation at the surface or at a shallow depth will result in their precipitation. This process closely resembles that occurring in supratidal sabkha environments where progressive brine enrichment results in evaporite sedimentation. In the pans of the Central Namib, salts more soluble than gypsum have been removed through leaching by meteoric waters undersaturated in soluble salts.

The processes involved in the development of these crusts are the same, in some respects, to those forming desert rose croûtes de nappe in that precipitation and cementation occur at the top of an evaporating or fluctuating groundwater table. A distinction must be drawn between sabkha evaporites and croûtes de nappe. The principal difference is that of local topography rather than structure or process, hence it is arbitrary. Sabkha evaporites are confined to the lowest lying areas of closed depressions or littoral zones while croûtes de nappe, as defined here, are less restricted, developing in areas where the water table approaches the land surface and not necessarily within distinct topographic basins. In this sense, subsurface sabkha deposits are varieties of croûtes de nappe.

b. Per ascensum models.

i. Capillary rise (see fig. 6.2A).

Models of desert crust development invoking the upward movement of groundwater and soil moisture by capillarity, and the precipitation of soluble salts at or near the surface as the water evaporates, have a long history. Many of the hypotheses involving this process were formulated in north Africa to explain the evolution of calcretes. Agafonoff (1936 a.) and Flandrin, Gautier and Lafitte (1948) accepted the veracity of Pomel and Pouyanne's (1882) capillary rise model to account for the calcretes of Algeria and Tunisia. Furthermore, while Durand (1949 b., 1951) and Despois (1954) identified a wide variety of calcrete types with different genetic mechanisms, they agreed that the hypothesis was valid for explaining the origin of the most prevalant forms of encroûtement. Scholz (1971) advocated the model in relation to the surface calcretes of the Central Namib. Opinions were not, however, unanimous by any means. Jaranoff (1937) working in Morocco, Gaucher (1943 a.b.c., 1958) and Gigout (1958, 1960) in Algeria and Fedoroff (1961) in France argued that the model was inadequate in many cases and that the process was perhaps unfeasible.

Several workers have attributed the formation of gypsum crusts to the same processes as calcretes, Glinka (1927) and Nikiforoff (1937) advocating capillarity in most cases and Durand (1963) in the case of powdery crusts. Detailed field studies, such as those of Stainier (1912) in Algeria and Gobert (1948) in southern Tunisia showed, however, that a variety of processes are involved in their formation, some not comparable with those of calcrete development. Durand's (1949 a.) assertion, quoting Flamant, that

"la croûte est le résultat d'une incrustation superficielle par suite de l'évaporation des eaux plus ou moins saline et séleniteuse qui suintent dans les capillaires"

has been supported by Dalloni (1953) in north Africa, Fathi et al., (1972) in Egypt and Petrov (1976) in Central Asia, despite increasing opposition to the

model. Tricart and Cailloux (1960, pp. 147-151) pointed out that in order to precipitate a gypsum crust 0.50 m thick, a depth of 200 m of water would have to evaporate. Even assuming maximum gypsum solubility in a sodium chloride-rich solution, which Tricart and Cailloux did not, groundwater replenishment of 40 m is much greater than could be expected in areas with annual deficits of potential evaporation over precipitation. Without replenishment water from about 100 m of underlying sediment would have to be drawn up (Holmes, 1969). While their observation that predominantly German workers have advocated the capillary rise model is supported by the conclusions reached by Kaiser and Neumaier (1932) and Knetsch (1937) in the Namib Desert, Meckelein (1959) in Libya and Beckmann, Scharpenseel and Stephan (1972) and Schwenk (1977) in southern Tunisia, many French workers have also endorsed it. In order to assess its applicability, it is necessary to examine the mechanism that has been proposed.

Using a 1.90 m high lysimeter with a surface area of 4.0 m^2, Roseau (1953) found that water evaporating at the surface is replaced from beneath. Plet-Lajoux, Monnier and Pedro (1971) simulated gypsum precipitation from groundwater saturated with gypsum by subjecting columns of quartz sand to surface temperatures of 44^{o}C to 68^{o}C, using infra-red lamps, equivalent to potential evaporation of 10 mm/day. They reported the development of a 5.0 mm thick crust with 32% gypsum at the top of the capillary fringe after 49 days. Since the water supply was replenished from beneath, maintaining the head of pressure, and the height of any upward seepage was not recorded, it is impossible to assess the role of soil capillarity. The mechanism requires the maintenance of porosity for the solutions to be drawn upwards along the negative pressure gradient produced by surface evaporation. This is achieved because crystallization is at quartz grain contacts resulting in displacive accretion. Nevertheless, the workers noted a decrease in evaporation over time which suggests that the process may only account for thin crusts since eventually a seal is created. In contrast to the conclusions of Roseau (1953) and the assumptions inherent in the work of Plet-Lajoux, Monnier and Pedro (1971),

Keen (1936) pointed out that capillaries do not exist in soil horizons. Voids between grains form a cellular structure and upward water movement is limited to that resulting from tension between the surface of soil particles and the solution. This is never more than 20 mm to 30 mm, though during experiments conducted in narrow cylinders, surface tension at the walls may increase this figure. He stated that

"evaporation from the surface proceeds by the progressive drying out of the top layers of the soil rather than upward movement of water from below to the surface."

Even if potential evaporation is high, 3650 mm annually in southern Tunisia (Schwenk, 1977), and the soil materials porous, extensive gypsum crust development cannot be attributed to capillary rise. A further argument against the capillary model was presented by Coque (1962,pp. 91-93) who pointed out that many of the surface gypsum crusts of southern Tunisia are located on slopes of up to 20^{o} and upon hills and high sand dune crests. Groundwater, in sufficient quantity to allow these crusts to develop, cannot form a water table which parallels the land surface so precisely over such varied topography.

ii. Crystallization from evaporating soil moisture (see fig. 6.2B).

In order to account for the development of gypsum crusts on steep slopes and in areas of high relief while accepting the capillary rise model, Bureau and Roederer (1961) proposed an alternative source of gypsum saturated water. They held that the gypsum-rich Mio-Pliocene sediments of southern Tunisia developed crusts at their surface through the rise, by capillarity, of soil moisture. The water, derived from antecedent rains, which replenish the soil moisture deficit, is drawn up to the surface by capillarity in the succeeding dry phase. During its residence in the soil and surface bedrock, the moisture becomes saturated with gypsum, especially if some lateral flow occurs, and this is precipitated at or near the surface when evaporation takes place. Gibson (1962) envisages a similar process on the gypsum encrusted moraines of South Victoria Land, Antarctica. Once

again, however, the validity of the model is put in doubt by the absence of any convincing evidence of significant upward moisture movement resulting from soil capillarity.

It is feasible that efflorescences of gypsum on surfaces adjacent to chotts and sabkhas, where groundwater is near the surface, result from a small amount of capillary rise. Russell (1885) described extensive examples around the playas of Nevada and Stephan (1972) mentioned gypsum beneath glauberite ($Na_2Ca(SO_4)_2$) in the Gafsa region of southern Tunisia. It is impossible, however, to determine whether the saturated waters arrive at the surface by capillarity or a rise in the level of the water table to the land surface through recharge. The sodium chloride efflorescences on the silts of the Kuiseb at Natab (see chapter four) develop when laterally flowing soil moisture seeps to the surface on the sides of the terraces. Whilst highly soluble salts such as these will be dissolved by the next rainfall, gypsum efflorescences are more resistant to solution. The gypsum crystals found at the surface in areas adjacent to Chott Djerid between Tozeur and Nefta in Tunisia (TN 1.3) may be efflorescences since the absence of inclusions is unusual in large crystals of subsurface origin. Similarly, the 5.0 mm thick pellicles of crystals consolidating the surfaces of gypsum dunes in the Nefzaoua are the products of continual solution of gypsum by dew followed by migration of gypsum ions to growing crystal faces during evaporation. In all these cases, the crusts are thin accumulations found in areas with very shallow water tables or upon gypsiferous sediments. They cannot be compared with gypsum crusts of greater dimensions which mantle a great diversity of deposits and terrain. Indeed, the mechanisms involved cannot account for accumulations other than thin, often ephemeral, features.

iii. Groundwater evaporation and fluctuation.

Accepting the limitations of models of crust development involving precipitation

of soluble salts from groundwater, owing to the topographic distribution of some forms, some crusts are closely related to the groundwater table. The gypsum croûte de nappe forms described in chapter five are nearly all located in areas of low relief where the water table approaches within 1.0 m to 5.0 m of the surface. Durand (1949 a.,1963), Bellair (1954), Le Houérou (1960), Coque (1962), Conrad (1969) and Kulke (1974) advocated groundwater evaporation as the mechanism involved in subsurface calcrete and gypsum crust formation. While some workers invoked small amounts of capillary rise (Tolchel'nikov, 1962; Pouget, 1968; Kulke, 1974), generally less than 1.0 m, the process involves loss of moisture by evaporation at the groundwater/soil atmosphere interface with resultant salt precipitation. Kondorskaya (1967), Vieillefon (1976) and Horta (1979) attribute all hydromorphic soils to this process. It would appear to be a suitable model for gypsum croûtes de nappe development including coastal sabkha evaporites (Shearman, 1963; Kinsman, 1964 a.b., 1965, 1969; Butler, 1969; Kendall and Skipwith, 1969).

Several problems are encountered, however, when the empirical evidence for groundwater evaporation is examined. While Ambroggli (1966) held that, in the Sahara, groundwater evaporation occurs at depths of up to 20 m, detailed studies of sand reservoirs in the Namib Desert contradict this. Owing to high rates of evaporation open reservoir storage is insufficient in warm desert areas. To overcome the problems, dammed rivers are allowed to fill with sand and silt and the resulting aquifer is tapped. Wipplinger (1958) found that while open storage dams lost 3.0 m of water/year through evaporation and seepage, artificial sand reservoirs experienced little evaporation after the first 40 days following flooding. Stengel (1968 a.b.) found that in fine sands and silts, water at a depth of more than 1.0 m did not evaporate. This is because in fine sediments the connections between the interstices are too narrow and circuitous to allow access of surface air and free circulation of the soil atmosphere. Schulze (1968) found that the diurnal temperature range at 1.2 m beneath the surface is never

more than 0.1°C. The annual fluctuation is only 4.7°C, about an average of 27.1°C. This is insufficient to allow all but minor amounts of evaporation. In coarse sediments, such as gravels, the maximum depth of evaporation will be greater owing to the larger dimensions of the interstitial cells but Keen (1936) held that in most soils devoid of vegetation, evaporation will not proceed to a depth greater than 1.5 m (see fig. 6.2C).

An alternative model to account for the precipitation of soluble salts from the surface of a groundwater table was developed by Pouget (1968). Rather than invoke evaporation of groundwater, Pimienta (1950) and Akhvlediani (1962) proposed that soluble salts crystallized from fluctuating groundwater tables, in the former case calcium carbonate in the vicinity of Tunis and in the latter gypsum in Central Asia. The problem of accounting for precipitation remains. Either the net volume of the solvent, that is water, is reduced thereby liberating the excess solutes when their maximum solubility is reached, or the solubility of the solutes is lowered by changes in water chemistry, temperature or pressure. It has been shown in chapter one that minor fluctuations in temperature and pressure, such as those which may be anticipated within the soil zone, do not greatly affect the solubility of gypsum. Furthermore, as pointed out above, evaporation of the solvent will only occur within about 1.0 m to 1.5 m of the surface. In effect, the only viable explanation for the precipitation of gypsum at the surface of a water table is of chemical reduction of its solubility. Pouget (1968) postulated two mechanisms to account for this. Firstly, within the soil zone plant roots may liberate carbon dioxide which, when dissolved in the groundwater, will increase the capacity for calcium carbonate solution. Dissolution of calcium carbonate will in turn reduce gypsum solubility owing to the common ion effect (Wigley, 1973, see chapter one). This process may account for the abundance of gypsum pedoforms in many surface and subsurface mesocrystalline crusts but they are not evident in croûtes de nappe owing to the large dimensions of the gypsum crystals. Bellair (1958) reported the development of crusts up to 1.0 m thick over 15 to 20

years in the Algerian Souf. He attributed their development to precipitation from a falling water table but the cause of this, overplanting of date palms, which may take up 1000 l of water daily, provides a possible mechanism for crust accretion. It should also be noted that Jaranoff (1937) and Boulaine (1961 a.) suggested that osmotic pressure around plant roots would result in calcium carbonate crystallization. This would increase gypsum solubility. The second hypothesis involves a rise in groundwater table into soil horizons with low sodium chloride concentrations. When this occurs, the dilution of the solution with respect to halite decreases gypsum solubility and results in its precipitation (see fig. 6.2D). This model, however, rests on the assumption that the waters recharging the groundwater have low sodium chloride concentrations but are saturated with gypsum. Since sodium chloride levels are low, so also will be the amount of gypsum in solution and the resident groundwater will be diluted in terms of both solutes. Though Pouget (1968) did not take this into account, at some point during dilution the waters will become supersaturated with gypsum since the gypsum/sodium chloride ratio increases (see fig. 6.3). Risacher (1978) described the reverse process in the Jayu Kkota saline in Bolivia where gypsum precipitation from laterally flowing groundwater ceases in the zone peripheral to the pan where sodium chloride is present in the soil zone. Krupkin (1963) found that in some cases the rise of sodium chloride may result in the precipitation of gypsum. The sodium chloride will react with calcium in the soil adsorption complex forming calcium chlorides which may in turn react with available sodium sulphate forming gypsum (see chapter two).

The structural and chemical characteristics of gypsum croûtes de nappe provide support for this second hypothesis. Pouget (1968) argued that the development of desert roses depends not on precipitation from highly saturated solutions but rather from waters which do not experience major chemical changes over annual cycles. A fluctuating water table which precipitates gypsum only during high

Figure 6.2 (overleaf)

Models of gypsum crust development invoking upward movement of saline moisture.
A. Capillary rise of saline groundwater resulting in the accumulation of a surface or near-surface gypsum crust.
B. Capillary rise of saline soil moisture resulting in the accumulation of a surface gypsum crust.
C. Restricted capillary fringe of soil moisture immediately above the water table. Evaporation occurs from the surface downwards and not by evaporative pumping and capillary rise (after Keen, 1936).
D. Precipitation of gypsum as a result of reduction of gypsum solubility through a decrease in sodium chloride content (after Pouget, 1968).

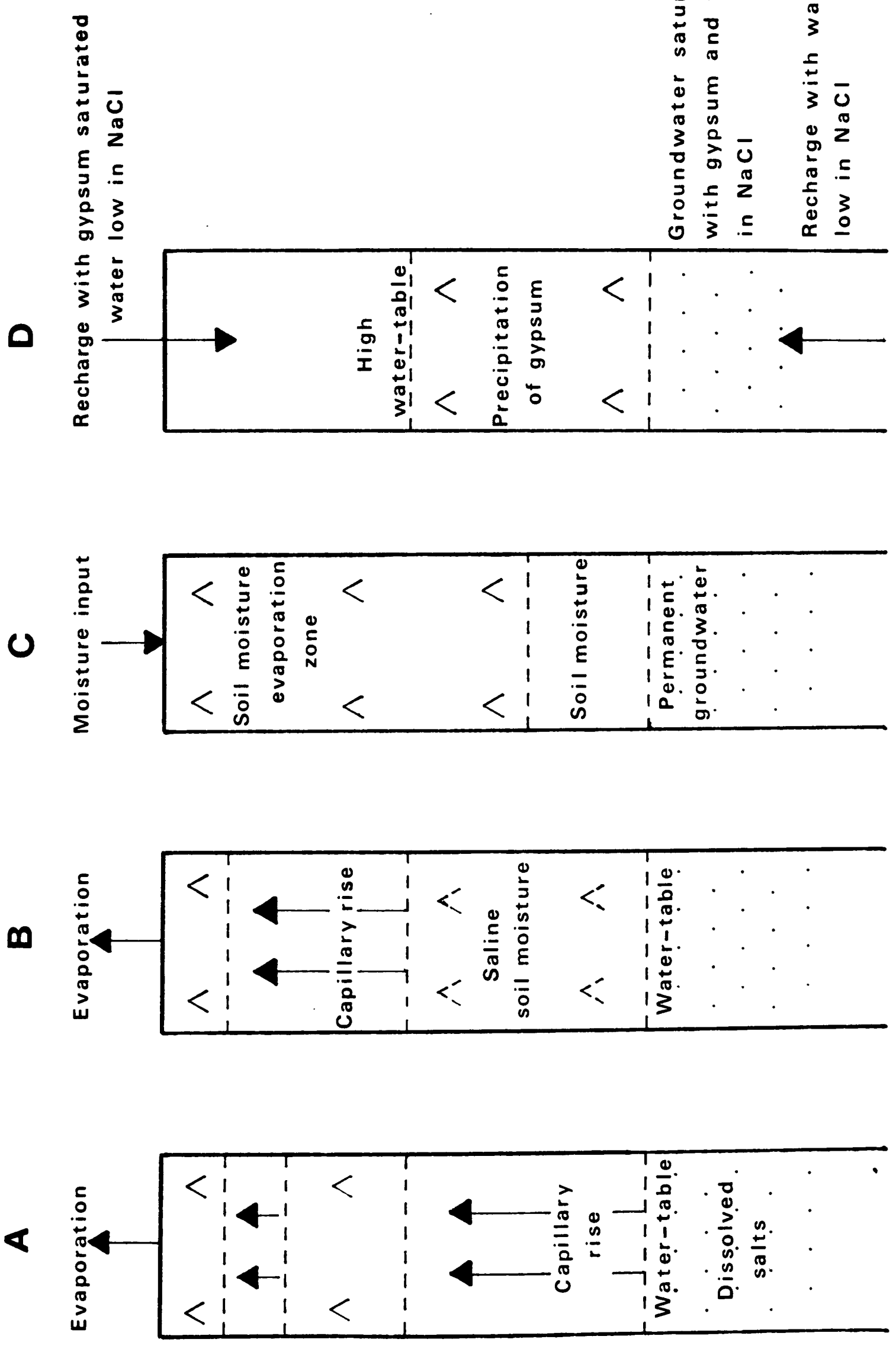
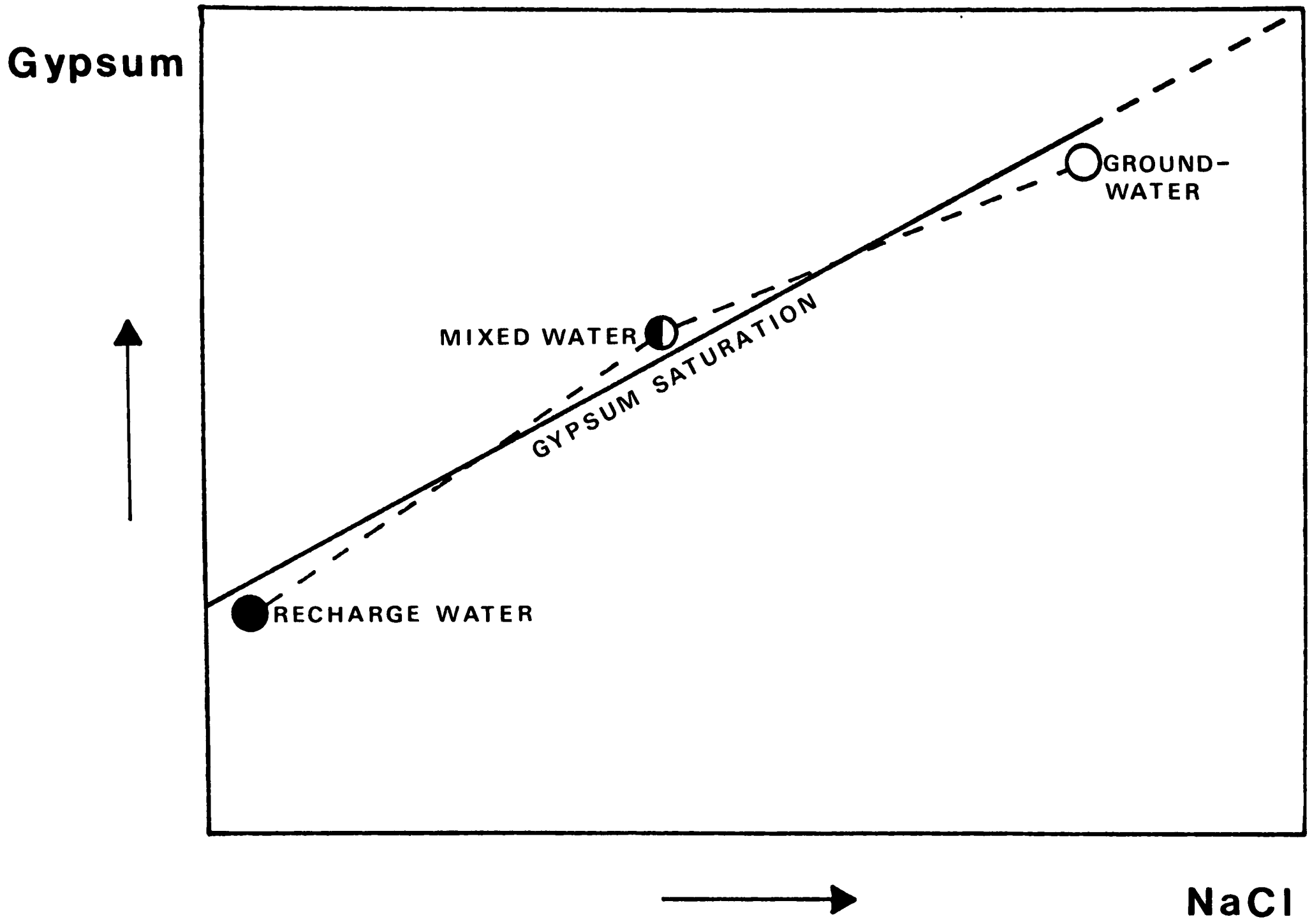

A
Evaporation
Capillary rise
Water-table
Dissolved salts
B
Evaporation
Capillary rise
Saline soil moisture
Water-table
C
Moisture input
Soil moisture evaporation zone
Soil moisture
Permanent groundwater
D
Recharge with gypsum saturated water low in NaCl
High water-table
Precipitation of gypsum
Groundwater saturated with gypsum and high in NaCl
Recharge with water low in NaCl

Figure 6.3 (overleaf)

Hypothetical consequence of the mixing of two gypsum-rich waters, one low in sodium chloride the other high. The mixed water is saturated with gypsum and should precipitate the mineral (after Pouget, 1968).


Gypsum
GROUND-
WATER
MIXED WATER
GYPSUM SATURATION
RECHARGE WATER
NaCl

stands will produce small crystals because growing surfaces will lose much of their ionic attraction during intervening dry conditions. This attraction will not be renewed on the same faces during the next period of saturation. This may explain the mesocrystalline and sometimes alabastrine textures of the crusts at Zaafrane (ZAA) and El Faouar (EF). While Cody (1979) held that the lenticular gypsum habit results from the presence of soluble green plant material in the solutions, this would seem an insignificant factor in the development of desert roses from groundwater in desert areas. Other ions must contribute to the differential growth of crystal faces. Sodium ion adsorption on the calcium ion face (111), the unit hemipyramid, would account for its underdevelopment in lenticular crystals. The lack of growth on the cleavage plane face (010), the clinopinacoid, is attributable to the fact that it is composed of relatively inert water molecules. The preferential growth of ($\bar{1}02$) is, however, enigmatic. The presence of quartz grains included poikilitically within lenticular crystals making up desert rose croûtes de nappe lends further support to Pouget's (1968) model of growth from chemically stable waters. Kastner (1970) found that inclusions of clasts occurred during rapid crystal growth when ionic migration exceeds the rate below which grains are pushed ahead of the face. In effect, poikilitic inclusions attest to rapid crystallization under saturated conditions which allow free ionic migration. The chemistry of desert rose croûtes de nappe reflects the abundance of insoluble residue often in the form of poikilitic inclusions. Beyond this, however, higher sodium levels compared with all other forms of gypsum crusts, except the evaporitic type, suggest that the influence of sodium chloride is significant. Strontium levels are relatively high. This is also a trait of gypsum crystals precipitated rapidly from chemically stable solutions which facilitate continuous free ionic migration for long periods (Kulke, 1974).

These characteristics suggest that the desert rose form of croûte de nappe

originates through gypsum crystallization near the surface of a groundwater table. To assess the validity of the model further it is necessary to examine the chemistry of groundwater in the field areas. Both regions are well endowed with groundwater chemistry data. In southern Tunisia, subsurface waters are exploited for agriculture and to some extent for human consumption. In the Namib, most work has focused on exploitation of water suitable for domestic supplies.

In southern Tunisia, Domergue (1953) identified three artesian basins, the Campanian in the Nefzaoua with water originating in the Grand Erg Oriental, the Turonian in the Djeffara as far north as Gabès with water from the south of Tunisia and, thirdly, the Infra-Cretaceous around Chott el Fedjedj which connects with southern Algeria. Pouget (1968) added three more zones, the Miocene waters of Djerba and Zarzis, the Senonian of the Chott Djerid area and, finally, the Kimmeridgian around Foum Tathouine. Pouget (1968) found that the chloride:sulphate ratios (milliequivalents) were less than 1.0 in the Turonian and Infra-Cretaceous groundwater but greater than 1.0 in the others. Though the ratios provide a rough indication of the solubility of gypsum in terms of the anticipated sodium chloride content, Clarke and Jones (1972) pointed out that Campanian water, with a ratio of about 1.45 indicating undersaturation of gypsum, readily precipitates gypsum after very little evaporation. The index is of little value. The chemistry of deep groundwater provides a rough assessment of the character of the water at the top of the water table but in this zone the influence of meteoric water and soil chemistry comes into play. Within any single artesian basin, near-surface waters exhibit considerable chemical variability over small distances. Table 6.3, for example, presents chemical data for Campanian water from 20 m below the land surface at Bir Zefazfa (OT) and for water from standing pools in an oued near Djemma (DJ), also in the Nefzaoua. It is evident that the surface waters are considerably enriched in all ions except iron. The minerals have been obtained from the upper soil zone when meteoric waters moved through it. Evaporation of the standing water further concentrates the salts and some mineral-enriched water

may percolate to the artesian groundwater. Similar enrichment of surface waters relative to artesian groundwater is apparent when the chemistry of water in standing pools in oueds near El Hamma de Gabès (EH) and near Sebkhret Mekracherma are compared with spring water from the Nefta (N) and Tozeur (T) oases (Senonian?) and well water from the Gabès (G) oasis (Turonian) (see tables 6.4 and 6.5). Unfortunately it is impossible to ascertain whether the desert rose croûtes de nappe at Sebkhret Mekracherma (SM 2) and Djemma (DJ 4.1) are precipitated from artesian water or surface derived water since the accessory elements in the crusts are derived predominantly from incorporated clay minerals (DJ 6 and SM 1.1 and SM 1.2, tables 5.8 and 5.9).

Despite the great aridity of the Central Namib Desert, there are large reserves of groundwater (Kaiser and Beetz, 1919) particularly along river valleys where often there is subterranean flow. This amounts to 66 m^3/hour in the Ugab and 88 m^3/hour in the Omaruru (Hart, 1964). In areas of maximum flow, soluble salt content is low, total dissolved solids are often below 1500 ppm (Hart, 1964) and rarely rise above 5000 ppm (Wipplinger, 1958). Stengel (1968 a.) pointed out that calcium levels are generally low, presumably owing to the absence of the element in the majority of the bedrock forming the Namib Platform. Nevertheless, Gevers and Westhuyzen (1937) reported that calcium levels in the river groundwaters increase towards the coast. It is impossible to assess whether this is a cause or effect of the extensive gypsum crusts in the coastal zone. In the groundwaters of the River Swakop, sodium, magnesium and sulphate levels also increase down-stream, which indicates that surface salt deposits from fog moisture are leached into the groundwater zone. Silica content also increases towards the Atlantic and probably results from high pH. Siliceous sinter is deposited around some thermal springs in the Swakop system (Gevers, Hart and Martin, 1963) but such sources of dissolved minerals are rare in this area. Gypsum croûtes de nappe of the desert rose form are usually localized in dry stream channels and river valleys in the Central Namib (PF 2.5 and 1MET). Groundwater chemistry suggests

Tables 6.1 to 6.5 (overleaf)

Chemical analyses of water samples from wells, springs and standing pools in southern Tunisia (see fig. 4.1 for locations).

6.1. Chemistry of aïoun water in the Sabkha el Melah near Zarzis (CM); chemical and mineralogical analyses of sabkha deposits from the Sabkha el Melah (CM) and from the White Lady Salt Pan (CC) in the Central Namib Desert.

6.2. Chemistry of waters of Chott Djerid. From Blidet on the eastern shore and from the centre of the chott.

6.3. Chemistry of waters from the well at Bir Zefazfa (OT) and from a trench in the oued bed at Djemma (DJ) both in the Nefzaoua of southern Tunisia.

6.4. Chemistry of waters from the Oued Magroum (EH) south of Chott el Fedjedj and Sebkhret Mekracherma (SM) to the east of the chott.

6.5. Chemistry of waters from springs in the Tozeur and Nefta oases and of well water from Gabès oasis.

TABLE 6.1

	CM(W) 1.1	1.2	1.3
Ca^{++} ppm	1885	1880	2000
Na^{+} "	34500	37500	38000
K^{+} "	1250	1150	650
Sr^{++} "	37.8	37.0	27.0
Mg^{++} "	8795	7030	3180
Fe^{tot} "	4.0	3.8	4.4
Cl^{-} "	100000	100000	120000
Cond. umho	>20000	>20000	>20000

	CM 3	CM 4.1	CC 2.1	CC 2.3
IR%	0.24	1.94	1.88	0.22
Gyp%	36.99	0.46	2.25	1.70
$CaCO_3$%	*37.93	-	-	-
Na^{+} ppm	27000	264000	365000	370000
NaCl%	6.86	67.11	92.78	94.05
K^{+} ppm	1200	650	600	850
Sr^{++} "	3200	tr.	100	-
Mg^{++} "	*90600	3800	1300	1900
Fe^{tot} "	80	140	900	100
Al^{+++} "	400	-	600	100
xsCa $^{++}$ "	* -	-	800	-

*24.79% dol. with 57900 ppm $xsMg^{++}$

TABLE 6.2

See fig. 4.1 for locations

	Chott Djerid Blidet 1.1	1.2	Standard 1.3	Centre 2.1	2.2	Standard 2.3
Ca^{++} ppm	605	602	2.0	831	1325	3.0
Na^{+} "	660	610	1.0	125000	60000	19.0
K^{+} "	47.0	69.0	-	4300	3700	17.5
Sr^{++} "	8.4	8.0	-	46.0	42.0	2.0
Mg^{++} "	120	153	-	6694	4110	34.0
Fe^{tot} "	1.0	0.80	0.6	6.0	8.0	-
Cl^{-} "	6000	5000	1000	210000	210000	2000
Cond. umho	6080	6080	N/A	> 20000	> 20000	N/A
pH	8.2	8.2	N/A	N/A	N/A	N/A

TABLE 6.3

Bir Zefazfa water (OT-W)	1.1	1.2	Standard 1.3
Ca^{++} ppm	395	305	4.7
Na^{+} "	1700	1200	6.5
K^{+} "	40.0	60.0	3.0
Sr^{++} "	10.8	19.2	-
Mg^{++} "	380	420	7.0
Fe^{tot} "	1.0	0.6	0.1
Cl^{-} "	4000	4000	2000
Cond. umho	12000	12000	-
pH	8.30	8.30	-

Djemma water (DJ-W)	1.1	1.2	2	Standard 3
Ca^{++} ppm	975	940	925	4.0
Na^{+} "	7000	6700	6900	6.0
K^{+} "	480	470	330	0.5
Sr^{++} "	19.4	19.0	21.0	0.4
Mg^{++} "	1000	965	960	8.0
Fe^{tot} "	0.1	0.1	0.2	-
Cl^{-} "	22000	22000	20000	1000
Cond. umho	N/A	N/A	N/A	-
pH	8.30	8.30	7.90	-

TABLE 6.4

Oued Magroum water El Hamma (EH-W)	1.1	1.2	1.3	1.4	1.5	1.6	Standard 1.7
Ca^{++} ppm	555	565	550	645	540	565	3.2
Na^{+} "	1500	1500	1700	1000	1700	17.50	-
K^{+} "	102	95.0	96.3	102	102	88.0	-
Sr^{++} "	14.4	14.2	14.0	16.8	1.42	15.0	0.8
Mg^{++} "	475	460	470	685	535	555	2.0
Fe^{tot} "	-	-	-	-	-	1.4	-
Cl^{-} "	4500	5000	4000	5000	4000	5000	1000
Cond. umho	11200	11200	11000	5600	2400	10600	-
pH	8.70	8.60	8.30	8.60	8.50	N/A	-

TABLE 6.4 (cont.)

Sebkhret Mekracherma water (SM-W)	1.1	1.2	1.3	Standard 1.4
Ca^{++} ppm	830	860	1180	20.8
Na^{+} "	3050	5050	9000	8.3
K^{+} "	340	490	580	4.5
Sr^{++} "	30.4	26.8	31.2	-
Mg^{++} "	1110	1310	1480	8.0
Fe^{tot} "	2.0	2.0	2.8	1.0
Cl^{-} "	16000	19000	22000	1000
Cond. umho	> 20000	>20000	>20000	-
pH	8.55	8.60	N/A	-

TABLE 6.5

Spring water, Tozeur oasis (T-W)	1.1	1.1	1.3	Nefta oasis (N) 1.1
Ca^{++} ppm	135	160	150	32.2
Na^{+} "	280	360	305	250
K^{+} "	24.3	27.5	28.0	20.0
Sr^{++} "	7.2	5.5	5.3	8.4
Mg^{++} "	800	76.0	94.2	127.0
Fe^{tot} "	0.1	-	-	1.0
Cl^{-} "	1000	1000	1000	1000
Cond. umho	N/A	2400	N/A	N/A
pH	N/A	N/A	N/A	N/A

Well water, Gabès oasis (G-W)	1.1	1.2	1.3 (standard)
Ca^{++} ppm	265	260	1.2
Na^{+} "	420	430	0.5
K^{+} "	20.0	20.0	-
Sr^{++} "	5.4	5.0	-
Mg^{++} "	92.2	100	-
Fe^{tot} "	0.8	0.6	0.4
Cl^{-} "	1500	2000	1000
Cond. umho	2700	2700	-
pH	7.80	7.80	-

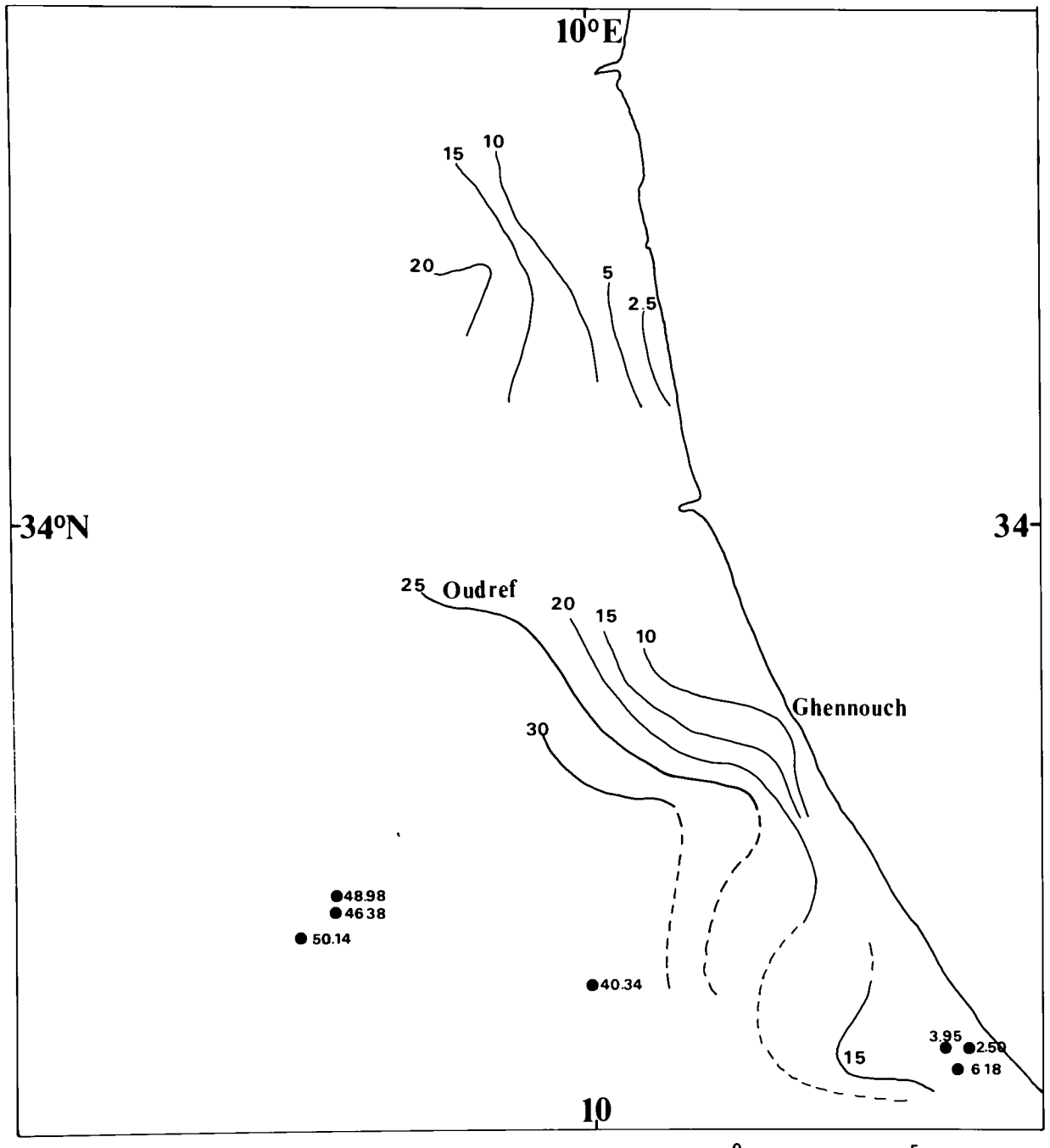
that the soluble salts are derived from surface deposits alone and not from bedrock as in southern Tunisia. The anomalous chemical compositions of the desert rose croûtes de nappe near the Hamilton Range (HA) and at Wlotzkas Baken (WB) may be tentatively attributed a lacustrine origin in the former case and to supratidal sabkha sedimentation in the latter.

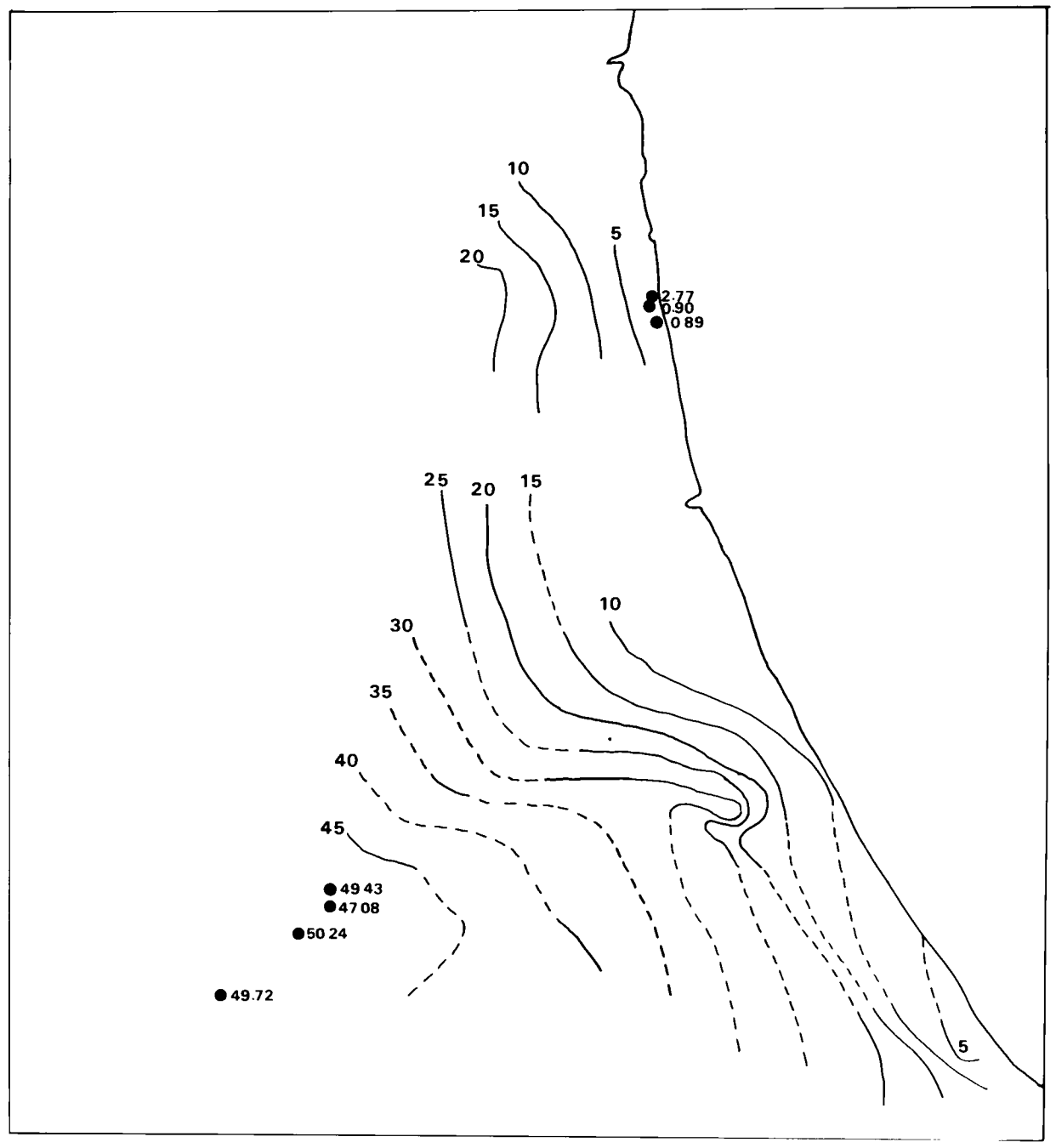
The potential role of groundwater in the development of gypsum crusts was examined in greater detail in a small area of southern Tunisia where a dense network of wells has provided the opportunity for intensive analysis of the hydrochemistry. The area covers 12,000 km^2 around Ghennouch and includes to the south-east the oases of Djara in the Gabès district, Oudref to the north-west of Ghennouch and several scattered wells and springs between La Skhirra and Gabès. Mekrazi (1974) collected water samples from eighty sites during September 1973 and April 1974, groundwater levels were measured during September and October 1973 and in April 1974 (see figs. 6.4A and 6.4B). The water analyses undertaken by Mekrazi (1974) provide measurements of calcium, magnesium, sodium, sulphate, chloride, titration alkalinity, pH and total dissolved solids. His maps of variation in total dissolved solids are reproduced as figure 6.5A and of sulphate/chloride ratios (milliequivalents) as figure 6.5B. For the purposes of the present work the data were analyzed using the WATSPEC computer programme developed by Wigley (1977). WATSPEC is one of a long series of computer programmes designed to provide a detailed breakdown of the theoretical mineralogical composition of groundwater using thermochemical equilibrium constants. It has the advantage over many other programmes of being able to provide results from very scant input data as well as that from more detailed chemical analyses. The programme calculates the chemical composition of a water sample, that is the concentration of all free ions and ion pairs. from a knowledge of the concentrations of various species. Once this has been achieved, the programme determines the theoretical level of saturation of the solution with respect to all the minerals for which all the components are provided in the input data. It is critical to ascertain the saturation levels of individual species in

Figure 6.4 (overleaf)

A, (left). Height of water table above mean sea level around Ghennouch, near Gabès, September/October 1973, in metres (after Mekrazi, 1974).

B, (right). Height of water table above mean sea level April 1974, in metres (after Mekrazi, 1974).





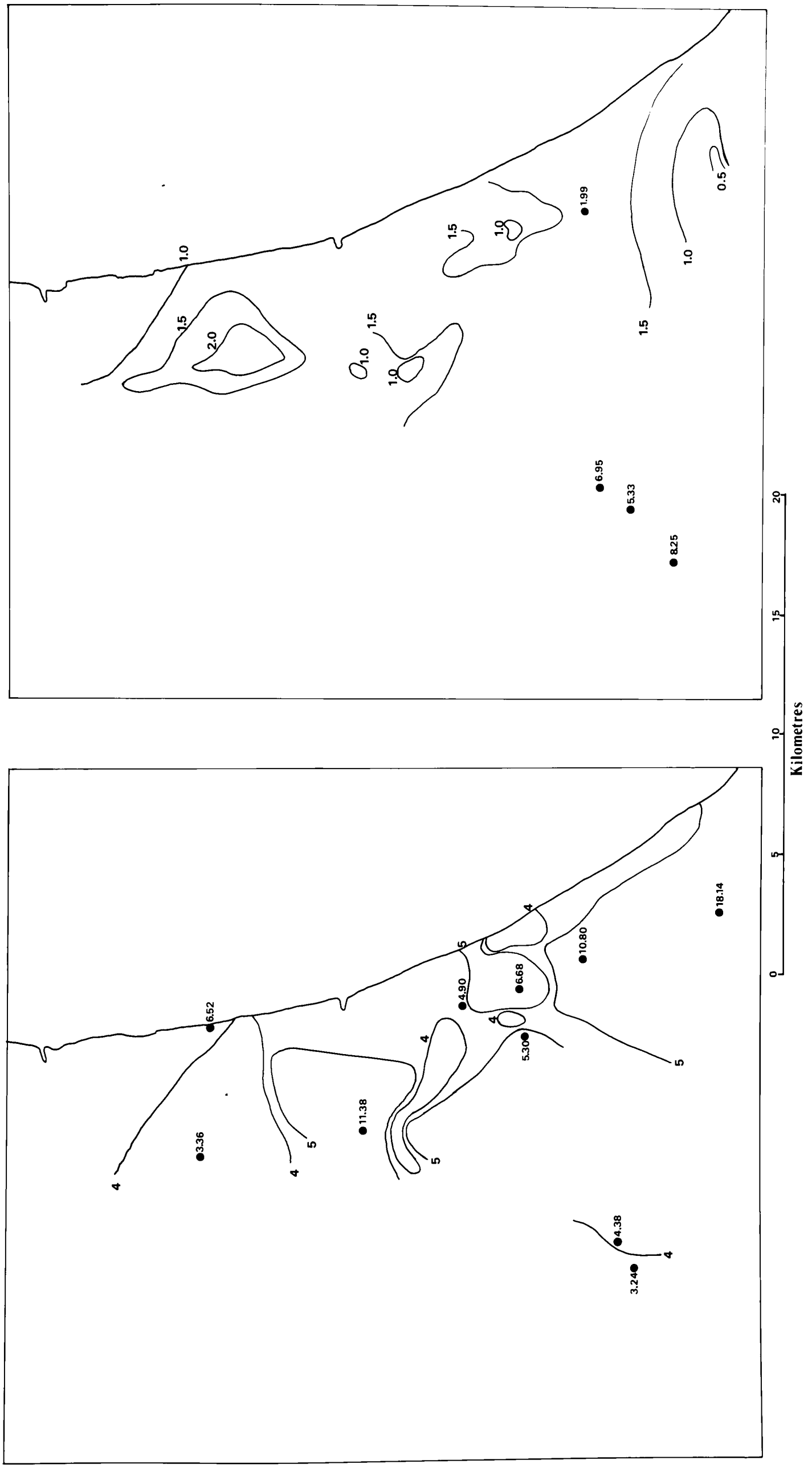
0 5 10 15 20

Kilometres

Figure 6.5 (overleaf)

A, (left). Total dissolved solids in groundwater from wells around Ghennouch, near Gabès, in grams/litre (after Mekrazi, 1974),

B, (right). Ratio of sulphate to chloride in groundwater from wells around Ghennouch, near Gabès, (rSO_4/rCl) (after Mekrazi, 1974).

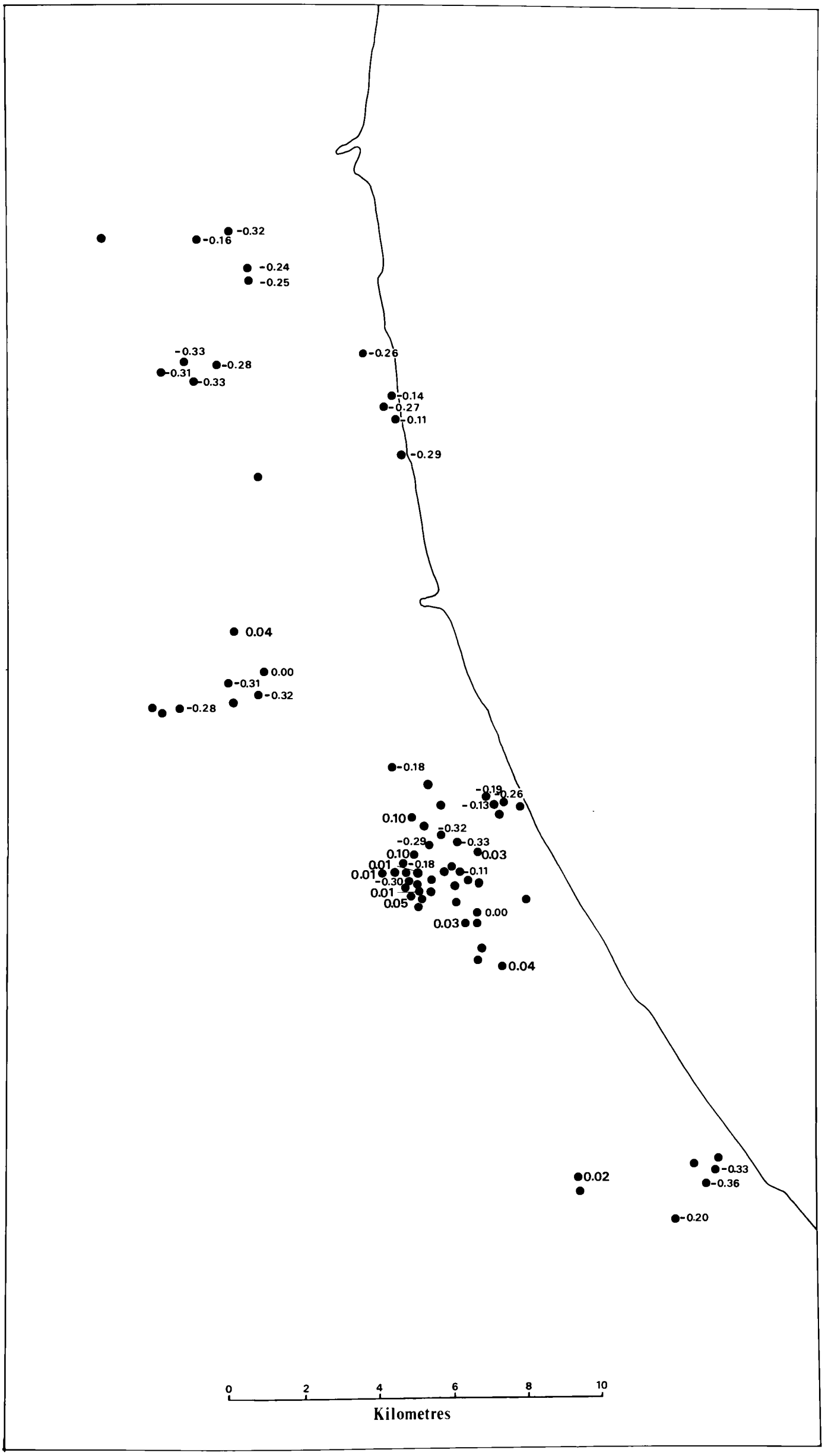

1.0
1.5
2.0
1.5
1.0
1.5
1.0
1.99
1.5
1.0
0.5
6.95
5.33
8.25
0
15
20
Kilometres
6.52
3.36
4
5
11.38
4
5
4.90
4
5.30
6.68
4
5
10.80
18.14
4.38
4
3.24
0
5
10

relation to not only the concentrations of the various ions it is composed of, temperature and pH but also other minerals which may form from the ions. By determining the activation energies of all the ionic constituents it is possible to calculate the proportions of individual anion species that theoretically will combine with the individual cation species that are available. The level of saturation of a specific mineral will be influenced by not only the concentration of its constituent ions but also the effects the other minerals present have on its solubility. Saturation level determines whether the solution has the potential for further dissolution or for precipitation of that mineral. In the case of gypsum, the theoretical saturation index will depend not only upon calcium and sulphate levels but also sodium and chloride speciation, the potential for calcium carbonate speciation, which itself is critically affected by pH and temperature, and to some extent upon magnesium speciation. Mekrazi (1974) did not provide temperature data for the waters collected so this has been estimated as $18^{o}C$, slightly below the mean annual air temperature for Gabès of $19.2^{o}C$ (Mensching, 1968, p. 39). The data from 78 samples were analyzed using WATSPEC and the results in terms of gypsum saturation indices are presented for 73 of these in figure 6.6. All the analyses except 14 showed undersaturation with respect to gypsum, a positive value indicating oversaturation. 3 samples had indices of 0.00 reflecting equilibirium in terms of gypsum saturation. In most cases, dolomite ($CaMg(CO_3)_2$) is the most saturated mineral with calcite and aragonite less so. In two samples with pH below 5.0 these minerals are undersaturated. Magnesite ($MgCO_3$) shows considerable variation above and below saturation level but is most often saturated. Hydromagnesite ($Mg_4(CO_3)(OH)_2.3H_2O$) is similarly variable but generally undersaturated. Nesquehonite ($MgCO_3.3H_2O$), huntite ($CaMg_3(CO_3)_4$) and brucite ($Mg(OH)_2$) are consistantly undersaturated as are the sodium salts, halite (NaCl), thenardite (Na_2SO_4) and mirabilite ($Na_2SO_4.10H_2O$). All the waters are undersaturated with respect to anhydrite which is always less saturated than gypsum.

The results of the WATSPEC computer analysis show that the groundwater in the

Figure 6.6 (overleaf)

Gypsum saturation indices of groundwater from wells around Ghennouch, near Gabès, determined using WATSPEC. Positive values indicate that the waters are saturated with gypsum, these are shown in bold type.


-0.32
-0.16
-0.24
-0.25
-0.33
-0.28
-0.31
-0.33
-0.26
-0.14
-0.27
-0.11
-0.29
0.04
0.00
-0.31
-0.32
-0.28
-0.18
-0.19
0.26
-0.13
0.10
-0.32
-0.29
-0.33
0.10
0.03
0.01
-0.18
0.01
-0.11
-0.30
0.01
0.05
0.00
0.03
0.04
0.02
-0.33
-0.36
-0.20
0
2
4
6
8
10
Kilometres

area north of Gabès, where surface gypsum crusts are very extensive, are predominantly undersaturated with gypsum druing the late summer and winter months. Croûte de nappe development through gypsum precipitation from these waters at this time of year would seem unlikely though, as pointed out above, conditions at the top of a groundwater table may alter the water chemistry considerably. It is possible that rapid achievement of equilibrium in terms of the saturation levels may result in negative saturation indices owing to gypsum precipitation. The presence of some oversaturated waters, as well as the chemical evidence of slow achievement of equilibrium in gypsum systems (see chapter one), suggest that significant hydrochemical changes would be necessary to promote gypsum crystallization from these groundwaters. While some gypsum saturated waters may not necessarily precipitate gypsum owing to slow rates of nucleation and crystal growth, undersaturated waters will never experience gypsum crystallization. The mineral speciation indices for one sample slightly saturated with gypsum at 18.0°C were computed at 10.0°C and 30.0°C. In the first case, all minerals except gypsum became less saturated, the gypsum index increasing from 0.04 to 0.07. In the second case, all except gypsum became less undersaturated, with huntite and hydromagnesite becoming saturated. The gypsum saturation index fell to 0.01 and that for anhydrite rose to 0.00. It is feasible therefore that under cool conditions during the late Pleistocene croûte de nappe development would be more pronounced. Not only would lower temperatures have affected gypsum solubility but lower potential evaporation rates would have raised the amount of groundwater recharge by meteoric waters, thereby diluting sodium chloride concentrations. Under such conditions Pouget's (1968) model of croûte de nappe development becomes more valid in terms of large scale gypsum precipitation from artesian waters. It is possible that exposed desert rose croûtes de nappe in the Nefzaoua at Zaafrane (ZAA) and El Faouar (EF) represent relic features which crystallized from groundwater under hydrochemical conditions more suitable for gypsum precipitation than those at present. This is probably also true of the crusts at El M'Dou (EM 3) which Pouget (1968) regarded as relic croûtes de nappe. The basal crust (EM 3.8)

at this site does, however, possess structural and chemical characteristics which suggest that it may be a recent feature (see chapter five).

Seasonal fluctuations in the level of groundwater in the area north of Gabès are shown on figures 6.4A and 6.4B. In most wells, the rise in the water level between late summer and spring is less than 1.0 m. Possible errors resulting from human exploitation of the well waters are likely to result in abnormally high seasonal variation since most utilization will be in the late summer when the agricultural requirements are greatest. Suffice it to point out that the minor differences lend further support to the suggestion that thick croûtes de nappe must represent gypsum accretions which have developed over long periods during which marked groundwater fluctuation has occurred. Furthermore, the depth of the water table beneath the land surface in this area is too great for it to be a source of gypsum making up surface or mesocrystalline subsurface crusts. It is necessary to examine further potential sources of gypsum and mechanisms of crust formation in order to ascertain their origin.

c. Lateral movement of gypsum.

i. Terrestrial waters.

In some cases, gypsum precipitated from groundwater is derived from bedrock at considerable distances from the area of crust development. Infra-Cretaceous groundwater in the Fedjedj basin may originate over 1,000 km away in southern Algeria. In supratidal sabkhas, oceanic and continental waters may seep through coastal sediments over distances of several kilometres before soluble salts are precipitated. Yet some surface crusts may also form by precipitation from laterally moving waters (Zimmermann, 1963). Salt crusts described by Conacher (1975) in Australia and Risacher (1978) in Bolivia form as waters flowing through the soil zone precipitate salts dissolved in them when they evaporate at or near the

surface around closed depressions and in valleys (see fig. 6.7A). Crystallization results from evaporation at or near the surface. A number of workers have attributed the development of laminated calcretes (croûtes zonaires) to such processes (Durand, 1963; Scholz, 1971). Throughflow is likely to occur on the glacis and pediment slopes of southern Tunisia dn the Central Namib Desert after rainfall. Since laterally moving waters will come in contact with greater amounts of soluble minerals than water stationary within the soil zone, enrichment will take place. The process may account for some near-surface crusts in low lying areas but many gypsum crusts are not this restricted in their distribution.

Models incorporating crystallization of minerals from water flowing laterally through the soil zone, in contrast to precipitation from surface slope-wash, represent a modification parallelled in hydrology. Durand's (1949 b., 1951) model of calcrete croûte zonaire development involved precipitation from waters flowing over the land surface, following precipitation in excess of the soil's infiltration capacity (Horton, 1933). The rarity of overland flow in many areas stimulated hydrologists to develop a series of models involving throughflow and pipe flow in the soil zone to account for storm runoff in river basins (Tennessee Valley Authority, 1965; Whipkey, 1965; Ragan, 1968; Dickinson and Whiteley, 1970; Dunne and Black, 1970; Weyman, 1970, 1973, 1974; Finlayson, 1977). These studies provide valuable insights into the mechanisms involved in runoff not only in temperate areas but also arid and semi-arid zones. Furthermore, the concept of a fluctuating soil-water saturated zone from which storm runoff is generated under suitable antecedent conditions is applicable in some ways to desert crust development. Whether surface runoff is generated or not, the soil water in this zone will evaporate during subsequent dry conditions precipitating the soluble salts which were dissolved during throughflow. Some crystallization or dissolution may result from chemical interactions affecting solubility but salt accretion will occur in the water accumulation and evaporation zone provided horizontal or vertical leaching does not predominate.

Gypsum deposits associated with springs have been described in Californa (Hess, 1909; Hunt et al., 1966) in Algeria (Stainier, 1912) and Australia (Madigan, 1930 Gypsum tufas are usually found in saline basins where mineralized waters evaporate as they reach the surface. Spring mounds, called chria or behour in north Africa, may reach heights of 20 m. Van Everdingen (1969) described gypsum tufa around thermal springs in the Canadian Rocky Mountains. In most cases, the material is made up of loosely cemented lenticular gypsum crystals up to 1.0 mm in diameter. The deposits described by Hunt et al. (1966) in Death Valley differ in that a consolidated columnar crust has developed, the surface of which has been dehydrated to anhydrite and hemihydrate. Again, deposits originating in this way are very localized and the process cannot account for the majority of surface gypsum crusts.

Transport of gypsum in suspension in surface waters is not widely recorded. Van Everdingen (1969) mentioned its occurrence in stream waters flowing from springs which experience rapid supersaturation with gypsum when their waters cool on reaching the surface. Butzer and Hansen (1968, pp. 105-107) tentatively suggested that gypsum crusts developed upon Nile silts near New Abu Simbel may be consolidated gypsiferous fluvial sediments. They also pointed out that the levels of gypsum saturation required for waters to carry gypsum in suspension without dissolution would also promote evaporite sedimentation.

ii. Detrital accumulations (see fig. 6.7B)

Transport of solid phase gypsum may occur by processes other than fluvial. Colluvial fan deposition in semi-arid areas involves the movement of large amounts of sediment in partial suspension in sheet flood or stream water. Transportation of sediments is rarely over large distances and the events are of short duration. In effect, considerable amounts of gypsiferous bedrock may be moved down-slope without appreciable loss through dissolution. When deposited on the glacis or

fans, the gypsum-rich sediments may be consolidated by meteoric waters causing some dissolution and recrystallization thereby forming surface crusts. This process has not been alluded to by other workers though the consolidation process is akin to that proposed by Bureau and Roederer (1961) for the development of gypsum crusts on gypsiferous Mio-Pliocene sediments in southern Tunisia. In the vicinity of the Oued el Melah (OMG) between Gafsa and Metlaoui, the surface of the middle of three glacis is composed of sands and gravels which extend to a depth of 1.0 m to 3.0 m, resting on columnar mesocrystalline gypsum crusts and Senonian (?) limestones and clays. Poor sorting and the absence of depositional bedforms suggest that the deposit is colluvial, the materials having been transported by slope-wash and in sediment laden streams from limestone escarpments forming the western end of the Gafsa chain. The percentages of different rock types making up pebbles between 10 mm and 20 mm in diameter are as follows (100 pebbles from each horizon).

	Gypsum	Limestone	Chert/Flint	Sandstone	Quartz	Shale
Surface	-	58.0	28.0	9.0	5.0	-
0.5 m to 1.0 m	39.0	31.0	18.0	11.0	-	1.0
Excluding gypsum	-	(50.8)	(29.5)	(18,0)	(-)	(1.6)

The absence of gypsum at the surface may be attributed to a different source of sediment but the similarity between the percentages of other rock types at the surface and at 0.50 m to 1.0 m, when gypsum is excluded from the latter, suggests otherwise. Gypsum has been leached from the surface by meteoric water infiltratin the soil zone. Beneath the surface, the water is saturated with gypsum and the fragments of gypsum bedrock remain intact, preserving their micropetrographic and chemical characteristics (see OMG 1 in tables 5.6 and 6.6). The well developed columnar subsurface crusts in this area (see chapter five) are a product of recrystallization of gypsum in the lower soil horizons. The processes involved in this accretion will be described in chapter seven. It is probable that similar processes occur at locations where gypsum bedrock is exposed at the surface. At Djebel el Aziza (DT) some of the subsurface crusts may be composed of gypsum

TABLE 6.6

Chemical and mineralogical composition of gypsum bedrock fragment from the Oued Magroum near Gafsa.

	OMG 1
IR%	0.42
Gyp%	92.48
$CaCO_3$%	-
Na^+ppm	1400
(NaCl%)	0.36
K^+ppm	200
Sr^{++} "	2300
Mg^{++} "	300
Fe^{tot} "	-
Al^{+++} "	100
$xsCa^{++}$ "	18700

leached from bedrock fragments scattered over the glacis slopes. It is feasible that some crusts consist of fragments of bedrock cemented by secondary gypsum. The classification of these forms as true gypsum crusts is complicated by the limitation proposed by Buringh (1968) that only those accumulations with 5% more gypsum than the underlying C horizon should be included. In these cases, micro-structural considerations overrule this qualification. In some instances, it is feasible that bedrock gypsum has undergone in situ structural alteration to deposits resembling gypsum crusts. The chemical similarities between the subsurface crust at Djebel el Aziza (DT 1.4, table 5.31) and some of the bedrock samples (DT 2.2 and 3.2, table 5.32) support this, as does the visible surface alteration of a fibrous gypsum sample DT 6.2 to lenticular crystals (see plate 5.34 and table 5.32). Nevertheless, in most of the cases described here stratigraphic considerations preclude such an origin. The majority of gypsum crusts are located within or upon sediments which are distinct from the local bedrock. At Oued Magroum (EH) near El Hamma de Gabès, inclusions of iron nodules (EH 7.5) (see fig. 5.8C and table 5.26) in the gypsum crusts are derived from bedrock yet the low gypsum content of the Wealdian clays, with the exception of discrete selenite strata, suggests that the gypsum making up the crust is allogenic.

iii. Aeolian transport of gypsum-rich sediments (see fig. 6.7C)

Coque (1955 b., 1958, 1960, 1962,pp. 93-101, 1969) recognized that many of the surface gypsum crusts found in the Nefzaoua cap sand dunes preserving their form. This occurs not only on large barchanoid dunes but also small nebkhas only 0.50 m to 1.0 m in height. He argued that these crusts represent superficially consolidated gypsum-rich wind blown deposits. Furthermore, he held that the majority of surface crusts in southern Tunisia could be attributed to the same process, the gypsum-rich sediments being distributed widely by deflation of evaporitic sediments from the chotts and smaller sabkhas. This would occur during periods of aridification when lake waters evaporated precipitating the soluble salts. When

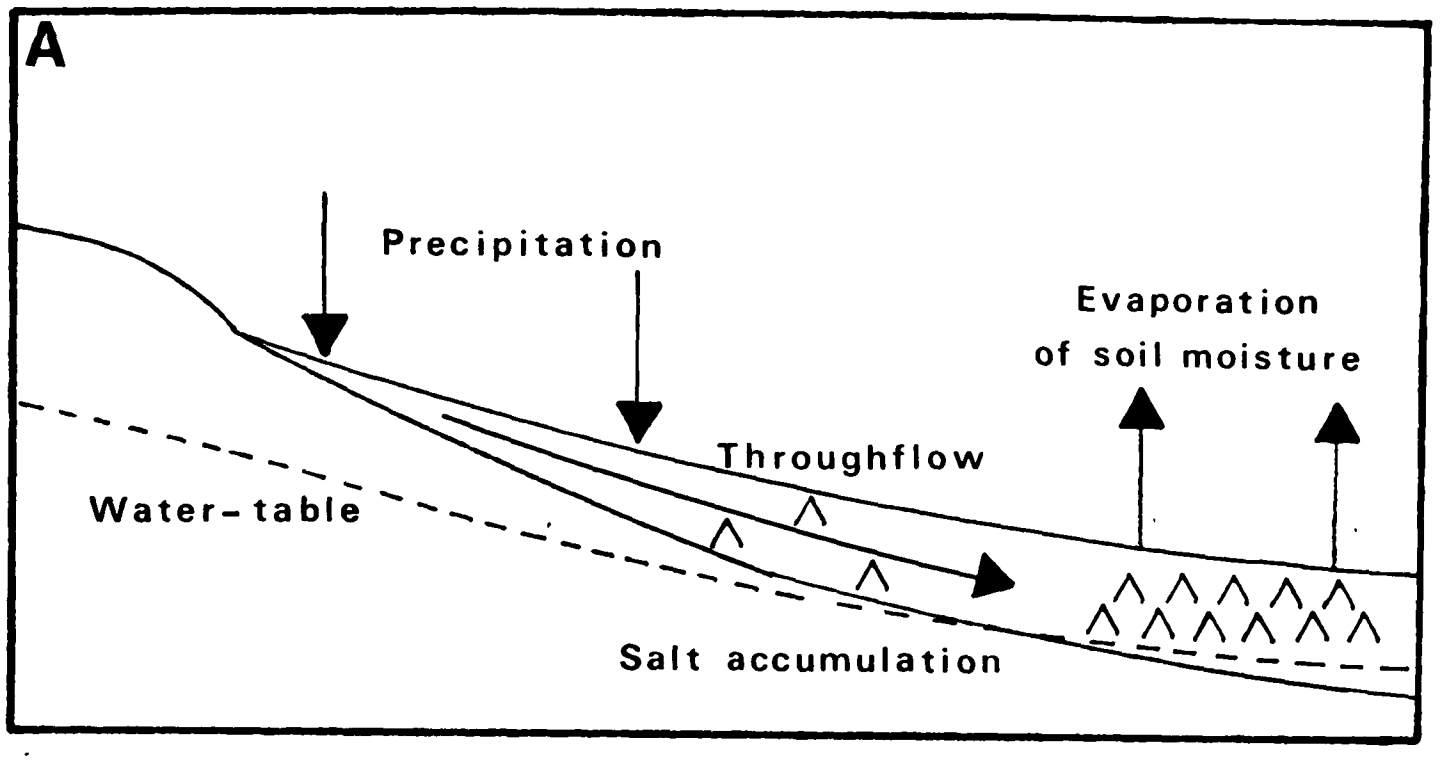
Figure 6.7 (overleaf)

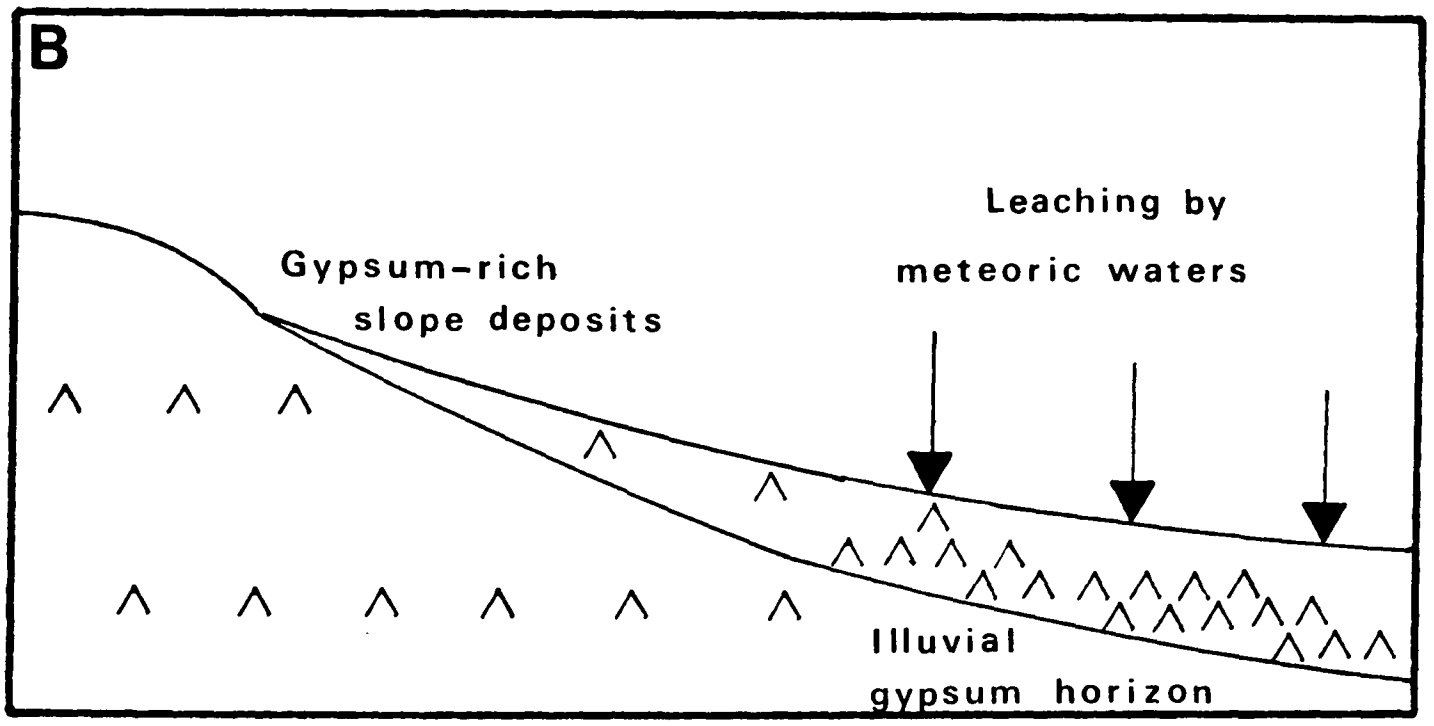
Models of gypsum crust genesis involving the lateral movement of gypsum in various ways.

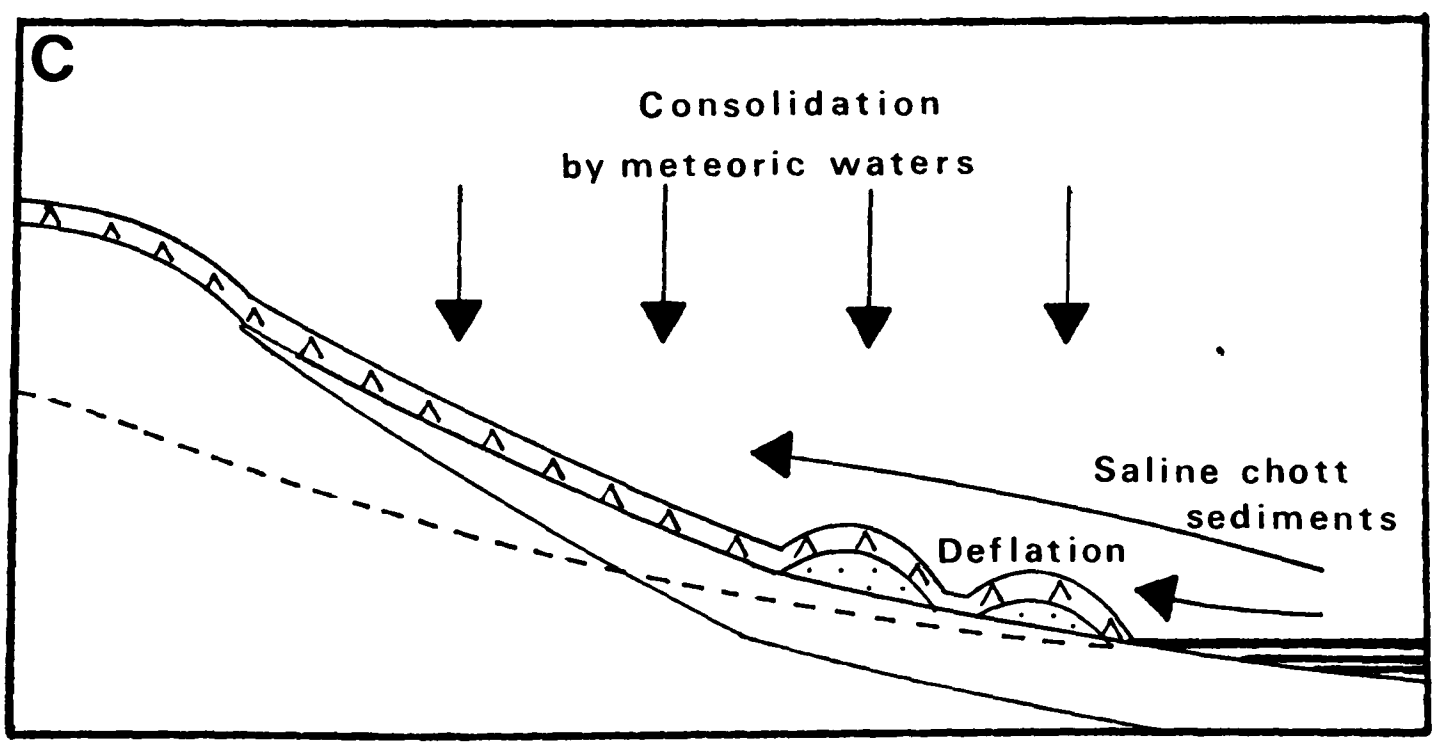
A. Accumulation of gypsum in closed depressions and valleys following dissolution of gypsum in slope deposits by throughflow.

B. Gypsum crust development in colluvial sediments derived from gypsiferous bedrock.

C. Consolidation of gypsum deflated from lacustrine and lagoonal evaporite deposits (after Coque, 1962).


A
Precipitation
Evaporation
of soil moisture
Throughflow
Water-table
Salt accumulation


B
Leaching by
meteoric waters
Gypsum-rich
slope deposits
Illuvial
gypsum horizon


C
Consolidation
by meteoric waters
Saline chott
sediments
Deflation

deposited on the surrounding land surface, meteoric waters would cause intermittent dissolution and recrystallization of the gypsum and bring about induration. The model has the advantage over those advocating groundwater influences of being able to account for crusts on steep slopes and hill crests as well as uniform thickness over varied terrain. Most significantly, it has the great advantage of apparently being able to explain the high gypsum contents of many surface crusts. Le Houérou (1960), Coque and Jauzein (1967) and Vieillefon (1976) supported the applicability of the model to most gypsum crusts in southern Tunisia. Horta (1979) in Algeria and Boulaine (1961 a.), Zimmermann (1963) and Page (1972) in Tunisia, felt that it was valid for certain forms. In other areas, similar models have been developed independently of those based upon work in north Africa. Jessup (1960 a. b., 1961 a.b.) held that the subsurface gypsum crusts of South and Western Australia were derived by deflation of evaporitic deposits from pans. He believed that the overlying sands and clays represented subsequent aeolian deposits deflated after all of the evaporitic gypsum had been removed from the pans. In Iraq gypsum crusts in similar stratigraphic positions were interpreted as aeolian deposits later covered by fluvial silts (Tucker, 1978).

The mechanisms and likelihood of deflation of gypsum-rich sediments from basins may be ascertained and verified from work undertaken in a number of areas where gypsum crusts are widespread. In Australia Jack (1921), De la Hunty and Low (1958) and Termier (1963) have described well developed gypsum lunettes downwind from deflation pans. Hills (1940) pointed out that often lunettes do not exhibit aeolian bedding structures and he suggested that in some cases these landforms may develop through the capture of atmospheric dust from other sources by wind blown spray from the pan. Jennings (1954), however, advocated vegetational capture of deflated sediments. A considerable amount of information is available from North America, where wind blown gypsum deposits are well developed in the White Sands Desert of New Mexico and around the Great Salt Lake of Utah. In the former area, McKee and Moiola (1975) held that the gypsum dunes of the Tularosa Valley

are deflation products from Lake Lucero playa. The well stratified dunes are composed of materials consisting of 94 % gypsum, 4.0% sodium chloride and 1.0% silicates (Jaeger, 1957). Talmage (1932) pointed out that there are fossil dunes beneath the present active deposits which are themselves 10 m to 15 m in height. Fossil gypsum dunes up to 40 m thick of Upper Jurassic age, in the Four Corners Region of New Mexico, are probably derived from coastal sabkhas (Tanner, 1965). In Utah, the playa deflation products were held to be derived from weathered selenite crystals within lake-bed clays rather than lacustrine evaporite sediments (Gilbert, 1895; Blackwelder, 1931; Jones, 1953). Eardley (1962) found the deposits consisted of about 65% gypsum and the remainder calcareous oölites (Artemia gracilis, after Jones, 1953). Dunes near Knolls, Utah are 3.0 m to 10 m in height and again overlie partially consolidated relic dunes. A sample collected by Dr. A. Goudie in 1978 has the following composition.

I.R.%	5.44	
Gyp.%	56.36	
$CaCO_3$%	17.44	
Na+ppm	2550	
(NaCl%)	0.65	
K^+	ppm	650
Sr^{++}	"	1000
Mg^{++}	"	4000
Fe^{tot}	"	1100
Al^{+++}	"	1100
$xsCa^{++}$	"	39000

Little detailed information on the mineralogical composition of gypsum-rich sabkha deflation products is available from north Africa. Kinsman (1969) and Amiel and Friedman (1971) described gypsum deflation from coastal and continental sabkhas in the Middle East and Termier and Termier (1963) from Chott Chergui in Algeria. Gypsum dunes up to 10 m in height are found inland from coastal sabkhas in Libya (Glennie, 1970). Around the Chotts of southern Tunisia deflated materials have a very variable composition. Trichet (1963) recorded 60% gypsum, 39% quartz and 0.5% of both calcium carbonate and sodium chloride. In contrast, Besler (1977) found only 19% gypsum in sands from Chott el Fedjedj. Trichet (1963) ascertained that the gypsum component of the dune sands comprised 10% fine material, 40% lenticular crystals and 50% crystals aggregated into pedotubules probably of floral origin. He argued that these materials undergoing botanic induration provide a modern analogue of the processes which consolidated the majority of the surface crusts in southern Tunisia, notably those at Oued Demna (OD) and Oued Ziqzaou (ZIG) and possibly some surface crusts such as that directly overlying Turonian limestone bedrock on Djebel ed Dissa (DI). Nevertheless, there are some shortcomings inherent in the model which make it inapplicable even to some powdery accumulations.

The first problem to arise is that of the incompatibility between the chemical composition of the gypsum crusts and present-day chott deflation products. Tables 6.7 and 6.8 provide chemical analyses of sands from dunes near Djemma (DJ) in the Nefzaoua, from the eastern end of Chott el Fedjedj between El Hamma de Gabès and Oudref (EH) and Seftimi (SEF) 7 km north-west of Limaguess at the southern extreme of Chott el Fedjedj (see fig. 4.1). The dune near Djemma (see fig. 6.8) consists of predominantly unconsolidated sands rising 8 m above a sand and clay plain, the surface of which is consolidated by sparry calcrete with abundant floating quartz grains (DJ 3). This overlies

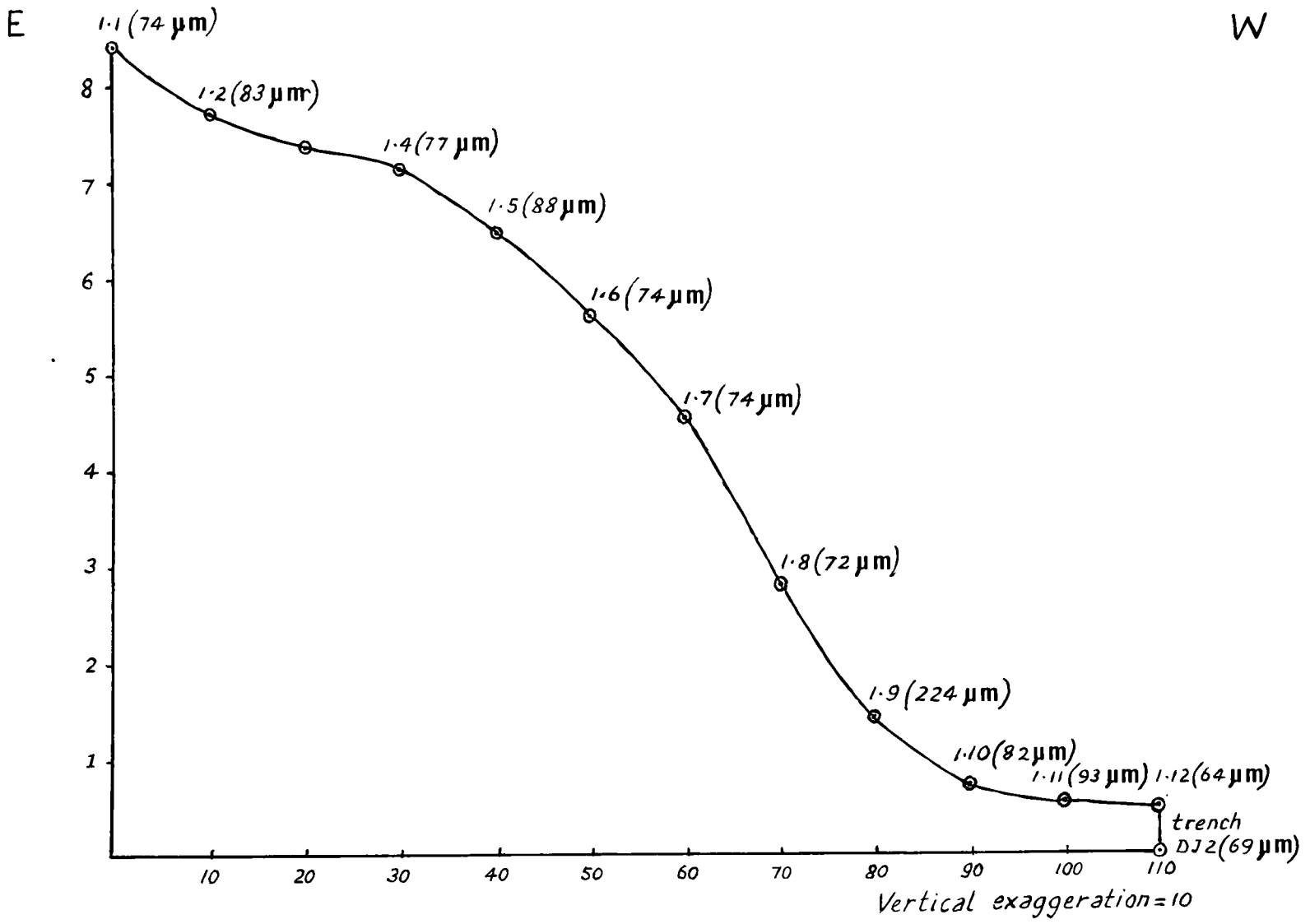


Figure 6.8

Profile of sand dune near Djemma in the Nefzaoua of southern Tunisia, scale in metres.
Figures in brackets are the median grain sizes, in microns, of the samples DJ 1.1 to DJ 1.12 and DJ 2.
Grain size distribution curves are shown in figs. 6.9 to 6.12.

fine sands (DJ 2, see fig. 6.12 for grain size distribution and table 6.7 for chemical analysis). All grain size analyses were carried out after clay dispersion of air-dried samples with sodium hexametaphosphate. The finer than 62.5 μm fraction was removed by wet sieving and the residue dried at temperatures below 35°C to avoid dehydration to hemihydrate before separation into 11 size classes at 0.5Ø intervals between 2.0 mm and 0.0625 mm using British Standard sieves. Some solution of gypsum will occur during wet sieving but examination of samples before and after the procedure reveals negligible differences. Dry sieving results in some abrasive breakdown of gypsum grains which may account for some of the angular fragments in the smallest grain size ranges. These, however, are also present in unsieved samples, indicating that such disintegration also occurs during aeolian transport. The grain size distribution graphs (see figs. 6.9 to 6.12) represent cumulative frequency curves for the coarser than 62.5 μm fraction. The graphic median grain sizes of the dune sand near Djemma range from 64.0 μm in the case of the basal sample (DJ 1.12), which contains 57.49% material finer than 62.5 μm to 224 μm for a sample from the toe of the dune (DJ 1.9). Even including this anomalously high figure, the mean median grain size for the dune 91 μm (3.45Ø), is unusually low as are all other southern Tunisian dune sands. Goudie and Watson (in prep.) quote a global mean median grain size, based on over 1000 samples, of about 180 μm (2.46Ø). The small size of Tunisian sands may result from the nature of the source materials or solution or salt weathering in highly alkaline and saline sabkha environments (Folk, 1978; Goudie, Cooke and Doornkamp, 1979).

In the zone where the dune meets the plain (DJ 1.11, DJ 1.12 and DJ 2), there appears to be admixing of silty and coarse sediments of fluvial origin. On the dune itself (DJ 1.1 to DJ 1.10), all of the samples with more than 20% gypsum also have striking bimodal grain size distribution curves. Sample DJ 1.1 has 19.84% gypsum and is slightly bimodal. DJ 1.2 and DJ 1.8

TABLE 6.7

Chemical and mineralogical composition of dune sand from Djemma in the Nefzaoua of southern Tunisia (figs. 6.9 to 6.12 provide grain size data).

See fig. 6.8

	DJ 1.1	1.2	1.4	1.5	1.6	1.7	1.8	1.9
IR%	69.90	64.92	55.94	39.00	60.60	68.84	73.34	40.02
Gyp%	19.84	3.73	33.02	39.60	28.22	21.57	6.54	40.72
$CaCO_3$%	5.06	4.14	4.83	5.98	6.44	4.37	2.00	6.21
Na^+ppm	2040	27600	1800	2200	1520	1780	4000	2760
(NaCl%)	0.52	7.02	0.46	0.56	0.39	0.45	1.02	0.70
K^+ppm	960	2500	600	600	800	600	820	980
Sr^{++}"	600	400	400	600	400	400	400	400
Mg^{++}"	5000	12400	4000	3800	4400	5400	4800	4600
Fe^{tot}"	1400	4000	1200	1000	1400	1400	1600	1200
Al^{+++}"	1000	2800	800	600	1200	1000	1800	1200
$xsCa^{++}$"	1600	26700	13800	25800	-	300	22800	18300

TABLE 6.7 (cont.)

See fig. 6.8

	DJ 1.10	1.11	1.12	2	3
IR%	41.78	55.16	61.18	90.86	51.40
Gyp%	30.41	32.77	5.88	4.82	2.01
$CaCO_3$%	5.52	2.76	12.41	1.38	*39.27
Na^+ppm	5600	4800	26700	2420	7000
(NaCl%)	1.42	1.22	6.79	0.62	1.79
K^+ppm	1000	1000	1680	720	1280
Sr^{++} "	400	400	200	-	200
Mg^{++} "	7600	6600	12800	1600	*17000
Fe^{tot} "	2000	2400	4200	1800	1600
Al^{+++} "	1600	1200	2200	200	1000
$xsCa^{++}$ "	5100	4700	-	-	* -

*12.90% dol.
29.79% $CaCO_3$

Figures 6.9 to 6.12 (overleaf)

Cumulative frequency curves showing grain size distribution characteristics of sand samples from a dune near Djemma in the Nefzaoua of southern Tunisia (see fig. 6.8).
Median grain sizes are as follows:

DJ 1.1 (11.31% finer than 62.5 um)
median 74 um, 3.74 ø

DJ 1.2 (5.56% fines)
median 83 um, 3.57 ø

DJ 1.4 (5.54% fines)
median 77 um, 3.69 ø

DJ 1.5 (4.85% fines)
median 88 um, 3.50 ø

DJ 1.6 (6.35% fines)
median 74 um, 3.74 ø

DJ 1.7 (7.33% fines)
median 74 um, 3.74 ø

DJ 1.8 (14.76% fines)
median 72 um, 3.78 ø

DJ 1.9 (5.53% fines)
median 224 um, 2.15 ø

DJ 1.10 (12.09% fines)
median 82 um, 3.60 ø

DJ 1.11 (13.95% fines)
median 93 um, 3.66 ø

DJ 1.12 (57.49% fines)
median 64 um, 3.91 ø

DJ 2 (17.51% fines)
median 69 um, 3.84 ø

DJ 4.2 (20.91% fines)
median 76 um, 3.70 ø

See table 6.7 for chemical analyses.

Figure **6.9**

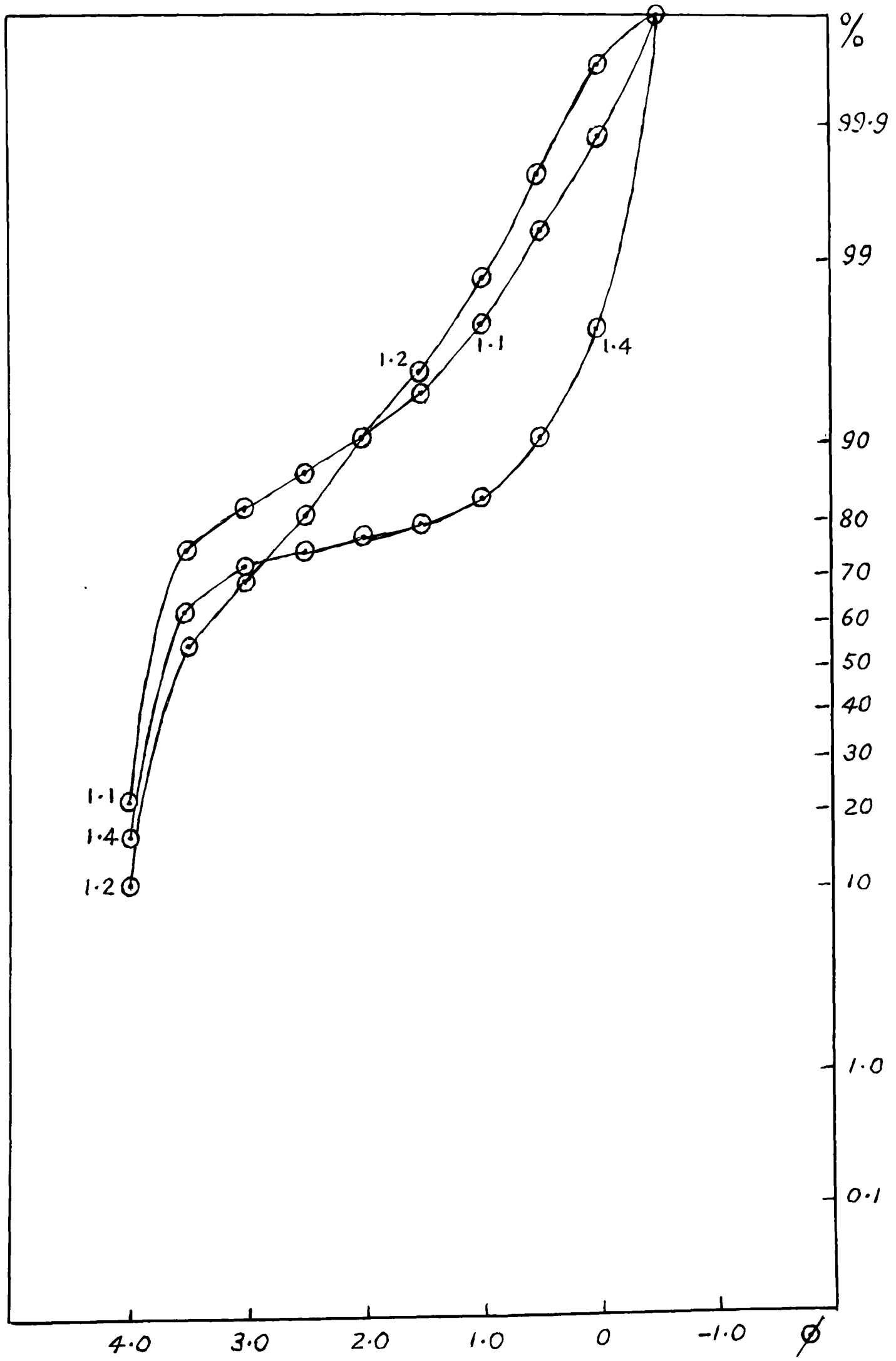

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Figure **6.10**

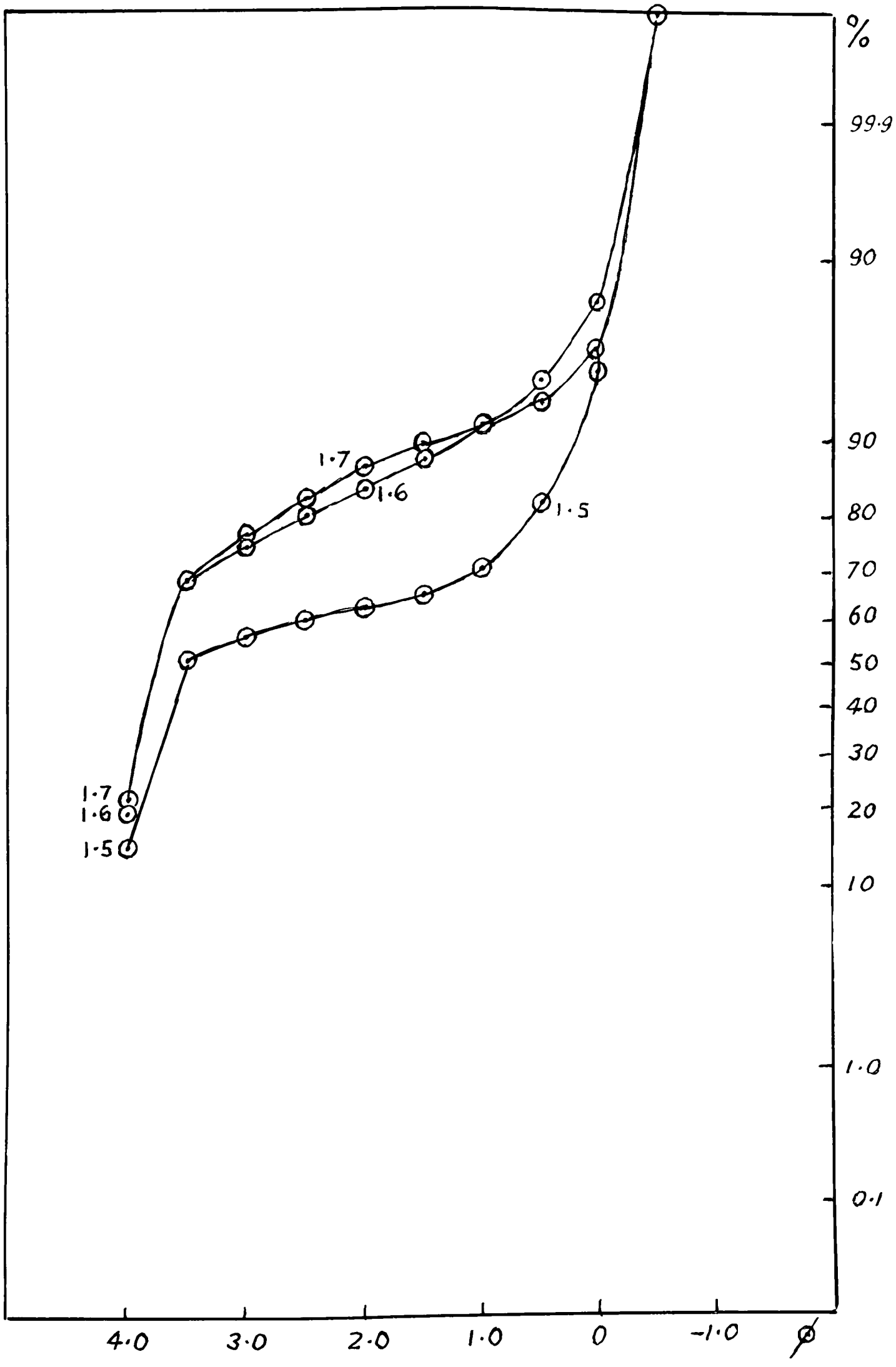

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Figure **6.11**

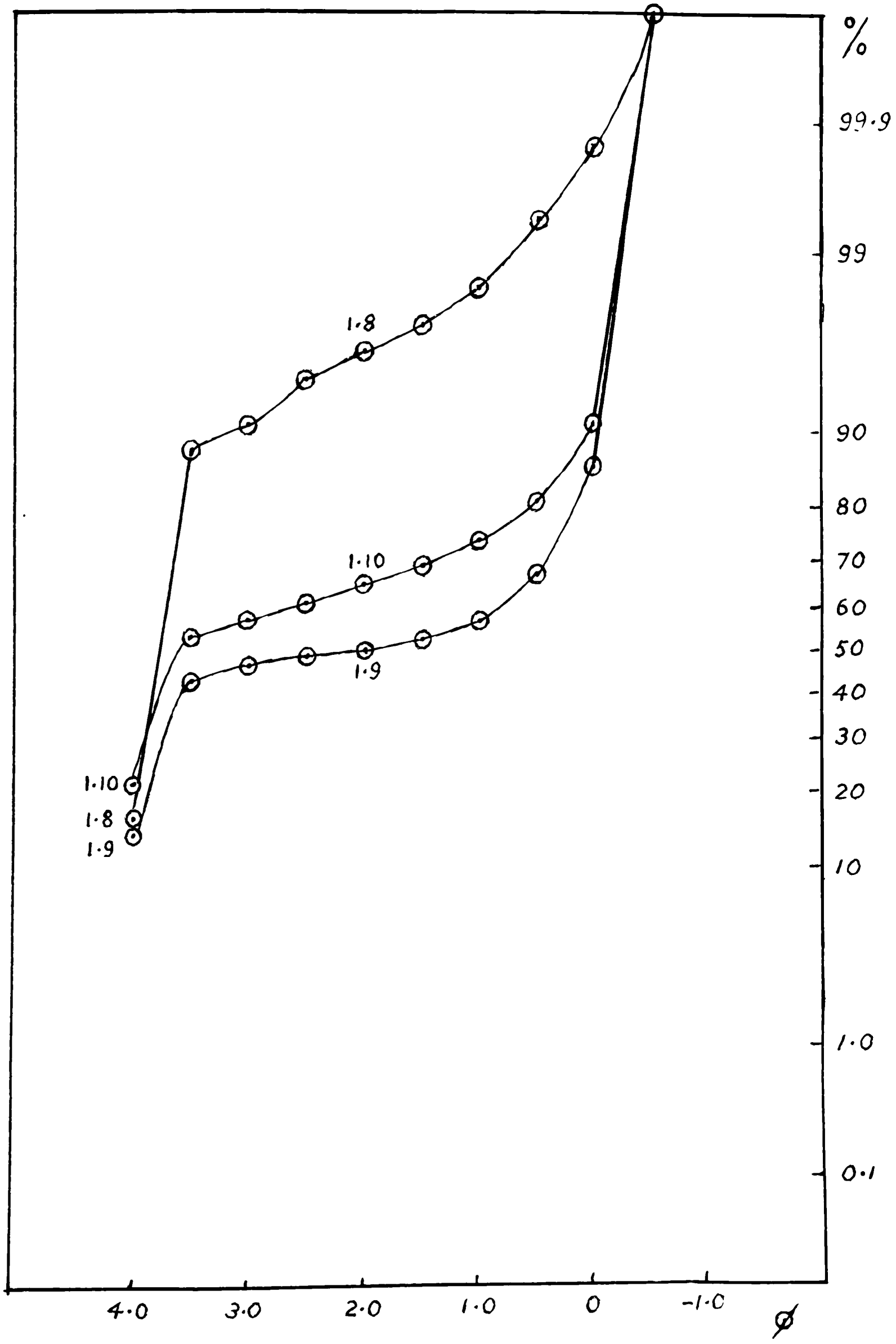
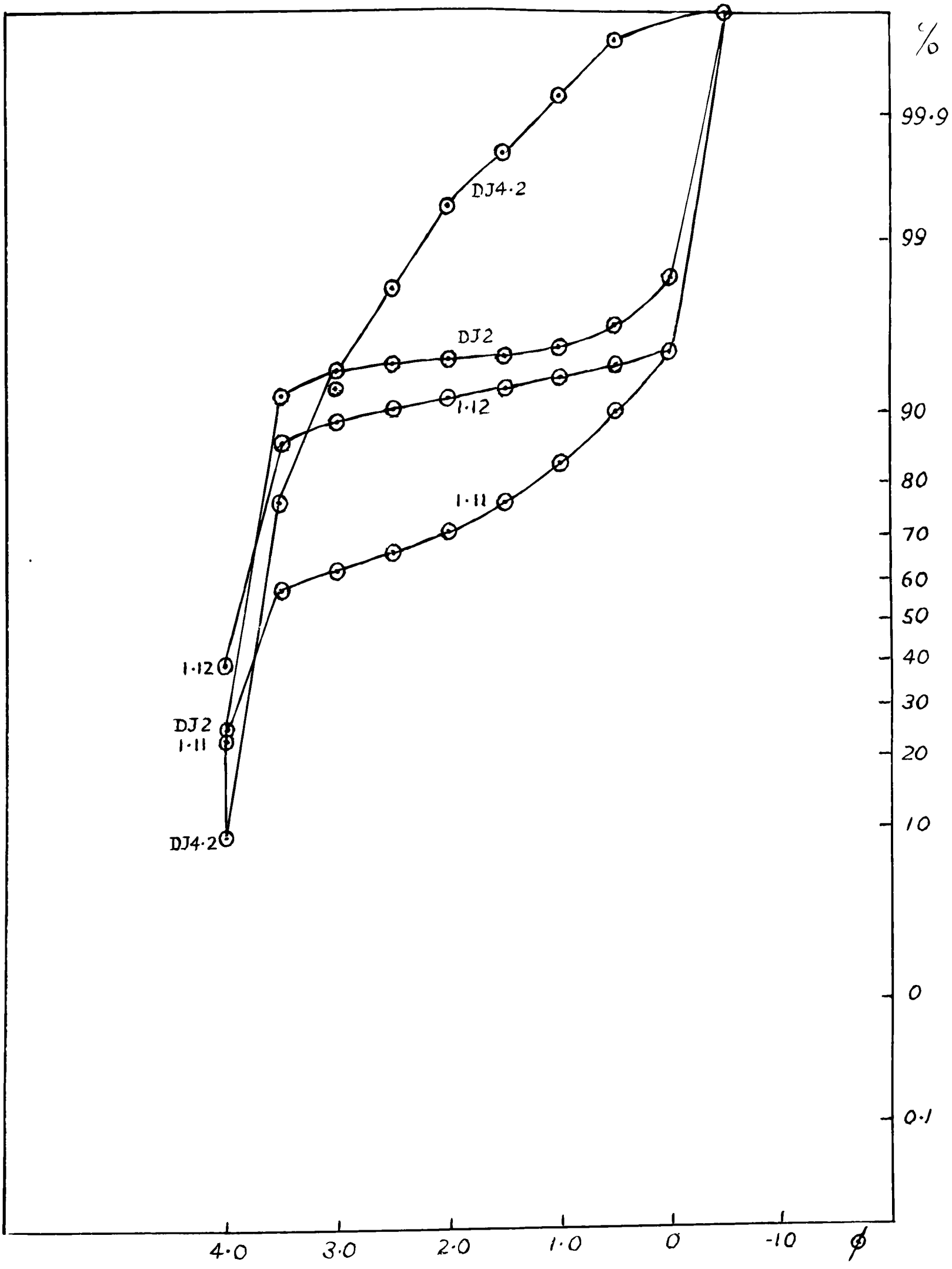

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Figure **6.12**


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have less than 10% gypsum and are comparatively well-sorted, these are small nebkha accumulations around plants growing on the dune. DJ 4.2 is from beneath a surface gypsum crust on another dune in the vicinity (see fig. 5.2C and chapter five). Once again, the material is well-sorted and has less than 15% gypsum (see table 5.8). Bimodality of these dune sands results from mixing of fine quartz grains with larger lenticular gypsum crystals. Those samples with the highest gypsum content (DJ 1.4, DJ 1.5, DJ 1.9, and DJ 1.10) have the strongest bimodal grain size distribution curves. It is possible that since gypsum is less dense than quartz, larger grains may be transported by the wind. Alternatively, the absence of large grains in the superficial nebkha deposits may indicate post-depositional crystal growth under the influence of wetting by meteoric water, though their gypsum content is low to begin with.

The gypsum crusts at the crests of dunes near Djemma (DJ 1.3, table 5.8) are chemically distinct from the gypsiferous sands. The gypsum content is in excess of 70% and there is less than 15% insoluble residue, suggesting that processes other than in situ consolidation of aeolian deposits are at work. Since these crusts cut across aeolian bedding structures, unlike the underlying calcium carbonate horizons, they do not constitute discrete gypsum-rich aeolian deposits. Nevertheless, when the dune sands and the crust are compared, there is a positive relationship between potassium, iron and aluminium content and the amount of insoluble residue and also between gypsum and strontium levels, while calcium carbonate, sodium and magnesium concentrations are fairly constant. It is possible, therefore, that gypsum enrichment relative to insoluble residue explains the chemical differences. A process which accounts for this is described in chapter seven.

At Zaafrane (ZAA, see chapter five and table 5.8 for chemical analyses) in-

terdune sands have a gypsum content of 35% to 65% and slightly bimodal grain size distribution curves (see fig. 6.13) though this is less pronounced in the subsurface sample (ZAA 3.2) and partially consolidated crust (ZAA 3.1). While it is probable that the gypsum crusts are croûtes de nappe, it is possible that they are aeolian deposits cemented under the influence of groundwater. Despite a gypsum content of over 40%, the dune sands (ZAA 2.2) at this site have a unimodel grain size curve. Since these are active dunes, while that near Djemma is partially fossilized, it is likely that bimodality is a result of post-depositional crystal growth or coalescence. At El Faouar (EF) and near El Hamma de Gabès (EH), static dunes with up to 50% gypsum (see tables 5.9 and 6.8) are also bimodal in their grain size distribution (see fig. 6.14) while at Seftimi, samples of sand from the crest, middle and base of the leeward side of an 8.0 m high active dune moving away from the edge of Chott el Fedjedj, are unimodal (see fig. 6.15). These sands have gypsum contents ranging from nearly 25% at the crest to 14% near the base (see table 6.8). Even though they are not bimodal, it is interesting to note that median grain size decreases with decreasing gypsum content. Microscope analysis of these samples shows that particle size groups below 125 μm never contain more than 20% gypsum while those above 250 μm always contain less than 20% quartz grains (100 grains were examined in each grain size range). The active dunes at Zaafrane (ZAA 2.2) contain less than 10% quartz grains among the particles over 177 μm in diameter and less than 10% gypsum grains among the particles below 88 μm. Evidently, the density of the minerals is critical but the scarcity of small gypsum grains may be a result of their greater susceptibility to solution than larger grains or of very light grains being winnowed out. In effect, bimodal grain size distribution curves may be attributed to two factors, post-depositional gypsum grain accretion and the different densities of gypsum and quartz. Microscopic examination of the wind blown gypsum grains also shows them to be essentially lenticular in habit but

Figures 6.13 to 6.15 (overleaf)

Cumulative frequency curves showing grain size distribution characteristics of sand samples from dunes in southern Tunisia.

6.13, dune sand (ZAA 2.1 and ZAA 2.2) and partially consolidated crust samples from Zaafrane (see table 5.8 for chemical analyses).

ZAA 2.1 (29.13% finer than 62.5 μm)
median 73 μm, 3.77 ø

ZAA 2.2 (10.76% fines)
median 115 μm, 3.10 ø

ZAA 3.1 (19.86% fines)
median 74 μm, 3.74 ø

ZAA 3.2 (18.05% fines)
median 70 μm, 3.83 ø

6.14, dune sand samples from El Fouar (EF) (see table 5.9 for chemical analyses) and El Hamma de Gabès (EH) (see table 6.8 for chemical analyses).

EF 1.1 (21.79% fines)
median 76 μm, 3.70 ø

EH 9 (19.15% fines)
median 86 μm, 3.52 ø

EH 10 (0.96% fines)
median 84 μm, 3.56 ø

6.15, active dunes near Seftimi, south of Chott el Fedjedj (see table 6.8 for chemical analyses).

SEF 1 (11.08% fines)
median 142 μm, 2.80 ø

SEF 2 (9.19% fines)
median 108 μm, 3.20 ø

SEF 3 (5.65% fines)
median 90 μm, 3.47 ø

Figure **6.13**

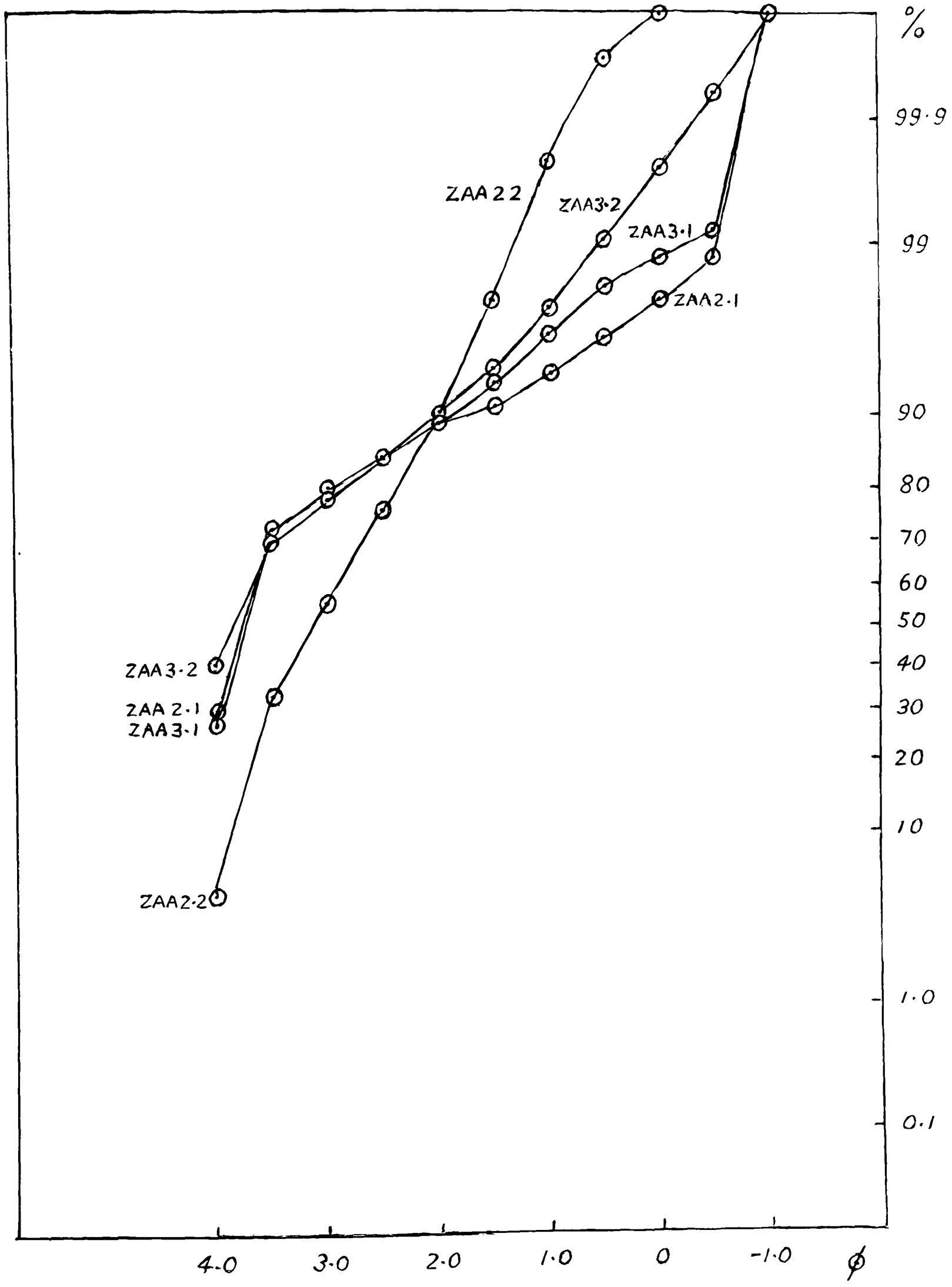

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Figure **6.14**

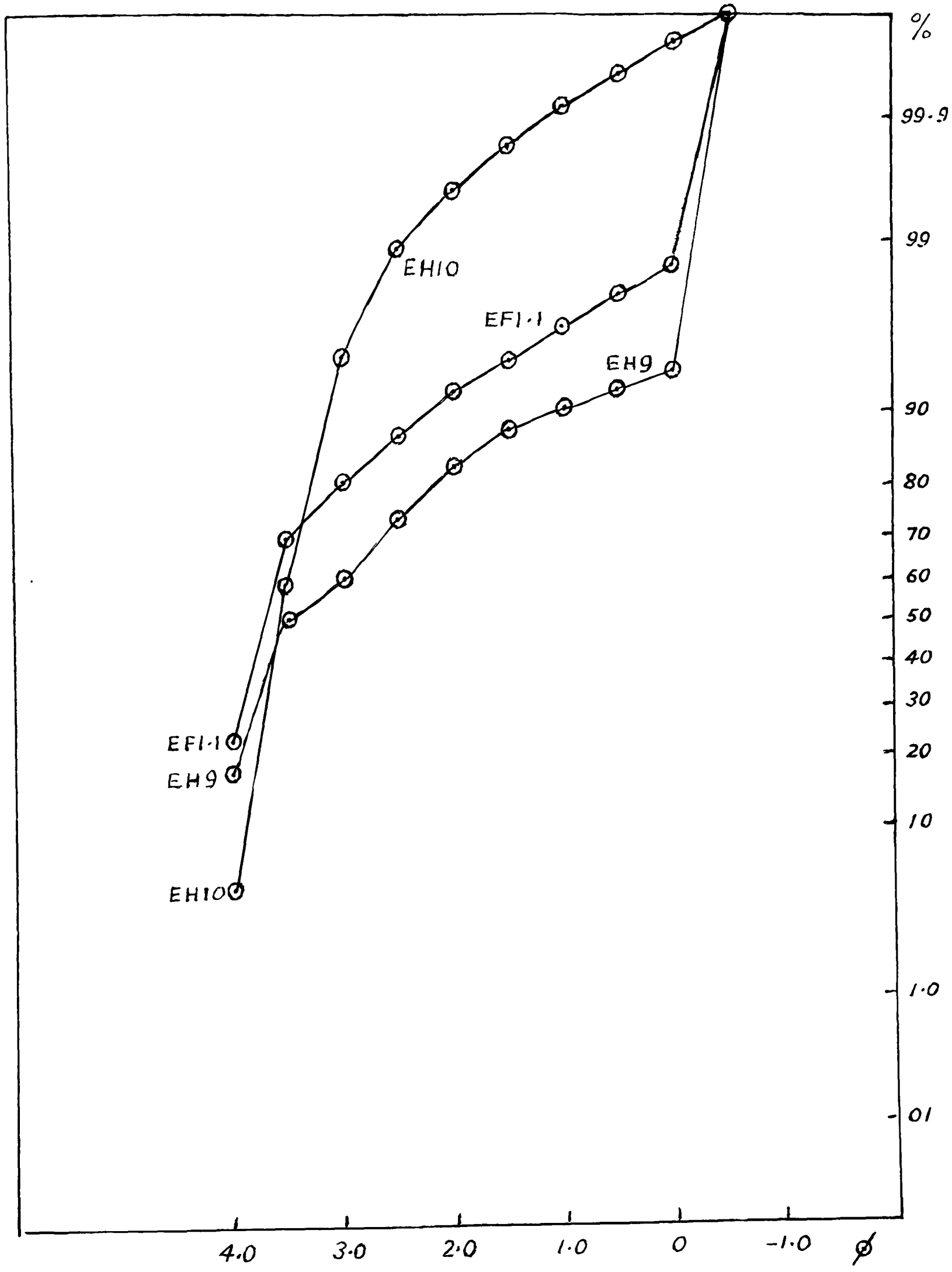
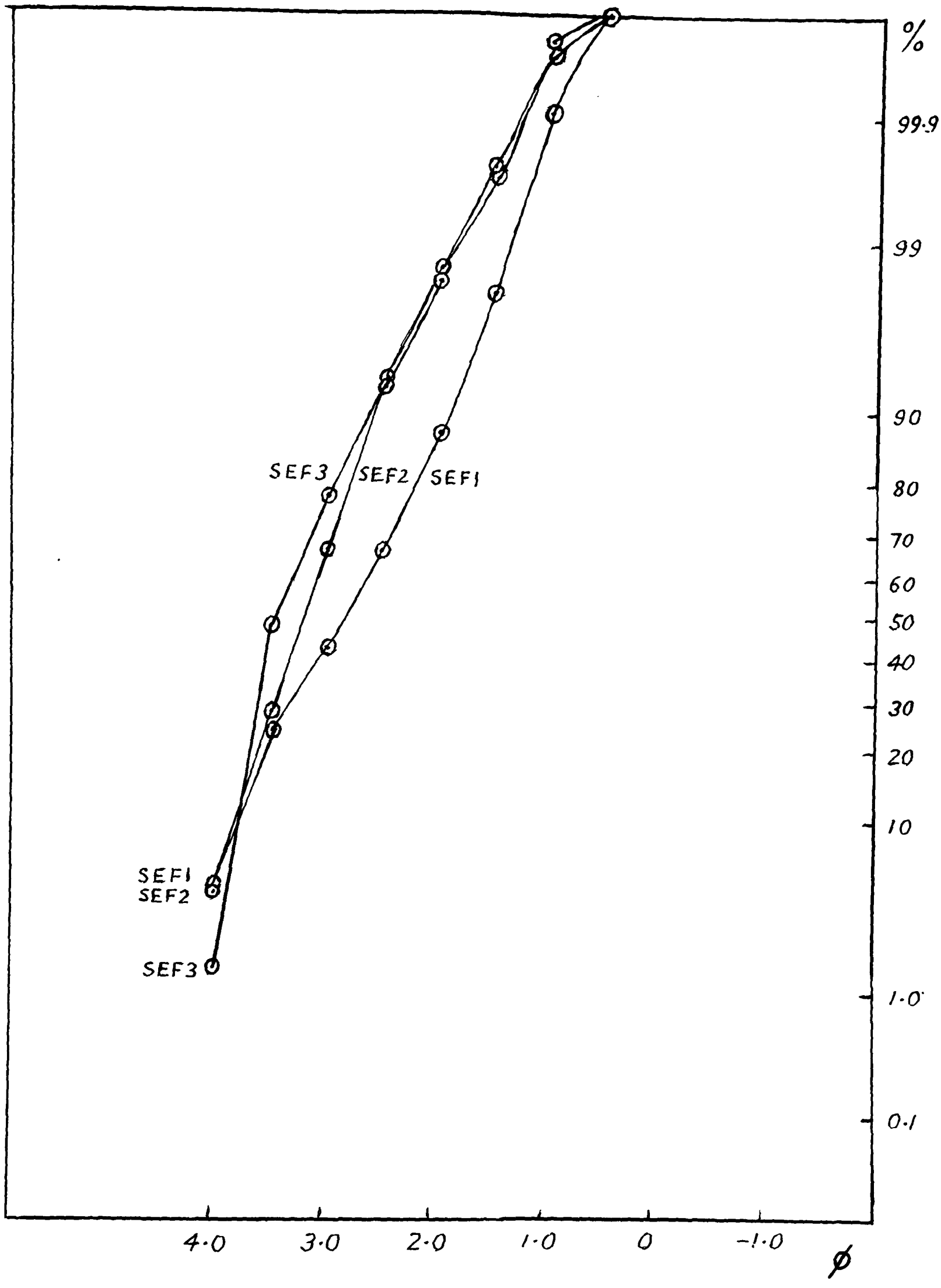

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Figure **6.15**


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ϕ

with considerable solutional pitting and impact fractures. The micropetrographic analyses presented in chapter five show that in most cases lenticular crystals are well formed and have distinct rims, though some larger porphyroblasts show signs of solution. It is impossible that such delicate crystals are primary aeolian deposits, rather they must be the product of in situ recrystallization (Rozanov, 1961, pp. 231-245; Page, 1972; Tucker, 1978). The most striking features of the quartz grains are the prevalence of solutional pitting and their tendency to be more rounded than many aeolian sand grains (Goudie and Watson, 1981). These characteristics may be inherited from the source materials in Cretaceous bedrock or may be the product of solutional etching in the alkaline Chott environments from which they were deflated (see chapter seven).

At Bir Zefazfa (OT) the sands overlying gypsum crusts in the river bed and on the surrounding land surface are not gypsiferous (see table 5.9). Surface sands are fine well-sorted materials, while bimodality of the basal sands (OT 4.2) suggests mixing of aeolian and fluvial deposits (see fig. 6.16). The high gypsum content of the crusts (OT 1.2 and OT 4.3) suggests that they are neither a desert rose croûte de nappe form nor consolidated aeolian sediments, unless gypsum enrichment has taken place. Stabilized dunes generally have higher gypsum contents than overlying or closely associated active dunes, which are younger, as at the eastern end of Chott el Fedjedj (EH 9 is a stable dune sample while EH 10 is active, see table 6.8). This lends support to Coque 's (1958) concept of cyclic deflation of gypsum-rich chott deposits, this having taken place only when lacustrine evaporation occurred after prolonged inundation. It must be pointed out, however, that the aeolian deposits examined here all have significantly less gypsum than most surface crusts.

Besides these reservations about the applicability of the aeolian transport

model, several stratigraphic inconsistancies also exist. Though the close relationship between some gypsum crusts and dunes in the Nefzaoua is evident, the absence of dune topography on the gypsum encrusted glacis slopes of the Chott region and Djeffara may suggest an alternative source of gypsum (Le Houérou, 1960; Bureau and Roederer, 1961; Page, 1972). It should be stressed, however, that aeolian deposits do not always exhibit well developed bedding structures or dune morphology. Blanket sand deposits are not an uncommon feature of many desert environments (Folk, 1978). Besler (1972) pointed out that in the north-west of the region, most sands are fluvial and lacustrine deposits, only in the Nefzaoua is aeolian sorting pronounced. Similarly, the location of many crusts is not compatible with an aeolian origin (Bureau and Roederer, 1961). It is possible that some of the crusts in the zone peripheral to the Chotts and the coast are altered wind deposited silts without well developed aeolian structures. The only evidence of these sorts of materials are, however, the loess deposits of the Matmata Highlands. These cover gypsum crusts and rarely contain more than 4.0% gypsum (Le Houérou, 1960), though calcium carbonate levels may reach 25%. Brunnacker (1973) recorded some gypsum horizons in loess deposits but indurated strata are usually composed of calcrete. Table 6.9 provides chemical analyses of three dominantly micritic calcrete horizons (M 1.1, M 1.2 and M 1.3) each about 50 mm thick from a 2.0 m section of loess south of Matmata and a sample of powdery calcium carbonate from the base of the section (M 2). The loess deposits are piled up against the western and south-western side of the highlands and reach thicknesses of 20 m. Thinner deposits are also found on the eastern flanks (Brosche and Molle, 1975, 1976). They represent Saharan dust deflated from the south-east, the Grand Erg Oriental (Grolier, pers. comm.).
The potential for such transport is supported by the evidence of Saharan dust in the Mediterranean and Central Europe (Goudie, 1978; Eriksson, 1979). The chemistry of these materials, however, precludes the likelihood that they

Table 6.8 (overleaf)

Chemical and mineralogical composition of dune sand from El Hamma de Gabès (EH) east of Chott el Fedjedj and Seftimi (SEF) south of the chott (figs. 6.14 and 6.15 provide grain size data).

Table 6.9 (overleaf)

Chemical and mineralogical composition of calcretes within loess deposits in the hills near Matmata (M) in southern Tunisia (see 4.1 for location).

TABLE 6.8

	EH 9	10	SEF 1	2	3
IR%	37.62	88.54	58.84	63.14	75.52
Gyp%	52.82	1.17	24.49	22.98	13.96
$CaCO_3$%	3.94	6.96	3.95	3.49	3.02
Na^+ppm	13.50	3.50	2000	2600	7000
(NaCl%)	0.34	0.09	0.51	0.66	1.78
K^+ppm	600	750	850	650	1050
Sr^{++}"	600	100	500	500	300
Mg^{++}"	2000	2000	2100	2300	3000
Fe^{tot}"	1200	2300	1500	1600	1900
Al^{+++}"	700	2500	1300	1300	1700
$xsCa^{++}$"	5900	-	31200	22500	7400

TABLE 6.9

	M 1.1	1.2	1.3	2
IR%	6.24	12.88	6.78	5.28
Gyp%	2.75	1.51	5.17	0.78
$CaCO_3$%	85.85	79.35	82.37	90.49
Na^+ppm	3500	3250	3400	5700
(NaCl%)	0.89	0.83	0.86	1.45
K^+ppm	500	670	500	260
Sr^{++}"	2300	600	900	100
Mg^{++}"	3200	3300	3500	200
Fe^{tot}"	1600	2900	1800	2100
Al^{+++}"	1700	3100	2000	1000
$xsCa^{++}$"	-	800		-

are a major source of gypsum making up the crusts of southern Tunisia.

Page (1972) presented a further reservation to Coque's aeolian model, arguing that many deposits overlying near-surface gypsum crusts cannot be explained in terms of sedimentation subsequent to the aeolian phase. While many of the accumulations of sands and silts overlying gypsum crusts are of aeolian origin, for example, the well-sorted sands above the crust at Bechri (CD 1.4, see fig. 6.17) and the large nebkha described at Oued Demna (see chapter five), others are more difficult to account for. The grain size distribution curves for sands above and beneath the surface crusts on the edge of Chott Djerid between Tozeur and Nefta (TN 1.2 and TN 2) are very similar (see fig. 6.17), their coarseness (medians 470 μm and 270 μm) indicating a fluvial origin. While it is possible that two periods of fluvial sedimentation have been interrupted by an aeolian phase, subsurface displacive accretion of the crust cannot be overruled. Even more engimatic are coarse sediments upon gypsum crusts. These may take the form of lag gravels (Hallsworth, Gibbons and Darley, 1951) as in the Central Namib Desert or large boulders as at Limaguess (L) in southern Tunisia and on the Lagunenberg near Cape Cross on the Namib Coast (CC) (see plate 5.23). In many cases, the topographic location of these materials precludes their deposition by gravity fall from bedrock outcrops or fluvial action. Jessup (1960 a.) held that coarse gravels overlying subsurface gypsum crusts in south-eastern Australia were moved upwards through the deposit by clay swelling and contraction during wetting and drying cycles. Springer (1958) and Cooke (1970) advocated this process to explain desert pavements in Nevada and Peru. The gypsum crusts of southern Tunisia and the Central Namib Desert usually have low clay contents, so upward translocation of coarse material is unlikely unless mineralogical volume changes resulting from thermal fluctuations occur. Even if this is the case, the volumetric variation between

Figures 6.16 and 6.17 (overleaf)

Cumulative frequency curves showing grain size distribution characteristics of sands from Bir Zefazfa (OT), between Tozeur and Nefta (TN) and Bechri (CD).

6.16, dune sands from the Oued Tarfa near Bir Zefazfa (see table 5.9 for chemical analyses).

OT 1.1 (2.66% finer than 62.5 μm)
median 69 μm, 3.89 ø

OT 4.1 (1.80% fines)
median 79 μm, 3.65 ø

OT 4.2 (17.13% fines)
median 68 μm, 3.91 ø

6.17, surface sands from Bechri (CD) and from between Tozeur and Nefta (TN) (see table 5.28 for chemical analyses).

TN 1.2 (3.37% fines)
median 470 μm, 1.08 ø

TN 2 (6.25% fines)
median 270 μm, 1.83 ø)

CD 1.4 (29.97% fines)
median 106 μm, 3.22 ø

Figure **6.16**

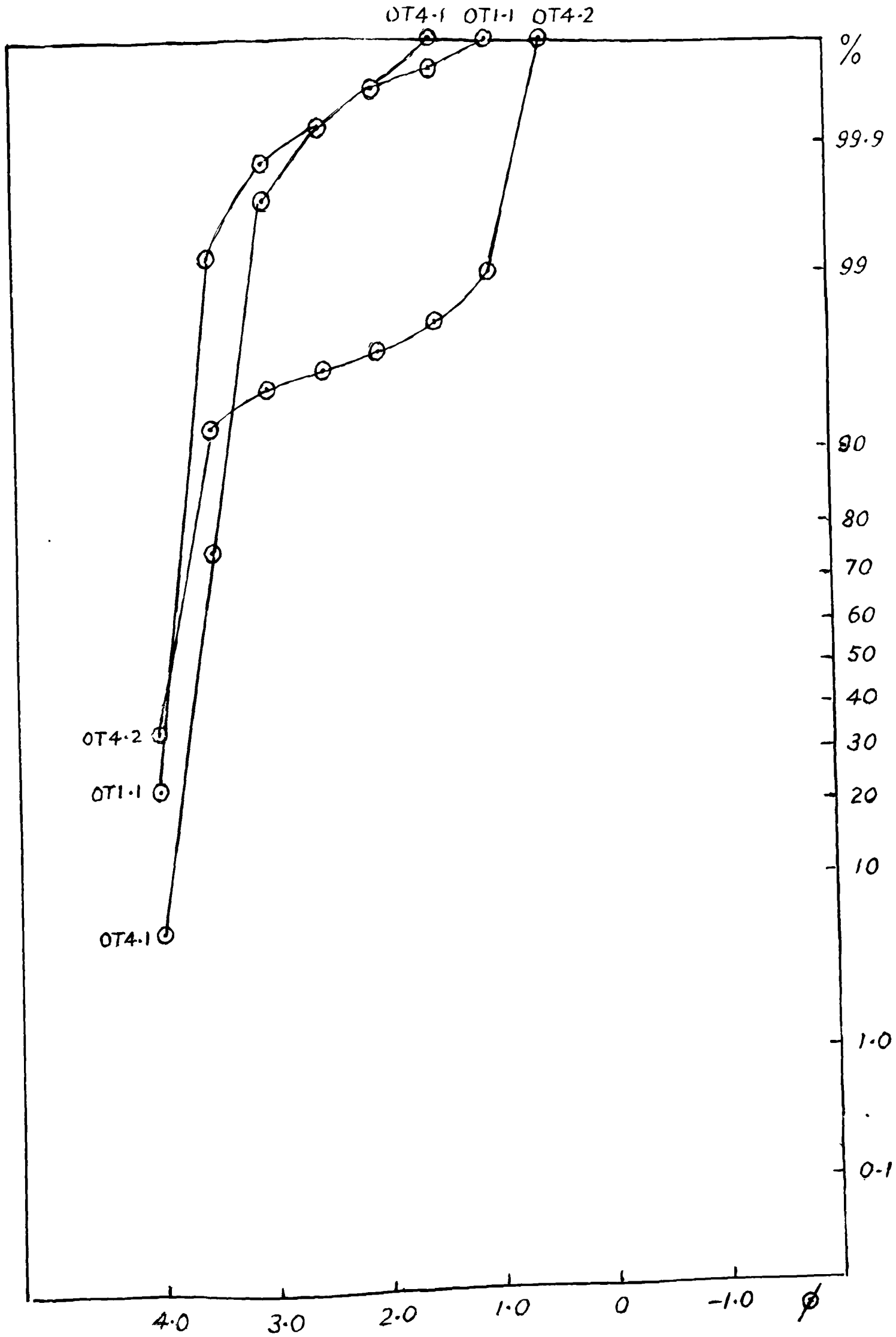
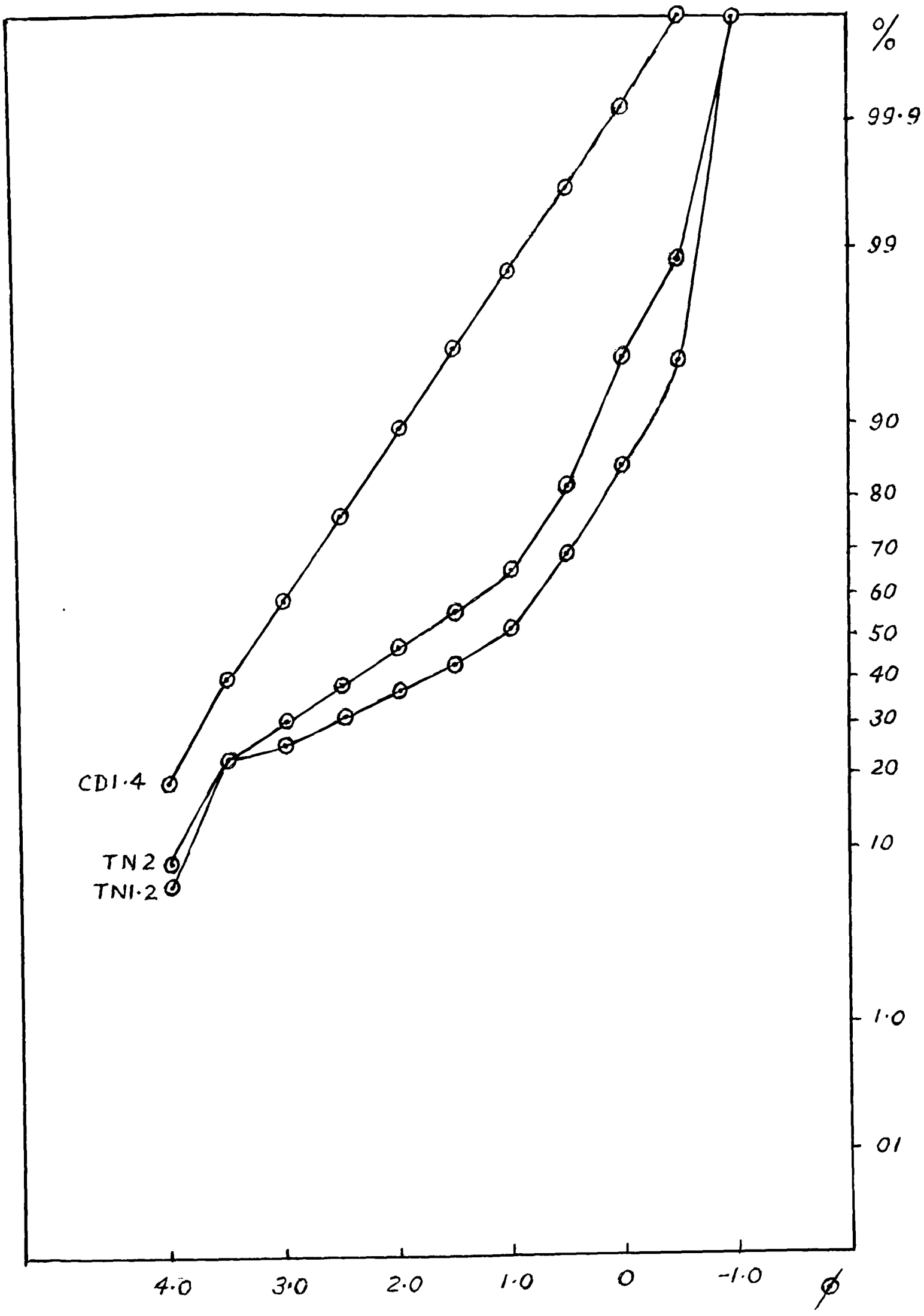

OT4-1
OT1-1
OT4-2
%
99·9
99
90
80
70
60
50
40
30
20
10
1·0
0·1
OT4·2
OT1·1
OT4·1
4·0
3·0
2·0
1·0
0
-1·0

Figure **6.17**


%
99·9
99
90
80
70
60
50
40
30
20
10
1·0
01
CD1·4
TN2
TNI·2
4·0
3·0
2·0
1·0
0
-1·0
ϕ

gypsum and hemihydrate is too small (Chatterji and Jeffery, 1963, see chapter one) and the surface clasts too large for it to be an effective mechanism. Surface gravels are usually aeolian lag deposits, residual after winnowing of finer particles, therefore underlying gypsum crusts may be accounted for only in terms of subsurface accretion. The greatest drawback of this hypothesis is the difficulty in explaining the high gypsum content of many crusts. As has been shown this also applies to the groundwater and aeolian models.

A final point should be made regarding the aeolian model. The gravel plains north of the Kuiseb in the Central Namib do not exhibit any aeolian depositional landforms, save for a few along the coast where sand has crossed the Kuiseb downstream from Rooibank, owing to the low frequency of floods reaching the Atlantic. It is possible that at low sealevel stands during the Pleistocene, coastal sabkhas were more extensive than at present and evaporitic deposits were deflated inland. Present coastal sands, however, contain less than 10% gypsum (Nagtegaal, 1973) and active coastal dunes even less, those at Sandwich Harbour (SH) have under 1.0% (see table 6.10). It is unlikely, therefore, that aeolian materials play a significant role in the development of gypsum crusts in this area. Near Cape Cross, however, lagoonal pan deposits are often highly gypsiferous (CC 2.2, table 5.7) and are important potential sources.

iv. Atmospheric inputs.

A number of workers have argued that the deposition of aerosols on the land surface is a critical factor in the evolution of saline soils and crusts in desert areas. In New Zealand, Gibbs (1949) found that sodium chloride in the atmosphere near the coast was leached through the soil zone mobilizing

TABLE 6.10

Chemical and mineralogical composition of dune sand from Sandwich Harbour (SH) in the Central Namib Desert (see fig. 4.4 for location).

SH	1.1	1.2	1.4
IR%	93.54	95.18	95.92
Gyp%	1.01	0.70	0.96
$CaCO_3$%	1.15	0.69	0.23
Na^+ ppm	1000	1000	900
(NaCl%)	0.25	0.25	0.23
K^+ ppm	850	750	500
Sr^{++} "	-	-	-
Mg^{++} "	2900	1900	1500
Fe^{tot} "	5500	3200	2900
Al^{+++} "	4200	3200	2800
$xsCa^{++}$ "	-	-	-

clays. Similarly, Downes (1954) held that many Australian soils were characterized by features suggestive of previous salinity, though soluble salts are no longer present. He felt that deposition of atmospheric cyclic salts during arid periods permanently altered soil structures, even if they are removed by subsequent leaching. In semi-arid and arid regions, cyclic salts are present within the soil zone (Ulrich et al., 1959) where they accumulate at the maximum depth of leaching (Yaalon, 1961, 1964 a., 1971 a.).

The movement and deposition of airborne salts may occur in several ways but these may be subdivided into dry transport and fall-out and those in solution. Dry fall-out may be equated with aeolian dust deposition; Eriksson (1955) found that it was a major factor in river water chemistry. The particles may be derived from a wide variety of sources not the least significant being human influences. Whitehead and Feth (1964) found that considerable amounts of gypsum deposited from the atmosphere in southern California were derived from cement works. Lettau (1939) remarked upon the high levels of atmospheric pollution in Africa resulting from the burning of grassland pastures in the dry season. Dust fall-out rates are very variable on a global scale, Goudie (1978) quoted from 6.6 mm/1000 years to 2.01 m/1000 years, the latter figure being from Iraq. Though gypsum contents have only rarely been determined, calcium carbonate levels are frequently high, 66.1% to 69.5% in the case of the Iraqi dust. As is shown in chapter seven, thick gypsum crusts may develop in areas where gypsiferous dust is deposited provided there is sufficient moisture to allow mobilization in solution, and thereby cementation, and insufficient to cause leaching through the soil zone. Dry fall-out probably provides a significant source of gypsum in continental areas where marine influences are negligible. In most cases however, solid phase soluble salts in atmospheric dust are derived from coastal and lacustrine environments by evaporation of sea-spray or defla-

tion of saline sediments. Yaalon and Lomas (1970) recorded sulphate deposition in Israel amounting to 8.7 mg/day on a 0.01 m^2 muslin cloth held vertically 1.5 m above the ground at a distance of 10 m from the sea. 56.9% of this was dry fall-out. 6.0 km inland, deposition totalled 1.39 mg/0.01 m^2/ day, 65.5% of it as dry material. Yaalon (1964 b.) reported surface airborne salt deposition totalling 10 to 20 kg/ha/year 50 km from the coast, while in the littoral zone deposition from sea-spray may reach 50 to 80 kg of chloride/ha/year (Yaalon, 1961; 1971 b.).

Rainwater is also a potential source of atmospheric salt deposits. In the Lower Rio Grande area, 24.5 kg of sulphur/ha/year are received (Fanning and Lyles, 1964). Atmospheric sulphur is usually in the form of sulphur dioxide and hydrogen sulphide (Holt, Cunningham and Engelkemeir, 1978), sulphate comprises only about 10% of all gaseous sulphur (De Bary and Junge. 1963). Nevertheless, industrial pollution raises rainwater sulphate levels to about 2.15 mg/l over the United States of America, while over the oceans they average 0.5 mg/l (Junge and Werby, 1958). Since smaller water droplets tend to have a higher salt concentration than larger ones (Yaalon, 1968), probably because salt particles form the nuclei around which atmospheric moisture accretes, fog and mist have greater salinites than rainwater per unit volume (Duce, Woodcock and Moyers, 1967). In the coastal zone of the Central Namib Desert, fog water has an average sodium chloride content of 0.333 g/l and 100 kg/ha may be deposited annually (Boss, 1941). Walter (1937) recorded concentrations which could account for 1000 kg of sodium chloride being deposited per hectare per year. Boss (1941) also noted that Cl/SO_3 ratios are less than in seawater, which suggests a more efficient releasing mechanism. Eriksson (1958) and Mizutani and Rafter (1969 b.) attributed this to gaseous hydrogen sulphide emission from seawater, while chloride is derived through seawater evaporation (Cunningham, 1941). Hydrogen sulphide is oxidized to

sulphur trioxide via sulphur dioxide. The decrease in Cl/SO ratios from 7.1 near the coast to 2.73,60 km inland (Boss, 1941), suggests that this is a gradual process. Any airborne calcium carbonate would be oxidized to gypsum by sulphur dioxide hence there is a potential for gypsum deposition in the coastal fog belt. On the basis of the average Cl/SO_3 ratio of 7.1 it may be assumed that with a mean deposition rate of 120 kg of sodium chloride/ha/year, 22,05 kg of gypsum/ha/year would also be deposited if all chloride was as NaCl and all SO_3 converted to gypsum. The latter is unlikely owing to solubility limitations (see chapter seven).

Several workers have attributed the formation of silcretes, calcretes and gypsum crusts in the Central Namib to deposition of salts from fog moisture (Stapff, 1887; Kaiser, 1923; Jaeger, 1927; Range, 1927). The process overcomes the need to explain the absence of aeolian depositional landforms which would be present if the source material for the crusts was gypsiferous sand. It should be pointed out, however, that the mechanisms involved still cannot account for the development of crusts beneath coarse surface sediments which cannot post-date crustal accretion. In order to overcome this problem, Tolchel'nikov (1962) invoked downward leaching of salts deposited at the surface by mists. Near Elista in the Kalmyk Republic, to the west of the Caspian Sea, Shumakov and Mikhovich (1960) recorded deposition of 33.18 kg of water soluble salts per hectare over a 35 hour period. Annual rates reach 475 kg of salt/ha/year. Of this, 90.6% is sulphate, though 83.5% is sodium sulphate. An alternative explanation has been mooted in the Central Namib Desert. This involves authigenic gypsum formation under the influence of chemical compunds present in fog water.

d. Authigenic gypsum crust development (see fig. 6.18A).

Martin (1963) held that the gypsum crusts of the Central Namib Desert are nearly all developed upon the calcretes of Tertiary age or calcium carbonate-rich bedrock. While Scholz (1963, 1968, 1972) pointed out that this was erroneous, in that in the vicinity of Gobabeb calcretes overlie gypsum crusts which in turn lie upon mica schist or granite bedrock (see fig. 5.7C), he agreed with Martin's genetic model. This involves the in situ alteration of calcrete to gypsum under chemical and bacterial influences. Similar models were proposed by Passarge and Meinardus (1933, p. 89) in Egypt and Meckelein (1959) in Libya but, although the former workers held that "chemisch wäre der Vorgang recht wohl möglich", there are some inherent difficulties. The major of these is the source of sulphate. While this may be available through the action of bacteria on bedrock sulphides (Vialov, 1941; Jessup, 1960 b.; Gibson, 1962; see chapter two), these are not abundant in the non-gypsiferous deposits of the Central Namib. The same problem applies to Rozanov's (1961) model of gypsum crust development from limestones in Central Asia.

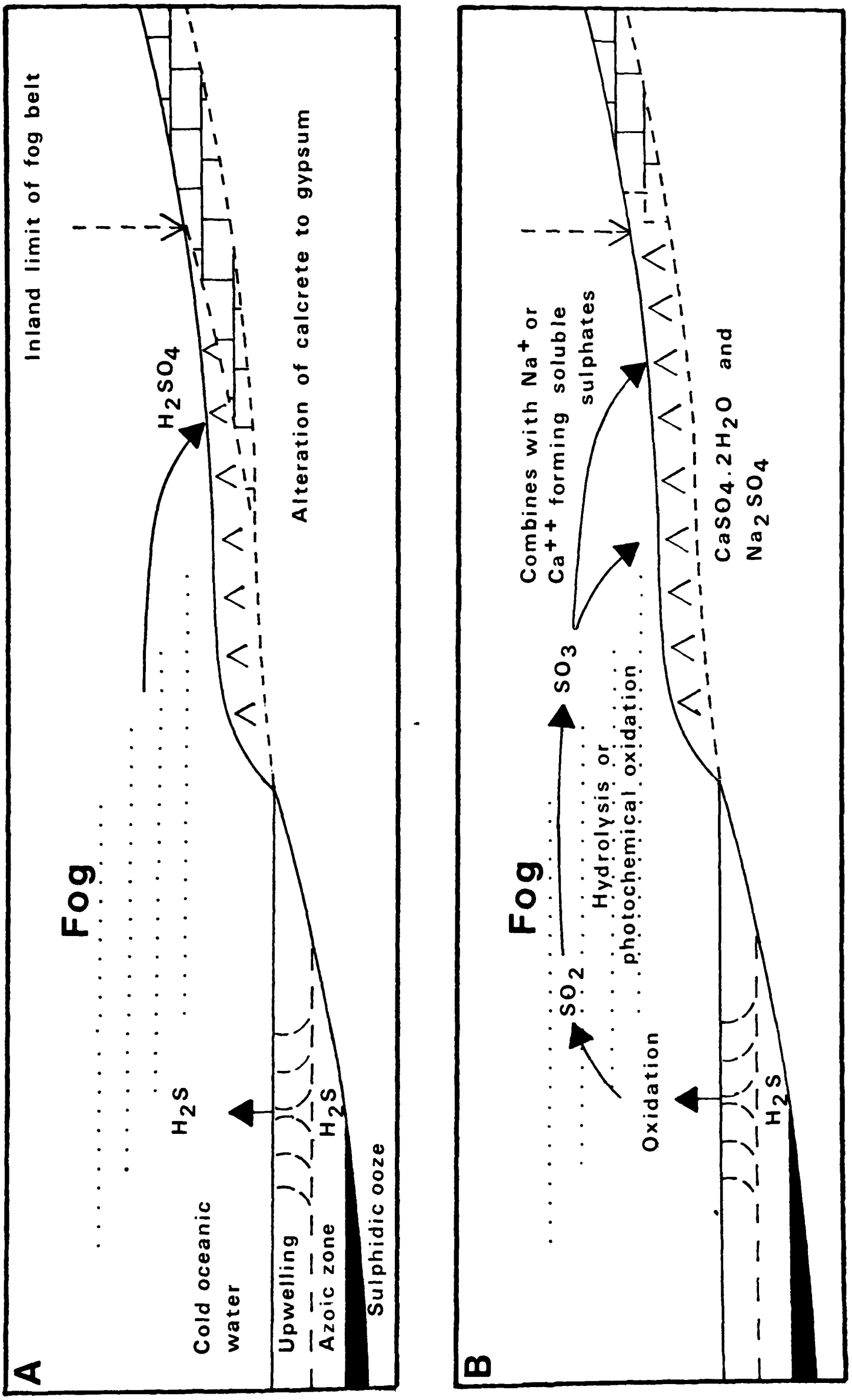
Martin (1963) suggested that the sulphate was derived indirectly from hydrogen sulphide. He pointed out that along the coast of the Namib the cold Benguela current is rich in plankton. Oceanic conditions are so unusual that Lüderitz has the lowest seawater salinity on the South African coast and Walvis Bay is the only tropical habitat of the kelp species Laminaria schinzii (Isaac, 1937). During the summer, warm saline waters move southwards into the Walvis Bay area killing the organisms which settle to the sea-bed. Their decay depletes the supply of oxygen in bottom waters and, in the anaerobic environments, sulphate reducing micro-organisms produce hydrogen sulphide which goes into solution. Though the azoic conditions are stable between 50 m and 150 m

Figure 6.18 (overleaf)

Models of authigenic gypsum crust development in the Central Namib Desert.

A. Liberation of hydrogen sulphide from bottom waters off-shore and alterations of calcrete to gypsum by sulphuric acid in fog moisture (after Martin, 1963).

B. Formation of soluble calcium and sodium sulphates in fog moisture following oxidation of hydrogen sulphide to sulphur dioxide and sulphur trioxide.


A
Fog
Inland limit of fog belt
H_2SO_4
Alteration of calcrete to gypsum
H_2S
Cold oceanic water
Upwelling
Azoic zone
H_2S
Sulphidic ooze
B
Fog
SO_2
Hydrolysis or photochemical oxidation
Oxidation
H_2S
Combines with Na^+ or Ca^{++} forming soluble sulphates
$CaSO_4.2H_2O$ and Na_2SO_4

below the surface, under certain oceanic conditions turbulence may result in upwelling of bottom waters and liberation of hydrogen sulphide which is carried inland dissolved in fog moisture. Martin (1963) argued that the formation of sulphuric acid from hydrogen sulphide would result in alteration of calcium carbonate surface deposits to gypsum. However, the reaction of sulphate and hydrogen sulphide species forming sulphuric acids has only been accomplished under low pH conditions at low temperature (Husain and Krouse, 1978). Sulphuric acid may be a by-product of pyrite oxidation (Broughton, 1934).

A considerable amount of work has been undertaken into the nature of sea-bed sediments in the Walvis Bay area and into the causes of hydrogen sulphide formation and liberation particularly in relation to fish mortality. Siesser and Rogers (1976) and Rogers (1977) found that under the influence of sulphate reducing Desulphovibrio desulphuricans bacteria, dissolved sulphates in bottom water form hydrogen sulphide which reacts with free iron to form authigenic pyrite and excess amounts react with dissolved calcium carbonate in the sulphidic ooze to form gypsum. Nevertheless, the biochemical environment is very different from that which could exist on the land surface. Indeed, Birch's (1975) work on bottom sediments off the western coast of the Cape Province shows that the conditions in Walvis Bay are very restricted. It has been suggested that fish mortality results from the mixing of hydrogen sulphide-rich bottom waters with surface water, resulting in death by anoxia or poisoning (Classen, 1930; Kaiser, 1930; Copenhagen, 1951, 1953). Upwelling may result from the intermittant discharge of subterranean water from the Kuiseb River (Waldron, 1901; Classen, 1930), which may itself be saturated with hydrogen sulphide and sulphates, into the bottom sediments of the bay. An alternative explanation was proposed by Brongersma-Sanders (1948) who argued that the extent of fish mortality was too great to be the result of Kuiseb discharge. She suggested that the upwelling of cold Benguela

waters resulted in poisonous dinoflagellate blooms. The fact that fish eating sea-bird and seal deaths have also been reported, supports this hypothesis. Copenhagen (1951) found no evidence of poisonous Gonyaulax and Gymnodinium dinoflagellates in the Walvis area, though their presence cannot be ruled out. Tectonic disturbances of bottom sediments, such as that reported by Captain Fitzroy (Narrative of the Voyage of H.M.S. Beagle, Stanbury, 1977 ed.) in the Bay of Concepción, after the earthquake of 1834, are not recorded in this area.

Periodic fish mortality and hydrogen sulphide liberation are probably the result of the combined effect of the upwelling of cold Benguela waters and related poisonous plankton blooms. Upwelling occurs when bottom waters replace surface waters blown offshore by strong winds (Bang, 1971), though the bathymetry of the continental shelf may localize the effects (Currie, 1953; Jones, 1971).

While Martin (1963) argued that hydrogen sulphide liberation and its mobilization in fog moisture was sufficient to facilitate calcrete alteration to gypsum, Besler (1972) suggested that such a chemical process was impossible. She held that carbonates would form polysulphides in the presence of hydrogen sulphide. Alternatively, sulphur in dew and fog moisture may form thiosulphates when it reacts with soil chlorides and nitrates. At high surface temperatures and in the presence of calcium carbonate and water, these may alter to gypsum. Whichever mechanism is invoked, Martin's observation that the calcretes and gypsum crusts have similar purities and chemistries should hold true. He found that the calcium carbonate content of calcretes was in excess of 90% while gypsum crusts had a similar percentage of gypsum, though Fockema and Fraser (1963) pointed out that over a large area purities may be reduced to 70% to 80%. While the calcrete and gypsum components of the section des-

cribed from near Gobabeb (GO) (see fig. 5.7C and table 5.21) show similarities in their contents of the two minerals, their juxtaposition is the reverse of that described by Martin (1963). Furthermore, while chemical comparisons are complicated by bedrock influences, micropetrographic analyses (see table 5.2, GO 4.4) show evidence of alteration of gypsum to calcite rather than the converse. Chemical analyses of micritic calcretes from near Ganab (GA), Homeb (HO) and Tsondab Vlei (TS) (see table 6.11) show variations in calcium carbonate content from about 55% to 80%, generally less pure than surface gypsum crusts in the coastal area (HA and SB), though these are associated with marble bedrock. Accepting the possible influences of insoluble residue content on potassium, iron and aliminium levels, other accessory elements show some differences. Strontium levels in calcretes are low while magnesium content is high when compared with gypsum crusts. Ionic bonding of these elements to the two minerals, magnesium to calcite and strontium to gypsum may account for this, but the source of the elements also indicates that the source of the gypsum is other than altered calcrete. Neither Martin's (1963) model nor Besler's (1972) allow for transport of ions in addition to sulphates or their source materials. Two micritic calcrete samples from a relic pan near Narabeb (NA 2.1 and NA 2.3, see table 6.11 and plate 6.1) show marked differences from gypsum deposits from a different depression in the same vicinity (NA 1.1 and NA 1.2, see table 5.7). Besides slight variations in strontium content, markedly different amounts of insoluble residue and high magnesium levels in the carbonate-rich samples testify that the gypsum is not an alteration product. The calcium carbonate deposit represents a former end-point of one of the "blind" rivers of the Dune Namìb, probably the Tsondab. High magnesium levels in the calcrete compare with those at Tsondab Vlei (TS), the element being derived from dolomites of the Naukluftgebirge in the upper catchment.

TABLE 6.11

Chemical and mineralogical composition of calcretes from Ganab (GA), Homeb (HO), Tsondab Vlei (TS) and Narabeb (NA) (see fig. 4.4 for locations).

	GA 1.1	1.2	1.3	HO 2.1	TS 1.1	1.2	NA 2.1	2.3
IR%	13.52	24.94	18.98	16.14	28.68	36.24	35.70	59.14
Gyp%	0.50	0.56	0.57	0.43	0.76	0.72	0.49	0.38
$CaCO_3$%	78.54	67.32	71.46	75.71	62.99	57.01	52.86	33.33
Na^+ppm	3000	2550	2700	3700	2650	2400	2200	2400
(NaCl%)	0.76	0.65	0.69	0.94	0.67	0.61	0.56	0.61
K^+ppm	1000	1800	1950	900	800	800	1650	780
Sr^{++} "	200	200	200	100	200	100	100	-
Mg^{++} "	5100	5200	6700	6200	10800	8300	10400	5000
Fe^{tot} "	3400	4000	5700	2500	6300	4100	10800	9300
Al^{+++} "	3000	4600	5700	2600	2500	2200	6300	4600
$xsCa^{++}$ "	10400	7100	4500	23900	12000	14200	7200	11700

Plate 6.1

The interdune valley at Narabeb (NA) in the Central Namib Desert. The light area to the right is a calcareous lacustrine deposit which consolidates the surface sands and has caused a minor inversion of relief.



Since gypsum is not present in bedrock in the Central Namib Desert and aeolian and authigenic origins appear inadequate to explain the wide distribution of gypsum crusts, an alternative source must exist. The presence of sulphur dioxide, oxidized hydrogen sulphide, dissolved in fog moisture has been mentioned above. Several mechanisms may account for the development of gypsum crusts through its influence without invoking the wholesale alteration of calcretes. As described in chapter one, sulphur dioxide may be altered to sulphate through hydrolysis by atmospheric water followed by oxidation in the presence of metal ion catalysts (Holt, Cunningham and Engelkemier, 1978). In some cases, hydrolysis may be replaced by photochemical oxidation. It is also feasible that gaseous sulphur trioxide reacts with sodium chloride in solution forming sodium sulphate (Eriksson, 1958).

$$SO_3 + 2NaCl + H_2O \longrightarrow Na_2SO_4 + 2HCl$$

This reaction may account for the high sodium sulphate levels in fog moisture reported by Shumakov and Mikhovich (1960) near the Caspian Sea. As described in chapter two, Krupkin (1963) found that sodium sulphate will react with calcium chloride in the soil adsorption complex to form gypsum. It is also possible that this occurs in fog droplets so any calcium present will be precipitated as gypsum when the moisture evaporates (Eriksson, 1958) (see fig. 6.18B). A further possibility is that free sulphur formed from hydrogen sulphide by Beggiatoa bacteria under aerobic conditions (Butlin, 1953) is deposited and later oxidized by Thiobacilli in the presence of calcium carbonate to gypsum. Native sulphur was described by Reuning (1925) near Walvis Bay. He felt that it was a marine deposit produced by the action of hydrogen sulphide on iron sulphates.

$$4H_2S + Fe_2(SO_4)_3 \longrightarrow 2FeSO_4 + 4H_2O + 5S$$

This is essentially the same process as Baas-Becking and Kaplan (1956) and Bonython and King (1956) proposed to account for native sulphur on the shores of Lake Eyre. It must be pointed out, however, that neither iron minerals nor native sulphur are plentiful in the Central Namib. In southern Tunisia, the iron nodules in the Oued Magroum (EH 7.5) are too diffuse to be of significance in crust development. It is more likely that in the Central Namib gypsum is available in fog and crystallizes at the surface when the moisture evaporates.

Isotopic studies of the sulphate and water crystallization of gypsum making up the crusts of the Central Namib Desert provide further insight into the origins of the various components. Sofer (1978) found that the $\delta D/\delta^{18}O$ ratios of the water of crystallization corresponded to meteoric waters suggesting that whatever the source of the calcium sulphate, it has been subjected to solution and reprecipitation by rainwater. In contrast, sulphur isotope studies show a close similarity between the sulphur in the sulphate component of gypsum making up the crusts and fog water, but not sea-bed oozes off the coast. Carlisle (1978) found that "older" basal gypsum crusts had $\delta^{34}S$ values (‰) of +12.9 to +17.7 (average +15.9), which does not uphold the concept of variable ages (Rust and Wieneke, 1976). The value obtained from one sample of fog-water was +13.41, while two from muds from Sandwich Harbour were -10.96 and -8.20. If Jensen and Nakai's (1961) observation that $\delta^{34}S$ enrichment occurs when biogenic hydrogen sulphide is oxidized is correct, it would seem likely that fog water is the main sulphate vector. Though Carlisle (1978) upheld this concept, his rejection of Martin's (1963) hypothesis on the grounds that bottom sediments have a low $\delta^{34}S$ level is not justified. The hypothesis may be invalidated on several grounds but since it also involves fog moisture transport, different isotope levels in gypsum crusts and sea-bed deposits, which contain other sources of sulphur, are

insufficient grounds for rejection. Sofer (pers. comm.) determined $\delta^{34}S$ levels in six gypsum crust samples from the Central Namib Desert. Though the locations are not specific, the results show a tendency for crusts at the coast (within 1.0 km) to have low levels, approaching those of marine sediments. Sample descriptions are not detailed, though one coastal example is composed of "gypsum roses". Without stressing the point, it is tempting to suggest that those samples with low $\delta^{34}S$ levels may be the same as the crusts at Wlotzkas Baken (WB) which micropetrographic and chemical analyses indicate are akin to supratidal sabkha sediments. It would follow, therefore, that some coastal crusts are composed of marine sulphates while inland the gypsum's sulphate component is derived from oxidized biogenic sulphide.

If the model of gypsum genesis in fog moisture or on the land surface in the presence of water and calcium carbonate is valid, the need to account for coarse sediments upon the crusts remains. Sufficient water is not available in fog precipitation to facilitate chemical changes at the depth to which some crusts extend. Hence, though the information presented here goes some way toward ascertaining the source of the gypsum, the precise processes of crust genesis are as yet undetermined. In the following chapter, a model is formulated which is able to explain the development of most of the crust forms which cannot be attributed to either lacustrine, supratidal or continental sabkha evaporite sedimentation or to gypsum crystallization from groundwaters.

CHAPTER 7

THE ORIGINS OF GYPSUM CRUSTS, CONTINUED

The problems inherent in many of the aforementioned models of crust genesis necessitate the development of an alternative hypothesis. While detrital, aeolian and atmospheric deposition provide adequate explanations of the source of gypsum in different areas, they are frequently unable to account for subsurface crusts because the overlying sediments are of greater age. Translocation of surface gypsum deposits into the soil zone, where accretion and consolidation take place, must be accounted for. Though the models which have been proposed vary in some of their details, they all involve leaching of water-soluble materials and illuvial accumulation in lower soil horizons (Barshad, 1955). Here they are termed the *per descensum* models.

a. *Per descensum* models.

Several workers have advocated the subsurface accumulation of calcium carbonate, leached from dust deposits at the surface, to account for pedogenic calcretes. In north Africa, Agafonoff (1936 b.) and Conrad (1969) suggested that some weakly developed calcretes may originate in this way and Gigout (1958) held that many examples of red limon soils represented carbonate depleted horizons above the indurated zone. The model's strongest propounders have based their conclusions on evidence from North America. Bretz and Horberg (1949) felt that in New Mexico, caliche (calcrete) developed when soil moisture containing calcium carbonate dissolved at the surface was drawn upwards by capillary rise and evaporated. Brown (1956) advocated the same processes but rejected the need for capillary rise of moisture to account for caliche genesis in the Llano Estacado of Texas. The model was refined further by Arkley (1963, 1967), working in California where Ulrich *et al*. (1959) and Gile, Peterson and Grossman (1966)

showed that the main sources of carbonate were aeolian dust and cyclic salts. In gravelly soils, leached carbonate thinly coats pebbles (Ku et al., 1979) and gradually accretes until all interstices are plugged. At this stage, laminar horizons (croûtes zonaires) develop as soil moisture evaporates at the perched water tables that form. In finer sediments, crystallization on grains results in nodule formation and these gradually coalesce to form a subsurface horizon upon which laminar accretion occurs. This horizon is not completely impermeable since it is built up of calcrete prisms about 0.50 m in diameter. These appear to be akin to columnar structures associated with mesocrystalline subsurface gypsum crusts, though less well defined.

Few workers have advocated the per descensum model to explain gypsum crust genesis. Tricart and Cailleux (1960) held that per descensum was a far more feasible mechanism for desert crust accumulation than per ascensum from groundwater by capillary rise or recharge. Nevertheless, the only case studies invoking the model are those of Tolchel'nikov (1962) in Central Asia and Scholz (1963) in the Central Namib Desert. Both these workers held that deposition of soluble salts at the surface, in fog moisture or formed by alteration of surface materials, may be a contributary factor in gypsum crust development. The strongest advocate of the illuvial model has been Page (1972) who argued that it accounted for the majority of Tunisian gypsum crusts.

The precise mechanisms involved in the precipitation of water soluble salts in the soil zone may be explained in terms of the cellular structure of the interstices. Keen (1936) stated that "when rain falls, each cell in turn levies toll on the percolating water and the surplus percolates downwards; none of it will reach the water table unless it is surplus to the equilibrium capacity of the superincumbent soil." In the subsequent dry period, soil water evaporation proceeds from the surface downward and any salts dissolved in the moisture will crystallize around the particles of soil to which the moisture adheres owing to

surface tension. There is, in effect, absolute accumulation of cement within the soil zone (Yaalon, 1971). Crystallization will not occur below the maximum depth of wetting (Yaalon , 1961) and salts will only be removed if there is a net downward flow of water undersaturated with respect to them (Yankovitch, 1961). Though Arkley (1963, 1967) proposed a mathematical model to assess the depth of soluble salt accumulations (equation 6, chapter three), it is difficult to extrapolate this to other areas because soil moisture storage capacity is variable. The empirical verification of the model shows that in principle it is possible to assess the depth of the zone of crust accretion on the basis of the monthly balance of available precipitation (precipitation minus actual evapotranspiration) and the storage capacity of the soil. In areas where the moisture storage capacity is exceeded, salt accumulation will not occur (Page, 1972). Yaalon (1964) and Drever and Smith (1978), however, pointed out that different salts have variable relative mobilities,

$$\frac{\text{distance travelled by ion}}{\text{distance travelled by moisture}}$$

. The implications of this will be dealt with below. Suffice it to point out here that both soil and water chemistry will have a critical effect on ionic mobility.

A number of examples of superimposed calcrete and gypsum or gypsum and halite crusts have been described in the literature. Gypsum crusts developed upon calcretes at Oued Zigzaou (ZIG) and Oued Zerkine (ZER) in the Djeffara of southern Tunisia appear to constitute later deposits since the underlying calcretes are extensive features predating alluvial sediments which may also have surface gypsum crusts, as at Oued Zigzaou. In the Central Namib Desert, similar sequences described by Martin (1963), were also interpreted as two temporally distinct deposits. In contrast, the opposite vertical juxtaposition, calcretes above gypsum crusts, have been observed more frequently. Furthermore, depositional

discontinuities are often absent. Near Gabès (Le Houérou, 1960), between Sfax and La Skhirra in Tunisia (Page, 1972) and in the region of Ghardaïa and Ouargla in Algeria (Horta, 1979), the sequences were interpreted as indicators of climatic fluctuations during the Pleistocene. Such changes may have resulted in the deposition of different salts at the surface, depending on the salts available in coastal or continental sabkhas. It is unlikely, however, that the earlier gypsum crusts would be preserved during the hiatus since the mobilization of calcium carbonate indicates wetter conditions under which gypsum mobilization would be more pronounced. Beyond this, an exposure near Sfax (see plate 7.1 and fig. 7.1) shows not only conformity between the horizons but also coeval deformation structures, these are tepee structures with breached peaks at 5.0 m to 10 m intervals. They represent one of the few large scale indicators of displacive crystallization encountered in either field region. The terrestrial location precludes the possibility that they are lacustrine desiccation structures, which those at Chott el Rahim (CR) may be. Calcrete overlying gypsum has also been described by Boulaine (1961 a.) in Algeria. An horizon 0.80 m thick containing 40% calcium carbonate and 3.0% gypsum underlies 0.50 m of unconsolidated silty sand and rests upon 1.2 m of material containing 30% gypsum and 10% calcium carbonate. Scholz (1971, 1972) reported similar sequences in the Central Namib Desert near Gobabeb.

Within 25 km to 30 km of the coast of the Central Namib Desert, gypsum crusts at or near the surface are underlain by halite crusts as at Rooibank (PF) (see figs. 5.4A and B and 5.9B). While Gevers and Westhuyzen (1932) felt that the sequences represented late Tertiary or early Quaternary sabkha or lagoon deposits, their extent and topographic location preclude this. It is likely that they are pedogenic features characteristic of dry areas where gypsum and sodium chloride are deposited at the surface. Further inland, sodium chloride is either less plentiful in fog water (Boss, 1941) or increased rainfall leaches it through the soil zone, here calcretes develop above gypsum crusts. Hunt et al.

Plate 7.1

Calcrete overlying gypsum horizon between Sfax and La Skhirra in southern Tunisia. Fig. 7.1 overleaf provides a schematic representation of a similar section in the same area. On the photograph tepee structures, developed in the platy calcrete, are evident on the left and above the field bag. The lighter coloured, less blocky material at the base of the section is gypsum.



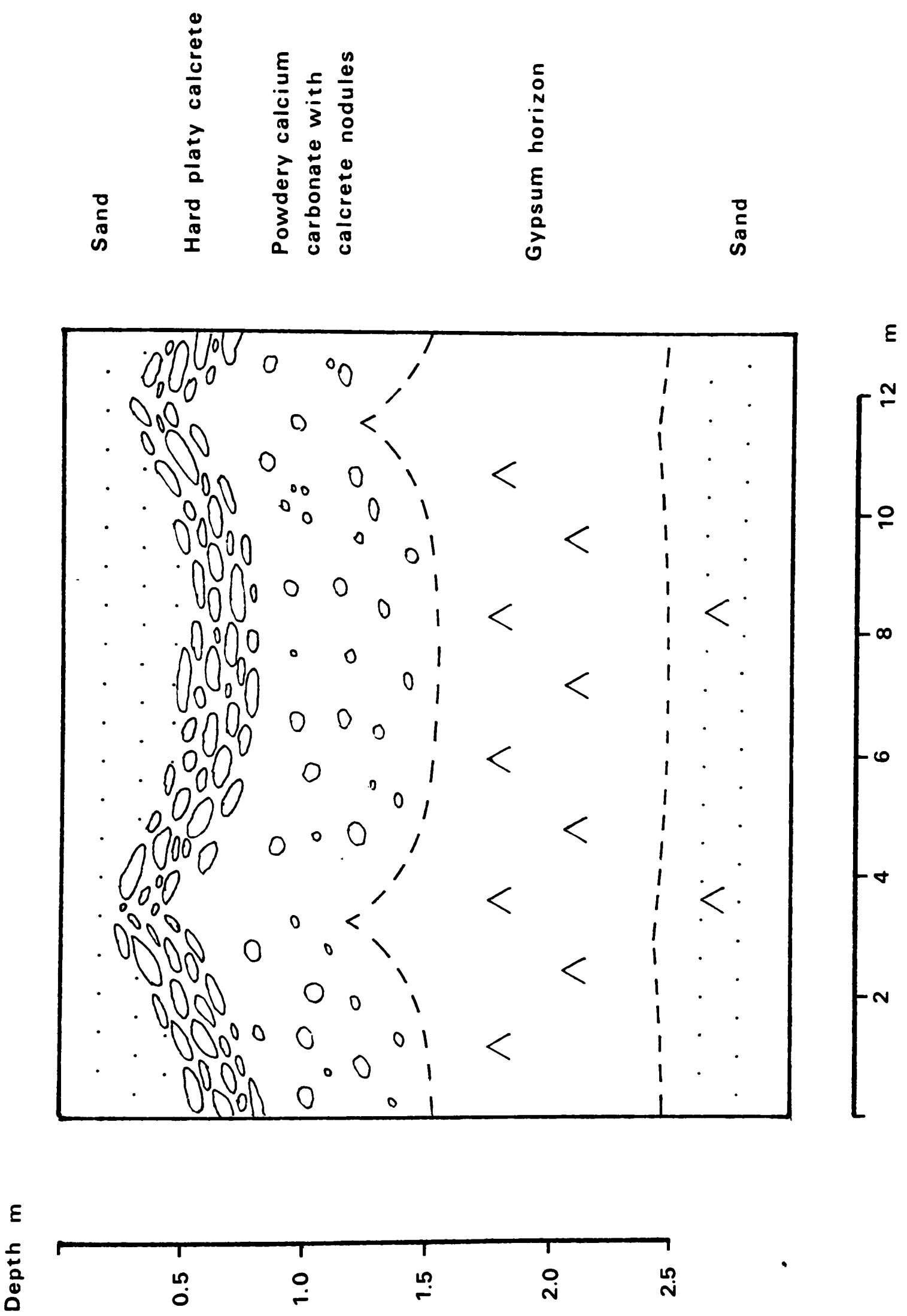


Figure 7.1

Schematic representation of a two-tiered soil crust, calcrete above gypsum. The upper 1.50 m to 2.0 m and the horizontal characteristics are based upon the section shown in plate 7.1. The lower part is based upon a section at Bir Ali, west of Sfax (after Page, 1972).

(1966) described similar sequences in Death Valley, California but interpreted them as spring deposits. In the southern Namib Desert near Lüderitz, Kaiser and Neumaier (1932) recorded examples of surface calcretes underlain by gypsum crusts which in turn rested upon halite accumulations. They held that the soluble salts were precipitated from groundwater by capillary rise. This is unlikely not only because capillary rise is ineffective over distances of more than a few centimetres but also because where both calcium carbonate and gypsum are found in crusts developed at groundwater table, the former underlies the latter (Kulke, 1974). This may be because carbon dioxide liberated from plant root systems increases carbonate solubility near the surface by lowering pH. Beyond this, saline groundwater, rising by any mechanism, will precipitate the least soluble salts first. Whether crystallization results from evaporation or the solution of additional salts in the soil zone, it is very unlikely that gypsum will become saturated before calcium carbonate. Progressive evaporation from a falling groundwater table could produce the sequence described by Kaiser and Neumaier (1932) but this assumes that there is unrestricted ionic migration through the aquifer. This is not the case in fine sediments where the connections between interstitial cells are narrow. Differential leaching of salts derived from the surface is a much more viable explanation.

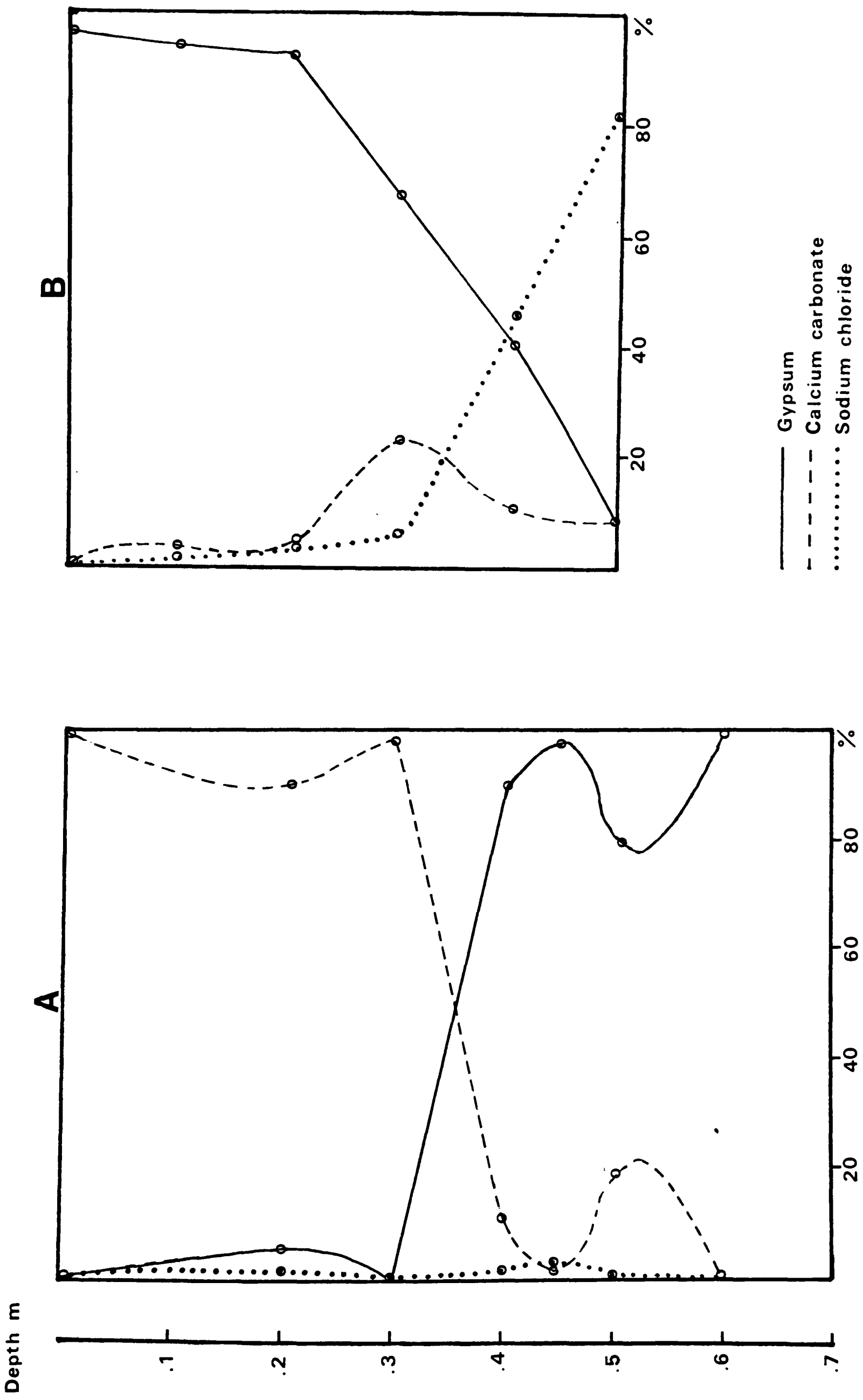
Two examples of two-decked soil crusts were examined in detail. These are near Gobabeb (GO) where calcretes overlie gypsum crusts (see fig. 5.7C) and near Rooibank (PF) where gypsum crusts overlie halite accumulations (see fig. 5.9B). The chemical data are presented in tables 5.21 and 5.29. The variation with depth in the amounts of calcium carbonate, gypsum and sodium chloride as percentages of the total amount of all three are shown in figures 7.2A and B. The total amount of these salts is between 56% and 72% in the Gobabeb section and 38% and 87% in that form near Rooibank. At Gobabeb, the low sodium chloride content throughout results in the mirroring of the gypsum and calcium carbonate values about the 50% line. There is a sharp cross-over point at about 0.35 m

Figure 7.2 (overleaf)

Two-tiered soil crusts from the Central Namib Desert.

A. Variation in salt content with depth through a calcrete-gypsum crust near Gobabeb (GO) (see fig. 5.7C). Gypsum, calcium carbonate and sodium chloride contents are plotted as percentages of the total acid soluble portion of the samples.
Samples were boiled in 1.0 N hydrochloric acid for 30 minutes.

B. Variation in salt content with depth through a gypsum-halite crust near Rooibank (PF) (see fig. 5.9B).


Depth m
.1
.2
.3
.4
.5
.6
.7
A
B
20
40
60
80
%
Gypsum
Calcium carbonate
Sodium chloride

where gypsum becomes more abundant down the section. At Rooibank the presence of all three minerals complicates the picture. Calcium carbonate is not abundant except for one sample (PF 1.4), in which it makes up 29.1% of the total salts. Again, however, a cross-over point is evident at about 0.40 m where halite becomes more abundant than gypsum down the section.

While the leaching model described above explains how soluble salts will accumulate in the soil zone, the sharp contrasts between horizons of different salts necessitate its modification. Water containing soluble salts dissolved at the surface will remain essentially stationary within the soil zone at the loci where it replaces the soil moisture deficit created during the antecedent dry conditions. Any single volume of water will contain equal proportions and amounts of salts as any other similar volume anywhere in the profile, assuming the antecedent soil chemistry is uniform. Therefore when evaporation occurs there will not be differential accumulation of salts. Experimental leaching of natural soils with solutions of sodium sulphate and sodium chloride results in reactions between the sulphates and calcium as chloride in the soil adsorption complex (Krupkin, 1963). Gypsum accumulates in the upper horizons, while free chloride and sodium are leached into the lower zone crystallizing as halite or moving out of the system. In effect, if there are undifferentiated salt accumulations initially, further increments of solute may result in bonding of certain ions in the leachate to resident salts as it moves down the profile. This will result in differential salt accumulations since the relative mobility of the ions varies. Yaalon (1964 a.) defined relative mobility as the distance travelled by an ion divided by the total distance travelled by moisture introduced at the surface, that is the depth of the wetting front beneath the land surface. He proposed that the depth of accumulation (D) of any salt could be determined as follows:

$$D = \frac{Pef.\ Rm}{Vf} \quad \ldots\ \ldots\ \ldots\ \ldots\ \ldots\ \ldots\ \ldots\ \ldots\ \ldots\ 7$$

(Pef) is the amount of effective precipitation after actual evapotranspiration, (Rm) is the relative mobility of the ion and (Vf) is a measure of soil texture which takes the soil moisture storage capacity into account. Fine sediments should exhibit more diffuse accumulations since flow paths are more variable. Yaalon does not, however, point out that coarse sediments will experience greater depths of leaching owing to their permeability and that they have a greater potential for ionic migration through resident moisture.

Drever and Smith (1978) found that leached soil columns liberated ions in the following order from most mobile to least:
Cl^-; Na^+; SO_4^{--}; K^+; HCO_3^-; Ca^{++}; Mg^{++} and SiO_2. This suggests that while bonding with the soil adsorption complex may be significant, the solubility of the mineral species is also critical. As would be expected, sodium chloride will dissolve most rapidly and in most abundance relative to the total amount present. This will be followed by gypsum and magnesium and calcium carbonates and finally silica. On the basis of this evidence it is possible to propose an alternative mechanism for differential salt accumulation in the soil zone. If at some time (t0) the soil zone does not contain any soluble salts but these are present at the surface, the first wetting phase (t1) will leach the salts into the soil where they will crystallize in amounts proportional to their solubility, assuming saturation is achieved with respect to each. At t2, water moving down the section will be more likely to become saturated with the salts because it will be in contact with the previous diffuse precipitates as it percolates. During this wetting phase, the more soluble salts will be leached from the upper zones if the water is undersaturated with them. Furthermore, dissolution of gypsum in the upper zone will result in calcium carbonate precipitation, if it is saturated, owing to the common ion effect (Wigley, 1973). Hence a calcium carbonate accumulation will develop though small amounts will move still further down the section where it crystallizes along with admixed gypsum and sodium chloride. At t3, water moving into the lower soil zone will dissolve

sodium chloride even when it is saturated with gypsum because halite is more soluble and it will move further down the section. Some halite will be present in the water replacing soil moisture in the upper horizons but this will always be dissolved and leached in subsequent wetting phases. At t4, water moving through the gypsum-rich horizons will dissolve increasing amounts of sodium chloride until levels reach about 150 g to 200 g NaCl/litre of water and gypsum solubility begins to decrease (Madgin and Swales, 1956; Zen, 1965) causing its precipitation. Predominantly sodium chloride will crystallize at greater depths, if it is not removed from the soil zone.

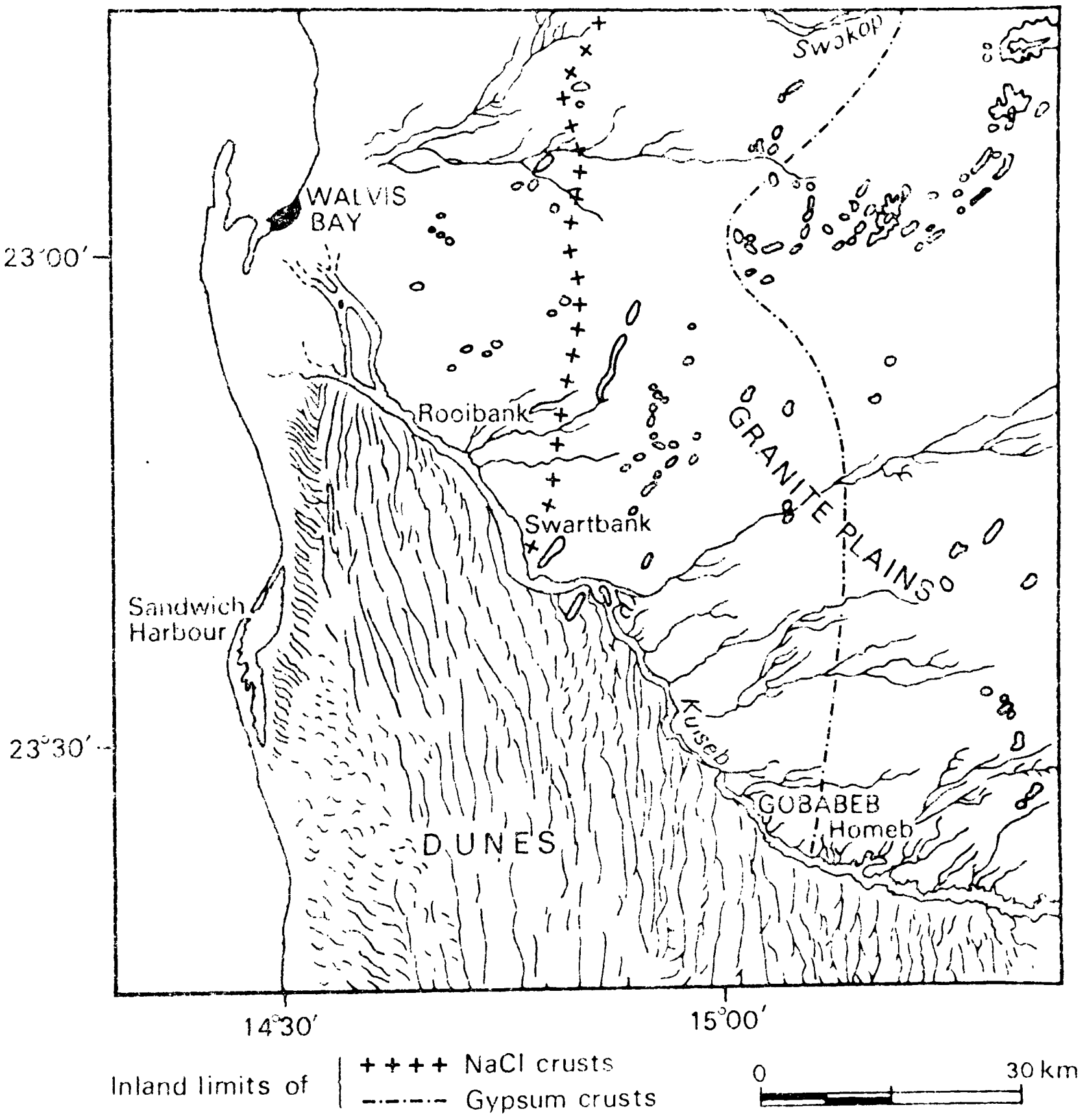
Though the number of wetting and drying cycles will be numerous before the differentiation of horizons is complete, the mechanism based on solubility of the salts appears viable. It is difficult to account for the absence of some salts in certain horizons, for example calcium carbonate in the gypsum zone and gypsum in the halite level. This may result from preferential bonding of calcium ions in the presence of abundant sulphate at the expense of carbonate in the former case and the formation of calcium chloride or sodium sulphate in the latter. These would be leached beyond the halite horizon. It is possible that in some cases, surface calcretes represent bacterially altered gypsum (Lattman and Laufenburger, 1974) but the absence or rarity of pseudomorphic structures after gypsum suggest that this is not a significant factor. The sharpness of the transition from one mineral horizon to another from nearly 100% calcium carbonate to 100% gypsum over 0.15 m near Gobabeb and from nearly 100% gypsum to nearly 90% halite over 0.30 m near Rooibank, is also enigmatic. It is possible that ionic bonding preferences vary markedly over small distances depending on the nature of the dominant salt. Alternatively, since the maximum depth of wetting will vary with each cycle depending on the amount of precipitation, the different mineral horizons may accrete toward each other. This would seem less feasible since host sediment would be sandwiched between the layers.

Two-tiered soil crusts are restricted to areas where both minerals are available at the surface from where they are leached and where climatic conditions are suitable for the mobilization and crystallization of both salts. Hence, in the Central Namib Desert, halite crusts are found only in the driest areas which are commensurate with the zone of sodium chloride deposition from fog (see fig. 7.3). In the Central Namib Desert and in southern Tunisia, calcretes overlie gypsum crusts in the areas where the maximum rainfall zone for gypsum preservation and the minimum for calcium carbonate mobilization overlap. In the Namib, this is the 25 mm to 50 mm rainfall belt and in southern Tunisia it is about 150 mm to 250 mm, the difference being attributable to the temperature régimes (see chapter three). In drier areas, there is insufficient available moisture to mobilize enough calcium carbonate to form a calcrete. If calcium carbonate is present at the surface, a pH of greater than 7.8 will greatly reduce its mobility since it is virtually insoluble. This may also be a factor in limiting the downward movement of calcium carbonate in two-decked soil crusts.

In most of southern Tunisia and the 50 km wide coastal strip of the Central Namib, surface carbonate dust is mobilized by the wind while gypsum dust and salt grains are dissolved by rainwater and fog moisture and move into the soil zone. The process can explain the development of most of the mesocrystalline subsurface crusts and, indirectly, many of the surface forms as well.

Micropetrographic examination of subsurface mesocrystalline gypsum crusts shows them to be composed of well developed lozenge-shaped crystals which rarely exhibit solutional features except near the surface. This suggests that the crusts are aggrading or at least static features. Some examples may represent surface alabastrine crusts which have been buried beneath aeolian deposits, such as that at La Skhirra (LS), but in most cases the crusts may be attributed to subsurface accretion. Page (1972) suggested that the source material for the crusts was the same as that proposed by Coque (1962, pp. 93 - 101), namely chott

Figure 7.3



Limits to the extent of different soil crusts in the Central Namib Desert. Two-tiered crusts occur predominantly in the vicinities of the inland limit of the underlying crust type.

deflation deposits. Rather than being consolidated in situ at the surface, Page (1972) felt that they were leached into the soil zone by rainwater and crystallized when it evaporated. While this observation is relevant, it must also be pointed out that many subsurface mesocrystalline crusts are found beneath surface crusts, either columnar, powdery or the intermediate cobble form (MET, L, OL, K, CC, SB). These deposits are usually composed of alabastrine gypsum and, when present, porphyroblasts of lozenge-shaped gypsum crystals exhibit solutional pitting. It is possible, therefore, that the subsurface crusts are the product of crystallization of gypsum leached from the overlying surface crusts. In support of this model, it is likely that surface cobbles represent the remnants of columnar surface gypsum crusts which have been destroyed by leaching. Solution would be most pronounced along the column walls which border the paths of downward water movement, the interstices between columns. Eventually, only the column cores from the indurated upper portion of the crust will remain. The cobbles have hard outer shells because continual solution and recrystallization result in the formation of a rind of well-cemented alabastrine gypsum.

Further, it seems likely that many powdery surface accumulations above subsurface crusts are deposits residual after this leaching process. It should be pointed out that some powdery gypsum accumulations, especially those with high gypsum contents, may represent aeolian deposits. The chemistry of these different crust forms lends further support to the model. The major differences between subsurface mesocrystalline crusts and columnar surface crusts are the higher levels of all the constituents, except gypsum, in the former. The variation may be accounted for in two ways. First, displacement of the host sediment is less complete in the subsurface crusts than the surface forms therefore elements associated with insoluble residue are more abundant.

Secondly, the surface crust has experienced a depletion in soluble salts through leaching while the underlying crusts are accumulation zones. The strontium levels are interesting since, being lower in the surface crusts than the subsurface, they do not parallel the variations in gypsum content. Differential leaching

probably explains this, enrichment occurring as a result of ionic migration through soil moisture to growing gypsum crystal faces. Powdery gypsum deposits not only contain more insoluble residue but are also depleted in sodium salts and strontium owing to leaching.

To take the model a stage further, it is likely that many surface crusts represent exposed subsurface mesocrystalline accretions. This is particularly significant to those surface crusts with coarse sediments which cannot post-date crust development upon them. In these cases, the gypsum has accumulated in the lower horizons after being leached past the materials in the upper soil zone. The chemical analyses of four profiles presented in figures 5.10 and 5.11 lend support to this view. While strontium levels are closely related to gypsum content, it is evident from the profiles at Metouia and Oudref particularly, that enrichment has occurred in the subsurface crusts. The same is also true of magnesium. These subsurface crusts have gypsum contents similar to the columnar surface crusts which overlie them, indicating that they may be of similar origin. While the surface crusts show micromorphological evidence of active solution of gypsum crystals, the subsurface crusts have well developed lenticular crystals which generally have sharply defined rim points suggesting little solutional activity and probably recent crystallization. Differential leaching of more soluble or more loosely bonded strontium and magnesium salts or free ions relative to gypsum may result in this subsurface enrichment once recrystallization occurs at the maximum depth of gypsum leaching. Subsequent deflation or erosion of finer surface materials results in the exposure of the gypsum crust mantled only by lag deposits. Under the influence of meteoric waters, textural and chemical modifications will occur owing to dissolution, recrystallization and some leaching. Some surface crusts may be aeolian deposits consolidated *in situ* through solution and reprecipitation by meteoric waters, as Coque (1962, pp. 93 - 101) suggested. This model is appropriate for crusts such as those at Djebel ed Dissa (DI) which lie directly upon Turonian limestone bedrock

(see fig. 5.8E) though Page (1972) held that these particular examples represented in situ alteration of limestone to gypsum. While the high gypsum content of some surface crusts may be explained in terms of being aeolian deposits, many subsurface crusts contain similarly large proportions of gypsum. Some contain as much as 90%. This great purity is the main reason for questioning the validity of the leaching model (Boulaine, 1958). It would be inappropriate to advocate the model without specifying the mechanisms by which the host sediment in the soil zone is displaced physically or replaced chemically.

b. Mechanisms of displacive and replacive accretion.

Several mechanisms may be involved in the displacement of pre-existing sediments during the formation of a gypsum crust. Clearly the filling of voids in the host sediment cannot account for the high gypsum and low insoluble residue contents. The amounts of insoluble residue expected ($C_{ir}\%$) if void filling alone is involved is:

$$C_{ir}\% = \frac{C_d \; S_{ir}\%}{S_d} \quad \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots 8$$

where (C_d) is the bulk density of the crust, (S_{ir}) the insoluble residue in the host material and (S_d) its bulk density (Goudie, 1973). In the case of the gypsum crust on dune sand at Djemma (DJ) in the Nefzaoua, the difference between the bulk density of the crust and the host material would have to be of a factor of about three to explain the differences in insoluble residue. In reality, the porosity of the crust and sand are about the same, 35% and 40%. Hence the sand will not have a bulk density of more than about 1.2 times that of the crust, the difference between the specific gravity of gypsum and quartz. The expected insoluble residue content of the crust will not be below 30% while the observed content is less than 15%. In other words, per unit volume the crust has about half of the insoluble residue expected, the remainder having been

removed chemically or physically.

Macrostructural evidence of physical displacement will take the form of disruption at the landsurface, assuming that volumetric expansion will be horizontal and vertical, in the upward direction. Compaction of underlying sediments will be limited relative to the weight of the overburden. Since 1.0 m of dry quartz sand will exert a pressure of about 16 kN/m^2 plus atmospheric pressure of about 101 kN/m^2, this may be readily displaced upwards by the crystallization force exerted by precipitating salts. These pressures may reach 4,000 kN/m^2 (Goudie, Cooke and Evans, 1970) during unidirectional crystal growth. As pressure counter to the crystallization force increases, ionic diffusion to the growth sites must be rapid enough to counteract the tendency to pressure solution (Phillips, 1974; Watts, 1979). Pressure increases resulting from volumetric expansion of hydrating salts may be even greater than crystallization forces (Mortensen, 1933; Cooke and Smalley, 1968; Evans, 1970) but mineralogical evidence of its occurrence in gypsum crusts is rare. Hunt et al. (1966) reported hemihydrate and anhydrite associated with columnar crusts in Death Valley. Tepee structures and pseudoanticlines are often the products of volumetric expansion and some testify to horizontal displacement of 25% to 30% (Watts, 1979). These features have been described in chapter five between Medenine and Zarzis (MZ) and at Chott el Rahim (CR) and above in the vicinity of Sfax. Domes developed in crusts probably of lacustrine origin as at Sabkha el Melah (SM), near Narabeb (NA) and near the Hamilton Range (HA 3), are structurally different. While they may be the result of small volumetric changes resulting from dehydration and rehydration (Breisch, 1978), there is no chemical or micropetrographic evidence of hemihydrate or anhydrite to support this. The blisters may be produced by small volumetric increases in surface deposits resulting from secondary gypsum crystallization or crystal growth. Coque (1962,p. 81) described gypsum nebkhas in the Nefzaoua of southern Tunisia. While he held that they were consolidated gypsum dunes, their form is similar to the domes described from sabkha environ-

ments and they may also be the products of displacive crust development.

The majority of mesocrystalline subsurface gypsum crusts and microcrystalline surface forms do not exhibit macromorphological displacive structures of these types. It is possible, however, that columnar structures attest to displacive accretion. Nodular accretion may result in the formation of a prismatic structure through vertical extension of closely packed nodules. Nodules are not a common feature of gypsum crusts in the two field areas. Pouget (1968) held that desert croûtes de nappe developed through coalescence of gypsum nodules and some small examples are found in the subsurface crust at Oudref (OU 2.1, see chapter five). Nodular accretion by host sediment displacement has been described by Wieder and Yaalon (1974) but for this to result in the formation of a regular columnar structure the nodules would have to be evenly spaced. Micropetrographic analysis of column walls shows evidence of accumulations of non-gypsiferous materials. However, these do not necessarily represent displaced host sediments. Both calcium carbonate and clastic material may be introduced from above dissolved and suspended in meteoric water moving down the spaces between columns. The presence of these fissures suggests that columniation is a product of volume reduction rather than expansion. Dehydration to hemihydrate resulting in a 5.0% volumetric decrease (Chatterji and Jeffery, 1963; Hunt et al., 1966) may be involved but there is no mineralogical evidence for this. Contraction of the crust may result from desiccation (Tucker, 1978). The lack of cohesion between gypsum crystals making up freshly exposed crust samples suggests that films of moisture cover the crystal faces. Induration, following drying of the crust, results from the precipitation of dissolved gypsum which then cements the crystals together. The volume reduction consequent on the evaporation of interstitial moisture will be localized along planes perpendicular to the direction of tension by partial cohesion of the crystals. Tensional stresses will be greater in the horizontal plane since the weight of the overburden will negate those in the vertical. Fractures will run vertically and

tend toward hexagonal nets assuming tension is equal in all directions. Progressive evaporation from the surface downwards will deepen the polygonal fissures which will be maintained by subsequent moisture seepage. Large-scale polygons developed on powdery gypsum crusts are structurally akin to those found on halite salt crusts in that the ratio of fissure depth to polygon diameter is about 1:10 (Lang, 1943; Willden and Mabey, 1961; Christiansen, 1963; Neal and Motts, 1967). The larger surface diameters of the polygons compared with those of the columns may be related to the sub-aerial origin of the former features. While gypsum, calcium carbonate and clay content affect the potential for contraction and cohesion, the reason for the different dimensions is an enigma.

Micromorphological features indicative of displacive crystallization are more abundant than large scale structures. The most common are floating grain textures. Since most mesocrystalline subsurface and microcrystalline surface gypsum crusts have less than 20% insoluble residue, often less than 10%, grain supported textures are rare. The maximum volumetric porosity of a grain supported mass of uniform spheres is 47.5% when each sphere is in contact with six others. The minimum porosity is obtained with an hexagonal close-packed structure and face-centred cubic structure. In both cases, volumetric porosity is 26% (Kittel, 1971, p. 28). Assuming the host material is pure quartz sand composed of spherical grains of uniform size, the percentage weight of insoluble residue in a crust with a quartz grain supported texture and complete gypsum void filling will be between 56.01% and 76.63%. Using these volumetric and gravimetric constants in equation 8, it is possible to determine the feasible levels of insoluble residue in a gypsum crust developed through non-displacive cementation of a sand composed of 50% by weight quartz and 50% gypsum. While it is incorrect to assume that such a sand occurring in nature would be composed of spherical grains of uniform diameter (see chapter six), it is useful to compute these theoretical constants. A loose packed grain supported texture would contain 24.39% by weight

insoluble residue and a close packed structure 34.38%. With 40% quartz and 60% gypsum making up the grain mass, the respective figures would be 19.24% and 27.12%. Hence non-displacive cementation of even the most gypsiferous sands encountered in southern Tunisia cannot account for the low insoluble residue contents of the majority of the crusts. It is possible that the crusts at Zaafrane (ZAA) and El Faouar (EF) in the Nefzaoua represent consolidated gypsiferous sand since the sands contain 40% to 50% insoluble residue and the associated crusts 20% to 30%.

Two mechanisms which account for displacive crystallization and floating grain textures may be postulated. First, Kastner (1970) showed that under conditions of slow gypsum crystal growth in sands saturated with gypsum-rich aqueous solutions, quartz grains are pushed ahead of the advancing crystal face. Acceleration of the crystal growth-rate eventually incorporates grains at the crystal face poikilitically. Cody (1976) pointed out that monovalent salts will alter the surface energies at different growth faces thereby facilitating inclusion of host sediment (see chapter one). While poikilitically included grains may float within the solid crystal, other masses of grains will be displaced but will nevertheless form zones of grain supported textures. The process is appropriate to the development of crusts composed of large lenticular crystals and possibly also to porphyroblasts in fine groundmasses (Misch, 1971).

Another mechanism is required to explain floating textures in gypsum crusts composed of small, lozenge-shaped crystals and alabastrine material. In these cases, the gypsum crystals and grains are too small to include host material but Watts (1979) pointed out that the loci of crystallization may account for displacement. Gile, Peterson and Grossman (1966) observed that incipient calcretes exhibited calcium carbonate crystallization around fine quartz grains and not as void filling. They held that the grain coatings were uniform on the particles and therefore displacive crystallization must also occur at grain contacts. Experimental simulation of crust development by crystallization

from gypsum saturated groundwater in sand columns (Plet-Lajoux, Monnier and Pedro, 1971) showed that the growth of small crystals occurs at grain contacts. If it is assumed that as a result of surface tension on the host grains solvent will adhere to them more effectively then simply filling voids, with the proviso that air is available to replace any reduction in the volume of solvent, evaporation will produce grain coatings. It follows that since surface tension will be greatest at grain contacts, causing weak adhesion in mois sands, evaporation will result in migration of solvent to these points. In effect, most of the available solute will be deposited near grain contacts. Providing surface tension at the crystal/grain interface is greater than the surface tension of the salt solution plus that at the grain/solution interface, a film of solvent is maintained (Evans, 1970). Crystallization, with molecular arrangements possibly aided by surface tension, will occur at the crystal/ grain interface causing physical displacement of the grains (Phillips, 1974). Providing the stresses resulting from this displacement are dissipated by lateral and upward vertical heave, subsequent saturation with salt-rich solutions may again result in displacive crystallization. It is also probable that even with a greatly reduced quartz grain population, crystallization will occur between existing salt crystals rather than at previous growth sites if their surface molecular arrangements have been altered and ionic attractive potential reduced. This is evident around aggregated clastic inclusions in some gypsum crusts where displacive crystallization occurs between the clasts and older gypsum crystals (see plates 5.63 and 5.64). Intermittent crystallization may be an important factor in the maintenance of porosity during cementation. If Plet-Lajoux, Monnier and Pedro's (1971) observation that evaporation rates decrease with continuous crystallization is indicative of reduction in porosity, recrystallization at different growth sites may allow for the preservation of voids. Watts (1979) invoked intermittent crystallization to account for displacive micrite in calcretes. While some surface gypsum crusts may represent consolidated aeolian or atmospheric deposits, it is also

possible that displacive crystallization occurs locally in cobbles for example, while overall the surface horizon is degrading through solution. This may explain the higher gypsum contents of surface crusts and cobbles than subsurface forms.

Further evidence for displacive crystallization is provided by brecciation structures in bedrock beneath gypsum crusts and in clastic material. Large-scale bedrock brecciation is encountered at two sites in southern Tunisia. At Oued Magroum, near El Hamma de Gabès (EH) (see fig. 5.8C), strata of Wealdian sandstone show signs of disruption near the base of the surface gypsum crust. At Djebel el Dissa (DI), Turonian limestones immediately underlying the surface crust are brecciated (see fig. 5.8E). Micropetrographic investigation of the breccia shows that the disrupting material is granular gypsum (see plate 5.51). It is possible that the gypsum has been precipitated in a previously fragmented horizon at the surface of the bedrock. The crystal textures do not in themselves provide conclusive evidence of displacive crystallization. The presence of floating fragments of micritic limestone and detached rinds along the sides of gypsum filled veins are indicative of displacive growth (Phillips, 1974), though the essentially two-dimensional nature of a thin section may give a false impression since the isolated areas of micrite may have been attached to the main mass above or below the plane of the slide. The coarse, granular texture indicates crystal growth over a long period in a saturated aqueous medium. This would provide the potential for displacive crystallization. Calcretes from the Central Namib Desert contain brecciated clastic material identical to that described by Watts (1979) in calcretes from north-west Scotland. Brecciation is exhibited best by fissile minerals such as book-leaf micas. Displacive crystallization in the cleavage planes produces an exploded, concertina-like structure (Watts, 1979) (see plate 5.68). Since the disrupted particles may be visually conjoined it is unlikely that thin section orientation has resulted in an erroneous impression. Clasts of less easily

cleaved minerals may also exhibit brecciation. Fragmented quartz grains with interstitial micrite show slightly variable extinction between the pieces suggestive of displacement. Rothrock (1925) showed that calcite crystallization was powerful enough to fracture quartz grains but other mechanisms may be involved. Brecciation of clasts in gypsum crusts is less common. Plate 5.41 shows a large quartz grain in a thin section of the crust at El Faouar (EF). While the detached plates and conchoidal fractures are indicative of weathering by salt crystallization (Goudie, Cooke and Doornkamp, 1979), this may have been achieved by sodium salts which were later replaced by the interstitial gypsum. It would be expected, however, that such a process would result in a greater disruption in the slightly variable extinction among the different components of the quartz grain.

The most striking evidence for displacive crystallization associated with gypsum crusts is provided by fibrous crystal textures. In chapters one and five the mechanisms for and indicators of displacive fibrous crystallization have been outlined. Fibrous gypsum textures are usually associated with bedrock where fissures created by hydraulic fracture (Shearman et al., 1972) or tensile strain (Phillips, 1974) are filled with fibrous aggregates. Matsuura (1925), Bundy (1956) and Holliday (1970) held that some displacive crystallization may occur but that this was related to the high pressure conditions associated with anhydrite hydration. While curved fibres, exhibiting undulose extinction, indicate tensile strain crystallization (Shearman et al., 1972; Phillips, 1974), bent crystals, with differential extinction along the fibre, testify to disruption subsequent to crystallization (Shearman et al., 1972). Furthermore, Phillips (1974) held that most clastic inclusions were at the centre of the vein suggesting syntaxial growth along the central line. This would argue against displacive crystallization. The observations of Aljubouri (1971) and Watts (1979) that the walls on either side of the vein fill fit together, are not strictly valid as indicators of displacive crystallization since any fracture

would exhibit such features. Watts lists three other points which support fracturing by crystallization. Passive growth would normally result in growth subperpendicular to the nucleation surface not parallel to the direction of crack widening, the cracks often do not exhibit tapered ends and square ends suggest displacive shear planes and, thirdly, wall dislocations suggest anti-taxial growth. While the calcite fibres described by Watts (1979) were curved which according to Phillips (1974) argues in favour of tensile strain crystal-lization, Watts felt that evidence of antitaxial growth, outward from the cracks, was sufficient to suggest displacive crystallization. Though he held that tensile strain would be insignificant in near-surface calcretes, the columnar structures associated with gypsum crust may testify to its occurrence. The argument rests on whether it is valid to assert antitaxial growth as indicative of displacement or curved fibres as proof of tensile strain.

Satinspar from Turonian bedrock in southern Tunisia (DT 3.1 and DT 3.2) has curved fibres and vein-centred clastic inclusions. It is therefore interpreted as non-displacive. In contrast, gypsum fibres from the surface crust near Rooibank (PF 2.2) and subsurface crusts from near Gobabeb (GO 4.4) and from Lagunenberg (CC 1.3) are generally straight and, even when slightly curved, exhibit uniform extinction. As in the case of fibrous calcite described by Watts (1979), the gypsum fibres are orientated perpendicular to the crack walls, parallel to the direction of crack expansion. The ends of the lens-shaped cracks are squared and inclusions are near the walls suggesting antitaxial growth. These features all suggest displacive crystallization. Still more convincing are large scale fibrous horizons developed on top of the near-surface gypsum crust at Rooibank (PF) (see figs. 5.4A and B and plate 5.34). This horizon, 0.10 m to 0.20 m thick, is developed between 0.10 m to 0.20 m of lag gravel at the surface and the underlying microcrystalline to mesocrystalline gypsum crust. The upper surface of the fibrous layer is irregular with small pinnacles of aggregated fibres developed beneath larger pebbles and pits filled with

finer sand. In this aspect the material resembles needle-ice. The stability of the land surface, protected by the desert pavement, and the freshness of the fibrous crystals argues in favour of recent crystallization at a shallow depth. Gevers and Westhuyzen's (1931) suggestion that these are ancient sabkha or lagoon facies is untenable on topographic grounds as well as the unweathered nature of the satinspar. Furthermore, Stapff (1887) reported fibrous gypsum in cemented dune sands more than 80 km inland and over 600 m above sea level.

The mechanisms of fibrous crystal growth are complex. Displacive crystallization can only occur from supersaturated solutions with free access of solute ions to the growth sites. An ephemeral perched water table may develop above the gypsum crust after rainfall. If the water has dissolved both sodium chloride and gypsum dust on the surface, it will have the potential to dissolve additional gypsum in the upper soil zone. Supersaturation with respect to gypsum will occur rapidly and will be sustained during the ensuing period of evaporation. Crystallization will occur at grain contacts and particularly on the undersides of grains where gravitational forces will pull the moisture and prolong crystallization. The development of gypsum beards beneath pebbles attest to this (Pias, 1972; Petrov, 1976). Intense evaporation with rapid crystal growth from an abundant and protracted flow of supersaturated liquid to the growth sites may produce the fibrous habit. Once it has developed, capillary movement along the intercrystalline boundaries may enhance the supply of ions in solution. Evidence of topotactic alteration of some satinspar to hemihydrate, not only testifies to intense evaporation but may also provide an additional source of ions in the form of hemihydrate dissolved in its former recondensed water of crystallization. Hemihydrate is not plentiful hence this is not considered a major contributing factor in the formation of near-surface fibrous gypsum textures. Fibrous textures are also developed in the halite crust underlying the gypsum horizons therefore such geochemical considerations are not a prerequisite of the fibrous habit.

It is felt that displacive crystallization at clastic grain contacts and in inter-crystalline spaces can account for the subsurface development of gypsum crust with exceptionally high purities. Most displacement will eventually tend toward the vertical since horizontal retaining pressures are greatest. Fibrous textures attest to highly efficient displacive crystallization and is certainly a factor in several examples of gypsum crusts in the Central Namib Desert. Owing to the solubility of gypsum it is unlikely that such textures will be long-lived. It is probable that dissolution and rapid recrystallization in the upper soil zone eventually produces a predominantly alabastrine texture. Hence in some cases fibrous textures may disappear, along with mesocrystalline lozenges and porphyroblasts, when a subsurface crust becomes exposed at the surface by deflation of the overlying sediments.

An alternative explanation for the purity of many gypsum crusts of subsurface, pedogenic origin is that the host sediment is replaced chemically by direct alteration or dissolution and crystallization of gypsum in its stead. The first of these has been advocated by Page (1972) to account for the surface crusts on Turonian limestone at Djebel ed Dissa (DI). Micropetrographic analysis of the basal breccia at this site shows no evidence of alteration of limestone to gypsum. It would appear therefore that the crust is either a consolidated aeolian deposit or a pedogenic feature which has had the overburden removed by slope wash or deflation. The presence of a mantle of powdery material on the crust on the lower slopes of the Djebel supports this contention. Other examples of <u>in situ</u> alteration of calcium carbonate to gypsum have been described in chapters two and six. Martin (1963), working in the Central Namib Desert, Passarge and Meinardus (1933) in Egypt and Rozanov (1961) in Central Asia, advocated subaerial replacive accretion of crusts and Krupkin (1963) showed experimentally that it occurred in calcium carbonate bearing soils leached with sodium salt solutions. The critical factor in the applicability of the models to subsurface crust formation by replacement is the presence of equal

amounts of calcium carbonate, in the host sediment, and gypsum, in the crust produced. Since most of the sediments above and below the gypsum crusts of southern Tunisia and the Central Namib Desert do not contain large amounts of calcium carbonate the models are inapplicable. The same applies to the authigenic development of gypsum from iron sulphides (Broughton, 1934; Vialov, 1946; Jessup, 1960 b.; Sass, Nathan and Nissenbaum, 1965; Buurman and Jongmans, 1975). The argument that all of the pre-existing material has been replaced is untenable since its mode of emplacement in such discrete strata is inexplicable.

Precipitation of gypsum in voids created by solutional removal of host sediment is an alternative mechanism for replacive accretion. Such minerals as calcium carbonate and soluble sodium salts are unlikely to have been present in quantity for the reasons presented above. Dissolution of quartz grains in the host sediment is a potential mechanism for relative enrichment of gypsum and possibly also for maintenance of porosity, the voids in the crust being the loci of former quartz grains. Experimental determination of silica solubility show that while 100 to 140 ppm of amorphous silica may dissolve in pure water (Krauskopf, 1956) the figure for quartz is 10 to 11 ppm at 25°C (Siever, 1972). There is linear increase in solubility with temperature but particle size has a more marked effect. Iler (1955, p. 9) showed that solubility of quartz is 7 ppm in the case of 100 millimicron particles and increases to 28 ppm, 120 ppm and 930 ppm respectively when particles are 10, 5 and 3 millimicrons in diameter. While this relationship may account for the removal of very small quartz grains, it has been shown in chapter six that quartz fragments in Tunisian dune sand is predominantly in the sand size fraction, greater than 62.5 μm. pH has a more significant effect on quartz solubility, than either temperature or grain size, but only above pH 9.0 (Alexander, Heston and Iler, 1954; Iler, 1955, p. 12; Krauskopf, 1956). Amorphous silica solubility at 25°C increases from about 100 ppm at pH 7.7 to 400 ppm at 10.26 and 1120 ppm at 10.6 (Iler, 1955). While Page (1972) held that gypsum crusts exhibited pHs in the range

8.0 to 8.5, Gaddas (1971) found values of 9.06 to 8.82 down a 1.05 m section with 70.87% gypsum at the base. While quartz solution will be enhanced under these alkaline conditions, there is little micromorphological evidence for it. Since soil pH levels decrease with increasing depth, evidence of reprecipitation of opaline silica should be present; it is not. Fersmann and Wlodawetz (1926) reported columnar gypsum crusts overlying silcrete in Central Asia but there was no evidence to suggest coeval development. Though some quartz grains show signs of minor solution (see plate 5.57), the rarity of reprecipitation (see plate 5.41) suggests that they may be relic features. Quartz grains deflated from alkaline sabkha environments, perhaps at the same time as gypsum deflation, would retain surface solutional features and overgrowths provided that abrasion during transport was not pronounced.

It is felt that the dominant process involved in the subsurface accumulation of gypsum crusts is displacive crystallization of gypsum derived from groundwater, in the case of desert rose croûtes de nappe, and from surface deposits leached by meteoric waters which replace the soil moisture deficit, in the case of mesocrystalline crusts. Many surface crusts represent subsurface accretions exposed by deflation of overlying unconsolidated sediments. Exposure results in some chemical and micromorphological modifications under the influence of meteoric waters.

c. Local implications and ages of gypsum crusts.

There are very few examples of geochronometric dating of gypsum crusts but a wide variety of stratigraphic evidence has been employed to correlate crust formation with different climatic phases of the Quaternary. Coque's (1955 a.b., 1958) assumption that the different glacis of the southern Tunisia Chott region correlate with periods of Chott inundation when increased rainfall or available moisture enhanced erosion, was not upheld by Zimmermann (1963). He argued

that since there was not any firm evidence in support of more than one Chott high-stand, inundation only occurred on one occasion. Nevertheless, his assertion that beaches at different heights represent one deformed strandline has been disputed (Mensching, 1968) on the grounds of minimal tectonic activity in the area during the late Pleistocene (see chapter four). Coque (1958, 1962, pp. 99 - 101) felt that the gypsum crusts on the higher glacis represent deflation deposits from the Chotts each correlating with a different phase of evaporation after transgression. Consolidation of the surface deposits would protect the relic land surface from erosion during the ensuing arid period. He held that where three encrusted glacis are represented, the highest is of Villefranchian age and the lowest Tyrrhenian (Mindel-Riss). Le Houérou (1960) similarly felt that many indurated surface crusts are palaeosols dating from between the Villefranchian and Tensiftian (Riss-Würm). Owing to the very limited climatic range in which gypsum crusts develop and are preserved, it seems unlikely that they would remain extant during the most recent major period of moisture availability, the Würm, when the lowest glacis is assumed to have formed. Even Dalloni's (1953) and Bureau and Roederer's (1961) suggestion that all gypsum crusts formed during the Tensiftian may be in error. Page (1972) felt that although three crust ages are evident the earliest, pre-Würm, is represented by only very degraded crusts. He proposed that the two main periods of Chott deflation resulting in crust genesis were, first, in the mid- to late Würm and, secondly, in the early Holocene. Though gypsum crusts are found on three different glacis in some areas (Steiner, 1963; Zimmermann, 1963; Mensching, 1964), this does not necessarily imply that they are contemporaneous with the facets of the landscape. Indeed, Page (1972) pointed out that crusts are often found on the break in slope between glacis, forming a continuous consolidated surface.

On the basis of Mousterian artefacts, Page (1972) dated the first of the major phases of gypsum crust development to greater than 28,000 years BP.

Two important qualifications must be acknowledged. First, such correlations between archaeological industries and specific time periods are tenuous. The end of the Mousterian period may extend to about 15,000 years BP. Secondly, and more significantly, the origin of gypsum crusts by leaching of surface deposits and recrystallization in the soil zone means that datable material within the horizon may predate the period of crust genesis by an indeterminate length of time. Associated deposits can only provide a maximum age for the crust. Hence a ^{14}C date of 8,000 to 9,500 years BP obtained from a bone in a Tunisian crust (Page, 1972) may be unrepresentative of the true age of the deposit. At Oued Ferd, near Kettana, shells in a crust were dated to 1,470 BP using ^{14}C (Page, 1972). These dates imply that gypsum crusts have developed at different periods during the Holocene. The absence of solutional features on gypsum crystals and the sharp definition of lozenge-shaped crystals making up mesocrystalline subsurface crusts supports the contention that these are recent and probably contemporary accretions. It may be tentatively postulated that in southern Tunisia, the present gypsum crusts have developed during periods of Chott regression succeeded by less arid phases. Hence, gypsum will have been deposited in the Chotts when they evaporated during a dry period before about 20,000 years BP. The deflated material may have been consolidated and leached into the soil zone during the less arid late Würm but when the Holocene commenced so did the main period of gypsum crust development. Whether as a result of increased precipitation or decreased evaporation during the late Würm, the Chotts were flooded with waters rich in soluble salts derived from bedrock *via* groundwater and probably from older surface or near-surface crusts subjected to solution by meteoric water. Evaporation of this water at the onset of aridity precipitated the salts in the Chott basins and the materials were deflated and deposited upon the surrounding land surface as dunes or powdery accumulations. Subsequent consolidation and leaching by meteoric waters has produced the present gypsum crusts. Owing to the shortage of detailed palaeoclimatic information in the Chott region, it is difficult

to establish whether the present surface crusts are mid-Würm or early Holocene subsurface crusts exposed following deflation of the overlying sediments. Regardless of their age, it is not necessary to postulate major cycles of arid and wet conditions to account for deflation and leaching of surface sediments since this occurs on a seasonal basis even under the present climatic régime. Furthermore, if removal of the superficial deposits is by slope-wash, it is likely that infiltration will remove some soluble material to the lower soil horizons. In most areas of southern Tunisia, the relic surface crusts are being destroyed at present through leaching by meteoric waters. The present rainfall is, however, insufficient to remove the water soluble salts from the soil zone and subsurface crust accretion is occurring. Indurated gypsum crusts may play an important rôle in preserving the glacis topography. Once slope-wash or deflation has exposed a crust, further erosion is greatly retarded. Solution will destroy the crust slowly as will deflation of small hemihydrate crystals produced by dehydration (Butzer and Hansen, 1968, pp. 105 - 107). Hence, while the present crusts may not be of great age, their predecessors protected a topography little changed from that of today. The greater slope angles of higher, older glacis may testify to minor modifications (Coque and Jauzein, 1967).

The absence of surface gypsum crusts on the lowest of the glacis in the zone peripheral to the Chotts suggests that the surface crusts elsewhere pre-date the late Würm transgression. It is possible, however, that the lowest glacis is a Holocene feature, not Würmian. If this is the case, subsurface crusts associated with the glacis, as at Bechri (CD), will be the product of leaching and subsurface recrystallization of recent surface deposits. More information is required about the Quaternary history of the Chotts before it is possible to deduce the precise palaeoclimatic implications and age of gypsum crusts in the area. On the basis of the stratigraphic and scant radiometric evidence presented here, it appears that gypsum crusts have developed and become

fossilized over several periods in the late Quaternary. Coque's (1962,p. 98) observation that "les croûtes ne sont plus que les constructions mortes" seems invalid. Rather, they are undergoing continual destruction and regenesis under the influence of meteoric waters. Kulke's (1974) ^{14}C date of 20,900 ± 850 years BP obtained from an Algerian desert rose croûte de nappe and Bellair's (1958) observation of contemporary development of the same features, suggest that this also applies to these forms of gypsum crust, though groundwater rather than meteoric water is the main influence.

The same difficulties occur when attempting to determine the ages of gypsum crusts in the Central Namib Desert. Wieneke and Rust (1976) obtained ^{14}C dates from gypsum crusts mantling coastal raised beaches in the vicinity of Swakopmund. On the basis of these they not only dated two marine transgressions to about 35,000 years BP and 26,000 years BP but also proposed equivalent ages for the crusts (see chapter four). They did not go into details of the processe of gypsum crust genesis but suggested that leaching may be an important factor producing polygenetic crusts (Rust and Wieneke, 1976), notably those over 1.0 m thick (Rust, 1975). In the same way as Coque (1955 a.b., 1958) held that Tunisian gypsum crusts developed during the transition from arid to wetter climatic conditions, Wieneke and Rust (1976) argued that in the coastal Namib they formed under arid régimes when sea level was high. They did not, however, distinguish between arid stable and arid active conditions as they did in the cases of other geomorphic features. The main problem in determining the geochronometric ages of the crusts is again that related to the contemporaneity of pedogenic crusts and the dated material. Isotopic examination, though not providing absolute dates, is preferable since the gypsum itself is the object of analysis. Carlisle (1978) found that $\delta^{34}S$ levels in supposedly older crusts approximated those of recent surface features, +15.9‰ and +15.8‰ respectively. Assuming that the $\delta^{34}S$ levels of the source material have not altered over time, the crusts must be of a similar age since isotopic fractionation will

progress with time. It seems likely that many of the non-evaporitic gypsum crusts are about the same age. Furthermore, the close correspondence between $\delta^{34}S$ levels in fog moisture and the crusts (see chapter six) suggests a recent age. It is impossible, however, to determine whether crust genesis has occurred in a specific time period, such as the Holocene, or whether it has been a prolonged process with older material being leached out of the soil horizons or combined with more recent material through dissolution and recrystallization.

Goudie (1973) outlined a theoretical method for estimating the ages of desert crusts on the basis of their thicknesses and input of source material. It assumes that surface dust is the sole source of salts, which produce the crust through per descensum, that dust deposition rates are constant and that none of the inputs of the constituent mineral are lost from the crust horizon. Using:

$$Ct = \frac{I}{Cg} \times \frac{D\ gyp.}{C\ gyp.} \quad \ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ldots\ 9$$

where (Ct) is the accretion rate in millimetres/1000 years, (I) the surface input in g/m^2/year, (Cg) the specific gravity of the crust, (D gyp.) the percentage of gypsum in the input material and (C gyp.) the percentage gypsum in the crust, it is possible to estimate the increase in thickness per unit time. Assuming the crust contains 80% by weight gypsum, 20% quartz, a rather high figure, and has a volumetric porosity of 40%, the specific gravity will be 1.422. Surface inputs of soluble salts in fog moisture are about 120 kg/ha in the Central Namib Desert (Boss, 1941; Eriksson, 1958) and the mean Cl/SO_3 ratio is 7.1 Hence the percentage gypsum which could theoretically be deposited is 18.37, assuming all chloride is combined with sodium and all sulphur trioxide forms gypsum (see chapter six). From equation 9 using these figures:

$$Ct = \frac{12}{1.422} \times \frac{18.37}{80}$$

$$= 1.94 \text{ mm/1000 years.}$$

This is a rate of less than 20 mm in the last 10000 years. Such a low figure argues that gypsum crust accretion in the Central Namib Desert has taken a considerable period of time, perhaps 250000 years for a 0.50 m thick crust to evolve. Alternatively, the salt deposition rate reported by Boss (1941) may not be representative. Walter (1937) reported a salt deposition rate ten times greater than this value. This still accounts for only 0.20 m of crust accretion during the Holocene. Other sources of salts cannot be ruled out. It should be stressed, however, that the limited atmospheric precipitation also restricts the amount of salt that can be mobilized into the soil zone. Assuming precipitation from mist of 40 mm/year (Boss, 1941) and a maximum gypsum solubility of 10 g/1 in an optimum sodium chloride-water solution (Zen, 1965), 0.1739 mm/year of gypsum may be mobilized. Free sulphate in solution will combine with cations other than calcium but may form gypsum through cation exchange in the soil adsorption complex in the presence of less salt-saturated rainwater. Hence, if gypsum was available, a crust accretion rate of 1.74 m/1000 years would be feasible, so accounting for the development of most crusts in the Central Namib Desert during the Holocene.

There are very few reported examples of non-evaporitic gypsum crusts in the geological record. Shallow-water and sabkha evaporite sequences have attracted considerable attention particularly during the last two decades (see chapter two) but most are associated with coastal facies. Some continental sabkha deposits may reach considerable thicknesses in tectonically active areas (Allen and Kramer, 1953) but non-sabkha groundwater crusts and pedogenic crusts have not been identified in ancient sequences even as isolated horizons. Gypsum-rich, fossilized sand dunes have been identified in the geological record in many areas (Tanner, 1965; Glennie, 1970) but zones of gypsum enrichment cutting across aeolian bedding planes, which would suggest crust developement, have not been reported. Pag and Adams (1940) described a 3.0 m thick horizon of anhydrite cemented quartz sand at the base of the Dewey Lake Formation overlying the Rustler Formation of the Ochoa Series of New Mexico. An unconformity at

the top of the anhydrite zone and bleaching of the surface may suggest pedogenesis on an aeolian gypsum deposit altered to anhydrite after burial. Termier and Termier (1963) described saccharoidal gypsum horizons in the Lédian (Upper Eocene) deposits of the Paris Basin near Cormeilles-en-Parisis. While they held that the material was an aeolian deposit, the micro-texture indicates post-depositional recrystallization. It is impossible to ascertain whether this was pedogenic or geologic. Grandjacquet and Haccard (1975) suggested a Mio-Pliocene pedogenic origin for certain gypsum deposits in the French Maritime Alps but this was dismissed by Vernet (1976) who argued that the structures were the products of tectonic disruption of Triassic evaporites. Gypsum concretions in an early Oligocene soil in Limburg were interpreted by Buurman and Jongmans (1975) as altered pyrite derived from decomposition of organic material. Though the scarcity of reported examples of gypsum crusts in the geological record may reflect a lack of identification rather than their rarity, the susceptibility to degradation may also account for it. The crusts develop near the surface probably over several thousand years. They will be preserved only if sufficient surface deposition occurs to bury the crusts so that they are beyond the range of the influences of undersaturated meteoric waters. Such burial must necessarily be by aeolian deposits since water-borne sedimentation will be accompanied by leaching. In areas of accelerated aeolian deposition, gypsum crusts will only develop during lulls in sedimentation because a relatively stable surface is a prerequisite. In effect, the likelihood of development, burial and preservation of gypsum crusts is very limited. Perhaps the most common indicator of gypsum crusts in the geological record will be calcite pseudomorphs after gypsum. Since calcium carbonate is less prone to dissolution these textures will be preserved more frequently than the materials they replaced. Though Lacroix (1897) reported pseudomorphs after gypsum from deposits in the Paris Basin, he held that they were a product of groundwater influences after burial of evaporitic gypsum. Nevertheless, gypsum pseudomorphs in calcretes such as those at Zerkine (ZER) in southern Tunisia provide valuable

indications of former gypsum crusts which are themselves very environmentally specific.

CHAPTER 8

IMPLICATIONS, SUMMARY AND CONCLUSIONS

The foregoing discussion of the distribution, nature and origins of gypsum crusts in semi-arid and arid environments has elucidated the processes of crust genesis and preservation, while the conclusions are based upon the examination of crusts in southern Tunisia and the Central Namib Desert, the descriptions of gypsum crusts from other deserts indicate that they are structurally and genetically similar. Here the implications of the models of crust formation will be assessed and the thesis summarized.

a. Geomorphic implications.

There is little evidence to support Coque's (1955 a.b., 1958, 1960, 1962) hypothesis that gypsum crusts of different ages mantle the glacis slopes around the Chotts of southern Tunisia. Indeed, since gypsum crusts also mantle the steps between the glacis slopes, it is likely that the crusts are of similar ages. Nevertheless, the gypsum crusts of southern Tunisia and the Central Namib play a significant rôle in the preservation of relic landscapes and landforms. The glacis of the chott region and fossil dunes of the Nefzaoua are protected from erosion by wind and water through the consolidation of the surface sediments by gypsum. Similarly, the erosion of raised beaches along the coast of Namibia is limited by the presence of relic gypsum croûtes de nappe. While gypsum accumulations will be dissipated very quickly when the climate becomes wetter than that under which they developed, under more arid conditions the crusts form a barrier to aeolian erosion.

The gypsum crusts of southern Tunisia, particularly the surface forms, have

important implications on the extent and degree of aeolian activity in the region. Around the chotts columnar gypsum crusts are exposed at the surface in many areas. Since these crusts represent exhumed subsurface accumulations, there has been considerable erosion of the overburden. This generally comprises 0.20 m to 0.50 m of silt and sand sized material. The presence of residual cobbles or pebbles on the surfaces of the crusts is attributed to the preferential removal of fines by the wind. The present wind régime would remove these airborne materials to the south-west, toward the Grand Erg Oriental. However, the loess deposits of the Matmata Highlands were probably deposited by south-westerly winds blowing from the erg (Grolier, pers. comm.) during the late Pleistocene. It is feasible that some of this deposit represents the leached powdery material found overlying many mesocrystalline subsurface gypsum crusts. It should be noted here that in some areas processes other than aeolian erosion have exhumed subsurface crusts. In the vicinity of Djebel el Issa, for example, only those crusts on the upper pediment slope are exposed. Further downslope the gypsum crust is mantled by a thick (1.0 m to 1.5 m) accumulation of silty sand. This material has been moved downslope by sheet-wash.

Another significant geomorphic implication of the model of gypsum crust genesis that has been proposed is exemplified at Djebel ed Issa. This is the ability of gypsum to cause mechanical weathering of rocks through displacive crystal growth in incipient fissures. Gypsum has rarely been considered as important as sodium salts, particularly chloride and sulphate, in the process of salt weathering. While Goudie, Cooke and Evans (1970) showed that large forces are exerted during unidirectional growth of gypsum crystals, most workers have stressed the importance of the volumetric changes during dehydration and rehydration (Mortensen, 1933; Cooke and Smalley, 1968). Mortensen (1933) held that dehydration of gypsum to anhydrite could take place readily in hot

deserts. While this is not borne out the chemical considerations discussed in chapter one, the transition from gypsum to hemihydrate and back to gypsum probably occurs in many warm deserts. The difference in unit volume between the two minerals is about 5% (Chatterji and Jeffery, 1963). Fookes and Collis (1975) held that both gypsum crystal growth and volumetric changes, resulting from thermal expansion of crystals and hydration process, are significant in salt weathering by gypsum. An interesting result of this is the occurrence of moving stones in southern Tunisia near Ghadamès (Di Cesare and Pratelli, 1967). Unlike the moving stones found on playas (Sharp and Carey, 1976) those in Tunisia move over undulating terrain, sometimes over several tens of metres. The mechanism involved consists of displacive gypsum crystallization from gypsiferous clay, over which the boulders travel. Crystallization occurs predominantly on the underside to the north out of direct sunlight. Here crystal growth is more prolonged and the boulder is pushed southwards.

The solubility of gypsum and its susceptibility to dehydration have long presented engineers with difficulties in areas where the mineral occurs in the bedrock. With increasing development of the warm desert regions of the world the presence of gypsum crusts at or near the surface is proving problematic. In the Souf of Algeria the construction of roads over gypsum encrusted terrain has given rise to a number of problems (Horta, 1979, 1980). Blistering and doming of the road-surface a short time after construction is attributed to displacive crystallization of sodium chloride and gypsum after removal of the surface sediments allows the evaporation front to penetrate to greater depths. Despite these difficulties, the roads are easier to construct and require less maintenance than those crossing sabkhas or sandy or calcreted terrair

It should also be noted that in some areas gypsum crusts provide potential resources for the plaster industry. This is particularly true in the case of

lacustrine and lagoonal evaporite deposits which may be very pure. In both southern Tunisia and the Central Namib Desert, surface crusts are exploited on a small scale for the plaster industry and as building-stone. In the coastal Namib particularly, where gypsum is absent from the geological record, the crusts provide an important potential source of gypsum (Martin, 1963). In India (Srivastava, 1969) and Australia (Jack, 1921; De la Hunty and Low, 1958) extensive areas of gypsum crusts are exploited for industrial purposes. Though gypsum crusts are rarely associated with other economically important minerals, Carlisle (1973) described associated uranium deposits in the Central Namib Desert which are currently being mined.

Agricultural development in gypsum encrusted areas is limited more by the local climate than by the presence of the crusts. In the Djeffara of southern Tunisia, shallow-rooting pistachio trees are cultivated on gypsum crusts. The roots penetrated to a depth of about 1.0 m where gypsum content is about 50% (Gaddas, 1971). They do not penetrate the underlying cemented zone where gypsum concentrations reach almost 90%. In some areas of Tunisia irrigated agriculture is undertaken utilizing saline water. The success of the technique rests upon the establishment of an efficient drainage system (Cointepas, 1965). Under such conditions an established gypsum crust will undergo dissolution provided that the irrigation water is not saturated with gypsum. In Namibia, gypsum obtained from the crusts along the coast is used to reclaim saline soils further inland where irrigated agriculture is possible. Gypsum which has been spread on the surface is leached into the soil zone by irrigation water. Here it combines with sodium chloride in the soil complex. Highly soluble salts such as calcium chloride and sodium sulphate are leached out of the upper soil zone (Krupkin, 1963; Drever and Smith, 1978).

These comments on the geomorphic and economic importance of gypsum crusts in

deserts are not intended to cover all the possible implications of the foregoing study. Many implications have not been considered. The value of gypsum crusts as palaeoenvironmental indicators has been referred to in chapter seven. There are many problems inherent in their dating and in reconstructing the climatic conditions under which now relic forms have developed. Isotopic studies of gypsum crusts have recently been undertaken by a number of workers (Sofer, 1978; Vieillefon, 1980). Though there are as yet many difficulties, particularly when dealing with the isotopic characteristics of the water of crystallization, in the future it may be possible to date gypsum crusts as accurately as it is to date calcretes at present. Vieillefon's (1980) study of the tritium levels of the water of crystallization in gypsum making up Tunisian crusts augurs well for the future. Further work is required to elucidate the precise processes involved in subsurface gypsum crust accretion. This study has been largely based on the examination of crusts from two field areas. While this has provided much data on the nature of the crusts, the genetic processes are possibly best monitored under laboratory conditions. It is hoped that the study undertaken by Plet-Lajoux, Monnier and Pedro (1971) will not remain the only attempt to simulate crust development in a laboratory.

b. Summary.

Gypsum-rich soils and crusts have been reported from all the continents, including Antarctica. They are usually restricted to areas with less than 250 mm of rainfall annually and an excess of potential evapotranspiration over rainfall in every month of the year. The crusts comprise surface or near-surface accumulations of calcium sulphate dihydrate which may reach thicknesses of over 4.0 m. The gypsum contents are very variable and may exceed 95% by weight in subsurface as well as surface crusts. Additional constitu-

ents are usually quartz and calcium carbonate. The gypsum accumulations show a great structural diversity, besides indurated microcrystalline crusts, desert rose crusts consisting of interlocking lenticular gypsum crusts, powdery accumulations and horizontally bedded crusts are common. Three main types of gypsum crusts may be distinguished on the basis of their stratigraphic and macrostructural characteristics. These are subsurface crusts, bedded surface crusts and a broad category of other surface forms. The subsurface type has two distinct forms. One consists of large gypsum crystals which are found at or near the water table in sabkhas and other areas where groundwater approaches within about 1.0 m to 2.0 m of the land surface. The second is made up of lenticular gypsum crystals about 1.0 mm in diameter and occurs 0.50 m to 1.5 m beneath a variety of terrain on slopes of up to 25^{o}. This form is frequently found beneath surface gypsum accumulations. Bedded surface crusts are restricted to lacustrine and lagoonal environments though relic forms are encountered forming areas of positive relief when surrounding unconsolidated sediments have been eroded.

i. Nature and origins of bedded surface crusts.

Horizontally bedded gypsum crusts are characterized by strata 50 mm to 100 mm thick which may be disrupted by tepee structures at intervals of 0.50 m to 2.0 m laterally. The upturned fractures are probably the product of desiccation at the land surface and are exaggerated by upward movement of saline groundwater. Individual strata comprise microcrystalline or alabastrine gypsum at the top grading downwards to lenticular gypsum crystals up to 5.0 mm in diameter. Gypsum contents are variable since horizons of silt and sand occasionally occur between the gypsum beds, about 70% by weight is common. In samples from the Central Namib Desert and southern Tunisia, in the vicinity of Chott el Rharsa, additional constituents are principally quartz grains

and calcium carbonate.

The graded bedding of the strata suggests that gypsum crystallization occurred in a shallow surface water environment. Surface evaporation from a shallow water body rich in dissolved salts will not produce the marked density stratification encountered in deep water. If the water is less than 1.0 m deep, sulphate reducing bacteria are unable to effect the alteration of gypsum to calcite or aragonite as crystallites sink from the surface. Hence the first gypsum crystals to precipitate will continue to grow on the bed as ions migrate to the crystal faces. Late in the evaporitive cycle not only is the time available for crystal growth reduced but also dissolved salts more soluble than gypsum, especially sodium chloride, may hinder growth and eventually nucleation. Assuming the water body evaporates to dryness, the final gypsum precipitates will form alabastrine material.

ii. Nature and origins of desert rose crusts.

Subsurface crusts composed of large interlocking lenticular gypsum crystals are usually confined to low-lying areas and are encountered at or just above the water table. Relic examples are characterized by corroded crystals which, when exposed at the land surface, may decompose completely to form microcrystalline crusts. The term croûtes de nappe has been adopted widely, the implication being that the crusts are the product of gypsum crystallization from saline groundwater. In areas which have experienced marked fluctuation in groundwater levels, thick croûtes de nappe develop. Kulke (1974) reported examples 5.0 m thick in the Algerian Souf.

The desert rose crusts usually have a markedly higher insoluble residue content than the other types of gypsum crust. The insoluble material (determined

after 30 minutes boiling in 1.0N HCl), which often constitutes 25% of the bulk sample, is in the form of quartz grains included poikilitically within the gypsum crystals. Poikilitic inclusions are rare in other gypsum crusts even when large porphyroblastic gypsum crystals are present. Poikilitic inclusion of host sediment within a growing crystal occurs during rapid and uninterrupted growth when ionic migration to the crystal faces exceeds the rate below which host grains are displaced. These conditions will be fulfilled at a near-surface water table where evaporation causes saturation with respect to gypsum and saline water is continuously replenished from below.

Several mechanisms have been proposed to account for croûte de nappe genesis. Pouget (1968) discounted the role of evaporation in cases where the crusts occur at depths greater than 2.0 m. He held that fluctuations in the amount of sodium chloride in the groundwater may cause periodic gypsum crystallization beneath the surface evaporation zone. If meteoric water percolating downwards has a lower sodium chloride/calcium sulphate ratio than the groundwater, the solubility of gypsum may be reduced sufficiently to promote its crystallization. Croûtes de nappe generally have higher sodium levels than most of the other types of gypsum crust except some bedded, lagoonal evaporites. Sodium probably stimulates the development of the lenticular crystal habit by retarding growth of the (111) face at the expense of $(\bar{1}02)$. However, the mixing model does not allow for long-term stability in hydrochemical conditions and may therefore preclude the development of large crystals with poikilitic inclusions of host sediment. Since most croûtes de nappe occur near the land surface, gypsum crystallization through groundwater evaporation is the most likely genetic process.

iii. Pedogenic crusts.

The other forms of gypsum crust, the mesocrystalline subsurface form and the surface forms which do not fall into the category of horizontally bedded crusts, are genetically related. They are essentially pedogenic, the surface forms representing exhumed and degraded illuvial accumulations.

Distribution:

Though references to surface and near-surface gypsum accumulations abound in geological, geomorphological and pedological literature, few workers have provided detailed information on the structures encountered. Generally the crusts are located in areas receiving 200 mm or less rainfall annually. The few exceptions, such as in Iraq and Rajasthan where annual rainfall may reach 300 mm, may be explained by high temperatures which maintain potential evaporation levels above actual precipitation throughout the year. Besides the climatic limitations on the distribution of gypsum crusts the other main prerequisite for crust formation is supply of calcium sulphate dissolved in groundwater or present in the atmosphere in fog moisture or in the solid phase as sand or dust. Crusts composed of authigenic gypsum produced bacterially or chemically have been reported from Central Asia and the Namib Desert. However, in both areas alternative mechanisms for gypsum accumulation are more probable.

Unlike the bedded surface crusts which are confined to ephemeral lake basins or lagoons and the croûtes de nappe which are restricted to areas with a near-surface water table, the mesocrystalline subsurface and other surface forms have a diverse local distribution. Though frequently found in the vicinity of large basins of inland drainage, such as Lake Eyre in Australia and the

Chotts of southern Tunisia, the terrain they underlie or mantle is very variable. Not only do they encrust extensive plains but also relic dunes, pediment slopes of 20^{o} or more, hill crests and interfluves. These characteristics apply to both the surface and subsurface forms which frequently occur together.

Structure:

The most striking structural feature of the indurated surface gypsum crusts is their tendency to columniation. Roughly hexagonal columns 0.25 m to 0.75 m in diameter may extend vertically through the full thickness of the crust which is usually 1.0 m to 2.0 m. The surfaces and walls of these columns are more indurated than their interiors being characterized by alabastrine crystal textures. This is the result of gypsum dissolution and rapid recrystallization under the influence of meteoric waters moving downwards in the narrow fissures between the columns. The origin of the columniation is enigmatic. It may be the product of lateral displacement of host sediment during gypsum crystallization but little non-gypsic material is found between the columns. It is more likely that tensional stresses caused by desiccation or partial dehydration of gypsum ($CaSO_4.2H_2O$) to hemihydrate ($CaSO_4.\frac{1}{2}H_2O$) result in fissuring. Downward water movement preserves the cracks. Mesocrystalline subsurface crusts also exhibit columniation.

Two additional forms of surface gypsum accumulation are common. One consists of unconsolidated powdery material which has a very variable gypsum content, from 20% to 90%. It is occasionally encountered overlying indurated mesocrystalline crusts. The second form comprises cobbles of indurated alabastrine gypsum studding areas of powdery gypsiferous material. The cobbles, which may be from 20 mm to 500 mm in diameter, are conical when small with

the apex on the underside, the flat upper surface is level with the land surface. The cobbles probably represent the remnants of a former columnar crust which has been degraded through solution, the powdery host sediment being residual material. Gypsum cobbles have the same superficial induration characteristic of columns.

Micromorphology:

In thin section mesocrystalline subsurface crusts are composed of lozenge-shaped gypsum crystals which represent lenticular crystals in two dimensions. The lozenges are generally less than 1.0 mm in length and only rarely contain inclusions of host sediment. Peripheral solution features are rare. In contrast, indurated surface gypsum crusts are made up of alabastrine material, individual particles having a diameter of less than 50 um. Larger lozenge-shape crystals occur as porphyroblasts which invariably show evidence of superficial solution. Porphyroblasts are absent from gypsum cobbles suggesting that infiltrating meteoric water partially dissolves gypsum crystals. Alabastrine textures are produced when rapid evaporation causes recrystallization but ionic migration to the embryonic crystal faces is limited by the availability of solvent.

Indurated surface crusts and the mesocrystalline subsurface form may also exhibit granular gypsum textures comprising interlocking irregularly-shaped crystals. These are confined to column edges and probably result from crystallization under water logged conditions. Two important crystalline textures which are encountered mainly in the subsurface form of gypsum are the fibrous habit and calcite pseudomorphs after gypsum. Fibrous gypsum has been attributed to two processes, tensile strain crystallization or displacive crystallization. While the latter has most often been invoked under high

pressures associated with anhydrite hydration, Watts (1979) described fibrous calcite which has formed near the surface. Fibrous gypsum crystals from subsurface crusts exhibit uniform extinction even when curved; undulose extinction would suggest tensile strain crystallization.

Lozenge-shaped calcite crystals, pseudomorphic after gypsum, have only very rarely been described from near-surface deposits. In examples where only partial replacement has been achieved, the calcite crystals grow into the gypsum crystals along cleavage planes, usually (010). There is no evidence of interstitial voids between the calcite crystal faces and the gypsum. The features therefore are not the product of solutional replacement but probably chemical alteration by sulphate reducing bacteria under aqueous anaerobic conditions.

Chemistry and Mineralogy:

While the forms of subsurface gypsum crust, the croûte de nappe and the mesocrystalline, may often be distinguished owing to the higher clastic content of the former, this is not the case with mesocrystalline form and the indurated surface crusts. The surface crusts contain slightly less insoluble residue and more gypsum than the subsurface crusts they overlie. Nevertheless, the gypsum content of either may be as high as 90% by weight, the remainder consisting of insoluble material in the form of quartz grains and generally less than 5% calcium carbonate. Minor constituents such as potassium, iron and aluminium are found in quantities proportional to the insoluble residue content, probably being derived from coatings on sand grains. Sodium, strontium and magnesium tend to be more abundant in the subsurface crusts than the overlying gypsum accumulations. It is unlikely that these cations are derived from the host sediment since the intermediate horizon between the gypsum crusts is often depleted with respect to them. It is probable that the sur-

face gypsum horizon has been leached by meteoric waters and the most soluble minerals translocated to the lower horizon. In the case of strontium ions, this process may involve the liberation of liquid inclusions in large gypsum crystals undergoing solution. Alternatively recrystallization of gypsum at the surface following partial dissolution of crystals may be too rapid to allow strontium ions to fill gaps in the crystal lattice. Free strontium ions will be leached into the lower horizons during subsequent wetting.

Powdery gypsum accumulations show a great variation in their chemical compositions not only in their gypsum and insoluble residue contents but also the accessory constituents. Gypsum cobbles have compositions very similar to indurated surface crusts lending support to the view that they represent remnants of surface crusts following solutional degradation. Their gypsum content may exceed 90% by weight.

iv. Origins of pedogenic gypsum crusts.

A number of models have been postulated to account for the development of surface gypsum crusts and the subsurface form which differs from the croûtes de nappe in terms of its topographic distribution and structure. Lacustrine sedimentation may be discounted on the grounds of the structural distinctions between the surface crusts presently encountered in lake basins as well as on topographic criteria. Many surface crusts occur on hill crests well above the possible zone of lacustrine transgression.

This topographic factor also precludes an origin related to groundwater influences. While it is true that gypsum croûtes de nappe may be structurally altered from aggregated large lenticular crystals to alabastrine gypsum through solution and recrystallization at the land surface, the occurrence

of surface crusts on steep slopes rules out the possibility that they are the product of evaporation of a near-surface water table. Furthermore, it is unlikely that capillary rise plays a role in their genesis. Keen (1936) argued that the cellular structure of soil voids precludes the upward movement of water by capillarity. Empirical evidence supports this. Moreover, in order for a 0.50 m thick gypsum crust to accrete, 200 m of water would have to evaporate. Assuming maximum solubility the figure would be 40 m. Without groundwater replenishment the water from 100 m of sediment would have to rise to the surface. Several alternatives have been proposed to overcome the shortcomings of the per ascensum models. Bureau and Roederer (1961) suggested that meteoric waters which replenish the moisture deficit in gypsiferous surface sediments are drawn to the surface by capillarity during subsequent dry periods, precipitating the dissolved gypsum at the surface. Risacher (1978) proposed that water moving laterally through the soil zone down slopes would rapidly become saturated with gypsum derived from the soil materials. Subsequent evaporation of soil moisture would result in gypsum crystallization and crust accretion. While these processes occur in some instances, they are not applicable in many areas where gypsum is not present in the local bedrock or lower soil zone.

A number of workers have suggested that surface gypsum crusts represent surface aeolian or atmospheric deposits which have become consolidated under the influence of meteoric water. Coque (1955 a.b., 1958, 1960, 1962), Le Houérou (1960) and Mensching (1964) held that the surface gypsum crusts found in the vicinity of the Chotts of southern Tunisia are evaporitic deposits deflated from the lake basins following evaporation of lake waters at the end of cooler periods during the Pleistocene. A similar model was proposed by Jessup (1960 a.b.c.) in South Australia. He suggested that subsurface gypsum crusts are deposits deflated from evaporitic pans, clays overlying the crusts representing materials subsequently deflated from the same pans. While the structural

and chemical distinctions between surface and subsurface crusts in the same section may be the result of diagenesis, the aeolian model has several inherent inconsistencies. Most significant of these is the presence of lag gravels and cobbles on top of surface crusts or above subsurface gypsum horizons. In many cases, the deposition of these gravels cannot post-date gypsum accumulation. Indeed, large bedrock fragments showing no evidence of lateral transport may overlie the crusts. The size of these boulders precludes the possibility that they were translocated upwards through surface sediments by cyclic wetting and drying causing volumetric changes in clay minerals or gypsum.

v. Per descensum accumulation of gypsum crusts.

Models of subsurface accumulation of water soluble salts have long been advocated to explain the genesis of calcretes. Page (1972) attributed the gypsum crusts of southern Tunisia to leaching of surface deposits of gypsum into the soil zone by rainwater. The water will replace the moisture lost during antecedent dry conditions and subsequent evaporation, from the surface downward, will result in the crystallization of salts dissolved in the water. Provided that the amount of infiltrating rainwater does not exceed the soil moisture deficit, the soluble salts will not be flushed out of the soil zone. Under such conditions accumulation of soluble salts will occur as long as they are available at the surface through aeolian or atmospheric deposition. In southern Tunisia, this takes the form of gypsiferous sand and dust deflated from evaporative basins. In the Central Namib Desert and parts of Soviet Central Asia, salts dissolved in fog moisture provide the principal source.

This model is applicable to only the mesocrystalline subsurface crusts. Indurated surface gypsum crusts are attributed to the exhumation of subsurface crusts by deflation or fluvial and colluvial erosion on slopes. The structural

differences between indurated surface crusts and the mesocrystalline subsurface form may be explained by the dissolution of lenticular crystals by rainwater and rapid recrystallization producing alabastrine material. Once exhumation has occurred, the exposed crust may act as a source of gypsum for another subsurface crust, provided that rainfall is sufficient to induce leaching. Under such conditions, salts more soluble than gypsum which are trapped in the subsurface accumulation zone, will be quickly leached out of the surface horizon accounting for some of the chemical differences between the crust types. Late in the cycle, the relic surface crust will consist of residual cobbles, remnants of the former columnar structures. While powdery gypsum accumulations may represent the final stage in the deterioration of an exhumed horizon, it should be stressed that the great variability in their chemical composition indicates that some may be primary aeolian deposits.

vi. Displacive crystallization.

This model of gypsum crust genesis provides a mechanism for the formation of crusts on hill crests and hill slopes and also explains the occurrence of lag materials above the accumulations. However, a subsurface origin gives rise to the problem of accounting for the purity of the gypsum crusts. The crusts have volumetric porosities of between 30% and 45%, essentially the same as most host sediments. Hence their accretion does not involve the infilling of voids between host grains but either displacement or replacement. There is little evidence for chemical replacement. In cases where the materials above and below the gypsum horizon are almost wholly quartz sand, the quartz grains incorporated within the crust only rarely show signs of solution. In contrast, displacive features are common. Pseudo-anticlines, the products of lateral displacement, are frequently encountered in subsurface crusts but they cannot account for the 50% to 80% volumetric displacement to which many crusts testify.

Micromorphological features such as floating quartz grain textures, brecciation of large clasts and the fibrous crystals habit are indicative of displacive crystallization. Soil moisture containing calcium sulphate in solution will adhere to the surface of host grains owing to surface tension. Crystal growth during evaporation will be most pronounced at grain contacts where water will tend to move owing to the greater surface tension. Plet-Lajoux, (1971) showed that the force of crystallization was sufficient to displace the host grains. Lateral displacement will be limited by the mass of host material and counteractive crystallization forces in the horizontal plane. The force of crystallization is well in excess of that exerted by an overburden of 1.0 m or 2.0 m. Upward displacement will therefore predominate. Since crystal growth is limited by the availability of moisture, small crystals will predominate. The ionic attraction of crystal faces will be greatly reduced once soil moisture has evaporated and subsequent crystallization will occur at contacts between crystals, displacing them rather than increasing their size.

vii. Differential leaching.

In areas where at any time of the year rainfall is sufficient to exceed the soil moisture storage capacity of the soil zone, subsurface accumulation of gypsum will be hindered by leaching of the salts into the groundwater zone. Moreover, in extremely arid regions insufficient moisture may be available to leach surface deposits of gypsum into the soil zone. It is in these areas peripheral to the main occurrences of pedogenic gypsum crusts that two-tiered crusts may develop. At the upper rainfall limit gypsum crusts occur beneath calcretes and at the lower limit above halite crusts in those areas where both salts are available for leaching. Page (1972) attributed the superimposition of calcretes on gypsum crusts in southern Tunisia to the leaching of two dis-

tinct aeolian deposits at different times. In the Central Namib Desert, halite horizons beneath gypsum were held to represent two different periods of coastal evaporite sedimentation. In both cases the two horizons are clearly differentiated, there being little mixing of the salts. This would be unlikely if leaching and displacive accretion occurred at two periods. A marine origin of the crusts in the Namib Desert may be discounted on topographic grounds, some occur more than 200 m above present sea level.

In areas where calcium sulphate and calcium carbonate or calcium sulphate and sodium chloride are deposited at the surface, both will be leached into the soil zone if sufficient rainfall is available to mobilize the least soluble. Both will accumulate if rainfall is insufficient to flush the most soluble mineral out of the soil zone. Periodically, above average rainfall will cause leaching beyond the main horizon of crusts accretion. At these times the salts will be removed to greater depths in quantities proportional to their solubilities. Chemical considerations come into play since gypsum solubility is reduced in the presence of calcium carbonate and enhanced by sodium chloride but the relative mobilities of the salts through the soil zone differ sufficiently to result in their accumulation within discrete horizons.

viii. Conclusions.

Gypsum crusts may be subdivided into three main types on the basis of genetic and structural criteria. Those formed by evaporation of shallow surface water bodies in lacustrine and lagoonal environments are characterized by horizontal, size-graded beds. Those which develop through evaporation of near-surface water table, croûtes de nappe or desert rose crusts, are composed of large interlocking lenticular gypsum crystals which enclose host sediment grains poikilitically. The clastic content of these crusts is frequently much higher

than that of the pedogenic crusts which may contain up to 90% gypsum. The pedogenic crusts have a number of forms. The variety is a product of exhumation and degradation of subsurface accumulations which are the only actively accreting form. Crust genesis occurs when meteoric water, which contains gypsum dissolved at the surface, is held in the soil zone owing to a soil moisture deficit until subsequent evaporation precipitates the soluble salts. Crystallization at host grain contacts and occasionally as fibrous gypsum results in displacive accretion. Most displacement is in an upward direction. Exhumation of the subsurface illuvial horizon exposes indurated columnar crusts which degrade to produce gypsum cobbles and powdery accumulations with a low gypsum content if sufficient rainfall is available to induce leaching. The dissolution of the surface crust may provide the gypsum necessary for the formation of another subsurface accumulation. Two-tiered pedogenic crusts, usually either calcium carbonate/gypsum or gypsum/halite, form when two salts with different solubilities are mobilized from the surface into the soil zone. Differential accumulation occurs if rainfall is insufficient to leach the most soluble salt out of the system.

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APPENDIX 1

The results of the chemical analyses of gypsum crusts, calcretes, bedrock and other associated sediments are presented in chapters five and six. A detailed description and assessment of the various analytic procedures adopted is provided here.

Methods of chemical analysis.

The most complex of the analyses undertaken was that of gypsum content. While gypsum solubility in water presents problems during thin section preparation, the mineral is not sufficiently soluble for any simple method of determination to be accurate. The problems have been tackled in a number of ways. These may be subdivided into chemical methods and electrometric methods.

Chemical determination of soil gypsum content usually involves the determination of calcium or sulphate. Though calcium content may be ascertained relatively simply by precipitation of the oxalate (Reitemeier, 1943) or more commonly using ethylene diamine tetra-acetic acid (EDTA) (Bodine and Fernalld, 1973), the presence of other calcium minerals complicates the procedure. Similarly, determination of the sulphate component possibly as benzedine sulphate (Reitemeier, 1943) or precipitation of barium sulphate ($BaSO_4$) using barium chloride ($BaCl_2$) will over-determine gypsum content if other sulphates are present. The barium chloride method is only suitable if the gypsum is in a solution (Pièce, 1952) since in the solid form an insoluble coating of barium sulphate forms over the gypsum particles (Garman and Hesse, 1975). Nelson, Klameth and Nettleton (1978) advocated determination of the amount of water of crystallization present in the soil by desiccating a preweighed sample for 48 hours.

This method also has several potential sources of error. Firstly, initial drying

to remove free moisture may result in partial dehydration of the gypsum to hemihydrate. As pointed out in chapter one, this may occur at temperatures as low as 35°C. Furthermore, desiccation to remove all the water of crystallization (20.91% by weight) would have to be at over 150°C, the lowest transition temperature between γ-$CaSO_4$ (soluble) and β-$CaSO_4$ (insoluble). The fact that Nelson, Klameth and Nettleton (1978) determined a maximum water content of 19.42% suggests that either there was partial dehydration of the sample before the experiment or complete dehydration was not achieved. Pouget (1965) found a correlation of + 0.951 between the humidity of soils and gypsum content in samples dried at 60°C. The presence of 0.90% by weight of water of crystallization, possibly zeolitic, in β-hemihydrate (Power, Fabuss and Satterfield, 1964) or soluble anhydrite, may account for part of the under-determination (see chapter one). Further error may result from the loss of water of crystallization from other hydrous minerals and clays, resulting in over-determination of the gypsum content.

Electrometric methods of gypsum determination suffer from similar problems as the chemical procedures. For example, Kovalenko (1972) proposed three different methods for chemical determination of gypsum, based upon leaching of the soil samples with water and determination of the calcium component of the dissolved gypsum chemically. While he realized that dissolved calcium carbonate would interfere, his suggestion of using calcium carbonate saturated solution in the leaching process, or chemically determining the carbonate component thereafter, are unsuitable. This is because in any solution saturated with either calcium carbonate or gypsum, further dissolution of the undersaturated mineral will lower the solubility of the other resulting in its precipitation (Wigley, 1973; Zozul, 1973). Though Bower and Huss (1948) and Jackson (1958) advocated conductimetric methods to overcome the problems of excess calcium and sulphate ions, and problems of cation exchange of calcium with sodium and magnesium in the adsorption complex, the difficulties remain. Lagerwerff, Akin and Moses (1965) compared

an electrical conductivity method and a chemical method of gypsum determination based on gypsum precipitation from a dilute soil extract using acetone. In the case of the conductimetric method, exchange error occurs. Dilution of any soil-water system containing both monovalent and divalent cations results in changes in cation-exchange equilibrium. The adsorbed phase is enriched with divalent cations at the expense of the monovalent (Krupkin, 1963). It is possible, however, to calculate the exchange error effect on conductivity. This method of gypsum determination is rapid but is unsuitable for accurate quantitative determination of samples with a high gypsum content. The chemical method examined by Lagerwerff, Akin and Moses (1965) avoids exchange error by determining sulphate associated with calcium but, since it is based on the solubility of gypsum, the presence of sodium chloride and other salts which increase solubility causes error. Since in many cases non-gypsic sulphate is less abundant than non-gypsic calcium. Golkin *et al.* (1960) advocated electrometric determination of sulphate in order to ascertain soil gypsum content. The method is again semi-quantitative.

Owing to exchange error and interference effects when calcium is determined and the limited accuracy and range of conductimetric methods, a chemical method of sulphate determination was preferred for the determination of the gypsum content of gypsum crusts. Pièce (1959) advocated the use of Amberlite IR-120 cation exchange resin to liberate sulphate from finely powdered anhydrite and gypsum.

$$2H - SO_3R + CaSO_4 \longrightarrow Ca(SO_3R)_2 + 2H^+ + SO_4^{--}$$

The amount of calcium sulphate may be determined by titration with sodium hydroxide to ascertain the sulphuric acid content. In preference to this very accurate method, which was developed for analysis of plaster which contains few interfering ions, a modification based on the same principle was adopted.

This method was developed by Coutinet (1965) and advocated by Vieillefon (1976) for the determination of gypsum crust gypsum content. It involves boiling of the soil sample in ammonium carbonate solution to facilitate a double decomposition.

$$(NH_4)_2CO_3 + CaSO_4 \longrightarrow CaCO_3 + (NH_4)_2\ SO_4$$

Since all the calcium is removed as a precipitate, all the sulphate in the gypsum is in solution as ammonium sulphate. The sulphate may then be precipitated as barium sulphate, by adding barium chloride, and thereby the original calcium sulphate content ascertained. It is assumed that other sulphates are either absent or will not react with ammonium carbonate. Sodium carbonate may be utilized in place of ammonium carbonate, this is only justified if the gypsum content is less than 60%, in which case the coefficient of variation is smaller (J. Vieillefon, pers. comm.). The method is as follows:

REAGENTS

0.05M ammonium carbonate solution.

0.20M barium chloride solution.

1.0N hydrochloric acid.

METHOD

Weigh 1.0 g of soil reduced to particles of less than 2.0 mm (N.B. powdering may result in partial dehydration) and place in 250 ml beaker.

Add 0.20 l ammonium carbonate (0.05M)

Boil for 30 minutes.

Cool slightly.

Adicify with hydrochloric acid (1.0N) to the point at which methyl orange turns red.

Filter and wash the residue with boiling water.

Boil the filtrant.

Add 0.03 l barium chloride (0.20M).

Leave standing for 12 hours.

Filter through ashless paper (Whatman 542).

Wash the residue with boiling water.

Place each filter paper and precipitate in a pre-weighed porcelain crucible.

Heat at 900°C for 2 hours.

Weigh the crucible.

Each gram of baked precipitate corresponds to 0.738 g of gypsum in the soil sample.

It is possible to modify the method to determine the amounts of hemihydrate and gypsum in the samples. Six 1.0 g fractions of each sample are boiled for different periods from 5 to 30 minutes in ammonium carbonate. When the results are plotted graphically against boiling time, the amounts of the different forms of calcium sulphate present may be ascertained owing to the variable rates of solution. This kinetic method was not adopted for these analyses.

Experimental determination of the accuracy of the method was undertaken using a powdered sample of satinspar from Upper Cretaceous bedrock in southern Tunisia. Chemical analyses showed that besides gypsum the sample contained the following:

Insoluble residue	0.40%
Calcium carbonate	1.61%
Na^{+} as NaCl	0.37%
K^{+} ppm	-
Sr^{++} ppm	200
Mg^{++} ppm	-
Fe^{total} ppm	-
Al^{+++} ppm	100

The analytical procedures used to ascertain these values are described below. The theoretical maximum gypsum content is 97.877%. Gypsum content was determined on ten samples weighing 0.10 g, 0.20 g, 0.30 g ... ... 0.90 g, 1.0 g, on a four decimal place balance. The percentage gypsum was above the theoretical maximum for the three smallest samples and below it for the others showing a decrease with increasing sample size. The mean was 96.455%, standard deviation 1.66 and

coefficient of variation 1.72%. While some variation will result from weighing errors the systematic decrease with increasing sample size is enigmatic. Over-determination may result from contamination of the precipitate by additional salts or the remnants of the incinerated filter paper. If this were the case, under-determination in the larger samples would be greater than the figures suggest. Under-determination may result from incomplete reactions of gypsum with ammonium carbonate or ammonium sulphate with barium chloride. Since the first of these reactions is time dependent it would appear to be the most likely source of error. It should also be pointed out that in addition to these analytical errors, the presence of sulphates other than gypsum will result in over-determination though only those sulphates that are water soluble or react with ammonium carbonate will influence the results.

Calcium carbonate was determined using a simple calcimeter (Allison and Moodie, pp. 1390 to 1392 in Black et al., 1965) calibrated using calcium carbonate. Though the presence of other carbonates that react with 1.0N hydrochloric acid will result in over-determination, their presence in the gypsum crust samples is unlikely. Dolomite is the most common mineral which causes interference. When its presence was ascertained, lactic acid was utilized in the calcimetry since it reacts with calcite, releasing carbon dioxide, but not with dolomite.

Gypsum and calcium carbonate were determined on 1.0 g and 0.20 g samples respectively. Since about 50 g of air dried material were crushed before aliquots were analyzed the results will reflect a uniform sample. All of the following analyses were conducted on a single 0.50 g aliquot of sample. Insoluble residue (I.R.) was ascertained by boiling the 0.50 g of crushed sample in 0.20 l of 1.0 hydrochloric acid for 30 minutes. It is impossible to dissolve gypsum in most cold acids over this time period. After filtering the residue was dried at 200^{o}C for 2 hours and reweighed. The filtrant was made up to 0.50 l with 2.0 ml lanthanum chloride (standard B.D.H. stock solution for A.A.S.) and distilled water. Aliquots of this

solution were then used for determining sodium and potassium by flame photometry and strontium, magnesium, iron, aluminium and calcium by atomic absorption spectrophotometry. Standards for calibration were prepared by diluting B.D.H. stock solutions with solutions of 0.20 l, 1.0N hydrochloric acid and 2.0 ml lanthanum chloride, made up to 0.50 l with distilled water. This solution was also used to dilute samples to less than 5.0 ppm sodium and 5.0 ppm potassium for flame photometry and less than 10.0 ppm of those elements determined using A.A.S. Within these ranges the relationship between concentration of the element and dilution is linear.

Response depression owing to ionization occurs when strontium, magnesium and calcium are determined using a hot nitrous oxide/acetylene flame for A.A.S. This is avoided when an air/acetylene flame is used but lanthanum chloride must be added to act as a releasing agent to overcome the response depression which occurs when elements such as aluminium are present (Whiteside, 1976). Total iron dissolved by the boiling in 1.0N hydrochloric acid was determined using an air/acetylene flame and a nitrous oxide/acetylene flame was employed for aluminium. While the presence of iron may depress the aluminium response, the effect was not felt to be significant enough to justify matching the concentrations in sample and calibration solutions. Similarly, ionization depression in the hot flame was assumed to be reduced by the presence of potassium in the samples though an ionization buffer was not added to the calibration standards.

Walsh and Howie (1967) assessed the experimental error in determining calcium and magnesium using A.A.S. as about 2.0% and found that the figures were generally greater than those obtained chemically. For this reason, the results of the analyses of gypsum crusts, tabulated below, contain a figure for excess calcium (xs Ca^{++}) in parts per million. The value is calculated as the amount of calcium determined using A.A.S. which is not accounted for by the total amount of calcium in the gypsum and the calcium carbonate as determined chemically. Where more than

1.0% (10,000 ppm) magnesium is present the theoretical dolomite content is also taken into consideration. This has been calculated on the basis of the maximum amount possible when magnesium and calcium are combined after subtraction of that present in gypsum. Notwithstanding the experimental errors and over-determination inherent in the analysis of calcium using A.A.S. and the under-determination of gypsum when using the chemical method described above, the excess calcium value is frequently quite large, of the order 0.50% to 4.0% and as high as 8.0%. Excess calcium is not usually present in those samples with a calcium carbonate content greater than the gypsum content, suggesting either over-determination of carbonate chemically, or under-determination of calcium associated with carbonate using A.A.S., or under-determination of gypsum chemically, or over-determination of its calcium component. Since excess calcium values greater than 2.0% (20,000 ppm) predominate when the gypsum content exceeds 40.0%, gypsum would appear to be the critical constituent. This becomes more apparent when the relationship between the excess calcium value and the weight percentage of sample not determined by the analyses is examined. The conversion factor is $((xsCa^{++})n)$ where $3.5 \leqslant n \leqslant 5.0$, while $\frac{CaSO_4.2H_2O}{Ca^{++} \text{ in gypsum}} = 4.2956$. This suggests that in many cases the excess calcium determined using A.A.S. is in the form of gypsum which has been under-determined in the chemical analyses by a factor of up to 10% in some cases. If this is the case, it is difficult to attribute the under-determination to any single factor. Other minerals in the samples may influence the reaction between gypsum and ammonium carbonate. The fact that in some samples which had a high chemically determined gypsum content the excess calcium value is low, even though this is not the case with samples analyzed in the same batch, supports this. It is also possible that some of the excess calcium determined using A.A.S. is derived from clay minerals therefore any deficits in the percentage composition of the samples may represent complex minerals dissolved during boiling in hydrochloric acid. Potassium, aluminium and in some cases, iron content provide a rough indication of amounts of certain

clay minerals and small iron contents may represent relatively large weight percentages of ferric and ferrous minerals.