Origin of fracture patterns and insoluble minerals in the Fort Dodge gypsum, Webster County, Iowa

by

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ABSTRACT

The Jurassic Fort Dodge Formation of Webster County, Iowa exhibits well-developed joints. The joints are generally planar, vertical, laterally extensive, and are limited to the gypsum bed, which is up to 10 m thick. Dissolution of gypsum by groundwater flowing along these joints has formed long, straight solution channels easily seen from the air permitting aerial photographs to be used for a quantitative study of the joint orientations. This study was done on three quarries, and the gypsum exposed there exhibited joint systems with orthogonal, random, and uni-directional strikes. The orthogonal and random sets are probably extension fractures produced by a stress field with a vertical σ_1, a horizontal, tensile σ_2 and σ_3, and a low deviatoric stress. These conditions were possibly the result of uplift which forced the crust to cover more area at a new, larger radius position within the Earth. Uplift was probably due to unroofing and isostatic rebound, either from erosion of now-missing Cretaceous strata during the Tertiary or from glacial ablation during the Pleistocene. The weight of the Cretaceous strata may have also caused partial dehydration of the gypsum, but the weight of the glaciers would not because of the lower temperatures involved. The uni-directional joint set is located near, and is parallel to, a northeast-southwest trending fault in the basalt of the Keweenawan basement. This fault, which is delineated by aeromagnetic data, was active throughout the Paleozoic Era and could have been reactivated by either of the above loading and unloading models. The uni-directional set of joints is thought to be the result of this reactivation.

The Fort Dodge gypsum is quite pure, with only 1-3.5% insoluble minerals. Soxhlet extractors were found to be effective in concentrating these impurities and leaving them intact. The most common impurity is very fine sand-sized to silt-sized quartz showing signs of eolian origin. Some of the quartz, however, is authigenic, forming euhedral crystals and spherical quartz clumps. Calcite is also common, and was probably deposited contemporaneously with the gypsum as an evaporite mineral. Other minerals include dolomite, orthoclase, kaolinite, illite, hematite, limonite, muscovite, and pyrite.

INTRODUCTION

Previous Studies

The gypsum member of the Jurassic Fort Dodge Formation is thought to be an evaporite unit deposited in a restricted marine basin under arid conditions (Bard, 1982; Dasenbrock, 1984). These researchers showed that the gypsum underwent considerable recrystallization. Several questions remain, however, related to the joint patterns and insoluble minerals in the gypsum. In particular, the insoluble minerals in the gypsum have not been fully characterized because their low concentration (1.0-3.5%) makes studying them difficult. Characterization of these impurities could give further clues about the climate and conditions at the time of deposition, as well as information about diagenesis. Another interesting problem is that of the joints in the gypsum and the conspicuous solution channels formed along them. These features have been described by Bard (1982) and Dasenbrock (1984), but the cause of the joints has never been determined. The stresses responsible for their formation might also have been a controlling factor in the extent to which dehydration and rehydration of the gypsum have occurred. Delineation of preexisting structures responsible for the joints might also have predictive value for the gypsum industry because in heavily jointed areas much of the gypsum has been subject to intense dissolution by ground-water and is not worth quarrying. Another important factor in the joint study was that it could take advantage of the large expanses of freshly exposed gypsum that occurred in 3 quarries simultaneously during June of 1985.

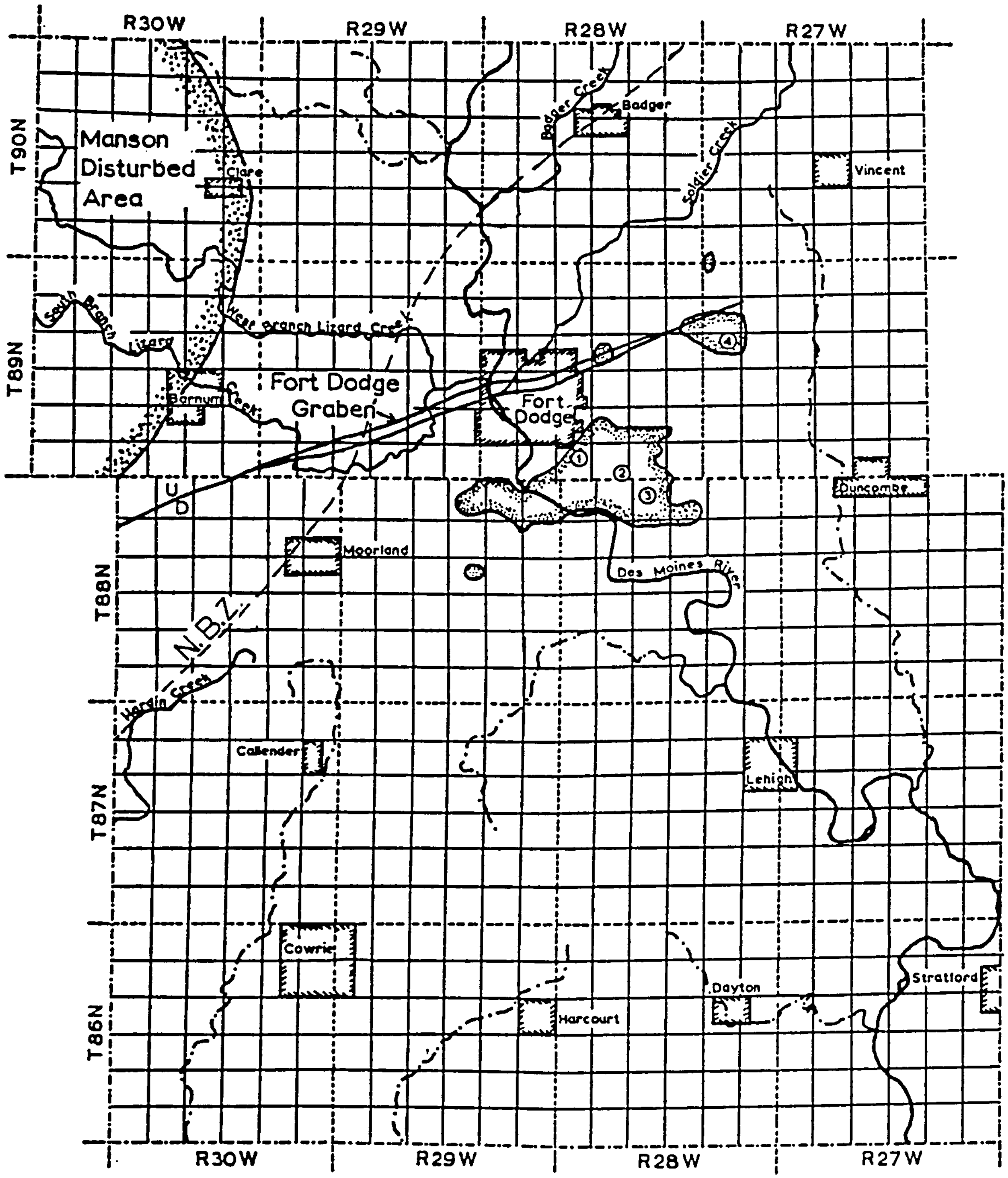
Local Geology

The Fort Dodge Formation is located near the city of Fort Dodge, in Webster County, Iowa. It consists of a very pure laminated gypsum member up to 10 m thick overlain by red beds of the Soldier Creek member. The areal extent of the gypsum member was determined by Dorheim (1978) and is shown in Fig. 1. Previous maps showing the extent of the Fort Dodge Formation (Hale, 1955; Wilder, 1917-1918) have included some areas where only the Soldier Creek member is present, and these maps were not used in this thesis. The formation unconformably overlies Pennsylvanian shales and sandstones of the Cherokee Group of the Des Moines Series in some places and the St. Genevieve Limestone of Mississippian age in others (Hale, 1955). It is overlain unconformably by about 20 m of Pleistocene glacial till. The formation has been tentatively dated as Late Jurassic in age by means of gymnosperm pollen (Shaffer, 1969). Marine microfossils have been observed in the Soldier Creek member (Johnson, 1986) but they are reworked from formerly adjacent rocks of Pennsylvanian age. More detailed descriptions of the origin and diagenesis of the deposit are presented by Bard (1982) and Dasenbrock (1984).

Figure 1. Location map of the study area in Webster County, Iowa.

The stippled patches near Fort Dodge indicate the known extent of gypsum in the subsurface, as determined by Dorheim (1978) from well cuttings and a resistivity survey. The dashed line labelled 'N. B. Z.' is the Northern Boundary Zone of the Midcontinent Geophysical Anomaly. Names and locations of quarries shown on the map are as follows:

1: North Welles (SW 1/4, NE 1/4, Sect. 33, T 89 N, R 28 W)
2: Celotex (SE 1/4, SE 1/4, Sect. 34, T 89 N, R 28 W)
3: Carbon (SE 1/4, NW 1/4, Sect.4, T 88 N, R 28 W)
4: National (location shown is approximate)


R30W
R29W
R28W
R27W
T90N
T89N
T88N
T87N
T86N
Manson Disturbed Area
Clare
Badger
Badger Creek
Soldier Creek
Vincent
South Branch Lizard
West Branch Lizard Creek
Fort Dodge Graben
Barnum
Creek
Fort Dodge
Duncombe
U
D
Moorland
Des Moines River
N.B.Z.
Hardin Creek
Callender
Lehigh
Cowrie
Dayton
Stratford
Harcourt
R30W
R29W
R28W
R27W

SECTION 1: JOINTS

General Description

Almost all exposures of the Fort Dodge gypsum exhibit well-developed joints. The joints can be seen in quarry walls and on the top surface of the exposed gypsum where overburden has been removed by quarrying operations. Spacing between the joints varies from 1 to 15 m. The joints themselves are generally planar, vertical, and laterally extensive. Their occurrence is vertically restricted to the gypsum layer. The joints were studied in three gypsum quarries, the North Welles and Carbon quarries belonging to U.S. Gypsum Company, and the Celotex quarry (Fig. 1). The National quarry was not studied quantitatively because the quarrying method used there leaves only a limited amount of gypsum exposed.

Almost all the joints have well-developed solution channels along them due to dissolution of gypsum by ground-water flowing along the joints. The channels are up to 2 m wide and up to 4 m deep and in some places extend all the way through the gypsum to the underlying strata. The channels usually extend laterally across the entire quarry exposure, and are straight, reflecting the nature of the joints that they follow. This relationship allows the use of solution channels to measure the strikes of the joints, a technique which eliminates the error of measuring recent joints caused by blasting or other quarrying activities because of the time it takes for a solution channel to develop.

Method of Study

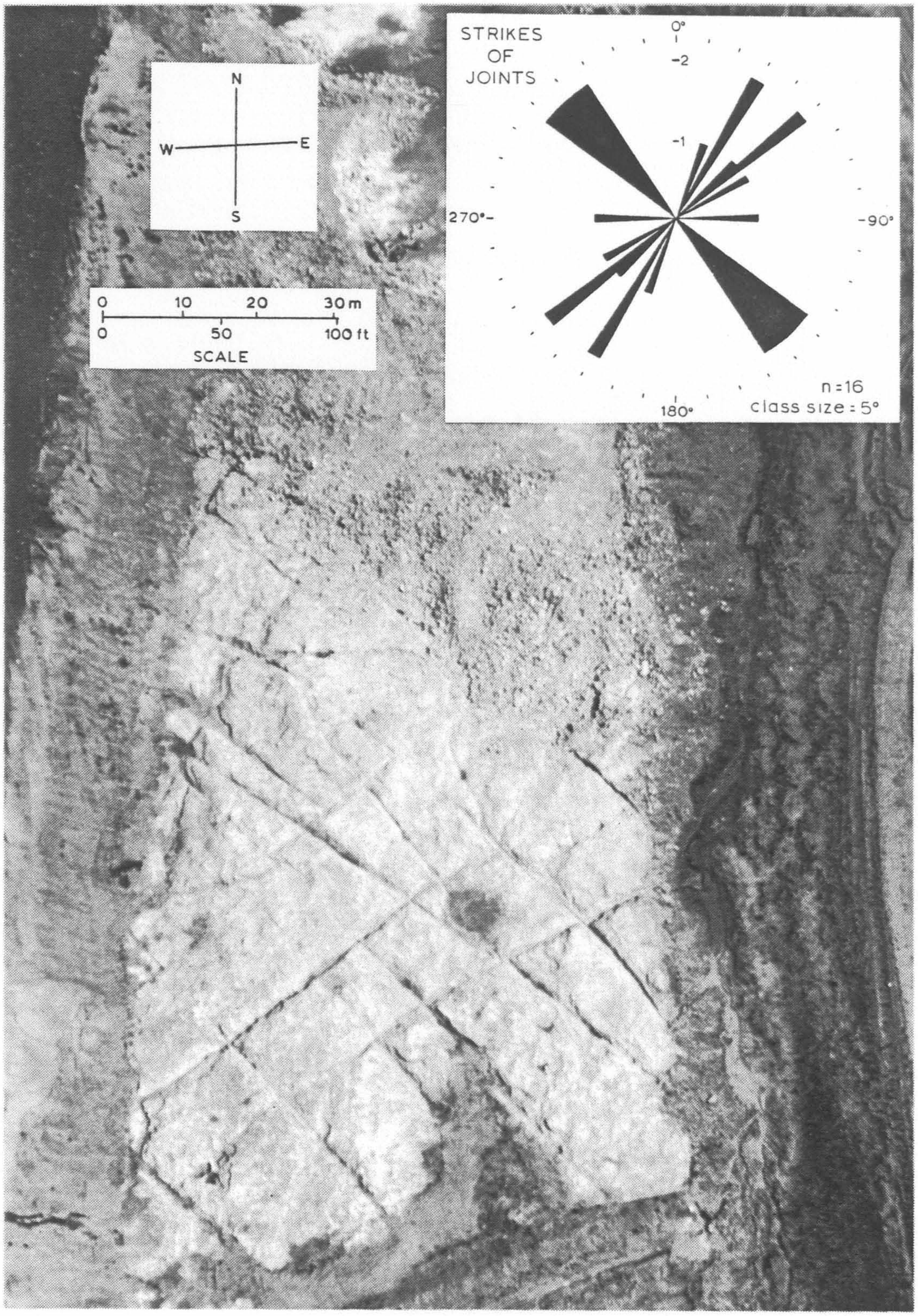
The orientation of joints in the Fort Dodge gypsum was studied quantitatively using aerial photographs. This method was effective for four reasons. (1) The solution channels that follow the joints are easily seen from the air, but are difficult to measure on the ground without an aerial photograph to help monitor the progress of the survey. (2) Except for a minor set of horizontal joints which will be discussed later in this paper, all of the joints are vertical or nearly vertical, so one need only measure the strike of a joint to get a good idea of its orientation. (3) The quarrying method used by some of the gypsum companies involves scraping the overburden off of large expanses of gypsum before blasting begins, thus exposing many joints at once. (4) The top surface of the gypsum is relatively planar. This enables a correction for parallax distortion to be applied to the aerial photograph in which strikes are to be measured. Without this correction, any departure from vertical of the camera angle will cause an apparent rotation of the joint traces in the final photograph, resulting in an inaccurate survey. The correction technique, described in Appendix A, produces a second-generation photograph in which the strikes of the joint traces have been restored to their true orientation and can be directly measured with a protractor.

Joint Orientations

Plate I is an aerial photograph of the North Welles quarry, accompanied by a rose diagram of joint orientations. Joints in this quarry exhibit an approximately orthogonal distribution, with sets

Plate I. Aerial photograph of the North Welles quarry, U.S. Gypsum Company

The rose diagram shows joint orientations taken from a parallax-corrected version of this photograph, as discussed in Appendix A


N
W
E
S
0
10
20
30 m
0
50
100 ft
SCALE
STRIKES
OF
JOINTS
0°
-2
-1
270°
90°
180°
n=16
class size = 5°

trending at about N 40° E and N 43° W. These two sets are believed to have formed contemporaneously, as shown by their intersecting, non-abutting relationship and similar development of solution channels.

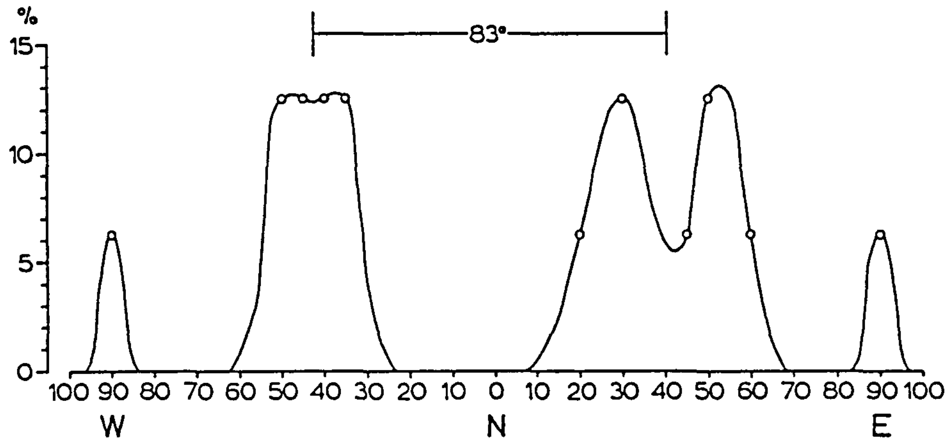
The data are replotted in Fig. 2A, using a weighted average method to locate the most common orientation (peak) within a set. This was accomplished by drawing a smooth curve to connect the data points (effectively introducing a continuously variable class size) and using the values along the curve to weight each orientation. The average orientation of each set was then calculated using the weighted values of the orientations within that set. This method is more accurate than picking the average orientation from a rose diagram by eye, because it allows the local skewness and kurtosis of a set to help define its average. Greater accuracy makes angles between the sets inherently more accurate as well, and aids interpretation.

Plate II shows the Celotex quarry and its joint orientations. An extra correction was performed on the data for the Celotex quarry (in addition to the parallax correction discussed in Appendix A) because an elongate outcrop shape, like that of the Celotex quarry, can bias the rose diagram. This bias is similar to that identified by Terzaghi (1965), whereby a scan line (or elongate outcrop) will cross many joints perpendicular to it, but will cross few that are nearly parallel to it, even though both sets may have equal frequency. The correction was made by dividing the quarry into two equidimensional halves, and the joint data from each half were added together. In this procedure, some of the longer joints (parallel to the long axis of the outcrop) are counted twice. This

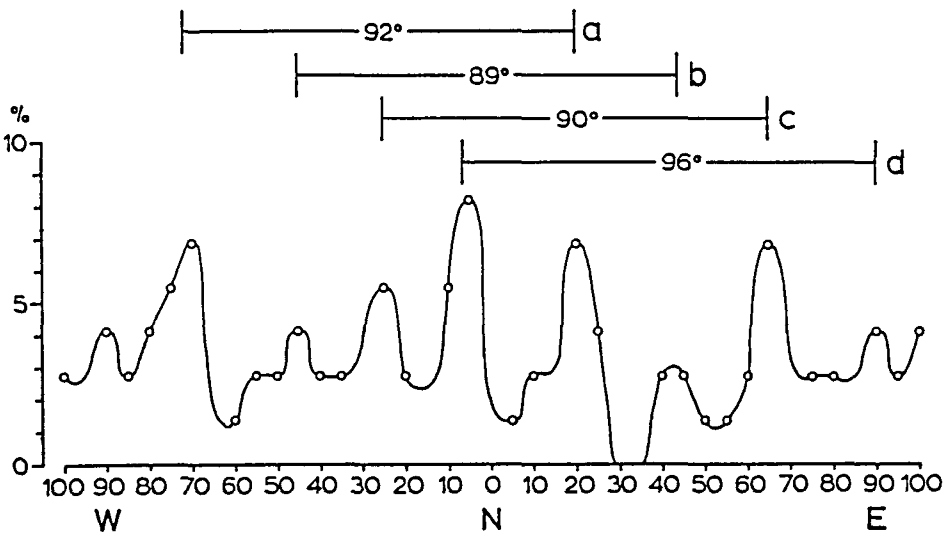
Figure 2. Joint orientation plots of the three quarries

1: North Welles quarry
2: Celotex quarry
3: Carbon quarry

1


83°
%
15
10
5
0
100 90 80 70 60 50 40 30 20 10 0 10 20 30 40 50 60 70 80 90 100
W
N
E

2


92° a
89° b
90° c
96° d
%
10
5
0
100 90 80 70 60 50 40 30 20 10 0 10 20 30 40 50 60 70 80 90 100
W
N
E

3

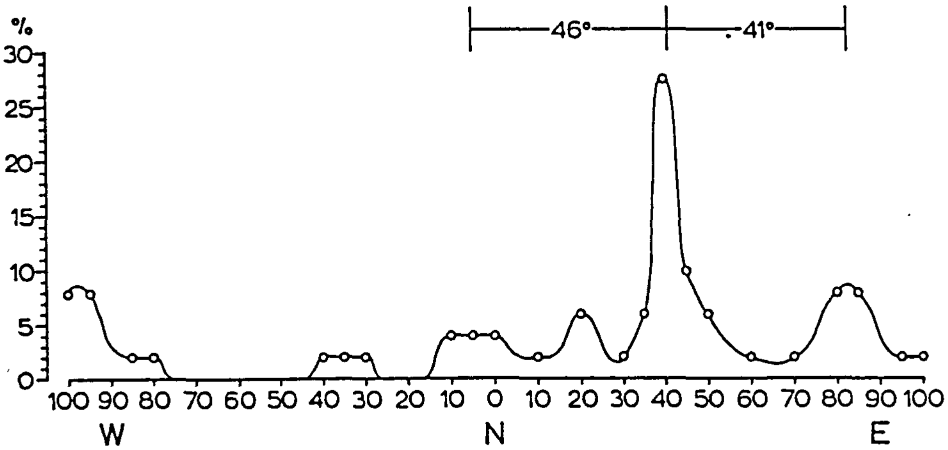
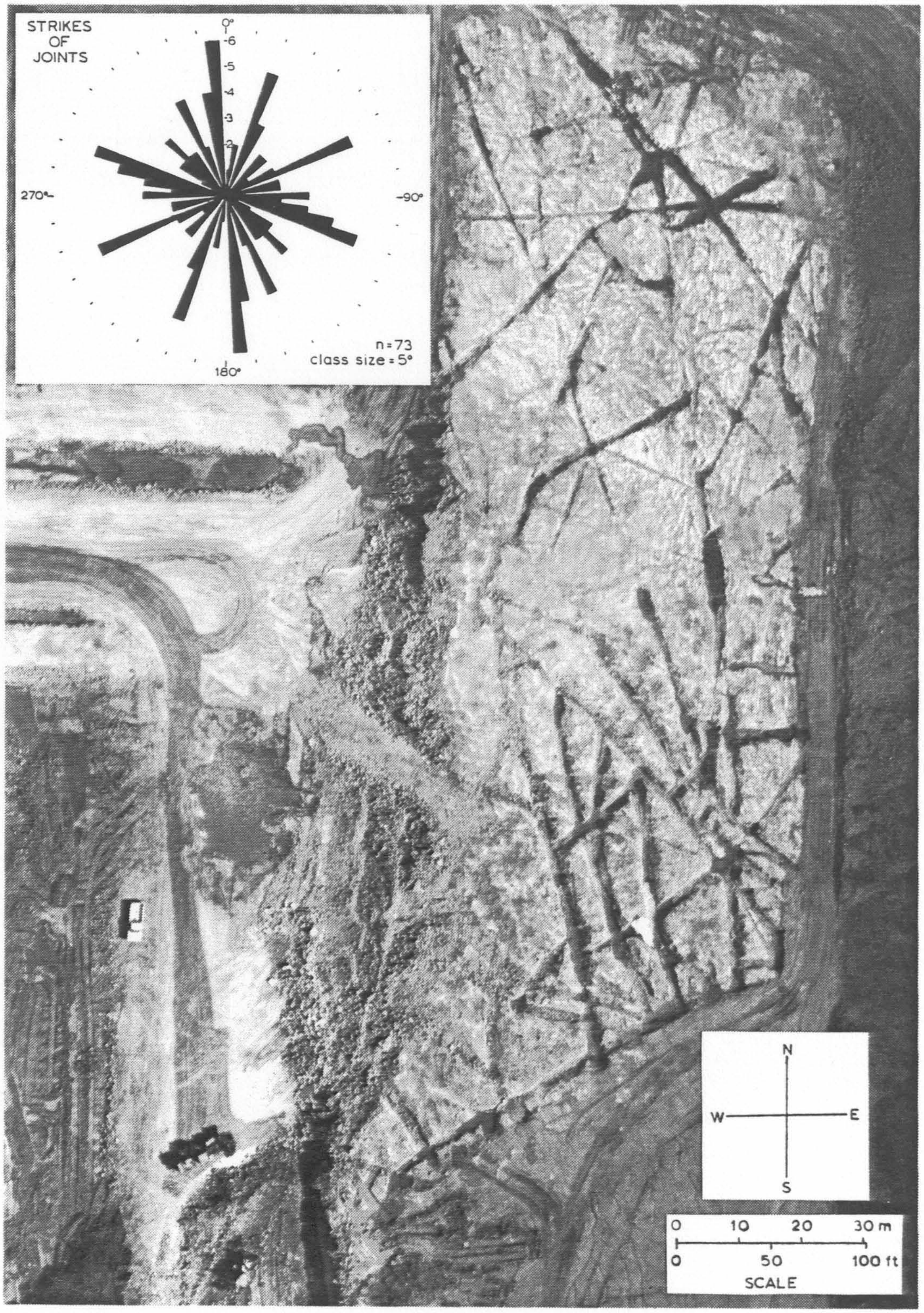

46°
41°
%
30
25
20
15
10
5
0
100 90 80 70 60 50 40 30 20 10 0 10 20 30 40 50 60 70 80 90 100
W
N
E

Plate II. Aerial photograph of the Celotex Corporation quarry

The rose diagram shows joint orientations taken from a parallax-corrected version of this photograph, as discussed in Appendix A


STRIKES
OF
JOINTS
0°
6
5
4
3
2
270°
90°
180°
n=73
class size=5°
N
W
E
S
0
10
20
30 m
0
50
100 ft
SCALE

is one way of weighting the joints by their length so as to eliminate the bias due to outcrop shape. The other two quarries' exposures (North Welles and Carbon) were equidimensional enough to preclude the need for this correction.

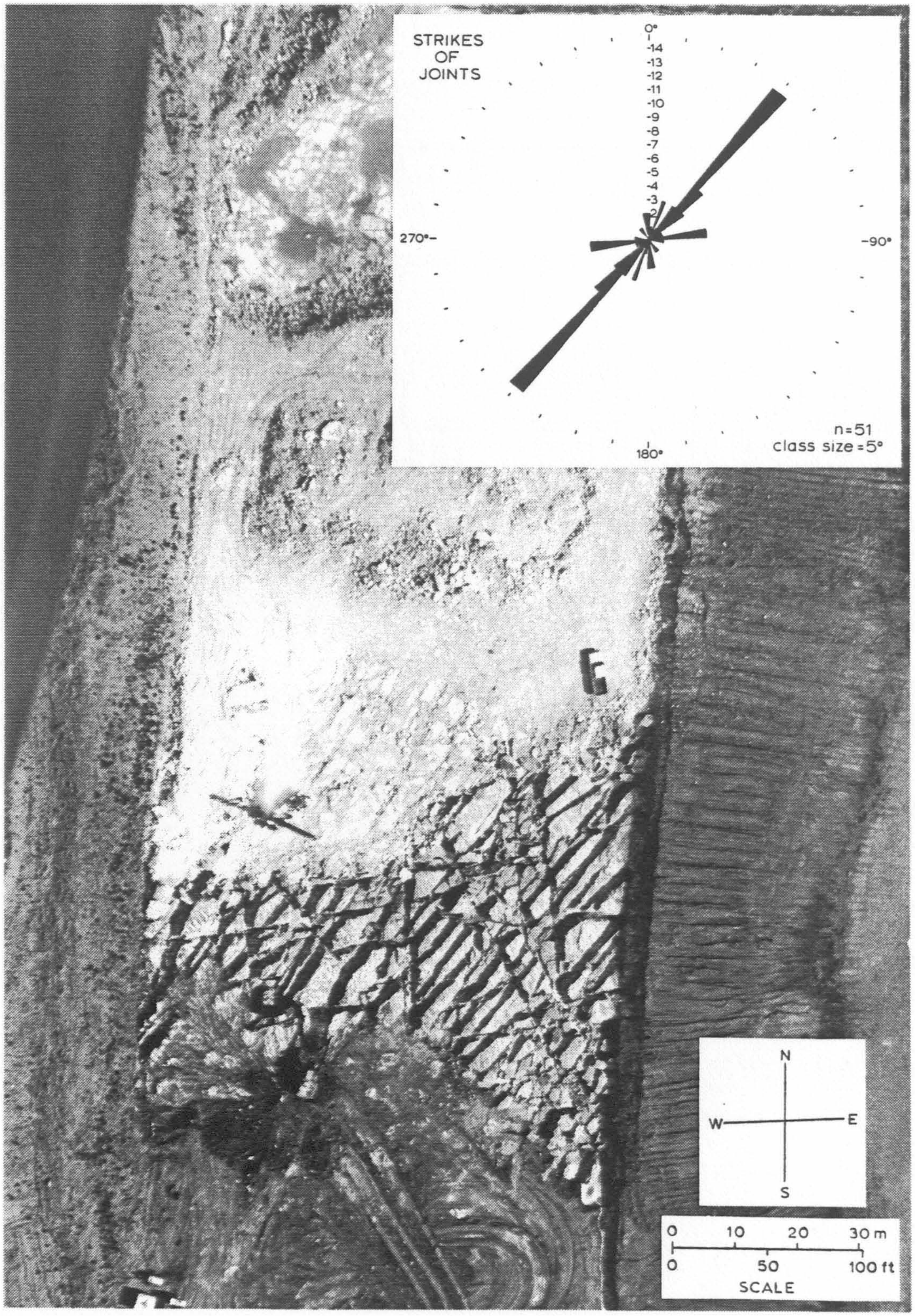
The abutting relationships between joints in the Celotex quarry indicate that not all the joints formed at once. The joint distribution appears to be random, but several orthogonal sets can be tentatively identified (Fig. 2-2). Whether the distribution is multi-orthogonal or random is discussed in a later section.

Plate III is an aerial photograph of the Carbon quarry with its rose diagram of joint orientations. The joints were probably contemporaneously formed, shown by their crossing, non-abutting relationships and similar development of solution channels. Orientations are replotted in Fig. 2-3. The Carbon quarry has a dominant joint set at N 41° E and subsidiary sets at N 83° E and N 5° W. The subsidiary sets are about 45° to either side of the major set and about 90° to each other.

A minor set of horizontal joints was found in the North Welles quarry. These joints occur along laminations with excess detrital material and decapitate some of the small topographic highs on the gypsum's slightly undulating upper surface. Two of the horizontal joints in the North Welles quarry could technically be called faults because they show signs of slight movement in the form of accretion steps, which are a chemical depositional feature found along some faults indicating direction of relative movement during their deposition (Hancock, 1985).

Plate III. Aerial photograph of the Carbon quarry, U.S. Gypsum Company

The rose diagram shows joint orientations taken from a parallax-corrected version of this photograph, as discussed in Appendix A


STRIKES
OF
JOINTS
0°
-14
-13
-12
-11
-10
-9
-8
-7
-6
-5
-4
-3
-2
270°
-90°
180°
n=51
class size=5°
N
W
E
S
0
10
20
30 m
0
50
100 ft
SCALE

Origin and Timing of Joints

North Welles and Celotex quarries

As a first approach, one can attempt to classify the joints in the Fort Dodge gypsum as extensional, shear, or hybrid, based on their angular relationship to the stress field which caused them. Fig. 3 shows the different stress conditions thought to be responsible for shear and extension fractures. The vertical joints of the Fort Dodge Formation most closely resemble the fractures in Fig. 3B, which are caused by a stress field with a horizontally directed, tensile σ_2 and σ_3. The vertical line formed by the intersection of the joint sets defines the σ_1 orientation, and the joints would be classified as extensional.

What is the origin of such stresses? The study area is located in the Interior Province of the Central Stable Region (Eardley, 1962). Orogenic deformation plays a limited role here, so the most probable cause of the orthogonal North Welles and random Celotex joints is epeirogeny from loading, unloading, or both, as a result of sediment deposition and erosion or glacial advance and retreat. Extension joints such as those in Fig. 3B can be created by loading if σ_1 is vertical and the sides of the block are unconfined. The sides of the gypsum were probably confined in the loading phase, however, because the substantial ductility of the overlying sediments, tills, and/or ice, would favor hydrostatic stress conditions that would confine the gypsum on the sides, as well as on top. Such confinement during loading would cause shear fractures to develop rather than extension fractures (Hancock, 1985). The lack of shear fractures in the North Welles and Celotex quarries reduces the possibility

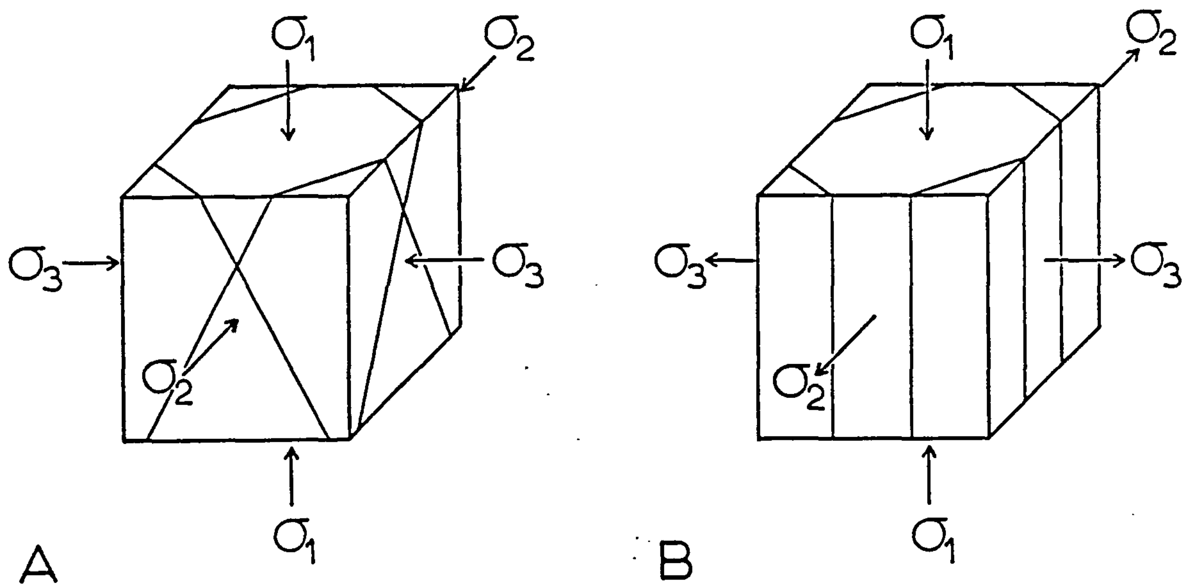


Figure 3. Diagrammatic fracturing of rock. (A: Shear fractures formed in general (triaxial) strain (after Reches, 1983). This mode of failure is favored when all stresses are compressive, the deviatoric stress (σ_1 minus σ_3) is high, and the average pressure is high (Hancock, 1985). B: Extension fractures. This mode of failure is favored when σ_2 and σ_3 are negative (tensile) and the deviatoric stress and average pressure are low (Hancock, 1985). The extension fracture model most closely fits the distribution of the Fort Dodge joints)

that loading caused the joints in the gypsum. The remaining option generally invoked to to explain vertical orthogonal and random joints is that of unloading (Price, 1966). A diagram of how this situation comes about is shown in Fig. 4. As overburden is removed by erosion or ablation, the rock mass rises isostatically and must cover more area at this new, larger radius position within the earth.

Fig. 4 also illustrates how far a rock unit needs to rise in the earth in order to make joints. In the North Welles quarry, for example, the joints are about 10 m apart and about 0.5 mm wide, so L in Fig. 4 is 10 and dL is 0.0005. Although isostatic rebound occurs by upward flexure with a radius of curvature much smaller than the radius of the earth, using the radius of the earth (6378 km) for R in the diagram will give us a maximum of how far the rock mass has risen. Solving for dR yields 320 m. The amount of overburden that would have to be removed to cause this amount of uplift, at equilibrium conditions, can be calculated from the following formula, derived from Archimedes' principle:

$$E = U \times \frac{d_m}{d_e}$$

E is the thickness of overlying material being removed by erosion, U is the amount of vertical uplift, d_m is the density of the supporting medium at the compensation depth (the mantle-- 3.3 g/cm^3, Ahnert, 1970), and d_e is the density of the material being removed by erosion. If the material being eroded is rock or sediment, then d_e is approximately 2.4 g/cm^3, and 440 m of strata must be eroded to cause a rebound of 320 m. If we assume that this overburden was due to Cretaceous (or other upper Jurassic)

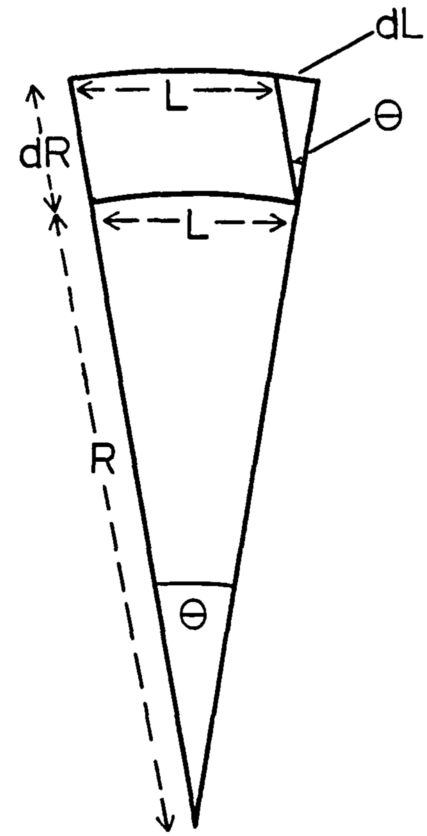


Figure 4. Conditions causing the formation of orthogonally and randomly striking vertical joints (after Price, 1966). (A bed of length L is extended to length L + dL after an uplift through distance dR)

sediments that have since been removed by erosion, then they need to have been approximately 440 m thick. The existence of this much Cretaceous strata in central Iowa has never been proven, but is a possibility since there is an unconformity between the Soldier Creek and the overlying glacial till, and the Tertiary period in Iowa was characterized by 65 million years of erosion (Anderson, 1983). If this was the case, the Fort Dodge gypsum probably also underwent extensive erosion during this time, and the gypsum that we see today may be a small remnant of a deposit that was very large at one time.

Alternatively, the major overburden may have consisted of ice rather than rock. We can consider this option because extensive Pleistocene glaciations have occurred in the area (Flint, 1971) (Fig. 5). If we use the density of ice, 0.92 g/cm^3, for d_e in the above equation, it follows that 1150 m of ice must be removed to get 320 m of rebound. The existence of this much ice during the Pleistocene can be confirmed by another, independent estimate. The thickness of an ice sheet can be roughly calculated from its lateral extent by the following equation, modified from Paterson (1981):

$$H = 3.3\ L^{1/2}$$

H is the maximum thickness of the ice sheet, and L is the lateral distance from the center of the sheet to the edge. The equation was derived by assuming a parabolic profile for the ice and a yield shear stress of 50 kPa (0.5 bar). An estimate of L for the above equation can be obtained from Fig. 5. The nomenclature and timing of the classical glacial

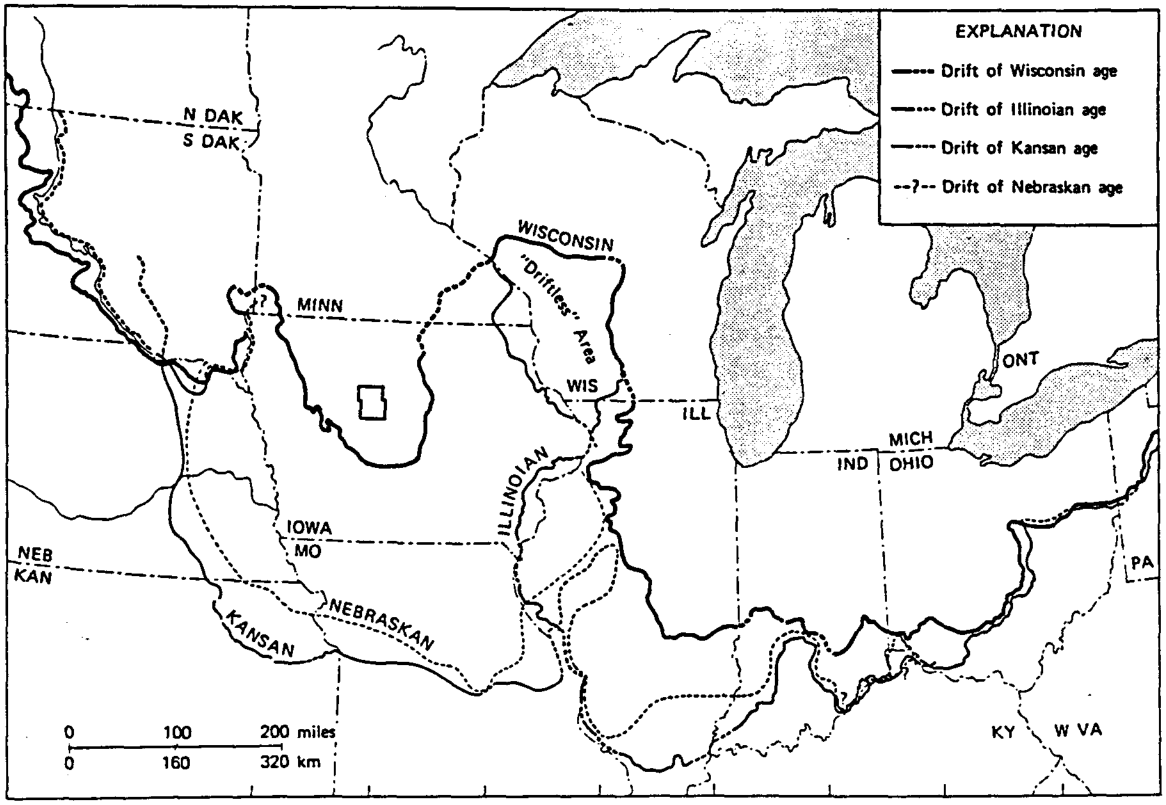


Figure 5. Extent of Pleistocene glaciation in the central United States (after Flint, 1971). (The outline of Webster County is shown in northwest central Iowa)

episodes are under some debate at the moment (Boellstorff, 1978), but the map is sufficiently accurate to give us an idea of the magnitude of the parameters involved. If L is approximately 275 km for the (classical) Nebraskan ice sheet at its maximum extent, H turns out to be 1730 m. This is more than the 1150 m of ice that we require, and shows that there was enough ice to cause a considerable amount of post-glacial isostatic rebound when it retreated.

An interesting side issue at this point is that the above estimates of ice and strata thicknesses are useful in deciding if the Fort Dodge Formation has been under enough pressure to cause the gypsum to dehydrate to anhydrite. Dehydration is favored at high pressures, high temperatures, low pore pressures, and high salinities (MacDonald, 1953; Hardie, 1967; Berner, 1971). These relationships are shown in Fig. 6. Observations by Blatt et al. (1972), also plotted in Fig. 6, indicate that the normal pressure-temperature gradient in the earth causes gypsum to dehydrate at about 305 to 610 m (1000-2000 ft.), so about this much Cretaceous overburden is required to dehydrate the gypsum. If we assume that the overburden was ice rather than rock, the argument is not as straightforward because the tendency of ice pressure to dehydrate the gypsum is countered by the lower temperatures involved. The pressure from 1730 m of ice is equal to about 660 m of strata (157 bars). Although this is well into the depths at which dehydration commonly takes place in the earth according to Blatt et al. (1972), heat from the normal geothermal gradient isn't there to help drive the reaction. At the bottom of a glacier, the temperature must be well below $0^{o}C$, or pressure would melt

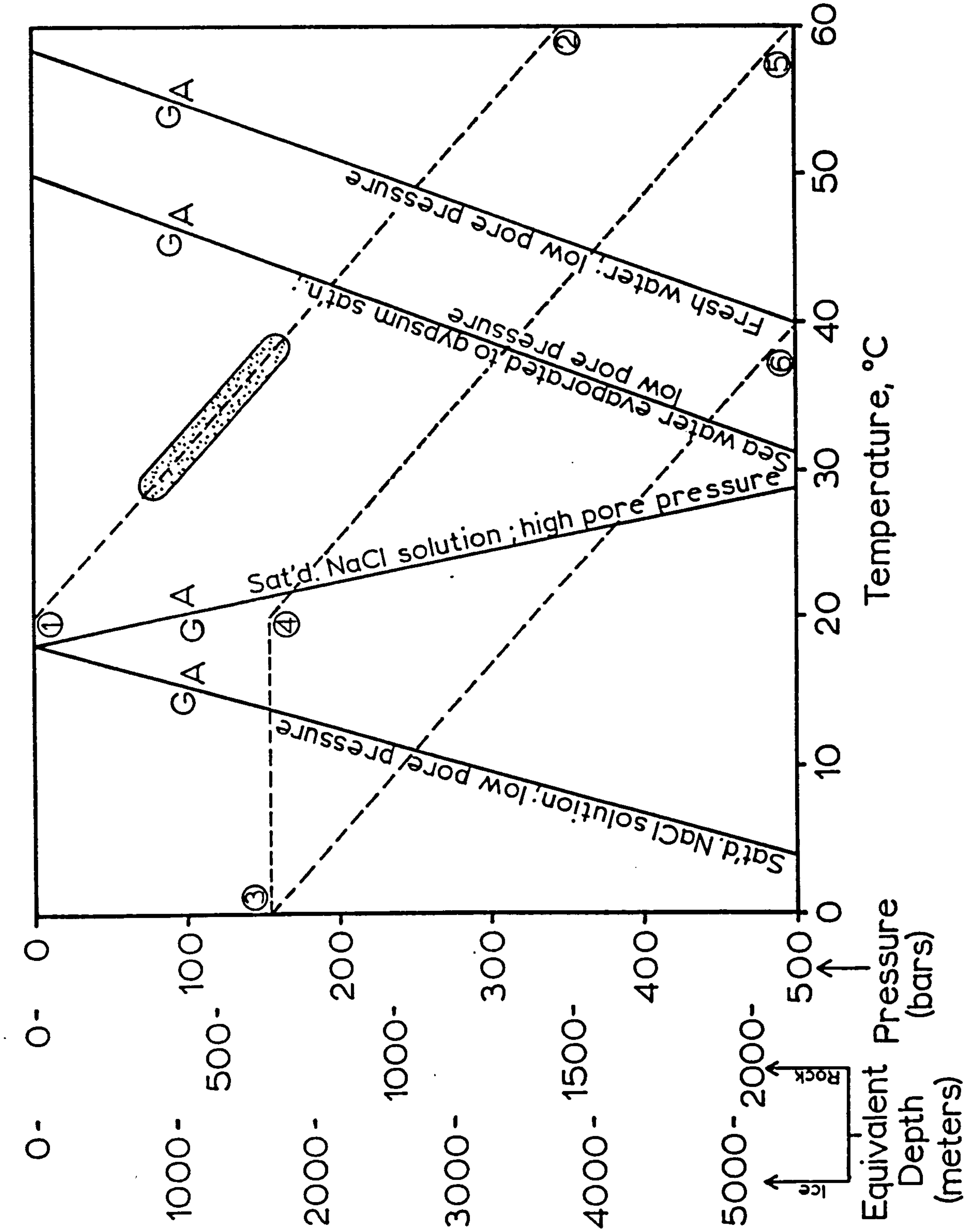
Figure 6. Gypsum-anhydrite equilibrium relations (after Berner, 1971)

G: gypsum is thermodynamically stable
A: anhydrite is thermodynamically stable

Stippled area indicates conditions at which gypsum is commonly dehydrated to anhydrite in the earth's crust according to Blatt et al. (1972)

Line 1-2: Normal pressure-temperature gradient in the earth.
Line 3-4-5: Pressure-temperature gradient under a glacier 1730 m thick, assuming instantaneous glaciation (no effect on the geothermal gradient)
Line 3-6: Pressure-temperature gradient under a glacier 1730 m thick, assuming the glacier has been there forever (total readjustment of the geothermal gradient)

The actual pressure-temperature gradient beneath a glacier will plot between lines 3-4-5 and 3-6. The gradient will plot closer to line 3-6 for an ice age of 10,000 years or more.



Fresh water; low pore pressure
Sea water evaporated to gypsum sat'n.; low pore pressure
Sat'd. NaCl solution; high pore pressure
Sat'd. NaCl solution; low pore pressure
G A
G A
G A
G A
1
2
3
4
5
6
0
10
20
30
40
50
60
Temperature, °C
0
100
200
300
400
500
Pressure (bars)
0
500
1000
1500
2000
0
1000
2000
3000
4000
5000
Rock
Ice
Equivalent Depth (meters)

the ice. Point 3 in Fig. 6 shows that at $0^{o}C$ and 157 bars, gypsum will not be dehydrated, even if pore pressure is low (open system-- hydrostatic pressure only) and enclosing solutions are saturated in NaCl. The only way that the Fort Dodge gypsum could have been partially dehydrated by ice burial is if there were enough strata overlying it to provide insulation from the ice, a higher geothermal gradient, and more overburden pressure. There are several problems with this. Even if there was a great thickness of Cretaceous strata over the gypsum at one time, most of it would have to have been eroded by the end of the Tertiary period because there is little time available to erode it afterwards. Also, the insulating ability of the strata turns out to be negligible because the cooling effect of long-wave climatic cycles, such as ice ages, penetrates fairly deeply into the earth's crust-- much more so than short wave cycles such as diurnal fluctuations. Cooling cycles on the order of 10,000 years penetrate about 2000 m into the earth before they are damped out (Garland, 1979). As a result, the geothermal gradient in the first 1000 m below a glacier will be radically altered toward lower temperatures, and the pressure-temperature gradient will be similar to that shown by line 3-6 in Fig. 6. We can follow this line to the first reaction curve to determine the first possible point at which dehydration can take place if optimum conditions are met. This occurs at a depth 390 m below the glacier (390 m below line 3-4, read off the equivalent depth rock scale), so at least this much strata is required between the glacier and the gypsum for dehydration to occur. Not only is this amount unlikely to have been there during the Pleistocene, but much more would be required if surrounding solutions were not saturated in NaCl or pore pressure was greater than just hydrostatic pressure. Observations by Blatt et al. (1972) (Fig. 6) indicate that optimum conditions for dehydration are not usually met. In conclusion, the glaciers did not help dehydrate the gypsum, even if there was an unlikely amount of strata between the ice and the gypsum. The only mechanism that could have dehydrated the gypsum is the existence of 300 to 400 m or more of post-Jurassic strata. Evidence for this has been reported by Dasenbrock (1984), who reported seeing boxy ghosts of anhydrite in some alabastrine gypsum nodules in the Fort Dodge gypsum, indicating that local conditions once favored partial dehydration.

Immediately after each ice sheet retreated, the gypsum would be subjected to a stress field with horizontal, compressive σ_1 and σ_2, and a vertical, compressive σ_3. Such a stress field was proposed by Dasenbrock (1984) to explain the vertical extension of some of the gypsum layers, which was accompanied by formation of horizontal veins with vertical fibers late in the diagenetic history of the deposit. As rebound in the area progressed, however, the σ_1 and σ_2 would diminish and eventually become the tensile σ_2 and σ_3 possibly responsible for joint formation. Glacial ablation could thus be responsible for both of the stress fields. Erosion of strata, however, would probably bypass the first stress field mentioned above (vertical σ_3) and just induce the second configuration (horizontal tension) because crustal rebound would be able to keep up with erosion, which would be much slower than ablation.

In proposing post-glacial rebound as a jointing mechanism, one must also take into account the dynamic aspects of rebound. Has there been enough time since the major glaciations for rebound to occur? Models and case studies indicate that post-glacial rebound is relatively rapid. The mouth of the Angerman River, Sweden, for example, has experienced 270 m of uplift (90% recovery from its initial 300 meter downwarping) in the last 10,000 years (Turcotte and Schubert, 1982). This suggests that the Fort Dodge area, which had its last glacial retreat about 12,000 years ago (Nilsson, 1983), has also had plenty of time for rebound to take place. It also suggests that there was enough time for complete rebound between the North American glacial episodes as well, as interglacial periods lasted on the order of 200,000 to 400,000 years each (Nilsson, 1983).

To summarize, the Fort Dodge gypsum probably underwent compression during the Nebraskan glaciation, expansion as a result of unloading during the Aftonian interglacial period, renewed compression during the Kansan glaciation, and expansion again during the Yarmouth-Sangamon interglacial period. The Illinois glaciation did not reach the Fort Dodge area, and the Wisconsin glaciation was not as extensive as the Nebraskan and Kansan, so its effect on the crust was relatively small. Boellstorff (1978) has discovered another till under the Nebraskan at its type section, so there may have been another, earlier loading-unloading cycle as well. The specific dates of the earliest glaciations will be discussed in the section on solution channels and dissolution rates.

Repeated glacial loading and unloading would not necessarily create new joint sets every time, because the older, original sets could accomodate new stresses. This probably happened in the North Welles quarry, as indicated by its single orthogonal system. In the Celotex quarry, however, repeated glaciations and rebounds may have produced multiple joint sets. This would partially explain why the linear joint frequency in the Celotex quarry is four times that of the North Welles quarry. Increases in joint frequency can sometimes be explained by a proportional drop in bed thickness (Price, 1966), but are ruled out in this case because the gypsum beds of the North Welles and Celotex quarries are of approximately equal thickness. Differential uplift as a cause for joint frequency variations is also unlikely, as both quarries are at similar elevations.

The question still remains as to whether the joint sets of the Celotex quarry are random or multi-orthogonal. The possibility of repeated phases of jointing is of little help, because repeated phases can create either random or multi-orthogonal patterns. The apparently equal spacing and nearly perfect right angle relationships of many of the joints sets (see Fig. 2-2) lend credibility to a multi-orthogonal distribution, but even random distributions have local, spurious concentrations that can be misleading. The following argument indicates that the temporal evidence between the proposed multi-orthogonal joint sets is contradictory, and that the safest conclusion to be drawn is that the joint distribution of the Celotex quarry is not multi-orthogonal, but random.

If the joint sets are truly multi-orthogonal, it should be possible to determine temporal relationships between the sets. Joint sets formed later should not be as well-developed as earlier sets (Price, 1966) because any initially large tensile stresses would have been relieved by the earlier joints. This, coupled with the fact that later orthogonal sets tend to bisect the angles between previous orthogonal sets, should enable us to determine the order in which the sets occurred. Application of the above logic to the data in Fig. 2-2, however, does not lead to definitive conclusions. The least well-developed set is set 'b', so it formed last. The 'd' set cannot be first or second because it is immediately followed by set 'b' (using bisecting relationships). This leaves a-c-d-b or c-a-d-b as possible permutations. Unfortunately, abutting relationships shown in Plate II tell a different story. Some of the joints from both 'a' and 'c' sets abut against other joints, and were thus formed later, and other joints abut against them. Part of the problem is that neither joint frequency nor abutting relationships are completely reliable as temporal indicators. Although the abutting relationship is accurate for the two joints involved, application to joint sets can be ambiguous (Hancock, 1985). Even if one set occurs first, residual stresses can cause late joint additions to the first set which abut against joints of the second or third set. These latecomers may locally have abutting relationships with the other sets that define a circular temporal order, such as a-b-c-a. To sum up, there is no conclusive temporal evidence in the Celotex quarry to suggest the occurrence of distinct, sequentially occurring orthogonal sets.

One possible cause for this apparently random pattern may be the modification of the ordinarily uniform stress field by variations in the compactibility of the underlying beds. The underlying Cherokee group has sandstone and limestone lenses supported in a shale matrix (Hale, 1955) and must exhibit a high degree of local variation in compaction characteristics. Such variation would favor random joint development, perhaps even initiated during the loading process. An area underlain by a uniform lithology would favor orthogonal development, formed strictly during the unloading process.

Carbon quarry

The vertical orthogonal North Welles and random Celotex joints can be adequately explained by the above mechanism of glacial loading, unloading, and rebound. The parallel joints of the Carbon quarry, however, indicate a different local stress field, and perhaps a different mechanism. Possible explanations include large scale exfoliation and influence from a local structure, such as a fault in the underlying rock formations. In the first case, a situation similar to exfoliation might come about because the Carbon quarry is close to the southeast edge of the gypsum subcrop. This would imply that the main set is extensional in origin and the two subsidiary sets are conjugate shear sets, all caused by a horizontal compressive force operating along N 40° E, parallel to the major joint set. σ_2 would be vertical, and σ_3 would be horizontal and possibly tensile. The fact that the main joint set is semi-parallel to the edge of the gypsum may help this argument, but the fit is not close. It is also unusual to have conjugate shears at 45° to σ_1. Stresses normal to the shear plane usually combine with the internal coefficient of friction to hold the rock together at this angle. Also, there are no hybrid shear joints between the extensional joints and conjugate shears, as would be expected in such a mix.

Possible causes other than exfoliation arise if one looks at nearby structures. The study area is located in the Interior Province of the Central Stable Region (Eardley, 1962), but there are several conspicuous structures in the vicinity which indicate structural activity at one time or another. Nearby structures include the Manson disturbed area, the Fort Dodge graben, and the Midcontinent Geophysical Anomaly, all shown in Fig. 1. The Manson disturbed area is a possible Tertiary meteorite impact site (Anderson, 1983) centered to the west of the area. The Manson disturbed area is probably too far away, however, to have affected the joints in the Carbon quarry while not affecting the other two quarries. The Fort Dodge graben is another structure. It extends through Fort Dodge proper at about N 70^{o} E (Hale, 1955). It too, though, is probably too far from the Carbon quarry to influence it while leaving the other quarries unaffected. The only other significant structures in the area that might have caused the joints in the Carbon quarry are basement faults in the rocks involved with the Midcontinent Geophysical Anomaly (M.G.A.).

The M.G.A. is located in the northern Great Plains of the United States and consists of a large gravity high and a change in the intensity of the ambient magnetic field. The anomaly is caused by a series of dense, magnetized basalt flows of the Keweenawan system (Thiel, 1956) which crop out in northwestern Wisconsin and continue in the subsurface to the southwest, through southeastern Minnesota, Iowa, southeastern Nebraska, and northeastern Kansas. These basalt flows are 1.1 billion years old (Goldich et al., 1966) and are related to rifting activity that occurred during Precambrian time (King and Zietz, 1971). Some of this basalt was encountered at 698 m (2290 ft.) while drilling a well in Fort Dodge (Hale, 1955). The edges of the rift were precisely determined by King and Zietz (1971) using gravity and magnetic data. The M.G.A.'s northwest edge, sometimes called the Northern Boundary Zone, passes just northwest of Fort Dodge, and is shown in Fig. 1. Gravity and aeromagnetic maps of Webster County are shown in Fig. 7 and Fig. 8.

The rift exhibits many faults. It is offset in a few places by transform faults (Chase and Gilmer, 1973), but none of these are in Iowa. In Iowa, the basalt blocks form a horst bounded by high angle faults with throws of about 5800 m (19,000 ft.) (Van Eck et al., 1979). These boundary faults trend approximately N 43° E in Webster County, parallel to the main joint set of the Carbon quarry. Additional faults, parallel to the boundary faults, occur throughout the 60 kilometer width of the rift (Van Eck et al., 1979). One of these northeast-southwest trending interior faults is probably responsible for the parallel joints in the Carbon quarry. The aeromagnetic map in Fig. 8 shows evidence for such a fault near the quarry. The northeast part of the map shows an abrupt, high magnetic gradient, the contours of which trend N 40° E. This gradient probably delineates a basement fault trending northeast-southwest which passes just southeast of the Carbon quarry and continues to the south-southwest. The main joints of the Carbon quarry are probably due to this fault. The Carbon quarry's minor sets at 45° to the main set would

Figure 7. Gravity map of Webster County, Iowa (after King and Zietz, 1971)

Contour interval is 10 milligals. N. B. Z. is the Northern Boundary Zone of the Midcontinent Geophysical Anomaly

1: North Welles quarry
2: Celotex quarry
3: Carbon quarry
4: National quarry

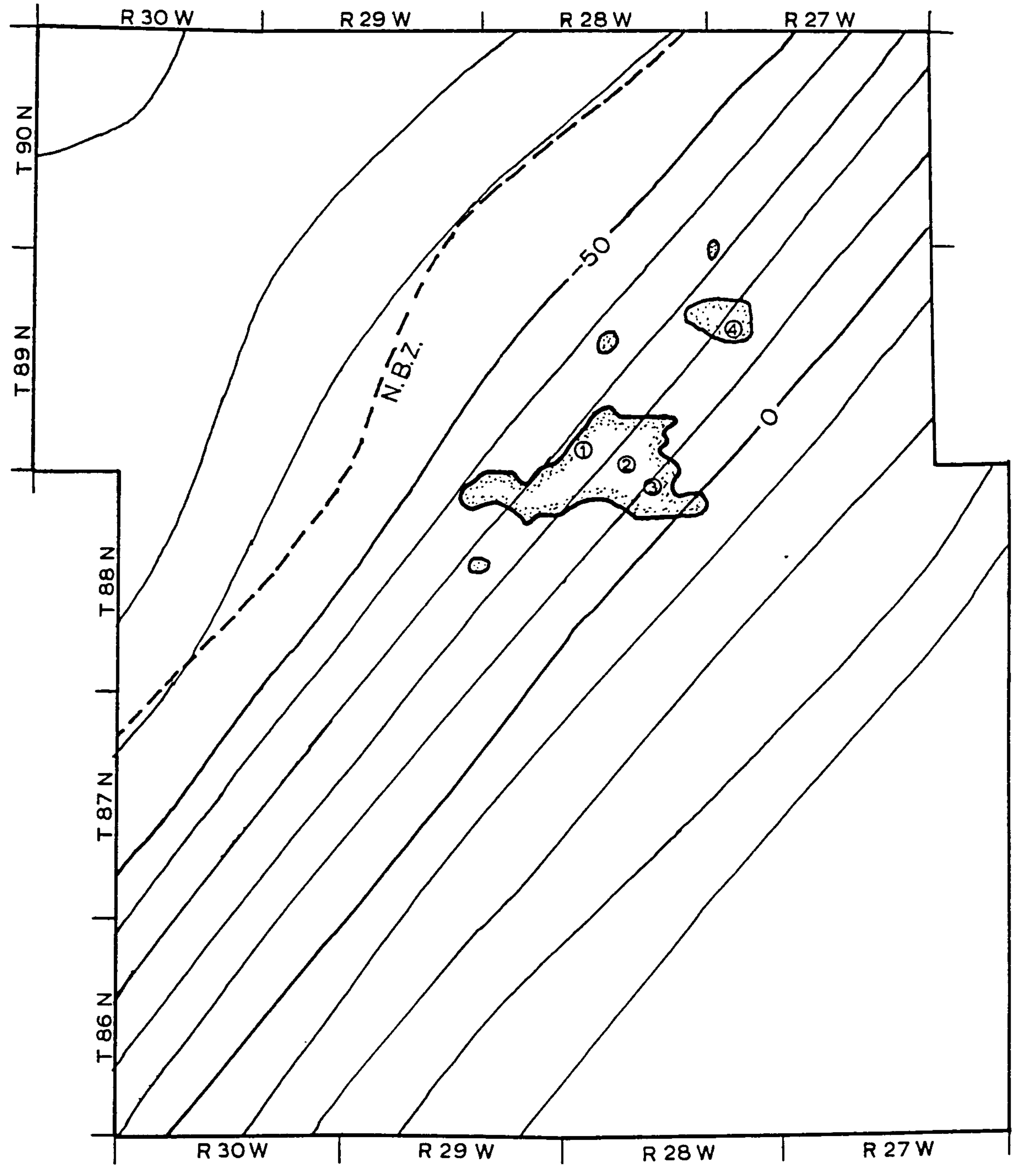
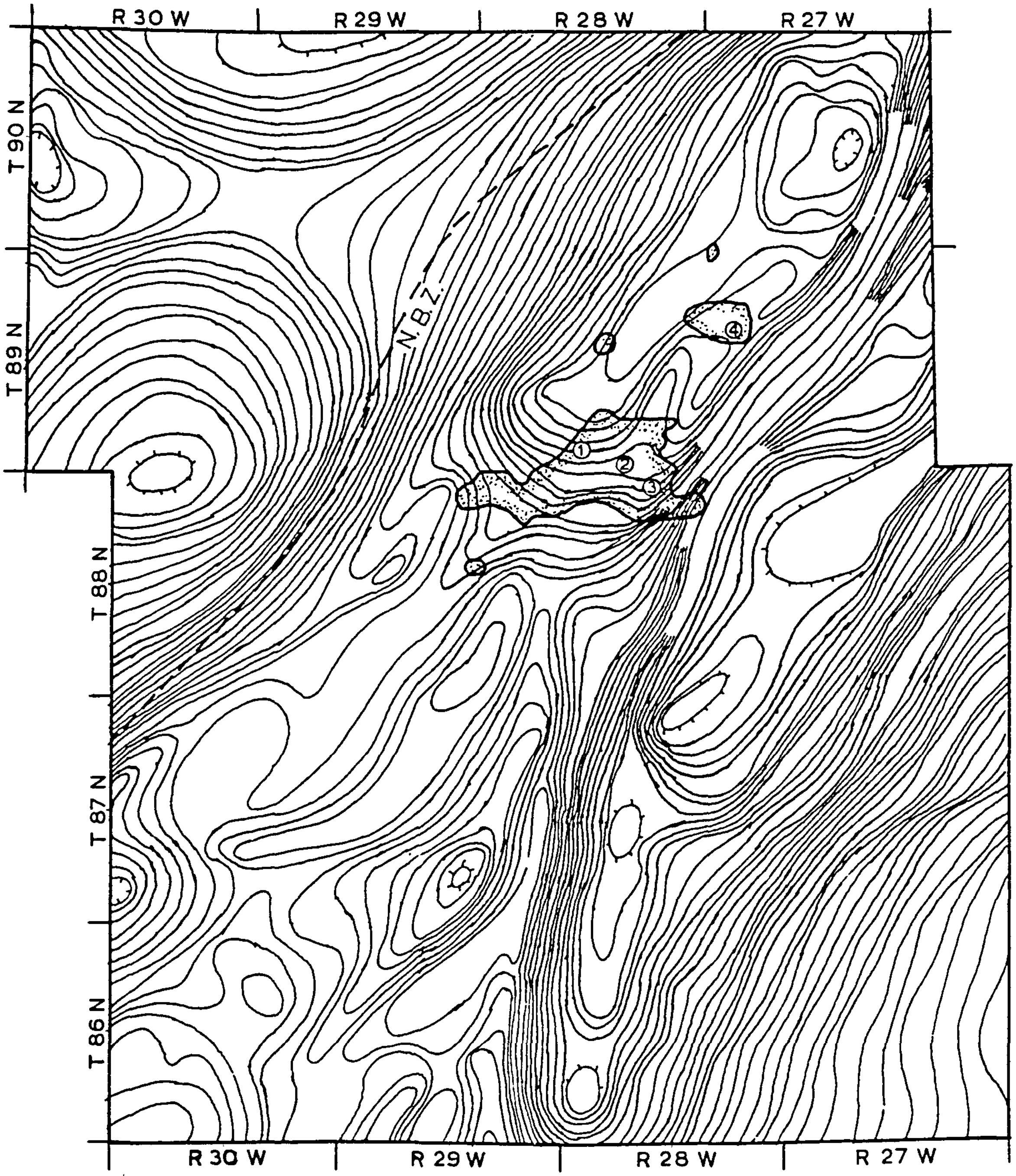

R 30 W
R 29 W
R 28 W
R 27 W
T 90 N
T 89 N
T 88 N
T 87 N
T 86 N
N.B.Z.
50
0
1
2
3
4
R 30 W
R 29 W
R 28 W
R 27 W

Figure 8. Aeromagnetic map of Webster County, Iowa (after Coons et al., 1967; King and Zietz, 1971)

Contour interval is 20 gammas. N. B. Z. is the Northern Boundary Zone of the Midcontinent Geophysical Anomaly

1: North Welles quarry
2: Celotex quarry
3: Carbon quarry
4: National quarry


R 30 W
R 29 W
R 28 W
R 27 W
T 90 N
T 89 N
T 88 N
T 87 N
T 86 N
N.B.Z.
1
2
3
4
R 30 W
R 29 W
R 28 W
R 27 W

then constitute a small orthogonal system developed independently of the major set.

Is there any evidence that this fault has moved since the Fort Dodge gypsum was deposited? Although the rift system is over 1 billion years old, there is ample stratigraphic evidence of movement along the boundary faults, as well as along interior faults, throughout the Paleozoic (Coons et al., 1967). The Thurman-Wilson fault in southwest Iowa, for example, defines the southeast edge of the rift in that area, and was active during the Pennsylvanian (Chamberlain, 1980). The deposition of the Pennsylvanian Cherokee Group in the Fort Dodge area was also largely controlled by the rift structure (Chang, 1984). The Cherokee Group strata are generally the ones that underly the gypsum. Even later, post-Jurassic, structural activity in the Fort Dodge area is confirmed by displacement of the gypsum beds by 15 to 23 m (50 to 75 ft.) along the Fort Dodge Graben (Hale, 1955). Although measurable post-Paleozoic movement along the major faults of the rift structure has never been confirmed, minor adjustment along the interior fault near the Carbon quarry could account for the parallel joint development found there. This adjustment was possibly induced by loading and unloading by Pleistocene glaciers or Cretaceous sediments. It is also possible that post-Jurassic movement along the fault created a structure responsible for protecting the gypsum from more extensive erosion, but this has not been confirmed. The interior fault also has important predictive value for gypsum exploration because of its control on jointing of the gypsum. Any gypsum closer to the fault than the Carbon quarry is likely to be heavily jointed (with N 40° E joints) and mostly or totally dissolved away by increased ground-water flow along the joints. Quarrying this gypsum would probably not be economic, and mining it would be plagued with roof support problems because the gypsum would be in the form of individual remnant blocks such as those found in the Carbon quarry.

Horizontal joints

The horizontal joints in the Fort Dodge gypsum are weakly developed in all quarries, but nonetheless warrant an explanation. They are probably exfoliation features following weaknesses along bedding planes with excess detrital material. The limited lateral extent of these joints indicates that exfoliation occurred after the vertical joint sets were formed, as they can not be traced across the vertical joints. Accretion steps on the faces of two joints in the North Welles quarry indicate 1-2 cm of horizontal movement of the upper blocks, with one moving toward N 12° W and the other toward N 14° W. Although these two directions are remarkably parallel, there are too few horizontal joints in the Fort Dodge Gypsum to fully document a recent systematic horizontal shear stress in the area. Overhanging relationships of the upper blocks on their supports indicate that movement occurred after the solution channels were developed. This suggests that perhaps the local topography of the gypsum surface controlled the movement.

Solution Channels and Dissolution Rates

If Tertiary erosion, rebound, and expansion are responsible for joint formation, then the joints and subsequent solution channels are probably less than 20 million years old. Based on this amount of time, a typical solution channel with a half-width of one meter would have experienced an average dissolution rate of about 1 mm per 20,000 years.

If, on the other hand, the joints have been formed by post-glacial rebound and expansion, they must have been formed since the first major Pleistocene glacial retreat. According to the classical glacial episodes, the first glaciation was the Nebraskan, which has been given an age of 1.0 million years by Boellstorff (1978), and an age of 1.2 million years by Cooke (1973). It was previously mentioned, however, that Boellstorff (1978) has found additional tills below the Nebraskan type section. These are interbedded with volcanic glasses that have been fission-track dated at about 2.2 million years. This limits the age of the (post-glacial) joints and subsequent solution channels to two million years. A typical solution channel with a half-width of one meter thus experienced an average dissolution rate of about one millimeter per 2000 years during this time. At this rate, the entire present thickness of the gypsum (10 m) could be dissolved in 20 million years. This would indicate that much of the of the Fort Dodge Gypsum has been dissolved and is now missing, and that the gypsum deposit must have been much thicker and more laterally extensive in the past. The small, isolated deposits of gypsum shown in Fig. 1 are probably remnants of this once large, thick, continuous deposit.

The solution channels are now filled with sediments from the overlying strata. In the North Welles and National quarries the overlying unit is the Soldier Creek member of the Fort Dodge Formation, and in the Carbon and Celotex quarries it is Pleistocene glacial till. The finer grained fraction of these overlying beds probably helped limit ground-water flow and protected the gypsum from more extensive erosion and development of solution channels. Bard (1982) suggested that the Soldier Creek member was more effective than glacial till in protecting the underlying gypsum, but deep (4 m) solution channels filled with Soldier Creek sediments in the National quarry indicate that the difference between the Soldier Creek and the glacial till is relatively minor compared to other controls, assuming that the solution channels were dissolved by ground-water in post-Soldier Creek time. One such control may be the permeability difference between the underlying Pennsylvanian shales and sandstones of the Cherokee Group. Occasional channel sandstones in the Cherokee Group would be subject to more ground-water flow than the surrounding shales. The greater dissolution of gypsum overlying such areas may be the mechanism responsible for the formation of large, vertical solution cavities locally found in the gypsum.

SECTION 2: INSOLUBLE MINERALS

Samples of the laminated gypsum were taken from the northwest corner of the North Welles quarry every 10 to 20 cm, starting 20 cm from the top and ending 370 cm from the top. This covered most of the gypsum member, which is 4 m thick here but is thicker in other parts of the same quarry. Some features were also sampled, including large and small gypsum nodules and tiers of fibrous gypsum. The nodules are most common in the upper levels of the gypsum and are white, fine-grained (alabastrine) features thought to be secondary growths which push impurities out of the way as they grow. They range in size from a few millimeters to tens of centimeters and have blocky ghosts which suggest an anhydrite precursor, indicating that local dehydration of the gypsum has occurred (Dasenbrock, 1984). The tiers of fibrous gypsum occur to at least some extent throughout the deposit but are most well-developed at the very top, where they were sampled. They are thought to be recrystallization features developed under vertical tensile stress. The long axis of the gypsum crystals in the tiers is perpendicular to the bedding and was parallel to σ_3 at the time of formation. In the case of the Fort Dodge gypsum, the long axes of the crystals do not necessarily coincide with the 'c' crystallographic axis, so the tiers are not true satin spar (Dasenbrock, 1984). The tiers have a brick-red color where they are overlain by the red beds of the Soldier Creek member, presumably because the tiers have incorporated some of the red material of the Soldier Creek into them when they were growing. The red color does not appear anywhere else in the gypsum.

Insoluble minerals make up a very small portion of the Fort Dodge gypsum and are thus difficult to see and identify. To obtain these impurities in reasonable quantities, a distilled water method using soxhlet extractors was used to dissolve away the gypsum and concentrate the insoluble minerals. This procedure, as well as sample preparation techniques used for X-ray diffraction (XRD) and the scanning electron microscope (SEM), is described in detail in Appendix B. Samples were weighed before and after gypsum dissolution to determine the percent of insoluble matter. The vast majority of the material is of very fine sand size (0.0625-0.125 mm) or smaller, and was examined with a binocular microscope (wet sample), XRD, and SEM.

Mineralogy and Morphology

The insoluble minerals identified by XRD consist of quartz, calcite, orthoclase, illite, and kaolinite (Fig. 9). These minerals occur throughout the gypsum, not only at all depths, but in all features except the nodules. Dolomite, hematite, limonite, pyrite, muscovite, and possibly tourmaline were seen under the microscope, but were too rare to be detectable by XRD. Dolomite occurs as disseminated anhedral particles and well-formed rhombohedra. Hematite and limonite occur as yellow-orange coatings on many of the quartz grains and as yellow-orange amorphous globs. Some of the larger globs (0.1 mm) have pyrite cores with a grainy, multi-colored metallic luster. The cores were more easily seen under the binocular microscope after the samples had been boiled in concentrated hydrochloric acid in preparation for SEM examination. Muscovite was also

Figure 9. X-ray diffraction data (molybdenum radiation)

Mineral peaks identified: Q = quartz, C = calcite, O = orthoclase, K = kaolinite, I = illite

A: Impurities in the laminated gypsum. Sample taken 320 cm from the top of the gypsum

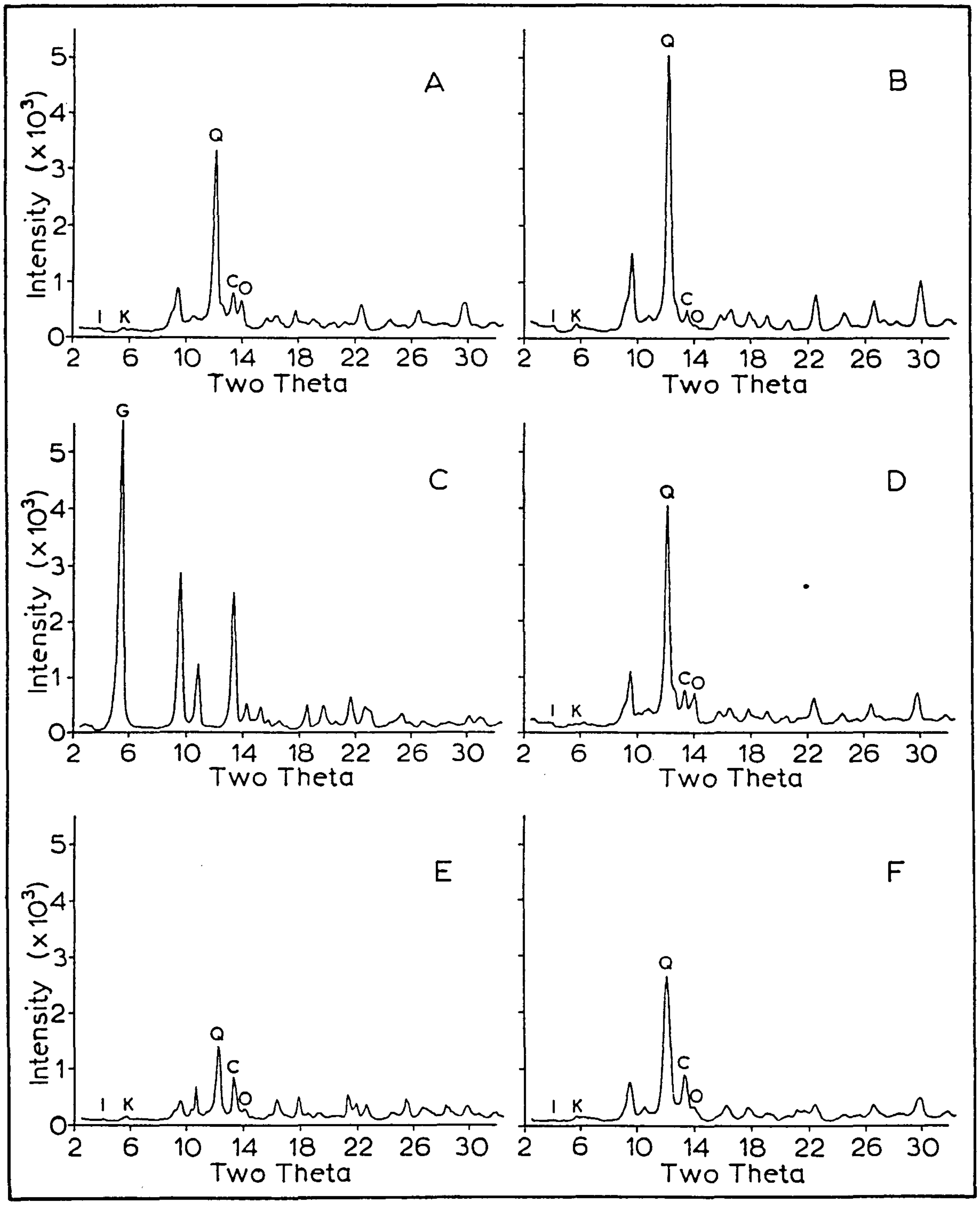
B: Soldier Creek sediments. Undissolved and untreated

C: Undissolved gypsum nodule. All peaks are from gypsum, indicating a very mineralogically pure feature

D: Impurities in the red tiered gypsum. Sample taken from the top of the gypsum. Compare with the overlying Soldier Creek sediments (B)

E: Impurities in the basal layer

F: Impurities in the tan tiered gypsum


A
B
C
D
E
F
Q
C
O
G
I
K
Intensity (x10^3)
Two Theta
0
1
2
3
4
5
2
6
10
14
18
22
26
30

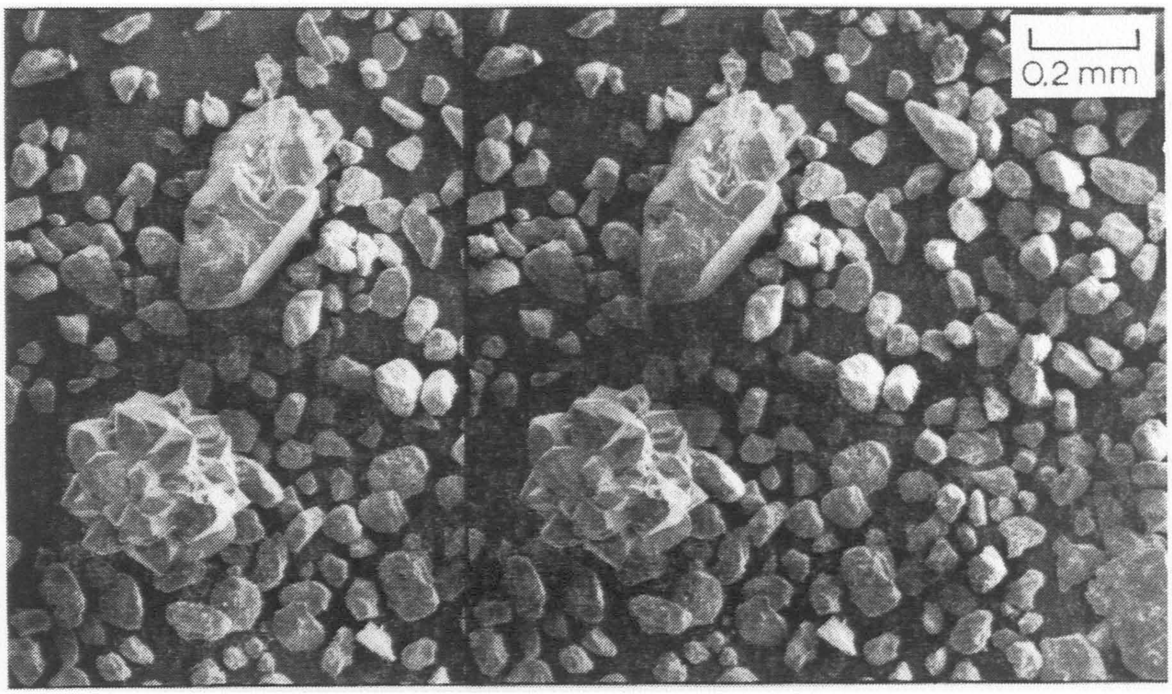
common, seen as tabular transparent grains exhibiting perfect cleavage when broken. Rare elongate crystals with rectangular outlines were tentatively identified as tourmaline. All of the above minerals except pyrite and limonite were previously detected by Dasenbrock (1984). XRD and microscopic examination were not able to confirm the presence of zircon, previously found by Dasenbrock (1984).

Quartz is by far the most abundant impurity, and has three distinct morphologies in the gypsum (Plate IV). Most of it occurs as very fine sand or silt-sized particles less than 0.1 mm across. These show poor sorting and a wide variety of roundness, ranging from angular to well-rounded. Most of them are frosted, and about 30 to 50% are stained yellow to orange, presumably from limonite or hematite. They occur throughout the vertical range of the deposit and, when concentrated along bedding planes, have the appearance of clay. This aspect has caused some researchers to mistakenly refer to them as clay minerals. Another quartz morphology is that of euhedral crystal form (Plate IV). These crystals tend to be doubly terminated and clear, though some have a small milky core. They usually range in size from just under 0.1 mm to 0.3 mm in length and 0.05 to 0.2 mm in width, but some crystals as long as 1.0 mm were seen. They are most abundant in the upper part of the deposit, and few are seen below the point 270 cm from the top except in the impure basal layer. The third type of quartz occurs as spherical growths of crystals (Plates IV and V). These 'quartz clumps' usually appear as clear bumpy spheres up to 0.3 mm across, and all of the ones examined had a small milky core. The 'bumps' or 'knobs' are euhedral to subhedral quartz

Plate IV. SEM photomicrographs showing different types of quartz particles found in the Fort Dodge gypsum (stereopairs, 15 kv acceleration voltage)

Top: A doubly terminated euhedral quartz crystal, probably authigenic, is in the upper left part of the photograph. A quartz clump, probably also authigenic, is on the lower left. All other grains are quartz of very fine sand to silt size thought to be of detrital (eolian) origin.

Bottom: Doubly terminated quartz crystal, probably authigenic.


0.2 mm

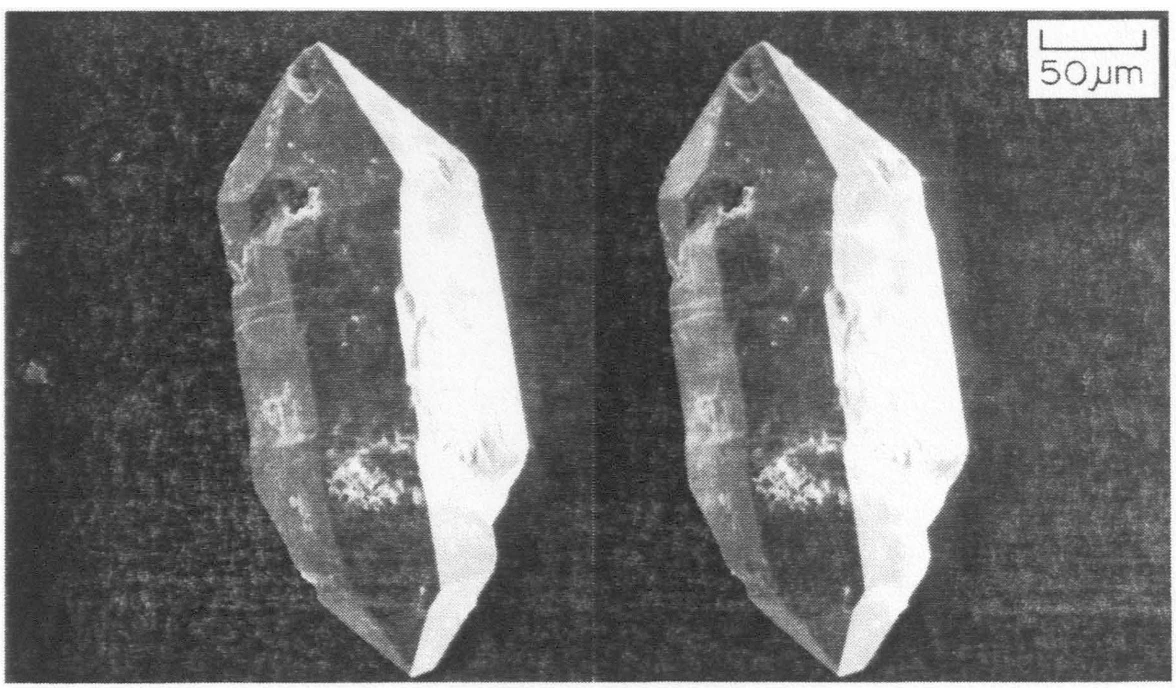
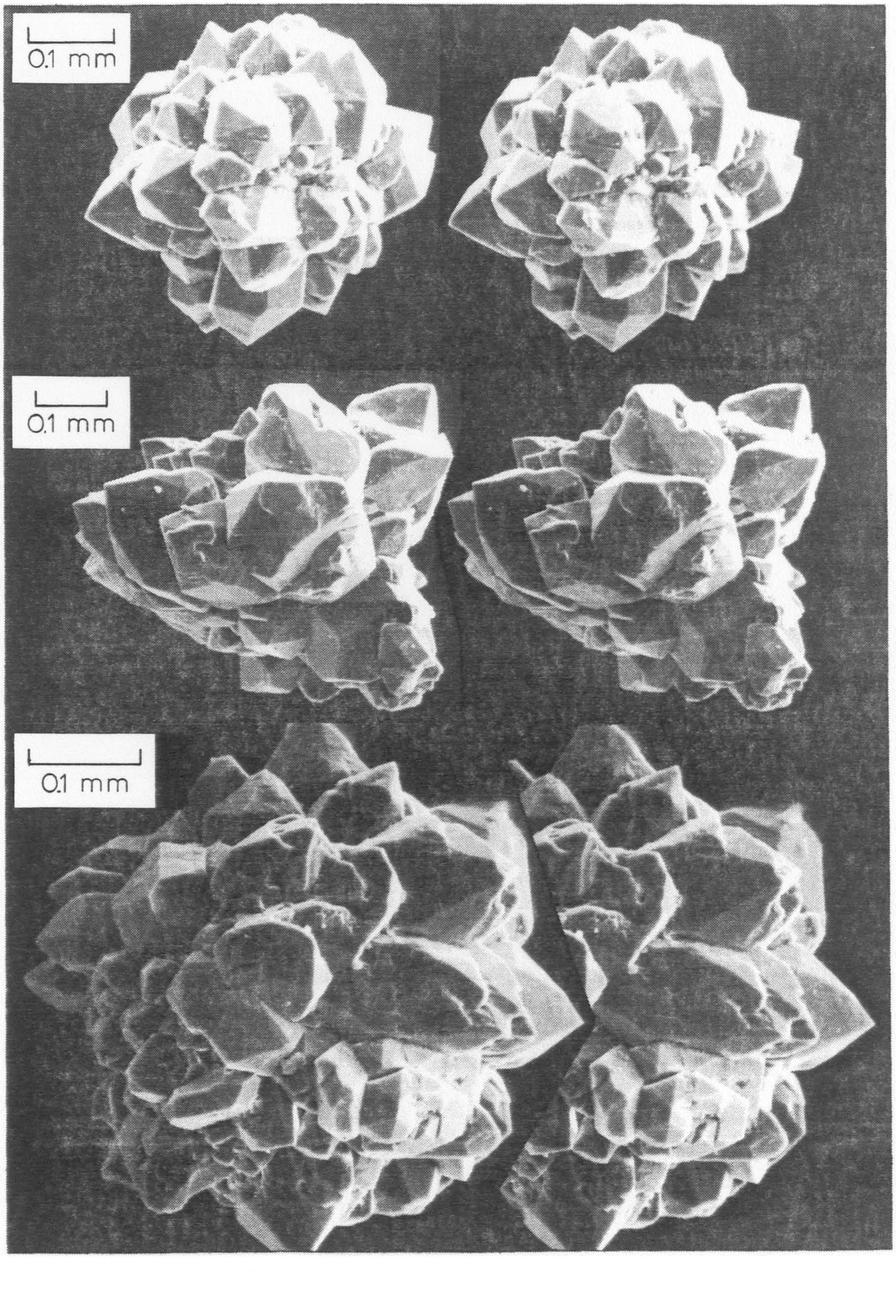

50 μm

Plate V. SEM photomicrographs of quartz clumps in the Fort Dodge gypsum (stereopairs, 25 kv acceleration voltage)


0.1 mm
0.1 mm
0.1 mm

crystal terminations, and the milky core in some of the broken ones appears to be calcite, as it effervesces in dilute hydrochloric acid. The quartz clumps occur throughout the vertical extent of the gypsum.

Calcite is second in abundance among the insoluble minerals. Only the lower half of the gypsum deposit has calcite crystals large enough to be seen (0.1 mm). The upper half of the deposit is also calcareous, but when its insoluble minerals are given the acid test under the microscope, the effervescence can not be traced to any particular crystals. The larger crystals have a euhedral to subhedral habit, forming multi-faceted, elongate crystals which are difficult to tell from the euhedral quartz without the help of acid.

Orthoclase peaks are found in the XRD data (Fig. 9), but the mineral was difficult to distinguish under the microscope, probably because of the apparently small size of the feldspar grains. Staining the orthoclase was attempted but not particularly useful because the small, loose nature of the grains made it difficult to obtain moderate yet thorough etching during the preliminary etching step. Also, the yellow color produced by the standard sodium cobaltinitrite method (Bailey and Stevens, 1960) is similar to the yellow stain possessed by many of the quartz grains. Kaolinite and illite were also detected by XRD, but the grains were too small to be seen under the microscope except for the faint glitter in the sample from light reflecting off clay particles in suspension.

Little difference could be seen in the mineralogy from feature to feature. Even the immediately overlying Soldier Creek beds have nothing mineralogically different from the insoluble minerals of the main gypsum deposit; they simply have a greater abundance of impurities than the gypsum. The tan tiers, however, had some well-indurated bits of very fine sandstone with calcareous cement.

General Abundance

The amount of impurities in the gypsum generally varies between 1.0 and 3.5% (Fig. 10). The upper half has slightly more impurities, but there is also erratic variation with depth. The sharp variations probably represent fluctuations in the rate of sediment influx compared to the rate of gypsum precipitation. The more general trend showing more impurities in the upper half may be due to authigenic growth of euhedral quartz crystals, as will be discussed later in this thesis. An additional sharp depth-related trend is shown by an impure basal layer near the Carbon quarry not plotted in Fig. 10. This is a macro-crystalline selenite layer 10 cm thick found below the bottom of the main gypsum sequence. It shows a much higher impurity content of 34.3% and is separated from the overlying laminated gypsum by a 4 cm thick mixture of soft brown fine sand and clay.

The abundance of insoluble material also varies from feature to feature. The nodules, for example, are nearly pure gypsum (Fig. 9C). Upon dissolution they give practically nothing (0.3%) for XRD analysis. This supports the theory that, if nodules are secondary, impurities are pushed out of the way as the nodules grow. The two tiers that were sampled have a higher impurity content than the laminated gypsum. The sample of the red tiers had 4.9% insolubles, and the tan tiers had 12.7%.

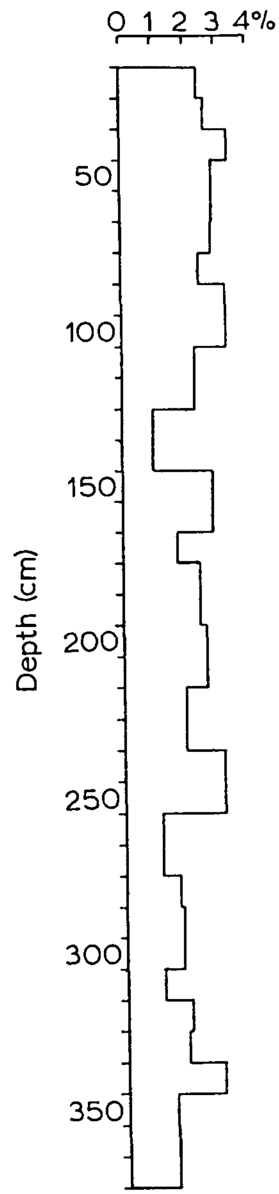


Figure 10. Vertical variation in the amount of insoluble minerals in the Fort Dodge gypsum

Visual inspection shows that most of this material is concentrated either at the boundaries of the tier or in the central parting between double tiers. The gypsum that makes up the tier is itself relatively pure.

Origin

The insoluble minerals show evidence of primary detrital, primary chemical, and secondary (authigenic) origin. The euhedral quartz and quartz clumps are probably authigenic because transport would probably have rounded the grains and imprinted features onto them that would have been seen on the SEM. The morphology of the quartz clumps suggests radial quartz growth from a calcite core, although no radial structures or patterns could be seen under the microscope.

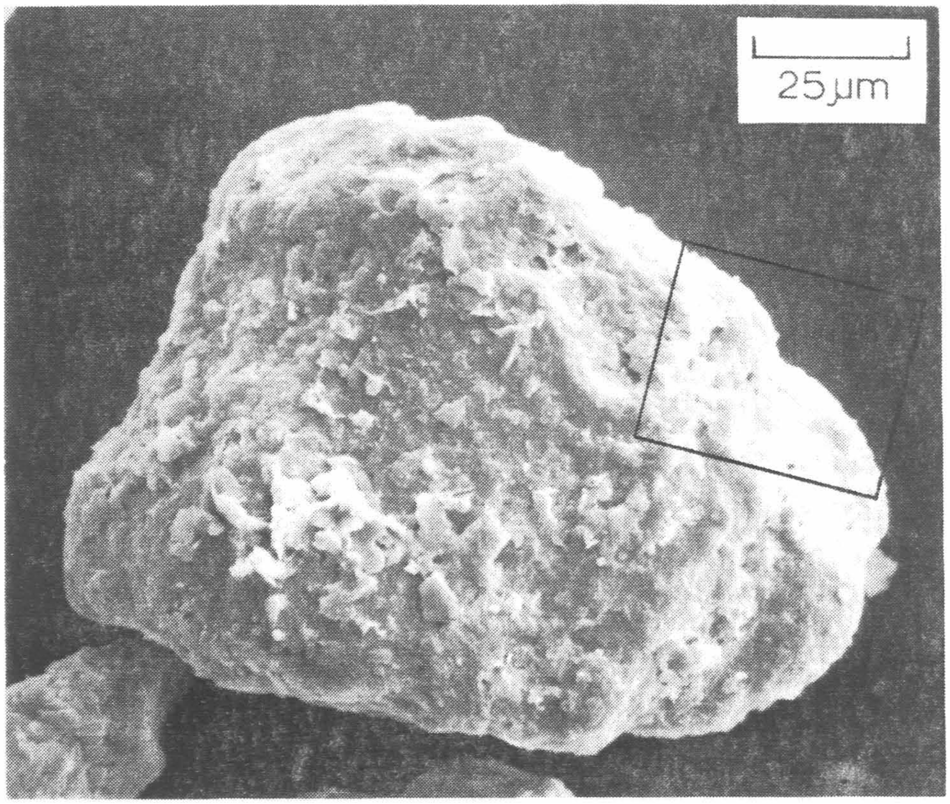
The remaining quartz particles probably have a detrital, eolian origin in an arid environment. This is indicated by several factors. A hot, arid environment is to be expected because it is the most reasonable way to obtain the high amount of evaporation required to concentrate sea water enough to precipitate gypsum. Iowa was at about 30^{o} north latitude during the late Jurassic, and this is the latitude which many of the major deserts of today occupy. In such environments, the wind often becomes the dominant geomorphic agent, and is quite effective in moving particles less than 1.0 mm diameter. The low concentration and evenly dispersed nature of the quartz particles would also not be expected if they were brought in by continuous or ephemeral streams. Streams would also tend to dilute the sea water concentrate and stop the precipitation of gypsum. Most of the quartz particles are also frosted, a common feature of eolian sand grains

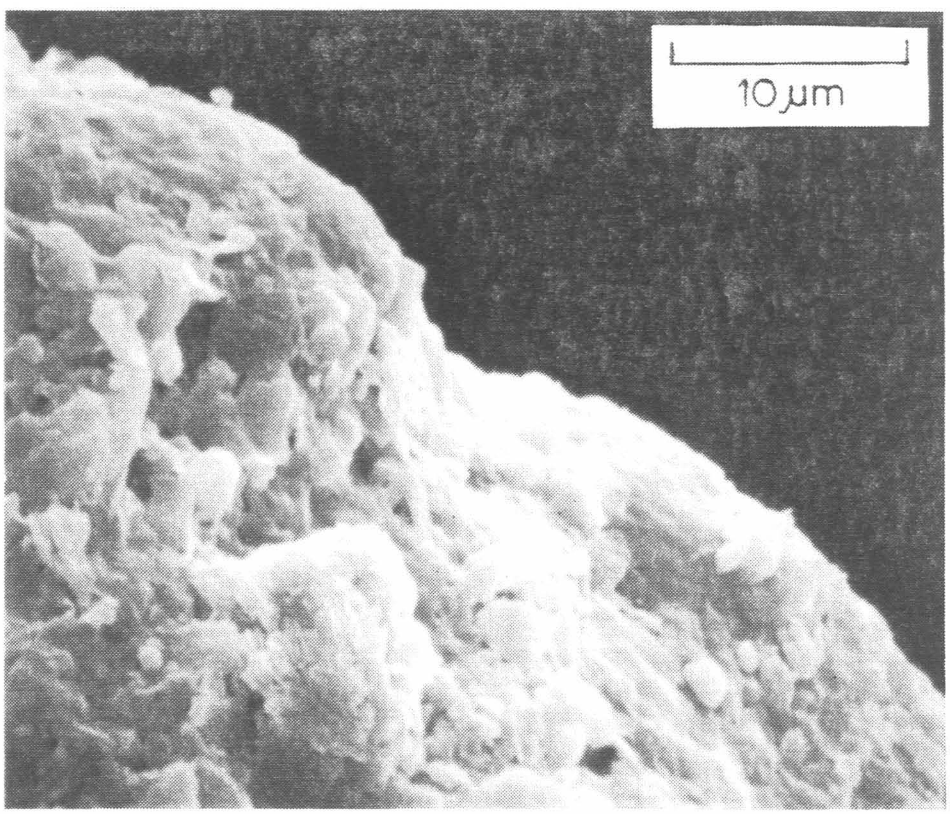
(Glennie, 1970). To further test the eolian hypothesis, the particles were examined with the SEM. Many microscopic surface features have proven to be accurate environmental indicators (Krinsley and Doornkamp, 1973). The (non-euhedral) quartz grains of the Fort Dodge gypsum have no features that indicate subaqueous transport, such as mechanically or chemically formed V-shaped indentations. They do, however, exhibit abundant cleavage plates (Plate VI), which are often produced in eolian environments (Krinsley and Doornkamp, 1973). The limonite and hematite coating on some of the quartz grains gives further support to the idea of a hot, arid, eolian origin, as deserts are well known for the production of red (hematite) and yellow (limonite) coatings on particles (Walker, 1979). The evaporite-red bed association in the geologic record is also common, and the relationship between the gypsum and the overlying red beds of the Soldier Creek member may be such an association. Some of the hematite and limonite was probably redistributed to form the amorphous globs and pyrite.

The orthoclase is probably also of eolian origin, based on its low concentration and wide dispersion. The kaolinite and illite could be either detrital (eolian) or authigenic, as breakdown products of the orthoclase.

The calcite and dolomite are probably of primary chemical origin, based on the fact that carbonates are often associated with evaporite deposition. Calcite is the first mineral to be precipitated in marine evaporite sequences (Clarke, 1924). As evaporation continues, gypsum begins to precipitate contemporaneously with calcite. Continued

Plate VI. SEM photomicrographs of a very fine sand-sized quartz grain showing upturned plates indicative of eolian origin (15 kv acceleration voltage; the boxed area in the top photograph is shown enlarged in the bottom photograph)


25 µm


10 µm

precipitation of gypsum may increase the Mg/Ca ratio enough to cause dolomite to be precipitated or calcite to be dolomitized (Dean, 1978). If this was the case for the Fort Dodge gypsum, the disseminated nature of the calcite and dolomite indicates that the carbonate facies of a marine evaporite sequence does not always have to be deposited in a geographically restricted area close to the mouth of the bay, but can be dispersed and interbedded with the other evaporite minerals.

SUMMARY

Evidence suggests that the orthogonal vertical joint sets of the North Welles quarry and random vertical joints of the Celotex quarry are extension fractures created by uniform horizontal tension. Such joints are usually explained by uplift of the crust, which is forced to cover more surface area at the new, higher radius position within the Earth (Price 1966). Uplift in the Interior Province of the Central Stable Region was probably caused by unroofing, and there are two equally probable unroofing mechanisms: erosion of about 450 m of Cretaceous strata during the Tertiary period, and ablation of about 1700 m of ice during the interglacial episodes of the Pleistocene. Both of these models account for enough isostatic rebound to cause joints, but only the Cretaceous strata model allows for possible dehydration of the gypsum to anhydrite because the weight of the glaciers would not be enough to compensate for the low temperatures involved. Formation of the random Celotex joints may have also been influenced by inhomogeneity of the underlying beds during loading. The parallel joints of the Carbon quarry may have resulted from late movement along an interior basement fault associated with the Midcontinent Geophysical Anomaly. The location and trend of the fault can be recognized on the aeromagnetic map of Webster County, and movement could have been triggered by crustal loading and unloading by either of the above models. The weakly developed horizontal joints are probably late exfoliation features that occur in the upper 2 m of the gypsum. They developed along bedding planes weakened by a relatively high amount of detrital material.

The Fort Dodge gypsum is a relatively pure deposit, with only 1.0 to 3.5% insoluble minerals on the average. The major minerals are quartz, calcite, and orthoclase. There are minor amounts of kaolinite, illite, limonite, hematite, muscovite, pyrite and possibly tourmaline. Quartz is by far the most abundant of these, and has three distinct morphologies: euhedral quartz crystals, quartz clumps, and very fine sand to silt-sized grains. The sand and silt grains probably have an eolian origin in an arid environment, while the euhedral quartz crystals and quartz clumps are probably authigenic. Calcite and dolomite were probably deposited contemporaneously with the gypsum, and could represent the carbonate facies that is ideally the first facies to be deposited in marine evaporite sequences. The orthoclase and muscovite are probably of eolian origin, and the kaolinite and illite could be either detrital (eolian) or authigenic, as breakdown products of the orthoclase. The iron minerals of limonite and hematite occur as coatings on eolian quartz grains, and were probably introduced into the basin as such. Some of the iron may have been rearranged to produce the orange-yellow globs observed, some of which have a pyrite core.

REFERENCES

Ahnert, F. 1970. Functional relationships between denudation, relief, and uplift in large mid-latitude drainage basins. Am. J. Sci. 268:243-263.

Anderson, W. I. 1983. Geology of Iowa: Over two billion years of change. Iowa State University Press, Ames, Iowa.

Bailey, E. H., and R. E. Stevens. 1960. Selective staining of K-feldspar and plagioclase on rock slabs and thin sections. Am. Mineralogist 45:1020-1025.

Bard, G. K. 1982. Petrology and diagenetic features of the Fort Dodge gypsum beds. Ph.D. thesis. Iowa State University, Ames, Iowa.

Berner, R. A. 1971. Principles of Chemical Sedimentology. McGraw-Hill, Inc., New York.

Blatt, H., G. V. Middleton, and R. C. Murray. 1972. Origin of Sedimentary Rocks. Prentice-Hall, Inc., Englewood Cliffs, New Jersey.

Boellstorff, J. 1978. North American Pleistocene stages reconsidered in light of probable Pliocene-Pleistocene continental glaciation. Science 202:305-307.

Chamberlain, R. E. 1980. Geology of the Forest City Basin in Iowa: Tectonic activity and influence on Pennsylvanian sedimentation. M.S. thesis. Iowa State University, Ames, Iowa.

Chang, S. C. 1984. A subsurface study of the Cherokee Group in a five county area southwest of Fort Dodge, Iowa. M.S. thesis. Iowa State University, Ames, Iowa.

Chase, C. G., and T. H. Gilmer. 1973. Precambrian Plate Tectonics: The Midcontinent Gravity High. Earth and Planetary Sci. Letters 21:70-78.

Clarke, F. W. 1924. The Data of Geochemistry. 5th ed. U. S. Geol. Surv. Bull. 770.

Compton, R. R. 1962. Manual of Field Geology. John Wiley and Sons, Inc., New York.

Cooke, H. B. S. 1973. Pleistocene chronology: long or short? Quaternary Research 3:206-220.

Coons, R. L., G. P. Woollard, and G. Hershey. 1967. Structural significance and analysis of midcontinent gravity high. Am. Assoc. Petrol. Geol. Bull. 51:2381-2399.

Dasenbrock, F. E. 1984. Petrology and diagenesis of the Fort Dodge Gypsum in the North Welles Quarry. M.S. thesis. Iowa State University, Ames, Iowa.

Dean, W. E. 1978. Theoretical versus observed successions from evaporation of sea water. In W. E. Dean and B. C. Schreiber (eds.). Marine Evaporites, SEPM Short Course Notes No. 4.

Dorheim, F. H. 1978. Gypsum Resources, Fort Dodge Area. Unpublished internal report. Iowa Geological Survey, Iowa City, Iowa.

Eardley, A. J. 1962. Structural Geology of North America. 2nd ed. Harper and Row, New York.

Flint, R. F. 1971. Glacial and Quaternary geology. John Wiley and Sons, New York.

Garland, G. D. 1979. Introduction to Geophysics: Mantle, Core, and Crust. W. B. Saunders Co., Philadelphia.

Glennie, K. W. 1970. Desert Sedimentary Environments. Developments in Sedimentology 14. Elsevier Pub. Co., Amsterdam.

Goldich, S. S., E. G. Lidiak, C. E. Hedge, and F. G. Walthall. 1966. Geochronology of the Midcontinent Region, United States. 2. Northern Area. J. Geophys. Res. 71:5389-5408.

Hale, W. E. 1955. Geology and ground-water resources of Webster County, Iowa. Iowa Geological Survey Water Supply Bulletin No. 4.

Hancock, P. L. 1985. Brittle microtectonics: principles and practice. J. Struct. Geol. 7:437-457.

Hardie, L. A. 1967. The gypsum-anhydrite equilibrium at one atmosphere pressure. The American Mineralogist 52:171-200.

Johnson, D. M. 1986. Reconnaissance study of the Soldier Creek beds, Fort Dodge, Iowa. M. S. thesis. Iowa State University, Ames, Iowa.

King, E. R. and I. Zietz. 1971. Aeromagnetic study of the midcontinent gravity high of the central United States. Geol. Soc. Am. Bull. 82:2187-2208.

Krinsley, D. H. and J. C. Doornkamp. 1973. Atlas of quartz sand surface textures. Cambridge University Press, Cambridge, England.

MacDonald, G. J. F. 1953. Anhydrite-gypsum equilibrium relations. Am. J. Sci. 251:884-898.

Nilsson, T. 1983. The Pleistocene: Geology and life in the Quaternary Ice Age. D. Reidel Publishing Company, Boston.

Paterson, W. S. B. 1981. The Physics of Glaciers. 2nd edition. Pergamon Press, Oxford.

Price, N. J. 1966. Fault and Joint Development in Brittle and Semi-brittle Rock. Pergamon Press, Oxford.

Reches, Z. 1983. Faulting of rocks in three-dimensional strain fields. II. Theoretical analysis. Tectonophysics 95:133-156.

Shaffer, B. L. 1969. Palynology of the Michigan "red beds". Ph.D. thesis. Michigan State University, Ann Arbor, Michigan. Dissertation Abstracts 19:114 6.

Terzaghi, R. D. 1965. Sources of error in joint surveys. Geotechnique 15:287-304.

Thiel, E. 1956. Correlation of gravity anomalies with the Keweenawan geology of Wisconsin and Minnesota. Geol. Soc. Am. Bull. 67:1079-1100.

Turcotte, D. L., and G. Schubert. 1982. Geodynamics: Applications of Continuum Physics to Geological Problems. John Wiley and Sons, New York.

Van Eck, O. J., R. R. Anderson, R. A. Black, and S. W. Daut. 1979. Regional tectonics and seismicity of southwestern Iowa. Annual Report FY79, NUREG/CR-0955. U. S. Nuclear Regulatory Commission, Washington, D. C.

Walker, T. R. 1979. Red Color in Dune Sand. In E. D. McKee (ed.). A Study of Global Sand Seas. U. S. Geol. Surv. Prof. Paper 1052.

Wilder, F. A. 1917-1918. Gypsum: Its Occurrence, Origin, Technology and Uses. Iowa Geological Survey Annual Report 28:47-537.

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APPENDIX A. METHOD OF PARALLAX CORRECTION FOR QUANTITATIVE JOINT ORIENTATION STUDIES USING AERIAL PHOTOGRAPHS

Background

The aerial photographs of the gypsum were taken with a 50mm lens from the side window of a 4-seat airplane at an altitude of about 400 feet. It was impossible to obtain a camera angle that was perfectly vertical, even under the best banking conditions. The photographs thus have inherent distortion of the joint trace orientations due to the parallax angle of the camera. This distortion can be categorized into two types which I will refer to as 'converging railroad track' and 'parallelogram'. In converging railroad track distortion parallel lines on the outcrop do not remain parallel in the photograph, but converge slightly toward the end that is farther away (Compton, 1962). This is more pronounced at the edges of the picture, and is exaggerated if a wide angle lens is used at a low angle of incidence. This type of distortion is assumed to be negligible in the present photographs because a 50 mm lens was used at a reasonably high angle of incidence, and the quarries are located in the center of the original pictures, where this type of distortion is least developed. The parallel nature of the joints in the Carbon quarry and in the aerial photographs indicate this is a valid assumption.

Parallelogram distortion is the second type of distortion, and in this study was significant enough that it had to be dealt with. In parallelogram distortion, initially parallel lines remain parallel, but squares are distorted into parallelograms because they are not being viewed face-on. All angular relationships on the surface containing the lineations change, so correction is required if the data taken from the photographs are to represent true directions.

Method for Correcting Parallelogram Distortion

The scale and compass orientation of each aerial photograph were field checked by surveying a large 90^{o} (N-S and E-W) cross on the top surface of the gypsum in each quarry, and plotting these crosses on an overlay at their identical locations in the respective photographs. Each cross was then extrapolated by drafting methods to construct a 61 X 61 m (200 X 200 ft.) parallelogram, whose size, orientation, and shape (departure from square) indicate scale, north direction, and amount of parallax distortion. If there were no distortion in the photograph, the plotted cross would have arms of equal length at 90^{o} to each other, and would project as a square.

To correct for parallelogram distortion, each photograph with its parallelogram overlay was mounted on a tilt board, which was then mounted onto a tripod equipped with a universal joint. The prints were then re-photographed in a tilted position using a Canon F-1 with a 420 mm lens. Each print was tilted in such a way that the parallelogram overlay appeared as a square in the viewfinder of the camera. A stereoscopic technique was used to accomplish this; one eye looked through the viewfinder of the camera, while the other eye looked at a perfect square drawn on white paper. Once the proper tilt angle was obtained, the parallelogram overlay was removed and a photograph was taken of the uncorrected print on the tilt board.

Each parallelogram was re-drawn on the new, corrected photograph using surface markings on the gypsum to locate it in the same position as in the first photograph. To determine the effectiveness of the correction technique, a comparison was made between the old and new parallelograms, which, ideally, should have four sides of equal length meeting at 90° and two diagonals of equal length at a 45° angle to the sides. Table 1 shows that the corrected photographs indeed have less parallax distortion than the original ones, and the parallelograms in them approach being perfect squares. Bearings of lineations on the new photographs are within 1° to 3° of true, and distances measured with the scale are 96% accurate or better, depending on their bearing. For comparison, a study based on the uncorrected photographs would result in errors on measured joint orientations of up to 5°.

Limitations

This procedure is the photographic equivalent of the electronic "tilt correction" feature found on some scanning electron microscopes. This feature restores lineations on planar, tilted specimen surfaces to their untilted lengths and angles-- seen if one's line of sight is perpendicular to the specimen surface. Objects above or below the planar surface, however, such as undulations or three dimensional objects, are distorted further from their true shape and orientation by tilt correction. Proper application of this correction method is thus restricted to cases where

Table 1. Comparison of parallax distortion on uncorrected and corrected photographs of the three quarries

Quarry	Scale Accuracy		Angle Accuracy in degrees	
	Uncorrected	Corrected	Uncorrected	Corrected
North Welles	> 91 %	> 99%	within 5	within 1
Carbon	> 91 %	> 96 %	within 4	within 3
Celotex	> 88 %	> 97 %	within 4	within 2

the surface containing the lineations is relatively planar. The top surface of the Fort Dodge Gypsum is such a case. Although there is considerable relief (up to 5 m) due to the solution of the gypsum along the joints, most of this relief is restricted to narrow channels that delineate the joints themselves, while the tops of these channels are at roughly the same height. This is especially true of the Carbon quarry, where deep, narrow channels run between flat-topped remnant blocks of gypsum. In such cases, the trends of the joints were taken from the top of the channel, level with the top surface of the gypsum. The correction technique is also restricted to photographs with a limited amount of 'converging railroad track' distortion, as it does not correct for this.

Another drawback to the photographic tilt correction technique is the degradation of picture quality in the corrected photograph because of the extra photoduplication steps involved. For this reason, the aerial photographs included in this thesis (Plates I, II, and III) are of the uncorrected photographs rather than the corrected ones. In this study, the higher resolution, uncorrected print was used to initially find the joints, and the lower quality, corrected print was then used to take their true strike.

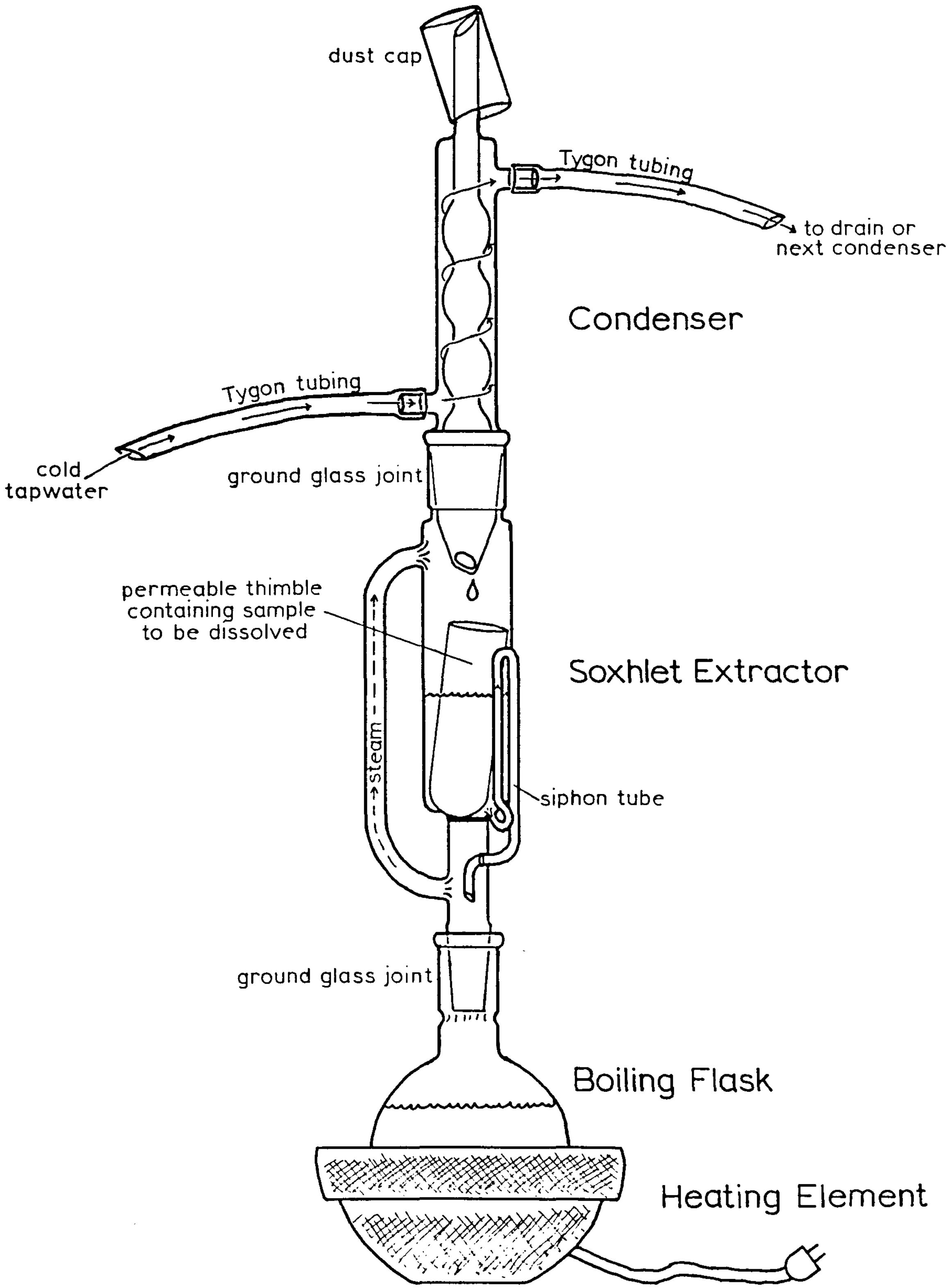
APPENDIX B. SAMPLE PREPARATION

In order to study the non-gypsum part of the Fort Dodge Gypsum in detail, a method was needed to dissolve the gypsum while leaving the remaining minerals and their surface features intact. A hot distilled water method using soxhlet extractors was used to accomplish this. The soxhlet extractor setup is shown in Figure 11. It consists of a boiling flask, condenser unit, and soxhlet extractor, all made of glass, and an electrical heating element. The heating element boils water in the boiling flask, and the steam travels up through the steam tube and into the condenser. Condensed steam (distilled water) drips from the condenser into the soxhlet extractor, where the sample to be dissolved is held in a permeable extraction thimble made of cellulose. The water level in the soxhlet rises until it reaches the top bend of the siphon tube. The water then siphons itself back into the boiling flask and the cycle starts over again. Regular tapwater is used to cool the condenser but does not enter the system.

Samples to be dissolved in the soxhlet extractors were crushed to get them to fit into the extractors and to speed dissolution. Unfortunately, crushing can contaminate the fine impurities and affect their surfaces. To prevent this, each crushed sample was dry-sieved, and only the clean coarser fraction (2.00-10.00 mm) was used, with its enclosed, fine-grained impurities still intact. The 0.590-3.360 mm size fraction was used for dissolution initially, but the 2.00-10.00 mm fraction was used after it was discovered that some of the euhedral quartz

Figure 11. Diagram of a soxhlet extractor setup

Arrows show the directions of water and steam flow


dust cap
Tygon tubing
to drain or
next condenser
Condenser
Tygon tubing
cold
tapwater
ground glass joint
permeable thimble
containing sample
to be dissolved
Soxhlet Extractor
steam
siphon tube
ground glass joint
Boiling Flask
Heating Element

crystals are as long as 1.0 mm. Most of the impurities, however, are of very fine sand size (0.0625-0.125 mm) or smaller.

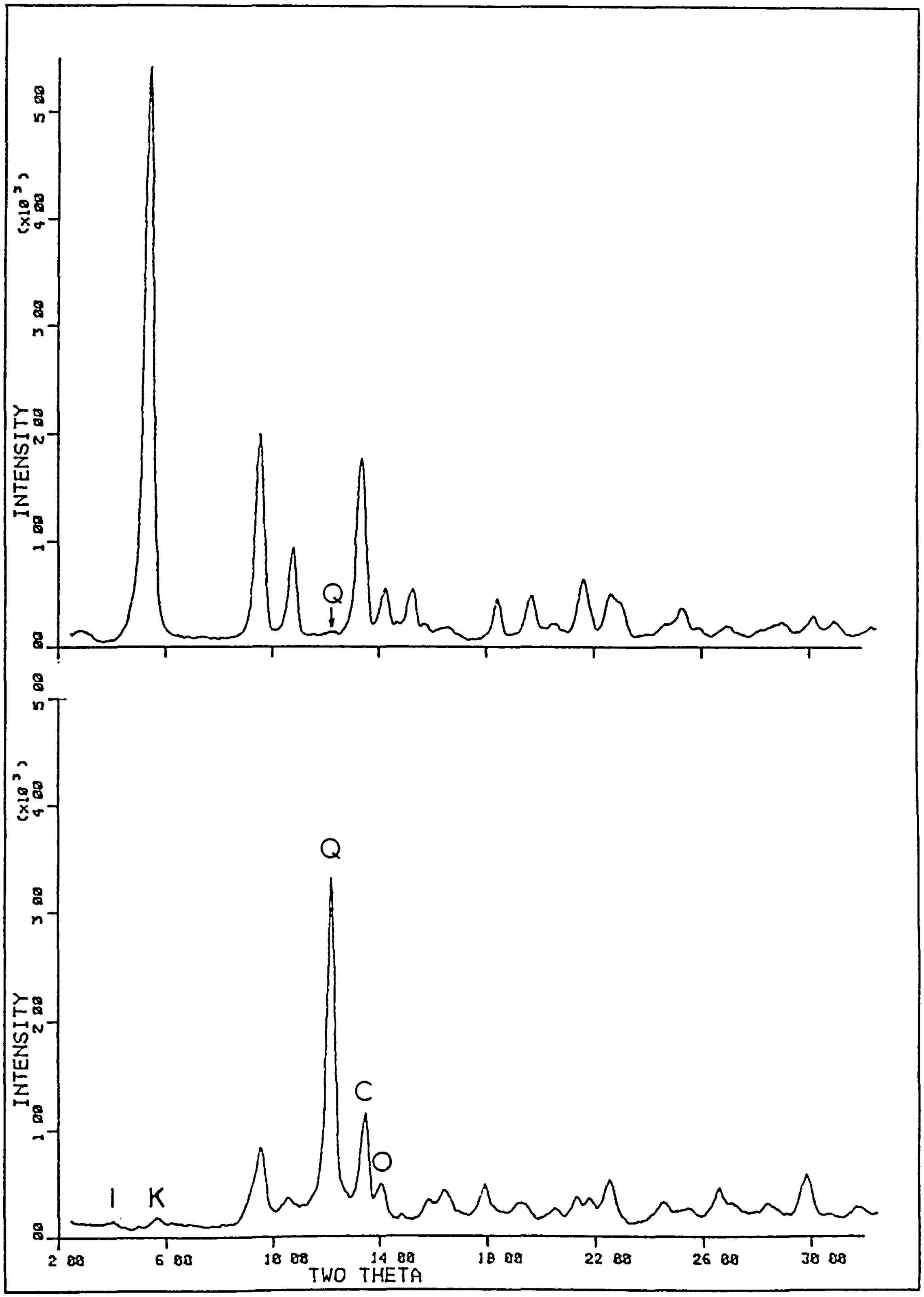
During dissolution, samples of the water in the soxhlet extractors were occasionally removed, cooled to room temperature, and measured for pH using a Markson Model 88 digital pH meter calibrated with a pH 7.00 buffer solution. The pH of water in the extractors ran between 8.30 and 9.52. It was thought that under these conditions, calcite would be preserved. To make sure, a test sample of 90% coarse gypsum (with a known impurity content) and 10% powdered calcite was mixed up. Fifty grams of the mixture were run in a soxhlet extractor for six weeks. All of the gypsum was dissolved in this time, but only about 3.5% of the calcite was dissolved.

A typical 30 gram sample of gypsum takes approximately 2 weeks to dissolve, depending on grain size, boiling rate, and condenser efficiency. Six banks of extractors were run simultaneously in this study in order to get a substantial number of samples in a reasonable amount of time. Once the gypsum in the sample was dissolved, the concentrated impurities were washed out of the extraction thimble and into a filter funnel equipped with a Gelman Metricel membrane filter with a 1.2 micrometer pore size. The samples were then filtered and dried using suction. The concentrated impurities to be run in the X-ray diffractometer were ground slightly with an agate mortar and pestle to mix the grains and break up any large clumps. Fig. 12 shows that using the extractors greatly increases the resolution of X-ray diffraction peaks of impurities with low concentrations, including that of calcite. Samples for the scanning

Figure 12. X-ray diffraction data (molybdenum radiation) showing resolution improvement after gypsum dissolution

Top: Fort Dodge gypsum, before gypsum dissolution. The sample was taken 150-160 cm below the top of the gypsum, and had 2.9% insoluble minerals. The quartz peak (Q) can barely be seen; all other peaks are from gypsum.

Bottom: Sample of the insoluble minerals from the above interval after gypsum dissolution. Major peaks indicate the presence of quartz (Q), calcite (C), orthoclase (O), kaolinite (K), and illite (I)


INTENSITY
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Q
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TWO THETA