Initiation and growth of gypsum piercement structures in the Zechstein Basin

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Abstract—The importance of tectonic processes in initiating halite diapirs has become much better understood in recent years. Less well understood is the development of diapiric structures involving rocks composed predominantly of gypsum. Below about 1000 m, gypsum dehydrates to anhydrite, which often obscures primary sedimentary textures. If the strain associated with diapiric rise in the rock induces the transition to anhydrite, obliteration of primary features in the gypsum can be expected. In our study, we infer that the diapiric movement in the Werra Anhydrite member of cycle 1 of the Zechstein Formation of Europe occurred before the initial transition of gypsum to anhydrite based on the presence of pseudomorphs of bedded primary gypsum crystals, the overburden lithologies and depositional environment, and the mechanical properties of gypsum, anhydrite and carbonate rocks. Faulting and differential loading of a shallow overburden were the key components in initiating the gypsum diapirism. The transition to anhydrite occurred after burial and after cessation of diapirism. In comparison, the diapirism of calcium sulfate of the Leine Anhydrite into the Leine Halite members of cycle 3 of the Zechstein Formation probably occurred much later after burial and appears to have been triggered by halite diapirism, which in turn triggered the dehydration reaction, causing the calcium sulfate to become the incompetent phase relative to the halite. Published by Elsevier Science Ltd.

INTRODUCTION

Besides the thick deposits of potassium-bearing rock salt, for which the Upper Permian Zechstein Basin of Europe is well known, there are several horizons of anhydrite. Petrography of the anhydrite horizons indicates formation in subaqueous depositional environments in various depths of water. Bedded anhydrite in the platform facies has been interpreted by previous workers to have precipitated as a shallow-water deposit of bedded selenite crystals (Richter-Bernberg, 1985; Langbein, 1987; Peryt *et al.*, 1993). The basinal, thinly-laminated anhydrite deposits are considered to be resedimented from the platform and shelf facies into gypsum turbidite deposits and laminated gypsum (Schlager and Bolz, 1977; Richter-Bernberg, 1985; Peryt *et al.*, 1993). Exposures of the Werra Anhydrite Member (Zechstein cycle 1) in gypsum quarries west of the Harz Mountains in northern Germany reveal numerous slump and slide features typical of soft-sediment deformation that have been described in detail by Herrmann and Richter-Bernberg (1955) and Langbein (1987). Although diapirism involving halite is well documented in many sedimentary basins, it was not until recently that Paul (1988) identified some of the more distinctive piercement deformation structures in the Werra Anhydrite at this location as gypsum diapirs.

The specific gravity of gypsum ($\rho = 2.32$ g cm^{-3})
makes it less dense than most sedimentary rocks, so gypsum diapirism should theoretically proceed analogously to diapirism involving halite ($\rho = 2.15$ g cm^{-3}). If buoyancy alone is considered the driving mechanism for diapirism, in most sedimentary sequences, the density inversion required for halite diapirism to initiate does not develop until the porosity in the overlying sedimentary rocks is reduced by compaction from the relatively high values at the surface of 50% or more to less than 30% (Lerche and O'Brien, 1987). The burial depth at which the critical porosity is reached for halite is 300–3000 m, depending on the rock type and the sedimentary basin (Athy, 1930; Rhodehamel, 1977; Sclater and Christie, 1980). For gypsum rocks, which are slightly more dense than halite but still less dense than sedimentary rocks, the porosity of overlying sedimentary rocks must be reduced to about 18%. The density inversion for a sedimentary overburden for gypsum diapirism should therefore occur at depths of 700 m to more than 5000 m.

The depth to which gypsum is stable theoretically is increased by high pore fluid pressure that can develop when an impermeable layer or rock is present (MacDonald, 1953; Zen, 1965; Blount and Dickson, 1973; Jowett *et al.*, 1993), although processes involved in diapirism are likely to aid dewatering of any dehydrating sediments. The dehydration reaction would eliminate the presence of any buoyancy force and make gypsum diapirism unlikely to be a viable mechanism for the development

of calcium sulfate piercement structures in most sedimentary basins.

Recent work has shown that diapirism generally begins at shallower depths due to differential loading or extensional or compressive tectonic stress, and that buoyancy forces alone are rarely sufficient to initiate diapirism (Jackson and Talbot, 1986). Jackson and Vendeville (1994) concluded that the most common initiator of salt diapirism is regional extension leading to brittle faulting in the overburden, and the flow strength of salt causes it to flow as a fluid over geological time into the thinned and weakened areas of the overburden. The results of rheological studies by Baumann (1984) indicate that the flow strength of gypsum is similar to halite at strain rates typical for diapirism, which indicates that the depth requirements for gypsum diapirism to occur can be met under geologically reasonable conditions. Diapirism in the geological record involving calcium sulfate as the buoyant component has been described in detail at only a few other localities besides the Zechstein Basin: the Sverdrup Basin in the Canadian Arctic (Gould and de Mille, 1964; Hoen, 1964; Schwerdtner and Clark, 1967; Stephenson et al., 1992), the Coahuila marginal folded province in northeastern Mexico (Wall et al., 1961; Weidie and Murray, 1967; Weidie and Martinez, 1972; Laudon, 1984) and in the Eastern Prebetic Foldbelt in Spain (de Ruig, 1995). This paper examines the conditions that were probably present after deposition of the Zechstein 1 Werra Anhydrite Member and the overlying Zechstein 2 Carbonate that led to diapirism involving gypsum, and the conditions leading to diapirism of calcium sulfate in the Zechstein 3 Leine Anhydrite into the overlying Zechstein 3 Halite, and draws some comparisons on the incidence of gypsum diapirism in other basins.

FLOW PROPERTIES OF GYPSUM

Deformation mechanism maps are constructed to predict the dominant flow mechanism and strain rates of a mineral in response to a given differential stress. Maps for halite indicate that, at strain rates characteristic for diapirism, strain in halite is accommodated by deformation in the transition between the dislocation creep and pressure-solution fields (Spiers et al., 1990; van Keken et al., 1993). The constitutive flow laws for halite were developed from deformation experiments done on natural rock salt from various localities and from synthetic halite samples. The data in the literature on material properties of gypsum are limited. Baumann (1984) experimentally determined the flow law of one type of gypsum from the Tafeljura of Switzerland to be

$$\dot{\epsilon} = A \exp\left(-\frac{Q}{RT}\right)\sigma^n, \tag{1}$$

where $A = 4.5 \text{ s}^{-1}$, $Q = 31\,500 \text{ cal/mol}^{-1}$, $R = $ gas con-

stant $(1.986 \text{ cal (mol·grad)}^{-1})$, $n = 4.5$ and $\sigma = $ differential stress (bars). Baumann's constant strain-rate experiments were performed under relatively low temperatures (20–100°C). The confining pressure of 150 MPa is not geologically reasonable for gypsum as it represents a burial depth of about 6500 m, but Baumann chose it in order to compare his results with results of experiments on halite and anhydrite carried out under similar conditions. The average grain size of the samples was 100 μm and experimental strain rates ranged from about 10^{-5} to 10^{-7} s^{-1}. Baumann determined by petrographic analysis that the dominant deformation mechanism was deformation twinning. The creep law for halite developed by van Keken et al. (1993) predicts steady-state flow of halite by parallel operation of dislocation climb and diffusive mass transfer or pressure solution. The total steady-state creep rate is given as:

$$\dot{\epsilon} = \dot{\epsilon}_{ps} + \dot{\epsilon}_{dc}, \tag{2}$$

where the subscript 'ps' denotes pressure solution and 'dc' denotes dislocation creep. The form of the diffusive mass transfer term is:

$$\dot{\epsilon}_{ps} = 6.95 V_m \times 10^{-15} \exp(-\frac{24530}{RT})\frac{1}{T}\frac{\sigma}{d^3}. \tag{3}$$

V_m is the molar volume of the solid phase, $2.693 \times 10^{-5} \text{ m}^3 \text{ mol}^{-1}$, and d is the grain size in meters (Spiers et al., 1990).

The form of the dislocation creep term is:

$$\dot{\epsilon}_{dc} = D \exp(-\frac{Q}{RT}\ln\frac{\tau_0}{\mu_0})\mu^{-n}\sigma^n, \tag{4}$$

where $n = Q/RT$. The constitutive parameters determined by Wawersik and Zeuch (1986) are $Q = 18.96 \text{ kJ mol}^{-1}$ (activation energy for cross-slip-controlled creep), $\tau_0/\mu_0 = 0.074$ and $D = 7.92 \text{ s}^{-1}$.

Diffusive mass transfer dominates the deformation of rock salt for geological stresses and strain rates, so deformation rate is therefore influenced by grain size. The temperatures of all the experiments conducted to determine the constitutive parameters in equations (1), (2), (3) and (4) were in the range of 20–100°C, and except for the halite pressure-solution experiments, the confining pressures were variable.

An attempt to compare the flow behavior of halite and gypsum given these considerations is shown in Fig. 1. The only deformation mechanism in gypsum identified by Baumann (1984) was deformation twinning, a creep mechanism not dependent upon diffusion. A plot of strain rate vs differential stress for gypsum (equation (1), and the dislocation creep term only (equation (4) from the halite flow law, suggests that at extremely low strain rates gypsum flow strength approaches that of halite (Fig. 1a). The strain rates generally considered to be important for halite diapirism are of the order of 10^{-14} s^{-1}, so for very low stresses neither halite nor gypsum can flow fast enough to form diapirs when
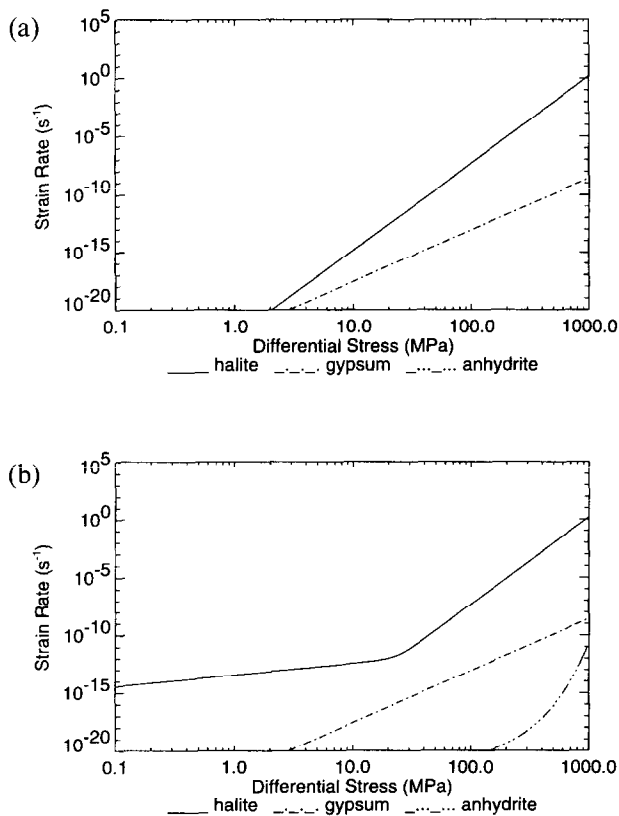
Fig. 1. Flow rate curves for halite and gypsum, with anhydrite plotted for comparison. All curves are calculated for $T = 30°C$. (a) The halite curve is for deformation by dislocation creep as predicted by equation (3) in the text (from van Keken et al., 1993). Gypsum creep is by twinning and cataclasis; the curve is calculated using equation (1) in the text (from Baumann, 1984). (b) Total steady-state flow rate for halite including creep by diffusional mass transfer (equation (4) in text). The gypsum curve is the same as in (a); the anhydrite curve is calculated using equation (5) in text which is based on an equation by Müller et al. (1981). The constitutive parameters used are measured from samples with average grain size of 100 μm, and the same grain size was used in the calculations of flow by diffusive mass transfer.

deforming by only dislocation creep or deformation twinning, respectively. The total steady-state flow law of halite that includes the diffusive mass-transfer term (calculated for $d = 100$ μm) brings the predicted strain rate for halite at low stresses up to values that are geologically significant (Fig. 1b). The very low slope of the halite curve for 0.1 to ≈ 30 MPa differential stress indicates near-Newtonian behavior for halite in this stress range; halite has effectively no creep strength at differential stresses below 30 MPa. The equation derived by Baumann (1984) for gypsum does not include a term for flow by diffusive mass transfer. However, as gypsum is also a high solubility mineral, it is probable that it would exhibit behavior similar to halite when deforming by diffusive mass transfer, possibly bringing the low stress-strain rate for gypsum into a range that is reasonable for geological conditions. More recent work on gypsum deformation by pressure solution by De Meer and Spiers (1993) shows that, indeed, gypsum undergoes creep at much higher strain rates in the presence of fluids. De Meer and Spiers (1993) measured strain rates up to

10^{-6} s^{-1}, and interpreted the rate-controlling step for pressure solution to be precipitation of gypsum in the pore spaces.

The flow law for anhydrite that was derived by Müller et al. (1981) is also plotted in Fig. 1(b) for comparison with halite and gypsum. The flow law is given as:

$$\dot{\epsilon} = A \exp[-\frac{Q}{RT}][\sinh \frac{\sigma}{\sigma_0}]^n. \tag{5}$$

The values for the anhydrite curve calculated in Fig. 1(b) are for a specimen of natural anhydrite with a grain size of about 100 μm. Constitutive parameters are $A = 2.07 \times 10^5$ s^{-1}, $Q = 36.4$ kcal mol^{-1} and $\sigma_0 = 80$ MPa. For deformation of anhydrite by dislocation glide and by twinning, diapiric strain rates are not predicted until differential stresses are of the order of 600–700 MPa. These stresses are nearly two orders of magnitude higher than the differential stresses considered to be important for salt flow in diapirs (Wawersik and Zeuch, 1986).

Heard and Rubey (1966) conducted deformation experiments on gypsum to investigate the effect of its dehydration to anhydrite on rock strength. In their high-confining-pressure experiments (200–500 MPa), they found a sharp decrease in rock strength at temperatures from 80 to 150°C at strain rates of $10^{-3.5}$–$10^{-6.4}$ s^{-1}, and petrographic evidence of dehydration upon examining the deformed samples. The experiments of Baumann (1984), carried out at somewhat lower strain rates ($10^{-4.3}$–$10^{-6.6}$ s^{-1}), found complete dehydration of specimens at 100°C. In experiments at 40°C, evidence of dehydration occurs only along grain boundaries, and Baumann interpreted that some of the deformation took place by grain-boundary sliding with the water acting as a lubricant. Because Baumann allowed the released water to drain from his specimens during deformation, the decrease in strength at the transition reaction was only temporary.

Hardie (1967) experimentally determined that the transition from gypsum to anhydrite occurs at a temperature of 58°C at 1 atmosphere. The experiments of Baumann (1984) provide direct evidence that strain can lower the transition temperature, while theoretical considerations suggest that the released water from the dehydration reaction can lead to overpressured pore fluids, which would inhibit the transition to anhydrite and increase the dehydration temperature (MacDonald, 1953; Zen, 1965; Blount and Dickson, 1973; Jowett et al., 1993). Calculations of hydraulic head over time for water released from the gypsum–anhydrite transition by Hanshaw and Bredehoeft (1968) indicate that, unless the layers confining the anhydrite are relatively impermeable, excess head will not be maintained. For a 15-m thick gypsum–anhydrite bed and a 1200-m thick confining layer, pore fluids would no longer be over-pressured after about 45 000 years. This is only on the order of one-tenth the time required for a diapir to rise through the confining layer at the fastest diapiric strain rates (Jackson et al., 1994). The rheology of the overburden in all cases is

critical for interpreting which conditions contributed to diapirism. That the overburden usually behaves as a brittle layer in geological conditions for diapirism is more consistent with observations of timing and depth for initiation of diapirs (Jackson and Vendeville, 1994; Jackson et al., 1994; Weijermars et al., 1993), and, based on experimental data, the flow strength contrast between gypsum and other sedimentary rocks is similar to that of halite. Two models of the rheology of rocks that undergo diapirism, fluid buoyant layer–brittle overburden layer and fluid buoyant layer–fluid overburden layer, may apply to gypsum diapirism in the Zechstein Basin.

GYPSUM PIERCEMENT STRUCTURES IN THE WERRA ANHYDRITE MEMBER

Exposures of calcium sulfate rock of the Werra Anhydrite Member (A1) of the Zechstein Formation piercing the overlying Stassfurt Carbonate Bed (Ca2) of the Zechstein are found in several quarries southwest of the Harz Mountains in Germany. The sediments were deposited on a paleogeographic high called the Eichsfeld Rise (Fig. 2). The wide range of deformation textures found in the calcium sulfate rock have been described by several workers (Herrmann and Richter-Bernberg, 1955; Richter-Bernberg, 1985; Paul, 1988; Peryt et al., 1993). The calcium sulfate facies (A1) is sandwiched between platform carbonate rocks of the Zechstein 1 horizon (Ca1 carbonate bed) below and the Stassfurt horizon (Ca2
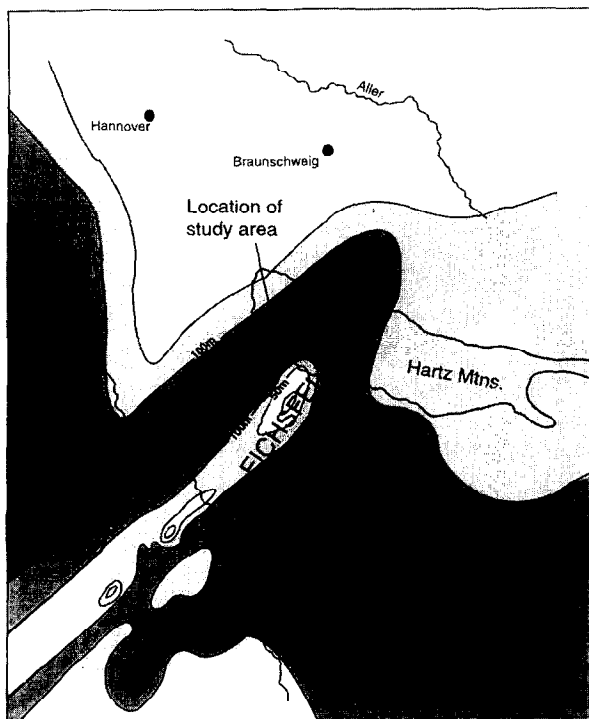

Fig. 2. Werra Anhydrite Member thickness and faces in the Zechstein Basin between the Elbe and Rhein Rivers, showing Eichsfeld Rise and the location of the outcrop study area.

carbonate bed) above (Paul, 1988). The Werra Anhydrite Member on the carbonate platform is interpreted to be a shallow-water, lagoonal deposit originally deposited as bedded selenite gypsum (Richter-Bernberg, 1985; Langbein, 1987). Peryt et al. (1993) identified textures in cores from the Werra Anhydrite Member in Poland as pseudomorphs of selenite of various crystal sizes as well as nodular gypsum or anhydrite that formed in the muddy facies of the primary depositional environment. The calcium sulfate in the Zechstein is always found in the form of anhydrite in the subsurface, although many of the exposures at the surface have been re-hydrated to gypsum (Richter-Bernberg, 1985). The gypsum is an alteration product of the secondary anhydrite and forms from exposure of the rocks to meteoric water. Bedding with similar appearance is preserved in both phases, and layer thickness in outcrop exposures ranges from 1 to 3 cm (Fig. 3a). The entire rock is fine-grained, nearly pure calcium sulfate, with the layering defined by a higher organic matter content (T. M. Peryt, personal communication 1993).

Calcium sulfate diapir structures are found on the platform facies of Zechstein 1. The sulfate is thickest around the margins of the platform and thins to 40–50 m in the basinal facies, where it contains alternations of finely laminated and mosaic anhydrite. On the shelf, the A1 sulfate is up to 250 m thick, and the overlying Ca2 carbonate bed can be as thick as 100 m, although it is more commonly about 60 m in thickness (Fig. 4). The piercement structures occur at the thickest part of the Werra Anhydrite. The shapes and sizes of the diapirs vary from round to elongate cupolas 50 m in diameter to 100-m high dikes only 100 m wide and with vertical walls extending for more than 2 km (Fig. 5). The piercement structure at the Rötzel Quarry is a linear feature more than 1 km long; a cross-sectional exposure of this diapir shows nearly vertically dipping Ca2 beds on the flanks of the piercement structure (Fig. 6a). The Ca2 carbonate bed is brecciated in its upper part, and the Werra Anhydrite is in contact with the anhydrite overlying it (A2, Stassfurt Anhydrite Member). Layering in the A1 diapir is subvertical with disharmonic folding (Fig. 6b). The brecciation of the Stassfurt Carbonate indicates a significant competency contrast when the brittle (lithified) carbonate was intruded by the very ductile (possibly only partially lithified) gypsum.

The structures in the diapiric gypsum at Rötzel show fluidal flow patterns indicating ductile deformation. Disharmonic folding in the cores of gypsum and anhydrite piercement structures exposed in quarries are similar to those observed in intensely deformed metamorphic rocks (Fig. 3b & c). In some places indistinct shear planes separate the intrusive gypsum from the non-diapiric gypsum. The intensity of folding associated with vertical curtain folds found in halite salt domes is not observed here, probably because the height of diapiric rise of the gypsum was only 100 m. The kinematic history of the Werra Anhydrite is impossible to resolve based on
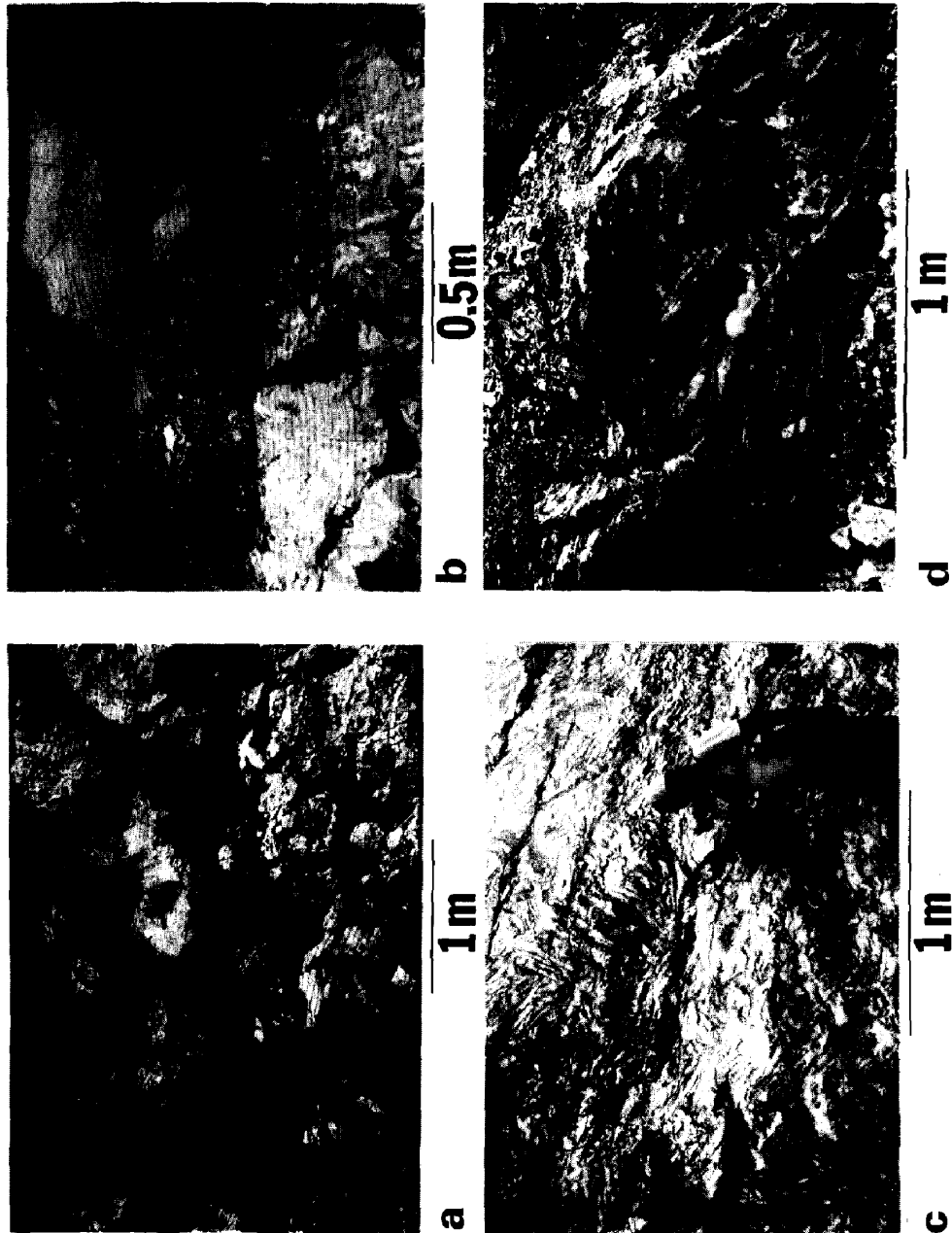
Fig. 3. Photographs of exposures of the Werra Anhydrite in the Harz Mountain area. (a) Exposures of well-bedded, non-diapiric gypsum at the Ührde Quarry, Germany. Height of exposure is 2 m. (b) Slump block of bedded anhydrite with a monoclinal fold in Schimpf Quarry. (c) Exposure of deformed, kink-folded gypsum bedding in the Werra Anhydrite Member at the base of a diapir exposed in the upper stope of the Schimpf Quarry near Osterode, Germany. (d) Same locality as (c), showing overturned folds associated with a higher degree of ductility. The lens cap (near the upper-middle portion of the photograph) is for scale.
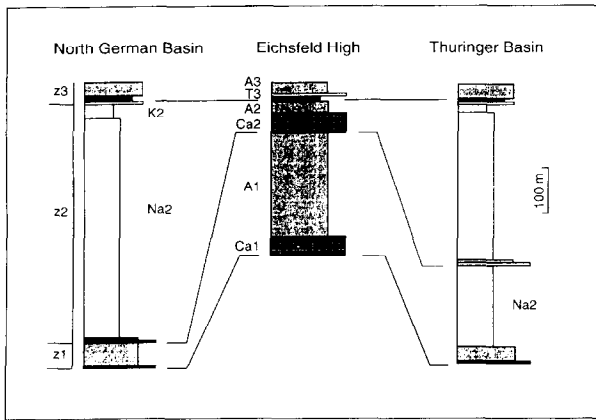
Fig. 4. Lithostratigraphic thicknesses of the Zechstein in a profile over the Eichsfeld Rise.

petrofabric analysis because all of the calcium sulfate has been transformed from gypsum to anhydrite, and much of it has undergone a second transition back to gypsum. In the Schimpf Quarry near Osterode, the overlying fine-grained dolomite layer (Ca2) was deformed by slumping, sliding and folding on the flanks of some of the piercement structures, suggesting that the sediments were incompletely consolidated when deformation occurred. Domal diapirs are observed in the Schimpf Quarry and the quarry near Ührde (Fig. 5) near the boundary of the carbonate and gypsum–anhydrite rocks,

and may form at these locations preferentially because the sediments of both lithologies are partially lithified. Slump blocks composed of anhydrite with monoclinal folds indicate partial lithification of what was at the time then bedded gypsum (Fig. 3d). The soft-sediment deformation features in Ca2 beds on the flanks of some of the gypsum–anhydrite piercement structures indicate that diapirism occurred very early, before burial to depths where temperatures would cause the conversion of gypsum to anhydrite.

An estimate of the time required for the calcium sulfate piercement structure to develop can be made using the time required to deposit the Stassfurt Carbonate Bed as a minimum. Using magnetostratigraphic correlation, Menning (1991) estimated the depositional period for the entire Zechstein to be less than 3 Ma, between 254 and 251 Ma ago. On the basis of these dates, an estimate for the maximum duration of the entire Stassfurt deposition is 0.5–1 Ma. Jackson and Talbot (1986) estimated the strain rate for Gulf Coast salt diapirs to be at least 10^{-16} s^{-1} in order for the structures to grow to their maximum height in 17–20 Ma. This strain rate would allow a diapir to develop about 30 m of relief in 1 Ma. If the Zechstein A1 calcium sulfate started to rise diapirically after at least half of the carbonate and some of the overlying sulfate (A2) was deposited, it would also have required a strain rate of at least 10^{-16} s^{-1}.
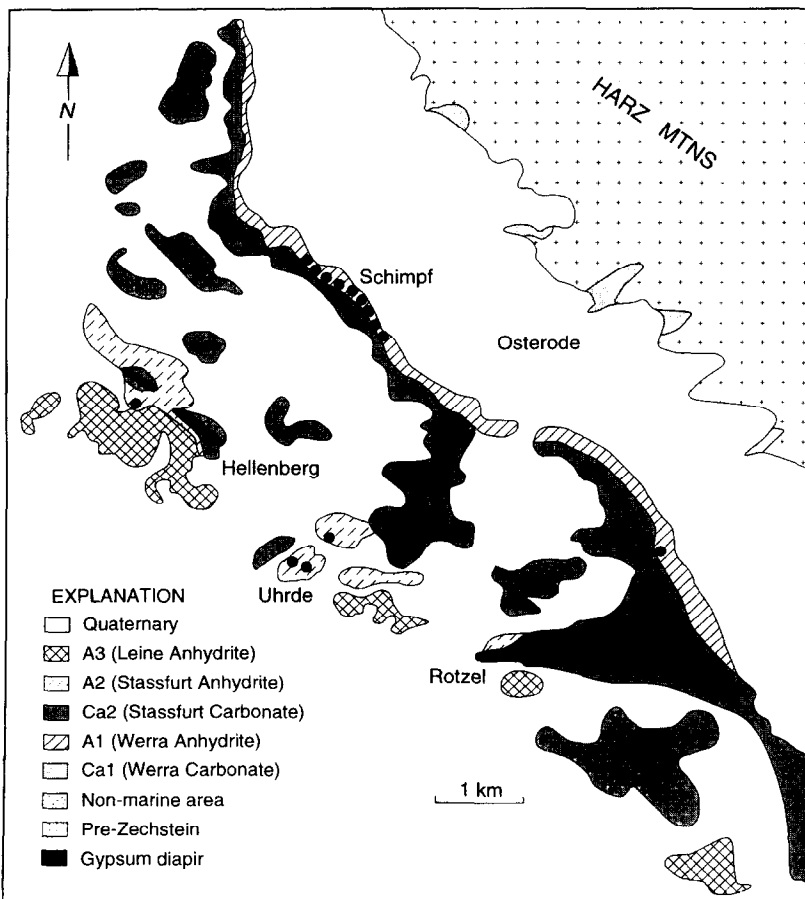

Fig. 5. Map of the locations of outcrops of the anhydrite piercement structures discussed in text. After Paul (1991).

Fig. 6. (a) Diapiric gypsum (white rock on the left side of the photograph) of the Werra Anhydrite Member intruding the nearly vertical Stassfurt Carbonate Bed. A rock hammer is given for scale. (b) Detail of deformed subvertical layering with disharmonic folding in gypsum. The diameter of the lens cap in the upper-right part of the photograph is 52 mm.

As was shown in Fig. 1, the flow law for gypsum (equation (1)) predicts a strain rate of $10^{-16} s^{-1}$ at differential stress of around 20 MPa. For halite, such a strain rate is dominated by fluid-assisted diffusional mass transfer. Given the fact that the sediments involved in the gypsum diapirism at this locality were partially unconsolidated, there was probably sufficient water present to allow flow of gypsum by the same mechanism, so that flow would occur at the required strain rate at even lower stresses.

The tectonic regime in which the diapirs formed may not have required gypsum to flow at lower stresses. In the quarry at Hellenberg, the slumping carbonates are deformed into a number of WNW-trending folds, subparallel to the trend of the Harz Mountains. The linear diapirs are also parallel to this trend. At the locus of the faults, gypsum could easily have protruded into the brittle carbonates by a differential loading process called 'reactive diapirism' by Vendeville and Jackson (1992). The general dip of the beds is to the southwest, which could indicate a transport direction to the northeast for the large slump block. Although thicknesses of the different facies vary in a complicated manner related to the development of carbonate and sulfate platforms, the basinal deep-water facies lie to the northeast, while southwest of the Harz Mountains the depositional facies of the Permian rocks grade from a shelf facies to a sabkha or subaerial evaporitic marginal marine environment (Richter-Bernberg, 1985). The large-scale regional tectonics of the Permian Zechstein involved syndepositional rifting as a result of NE–SW extension (Paul, 1994). The relatively small gypsum diapirs are a local expression of the regional tectonic regime. In addition, many of the major normal faults and grabens trend parallel to the axes of the linear diapirs at the Harz Mountain locality (von Schacl, 1991).

CALCIUM SULFATE DIAPIRS IN THE LEINE HALITE MEMBER

The halite diapirs in the Zechstein Basin are larger-scale structures than the gypsum diapirs at the Harz Mountain locality. The salt stock which occurs under the Allertal fault graben approximately 70 miles to the northeast is 250–300 m high and 1–2 km wide (von Schacl, 1991). Diapiric structures composed of the Main Anhydrite (A3) were observed in salt mines in the Leine Halite (Na3) of the Leine series of the Zechstein 3. These structures were first reported by Fulda (1929) who called them 'Anhydrit-klippen', which translates to 'anhydrite reefs'. Their appearance and possible origin were treated in detail by Hemmann (1972). The 'Anhydrit-klippen' are fine-grained anhydrite structures intruding recrystallized diapiric halite, and Hemmann attributed the origin of these structures to differential loading of the gypsum rocks due to variations in thickness. Hemmann (1972) interpreted the anhydrite structures to have formed

nearly syndepositionally, as halite accumulated on top of the gypsum and differential loading caused it to begin to rise diapirically into the halite. As the gypsum rose into the salt, the phase change from gypsum to anhydrite released water that intruded into Na3, resulting in the development of a recrystallization zone in the halite near the contact with the anhydrite 'reefs'.

The dependence of the stability of gypsum vs anhydrite on temperature and pore fluid pressure has already been discussed. However, the presence of Na3 halite as the overburden for the A3 calcium sulfate introduces another important factor for stability relationships: the salinity of pore fluids. The investigation of the transition temperature by Hardie (1967) was based on the activity of H_2O, which varies directly with the salinity of water. The dehydration of gypsum to anhydrite is observed in modern sabkha environments when gypsum is in contact with warm, high-salinity brines (Hardie et al., 1983). It is possible that the salinity of the gypsum pore fluids was high enough during deposition to initiate the dehydration reaction, thereby turning the gypsum layers into a water-laden anhydrite crystal mush with no flow strength. Possibly the released water flowed into the salt, dissolving and weakening it so that diapirism could initiate easily. Alternatively, because salt bounded the gypsum above, and clay below, the pore fluids in the gypsum layer could have been overpressured, inhibiting the transition reaction. In the second case, a metastable condition exists where a trigger is required to initiate diapirism. Kupfer (1989) considered gypsum dehydration to be an agent for halite diapirism. There is evidence in the salt stock below the Allertal Graben for the opposite to be true, that is, halite diapirism was the trigger for gypsum diapirism.

Hemmann (1968) reported that stratiform anhydrite banks up to 40 cm thick can be found some distance away from the Anhydrit-klippen. This indicates that conditions leading to the mobilization of what was probably originally gypsum did not occur ubiquitously in the basin, which might be expected for salinity-induced dehydration. During diapiric movement of the salt, the deformation of the gypsum beds within the salt triggered the dehydration reaction, causing the formation of mini-diapirs of gypsum within the main salt stock. In this case, the very low flow strength of the anhydrite–water mush allowed it to flow into the salt. The appearance of the anhydrite 'reefs' would suggest that the flow was influence by gravity, because all of the reefs form intrusions piercing younger salt. The presence of a clay layer just below the anhydrite may, however, have precluded the necessity for the salt to be in its original horizontal position when the anhydrite cliffs formed. The clay formed an impermeable layer, so that the predominant direction of movement of the released fluid was toward the top of the salt, or laterally as small stringers of anhydrite are found extending from the cliffs into the surrounding halite and anhydrite layers. Adjacent to the cliffs are zones of recrystallized halite aureoles, which

also extend laterally following bedding, indicating preferential paths followed by the released water. One of the more difficult problems encountered when attempting to find the trigger for diapiric movement of gypsum that has been converted to anhydrite is the timing of the dehydration reaction. Detailed petrography of the rocks can help but often, even with the addition of the rheological analysis of the rocks in question, the tectonic setting and some understanding of the depositional conditions, the timing of the transition reaction of the Leine Anhydrite remains equivocal.

GYPSUM DIAPIRISM IN OTHER SEDIMENTARY BASINS

It appears that in both examples studied in the Zechstein Basin the calcium sulfate rocks were the buoyant material involved in diapirism. Halite is not present under the diapiric structures in the Harz Mountain locality, and the depositional environments of the sediments indicate that halite was not deposited there. In the Sverdrup Basin of Arctic Canada, numerous diapirs originating in Carboniferous sediments composed of anhydrite or gypsum pierce the overlying sedimentary section (Gould and de Mille, 1964; Hoen, 1964). The diapirs are cored by halite, suggesting that the anhydrite was forced upward along with the halite as part of the diapir (Stephenson et al., 1992).

The question of what calcium sulfate phase was involved in the diapirism in the Sverdrup Basin is still a subject of debate. Because the relatively dense and brittle anhydrite should be pierced by diapiric halite, Schwerdtner and Osadetz (1983) proposed a diapiric model where the anhydrite was able to ride on top of the rising halite diapir because it was in the form of a crystal mush of anhydrite plus water. However, the preservation of primary gypsum depositional textures does not support the development of a crystal mush during diapirism. Similar to the anhydrite on shelf facies of the Werra Anhydrite Member of the Zechstein, much of the calcium sulfate in the Sverdrup Basin is interpreted to have been deposited as a shallow lagoon evaporite deposit, and still has textures that are very similar to those found in the Miocene Solfifera Series in Sicily (Hardie and Eugster, 1971). Preservation of the primary bedded selenite occurs when dehydration takes place after lithification of the gypsum, as would be the case for burial-induced dehydration. Stephenson et al. (1992) interpreted the halite–anhydrite diapirs to have initiated during syndepositional rifting during the Carboniferous and Permian periods. Growth of the diapirs then continued during following periods of horizontal compression. Stephenson et al. (1992) concluded that horizontal compression allowed continued diapiric movement even with high-density anhydrite as a major component, and that the ductility of anhydrite at temperatures above 100°C facilitated its flow along with the halite. Since the tectonic

history of the Sverdrup Basin after evaporite deposition starts with an extensional phase, it is likely that faulting above the evaporites initiated the diapirism (Jackson and Vendeville, 1994). The Carboniferous sediments overlying the evaporites are composed of limestones in some parts of the basin and shale, siltstone and limestone in others. The flow strengths of halite and gypsum are much lower than each of these sediments, and the early extensional tectonics probably allow both materials to flow into diapiric structures. The gypsum that cores the diapirs may never have been buried deeply enough to reach conditions for the transition to anhydrite. Rates of deposition estimated by Beauchamp et al. (1989a,b) indicate that in order for diapirs to have kept pace with sedimentation, they would have had to rise approximately 2800 m over 70 Ma. The minimum strain rate required is $10^{-16}\,s^{-1}$, which easily is attained in most diapiric environments (Jackson et al., 1994).

Diapirism involving gypsum and anhydrite in the Eastern Prebetic Foldbelt of southern Spain also appears to have been triggered by extensional tectonics (de Ruig, 1995). A significant difference in this basin is that the Triassic calcium sulfate deposits are associated with shales instead of with halite and carbonate rocks. de Ruig (1995) concluded that buoyancy could not have been involved because the anhydrite and shale of the diapiric material is denser that the overlying rocks. Re-sedimented Triassic gypsum and mudstones in Cretaceous rocks indicate the earliest time of piercement. The regional thickness of Cretaceous sediments is of the order of 1000 m, which de Ruig (1995) interpreted as an indication that the gypsum had been converted to anhydrite by mid-Cretaceous time. He concluded that the diapirism in the Eastern Prebetic Foldbelt was the result of the low viscosity of the Triassic evaporites and the fracturing of the brittle overburden. However, in this basin, as in the Zechstein and Sverdrup basins, it appears that diapirism could have been initiated early in its depositional history so that the gypsum in the diapirs could have remained relatively close to the surface and would not have been deep enough for the transition reaction.

DISCUSSION

The relationship of the gypsum–anhydrite transition to the development of gypsum diapirs can obscure the information needed to decipher their history. However, the timing of the transition, if it can be estimated, can actually be useful in determining the mechanisms of emplacement because of the additional constraints required on the system. Baumann and Stucky (1984) studied the undeformed gypsum horizons in the Bergwerk Felsenau of northern Switzerland and the deformed anhydrite of the same horizon in the Mettauer overthrust zone. They concluded that the differential stresses caused dehydration of the gypsum, that the water escaped into

clay layers and that the anhydrite formed a lithified rock which then continued to deform by faulting and fracturing in some places. Gypsum is intimately associated with the anhydrite of the same horizon, which Baumann and Stucky (1984) attribute to a higher halite content in the anhydrite layers. They suggested that after a certain amount of anhydrite was formed, the shear strain of the layers was halted and with it the transition to anhydrite. The maintenance of an anhydrite crystal mush proposed by Schwerdtner and Osadetz (1983) may only be short-lived under the conditions of a rising diapir before enough water is released for a dense, brittle anhydrite cap to develop.

The presence of the pore fluids is not as important for the Werra Anhydrite Member in terms of the dehydration of gypsum because the depths of burial are far below those required for the transition to anhydrite, but fluids are important in the facilitation of fluid-assisted creep. The rate-controlling deformation mechanisms for halite range from cross-slip of screw dislocations to climb of edge dislocations to fluid-assisted diffusive mass transfer at lower and lower stresses. In the Sverdrup Basin, the Eastern Prebetic Foldbelt and the Werra Anhydrite Member of the Zechstein Basin, the beginning phases of diapir growth were probably due to flow of gypsum by diffusion creep. As suggested above, if the diapirism in the central Sverdrup Basin was caused by differential loading, there is no critical overburden depth before a less dense, more ductile, layer can rise, and diapirism could have begun before gypsum reached conditions that would cause the transition to anhydrite. Because the difference in the flow stress of wet gypsum and wet halite under geological strain rates is negligible, it is very possible that the gypsum behaved exactly like halite during diapirism in the Sverdrup and Zechstein basins. Immediately overlying the calcium sulfate in all of these basins is a carbonate layer. This condition contributed to the likelihood of early diapirism because carbonate rocks tend to become cemented early in their depositional history, and are denser than most sediments. A significant density and flow strength contrast exists very soon after deposition, allowing diapirism of gypsum or gypsum plus halite to initiate and proceed close to the surface.

The case of the 'Anhydrit-klippen' in the Leine Halite of the Zechstein 3 is quite different to other instances of diapirism. Fluids released from the dehydration reaction from gypsum to anhydrite affected the rheology of both the halite and the gypsum–anhydrite. The fine-grained texture of the anhydrite could have formed during deformation and dehydration of the diapiric calcium sulfate, or it could have originally been fine-grained anhydrite or fine-grained gypsum. The location of the anhydrite layers within the basin suggest that the sediments were originally part of the finely-laminated basinal calcium sulfate facies. That the texture is massive in the anhydrite makes it difficult to determine the original texture and, consequently, the timing of the transition reaction. If the calcium sulfate were originally

fine-grained gypsum, dehydration may have taken place before diapirism if pore-fluid salinity was high enough, during diapirism if deformational stresses induced dehydration, or after diapirism and deeper burial if temperatures were not previously high enough. In any case, fluids played an important role, as indicated by the recrystallized halite aureoles around the anhydrite structures.

CONCLUSIONS

Study of the three sedimentary basins described here show that, although gypsum diapirism is not a common phenomenon, sets of geological conditions can exist where diapirism involving calcium sulfate can develop. The diapiric structures were the result of the upward rise of a buoyant layer possessing much lower flow strength than the structurally weakened brittle overburden. The geological conditions for the development of gypsum diapir structures require early extension of the basin so that diapirism can initiate before deep burial and dehydration of the gypsum. Except for the case where structures composed of anhydrite intrude halite, in all localities where calcium sulfate forms diapirs and pierces the overlying sediments the brittle overburden is composed of carbonate rocks. Early cementation of the carbonate may be the key, because the diapirs appear to have formed when the rock was gypsum and not anhydrite. Comparison of the rheological properties of gypsum and halite shows that the theoretical considerations of ductile flow of salt at geological strain rates may be completely applicable to the flow of gypsum rocks under similar conditions. However, because gypsum diapirism must occur relatively early in basin evolution history, when it is present it provides additional information regarding the timing of tectonic events, temperature and strain history within the basin.

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