

Badenian evaporite basin of the northern Carpathian Foredeep as a drawdown salina basin

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ABSTRACT:

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The Middle Miocene (Badenian) Ca sulphate-dominated evaporites of the northern Carpathian Foredeep (in Ukraine, Poland, the Czech Republic and Moldova) are interpreted as deposits of a giant and shallow salina basin developed in northern Central Paratethys during the Badenian salinity crisis. The predicted features of a salina basin model are discussed and compared with the actual geological record of the Badenian evaporites. The main depositional features of a salina basin, are: (i) evaporite drawdown; (ii) rapid and large fluctuations in basinal water level; (iii) presence of both shallowing-upward and deepening-upward depositional sequences; (iv) aggradational deposition; (v) variable marine, non-marine, and/or mixed characteristic of brine; all of which are recognised in this Badenian basin. Aggradational evaporite deposition was controlled by water or brine level fluctuations within the basin or subbasins. Because the basin was separated from the sea by some emerged barriers these fluctuations were only weakly dependent on world-wide sea-level changes but were rather controlled by regional climate.

Key words: Gypsum evaporites, Basin model, Salina, Evaporite drawdown, Badenian salinity crisis, Carpathian Foredeep, Paratethys.

INTRODUCTION

Models of evaporite basins are essential for understanding the geological record of chemical deposits precipitated from brines and permit understanding and prediction of the distribution, content and architecture of geological strata both within and around such basins. Many earlier models applied to ancient evaporite basins are abandoned or questioned as unrealistic in the face of new geological data and sedimentological studies of modern environments (see SONNENFELD 1984, GOODALL & *al.* 1992, KENDALL & HARWOOD 1996). Since the proposal that the Messinian evaporites could be a product of the nearly total drying out of the Mediterranean Sea (CHUMAKOV 1971, HSÜ & *al.* 1972), much investigation led to new interpretations of many ancient evaporite basins as drawdown basins similar to

the Messinian Mediterranean. Attempts were also made to create a reasonable model of such a drawdown basin based on the concept of sequence stratigraphy (TUCKER 1991, WARREN 1999), and observations in recent salt lakes (KIRKLAND 2003, BAŁEŁ 2004). The present paper compares predicted features of the drawdown basin model with the geological record of the Badenian Ca sulphate-dominated evaporites in the northern Carpathian Foredeep and suggests that they could also have been deposited in such a variable basin. The idea presented is crucial for future research and understanding of the geological history of the Carpathian Foredeep, particularly during the Badenian salinity crisis. The depositional dynamic and facies architecture of a drawdown basin differ from those in 'traditional' evaporite basins fully connected with the sea. The concept presented can better explain many fea-

tures of the Badenian basin and solve some controversies concerning the origin of the Badenian evaporites.

The following paper is divided into two parts. The first part is devoted to the theoretical characters of selected diagnostic features of a shallow drawdown evaporite basin. The second part shows that these features can be found in the sedimentary record of the Badenian evaporites. This second part is also a brief review of the present state of investigation of the Badenian Ca sulphate deposits in the northern Carpathian Foredeep.

SALINA BASIN VERSUS EVAPORITE LAGOON MODEL

So far, the most commonly accepted model of evaporite deposition is a classic barrier lagoon model that assumes the existence of a narrow strait separating the evaporite basin or salt lagoon from the ocean (see GRABAU 1920, DRONKERT 1985). This model is hydrologically open. The marine water flows into the basin as surface current and the brine can outflow as refluxing bottom current (Text-fig. 1). The size of the strait and the speeds and volumes of these brine and water currents are essen-

tial factors controlling the course of evaporite deposition in the basin (e.g. SONNENFELD 1984). The model is attractive because the assumed reflux brine current is the simplest explanation for the 'escape of salts' from the system - a necessary condition for the deposition of ancient evaporite sequences, which nearly always show different proportions of salts from those expected from complete evaporation of seawater (e.g. KLEIN-BEN-DAVID & *al.* 2004). This popular model of an evaporite basin was criticized as hydrologically unsound and unrealistic (SHAW 1977; KENDALL 1988; KENDALL & HARWOOD 1996; WARREN 1999; see also BRONGERSMA-SANDERS 1971). Recent literature focuses on the drawdown basin model, which assumes both inflow of seawater and escape of brine through underground seepage. This second model (Text-fig. 1), referred to here as a salina-type basin model or salina model (cf. GRABAU 1920, LOTZE 1938, BORCHERT & MUIR 1964), is hydrologically more restricted. Because the basin is not fully connected with the sea, the brine level does not necessarily coincide with the world sea-level fluctuations and the basin is, in fact, a saline lake.

GENERAL FEATURES OF THE SALINA BASIN MODEL

The salina basin model assumes that: (1) the evaporite basin was a nearly closed depression separated from the ocean by some topographical barriers; (2) evaporation was the main reason for water deficit and for chemical sedimentation; (3) the sea water entered the basin by seepage and/or occasionally by direct inflows over the barriers. Additionally, it can be assumed that (4) the salina basin (like the Badenian basin) was a system of interconnected or temporarily disconnected subbasins.

The salina model for evaporite deposition belongs within a large group of hydrologically closed or semi-closed models of evaporite basins (see reviews in HSÜ 1972, SONNENFELD 1984, DRONKERT 1985), known under various names (Tab. 1). The salina basin discussed is closely related to the model of interconnected evaporite subbasins introduced by BRANSON (1915, see also DRONKERT 1985) and developed by BORCHERT & MUIR (1964, fig. 5.1) and BORCHERT (1969). The variable processes operating in the salina-type basin were discussed by many authors (HOLSER 1979; LEPESHKOV & *al.* 1981; JAUZEIN & HUBERT 1984; TRASHLIEV 1984b; LOGAN 1987; WOOD & SANFORD 1990; TUCKER 1991; GOODALL & *al.* 1992; KENDALL 1988, 1992; WARREN 1999; KIRKLAND 2003; KLEIN-BEN-DAVID & *al.* 2004). A drawdown salina basin model was applied to many ancient giant evaporite basins (see HSÜ 1972, ROUCHY 1982, TUCKER 1991, WILLIAMS-STROUD 1994, KIRKLAND & *al.* 2000, ROUCHY & *al.* 2001, KIRKLAND 2003) and was already accepted for some

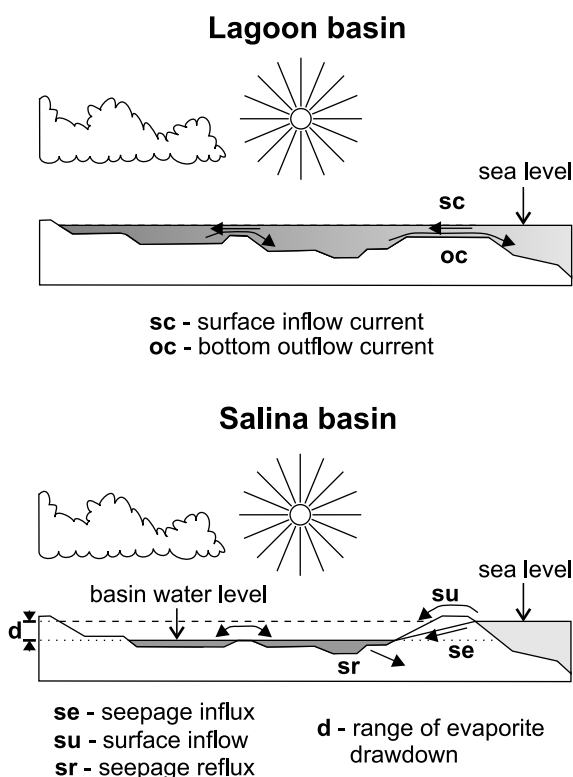


Fig. 1. Salt lagoon (barrier) and salina (drawdown) model of evaporite basins

Badenian evaporites in Ukraine (PANOV 1983, p. 101), Bulgaria (TRASHLIEV 1984a, fig. 2) and Romania (MARINESCU & MÁRUNTEANU 1990). The model has many recent analogues in coastal salinas around the world although they are very small in size in comparison with the ancient saline giants. The MacLeod salina, Australia is the best known example of a functional marine, seepage-inflow salina basin (LOGAN 1987), whereas the Kara Bogaz Gol, Turkmenistan, during the recent lowstand in 1947-1980 represented a non-marine, surface-inflow salina basin (LEPESHKOV & *al.* 1981, table 5).

The model discussed below is simplified but can be readily applied to the Badenian evaporites. It consists of a shallow basin containing selenite deposits and lacking extensive gypsum slope facies ('gypsum wedge'). The influence of tectonics on evaporite sedimentation is not considered. Like any other model, this one can be validated, improved or abandoned in the light of new data.

The salina basin shows several different features from a simple saline lagoon. They include: 1) evaporite drawdown; 2) water-level changes not (or poorly) dependent on sea-level fluctuations; 3) sedimentary dynamics controlled mainly by climate, and poorly or only occasionally by sea-level changes; 4) high accommodation not dependent on sea level; 5) aggradation reflected as 'autocyclic transgression' or an 'intrabasinal transgressive systems tract' (WARREN 1999, p. 95); 6) variability of possible brine types (marine, non-marine, and/or mixed). The features listed are discussed below in detail.

Evaporite drawdown

The first phase in the evolution of a salina basin is its separation from the sea and an ensuing *evaporite drawdown* (MAIKLEM 1971). The water level in the basin possibly falls at a rate similar to that of recently shrinking salt lakes, from several cm/year to over 0.5 m/year (WILLIAMS 1996). The basin, however, does not dry out completely. The water level falls down to a state of equilibrium when evaporation (and reflux) will be balanced by water influx to the basin (JAUZEIN & HUBERT 1984, KRUMGALZ & *al.* 2000). Evaporite drawdown in an enclosed basin is much more rapid than sea-level fall (TUCKER 1991). Thus, thick regressive clastic sediments apparently do not form as in the Lake Asal salina, Republic of Djibouti (GASSE & FONTES 1989). Most isolated basins are almost devoid of tides. A rise in the salinity and density of the water roughly coincides with the advance of drawdown, although this depends on many factors like inflow/reflux ratio (WOOD & SANFORD 1990, SANFORD & WOOD 1991) and air humidity (KLEIN-BENDAVID & *al.* 2004). The waves and wave base level become smaller and less pronounced in denser and shallower brine (KWIATKOWSKI 1972, p. 69;

SONNENFELD 1984, p. 54) and hence clastic deposition is gradually reduced. When the rising salinity of the brine reaches gypsum saturation deposition of this mineral begins. In shallow basins, like in Australian and Egyptian coastal salinas (WARREN & KENDALL 1985, LOGAN 1987, ORSZAG-SPERBER & *al.* 2001), the marginal 'gypsum wedge' predicted by models of deep-water salina basins (TUCKER 1991) does not form or is negligible (restricted to small subordinate lagoons or pans at the margin of the basin). Gypsum is deposited at the bottom of the lowest areas of the shallow basin on the surface inherited from the pre-evaporite marine stage. The margin of the basin is composed of emerged marine deposits which formed directly before the onset of evaporation. Gypsum is deposited in a similar manner to gypsum deposition in many recent salinas - by aggradation of horizontal layers covering the bottom of the depression (ARAKEL 1980, WARREN 1982, WARREN & KENDALL 1985, AREF & *al.* 1997, AREF 1998, ORSZAG-SPERBER & *al.* 2001). Variations in bottom relief, and the presence of deeper subbasins can complicate this scenario.

Water-level changes and sedimentary record

Unlike in marine lagoons, the water level in a salina basin does not coincide with, and is not directly controlled by the worldwide sea level. A salina basin is technically a lake that receives a substantial influx of marine water from the sea (WILLIAMS-STROUD 1994, KIRKLAND & *al.* 1995, KIRKLAND 2003). The behaviour of water level and depositional dynamic in a salina basin, especially in seepage-influx basins, are thus similar to salt lake systems (GOODALL & *al.* 1992, BABEL 2004). The water level and sedimentation are controlled rather by climate, not by sea-level changes. The salina basin probably responds to climatic cycles, and its water level and salinity rise and fall as commonly recorded in lakes. The water-level changes in a salina basin differ from those of the sea both in time and space. They are more rapid and larger (TUCKER 1991), and presumably also more frequent. During lowstand a salina basin is commonly divided into many subbasins by emerged barriers. The water levels controlling sedimentation in particular subbasins can thus differ, further complicating the sedimentary record. Nevertheless, even small sea-level changes can control the rate of marine water influx to the basin both through seepage (TUCKER 1991, GOODALL & *al.* 1992) and surface inflow through channels (LEPESHKOV & *al.* 1981). The seepage influx rate is highest during maximum drawdown, and gradually ceases during the basin infill (KENDALL 1992). Nevertheless, the seepage rate is probably more constant than inflow via open channels and can be stable through thousands of years (KIRKLAND 2003).

Sometimes catastrophic influxes of sea water into a given depression lead to rapid and unexpected water-level rises. Complete closure of barriers blocking the marine water influx can lead to rapid drying out of a shallow basin (in dry climate) or its transition into a brackish lake (in wetter climate).

The salina basin is presumably affected by the large-scale, climate-controlled fluctuations in water level as documented in many ancient salt lakes (e.g. MAINGUET & LÉTOLLE 1997). Alternate highstands and lowstands are crucial for chemical sedimentation in the basin and can be recorded by shallowing-upward and deepening-upward units and emersion surfaces. The hydrological structure of the basin is complicated because of the existence of subbasins and density stratification of the brine. Only during highstands are subbasins fully connected; during lowstands they are partly or completely separated. The hydrology of stratified brines, brine transport and brine mixture within and among such subbasins are the subject of modelling in a separate paper (BABEL 2004).

Accommodation

Accommodation is defined as 'the space made available for potential sediment accumulation' (JERVEY 1988 discussed in MUTO & STEEL 2000). Extremely shallow-water sediments observed in marine sequences suggest that accommodation for the depositional system is low and erosion or non-deposition dominate over accumulation. Therefore emersion surfaces in such marine sediments are very common. This is not true in the case of the shallow-water sediments of a salina basin. The accommodation in a salina basin, unlike in saline lagoons, is large, and erosion and emersions can be subordinate or lacking even during extremely shallow water deposition.

The shallow-water sedimentation in the salina basin is specific. The basin is not connected with the sea and sedimentation is controlled mainly by water or brine levels within the basin. The water levels are lower than the sea level. They can easily rise together with sedimentation and basin infill. Maximum water-level rises in the salina basin take place during catastrophic floods of marine water. Giant floods of meteoric water are less probable within an evaporite environment. In the deepest depressions, erosion caused by floods is practically absent. Erosion can operate there only during complete drying out of the basin, as deflation or dissolution. Total drying out of the deep drawdown basin requires, however, an extremely arid climate. Therefore, shallow water or subaerial deposits are recorded in a salina basin in thick continuous sequences like the tens of metres-thick deposits of some ancient Chilean salars, which are composed exclusively of

efflorescent and pedogenic halite (LOWENSTEIN & *al.* 2003), or the 25 m-thick pedogenic gypcrettes from the continental Calama Basin in Chile (HARTLEY & MAY 1998). The shallow water deposits of a salina basin do not represent marine lowstand but internal lowstand within the basin (WARREN 1999).

The vertical span of accommodation in a salina basin is open to discussion. Accommodation can be defined by the entire space within the depression, or the space limited by mean sea level (equilibration level by LOGAN 1987, pp. 8-9). The accommodation can be limited by groundwater-level or the water levels of saline pans, or by mean brine level (LOGAN 1987, p. 9). The 'minimum' accommodation can be controlled by the position of pycnoclines in saline pans and by subsurface brine tables in emerged margins of the pans. Such a limit is justified by the fact that evaporite salts above the pycnocline are normally subjected to dissolution and only those below the pycnocline are preserved (cf. WARREN 1999, BOBST & *al.* 2001).

Aggradation

The general pattern of gypsum deposition in a salina basin is aggradation (cf. LOGAN 1987, WARREN 1999) i.e. 'strictly vertical accumulation of sediment' (MATTHEWS 1974, pp. 71-72). The aggradation phase of deposition follows the initial drawdown phase in the shallow salina basin. Evaporite deposition starts in the shallow brine in the interior areas of a drawdown basin and salts infill the depression 'layer by layer'. The geometry and course of this aggradation is dependent on the mean brine levels and water table levels in the basin. The evaporite deposits thus generally show layer-cake stratigraphy. Aggradation is reflected by a general onlapping pattern of the younger evaporite strata. This onlap can be obscured by karst in the contact zone of evaporites with meteoric waters and groundwaters on the basin shores (KENDALL 1988). The aggrading pattern can be modified by autocyclic lateral migration of facies. The aggrading and onlapping gypsum deposits under discussion are an equivalent of 'autocyclic transgression' or an 'intra-basinal transgressive systems tract' distinguished by WARREN (1999, p. 95) within basinal halite deposits in deep salina basins.

Shallowing-upward and deepening-upward sequences

The rate of evaporite gypsum and halite sedimentation may be very rapid in comparison with average rates of subsidence and sea-level change. It is therefore expected that in evaporite environments controlled by sea level, like saline lagoons or coastal sabkhas, shallowing-upward

sequences are typical, and are either the only type of sequence represented or the predominant one. Indeed, such sequences were documented in a few ancient marine lagoon and sabkha deposits (e.g. BOSELLINI & HARDIE 1973, TUCKER 1999b) but appear to be absent, or scarcely recognizable, in the most of the giant evaporite basins. Shallowing-upward, up to 10 m-thick gypsum sequences were recognised in some recent salinas cut off from direct sea-level control (WARREN 1982, AREF & *al.* 1997). They were indicated as typical of salina basins, with a suggestion that some ancient shallowing-upward salina sequences can be as much as 30 m thick (WARREN & KENDALL 1985, fig. 6). However, many other recent and subfossil salinas contain only extremely shallow-water gypsum deposits and shallowing-upward sequences are poorly recognised (AREF 1998). Subaqueous salina deposits can show a record of rapid short-term emersions (ORSZAG-SPERBER & *al.* 2001). Salina basins are technically lakes and therefore their water-level rises can be extremely quick. It seems that in larger saline basins, because of the greater accommodation, the rate of water or brine level rise can easily keep pace with evaporite deposition rate and can even be much higher. Therefore, thick monotonous sequences composed of the same shallow-water facies as well as deepening-upward sequences should be recorded in the large salina basins. An example of a 12 m thick-, deepening-upward gypsum sequence supposedly deposited in a Messinian salina was documented by YOUSSEF (1988) and AREF (2003) in Egypt. The sequence also shows features of an upward increase in salinity. KENDALL

(1992, fig. 37) described other examples of such deepening-up and brining-up sequences. These sequences appear to be hardly possible in environments directly controlled by sea level within salt lagoons. Consequently, deepening-upward and brining-upward sequences seem to indicate a salina basin or salt lake environment. The water level rise associated with the salinity rise is observed in the Salton Sea in California (Cook & *al.* 2002).

Brine

In a salina-type basin the main source of salts is seawater and therefore both brine and evaporite salts should generally show the marine geochemical characters that are observed in recent coastal salinas (LOGAN 1987). Salinity rise should lead to deposition of evaporite salts following the well known Usiglio sequence (STEWART 1963). Modelling and field observations from the MacLeod salina revealed, however, that this depends on the inflow/outflow (seepage outflow in a salina basin) rate, which controls the maximum salinity level in the basin and thus limits the possibility of the precipitation of higher salts (VALIAEV 1970, SONNENFELD 1984, LOGAN 1987; WOOD & SANFORD 1990). Changes in the inflow/outflow rate influence the order of the precipitated salts and, for example, gypsum can be deposited after halite, as documented in the MacLeod salina (LOGAN 1987; KENDALL & WARREN 1988, fig. 2.35; KENDALL & HARWOOD 1996). The inflow/outflow rate also controls the thickness of deposit-

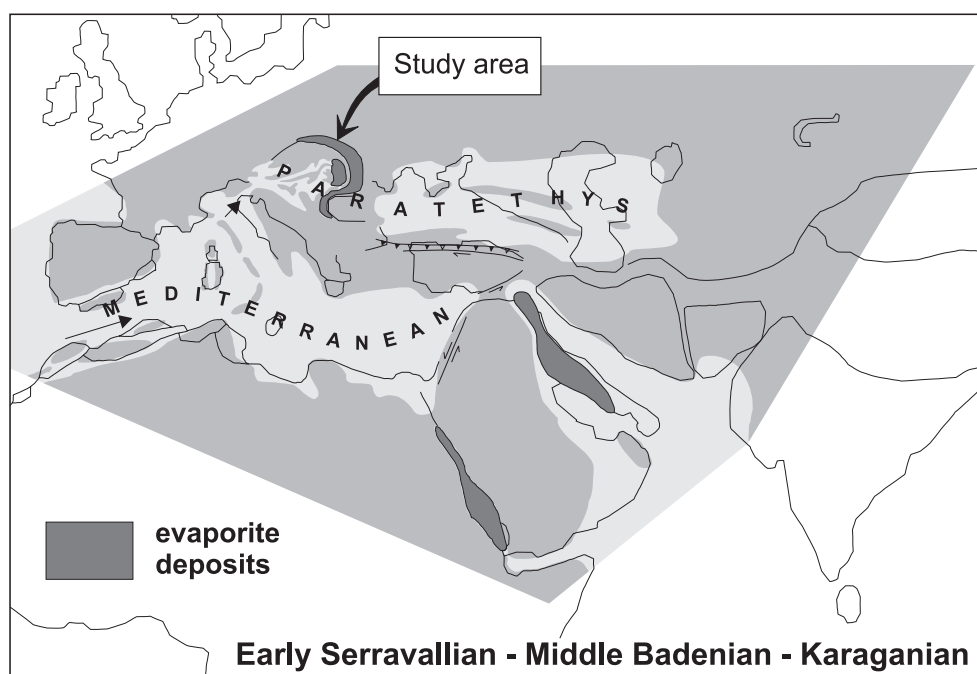


Fig 2. Palaeogeography of Paratethys and Mediterranean in the Middle Miocene; after RÖGL (1998)

ed evaporite salts, and sometimes there is only a small accumulation of gypsum before halite (SANFORD & WOOD 1991). In the case of limited seepage outflow, the Na^+ , K^+ , Mg^{2+} and Cl^- ions not involved in the initial Ca carbonate and Ca sulphate precipitation, and SO_4^{2-} not fully used for Ca sulphate precipitation, can accumulate in the brine, and the basin has a great potential for the deposition of NaCl and K-Mg salts (see HITE 1970).

Long-term chemical evolution of marine brine in a salina basin can lead to a brine showing the characters of a non-marine brine (KLEIN-BEN-DAVID & *al.* 2004). When inflowing continental or other non-marine waters show ionic composition and/or ionic proportions different from those of marine water for a sufficiently long time, mixing can change the marine proportions of ions and produce mixed brine (HARDIE 1984). The ionic composition of brine changes with time principally because of the precipitation of succeeding evaporite minerals and various back- and early-diagenetic reactions with chemical sediments.

During advancing evaporation all these processes selectively remove particular ions from the brine (see VALIASHKO 1962, EUGSTER & HARDIE 1978). The other causes of chemical changes are dissolution and re-precipitation (recycling) of earlier or older salts. In the case of variations in bottom relief, the brine can evolve in different ways in each subbasin or parts of the same salina basin. Refreshment of brine is possible due to increased influx of both marine water and meteoric water. In relatively wetter climate, brackish subbasins can develop on the landward side of a salina basin (as in the model by SELLI 1973, fig. 10). They also can appear in the final stages of evolution of the salina basin, when the water level in the basin is at sea level and the influx of marine water is minimum (see KENDALL & HARWOOD 1996, fig. 8.7). In summary, a salina basin can show various types of brines depending on place and stage of basin evolution – from strictly marine to non-marine; mixed and even brackish, in the case of wet climate.



Fig. 3. Present distribution of Badenian evaporites in the Carpathian area (after KHRUSHCHOV & PETRICHENKO 1979, GARLICKI 1979, modified)

Many ancient evaporite basins commonly show both marine and non-marine physical features of sediments, contain rare both marine and non-marine fossils, and reveal unequivocal geochemical characteristics of salts pointing to both marine and non-marine derivation of the brine. As shown by KIRKLAND & *al.* (1995, 2000) and DENISON & *al.* (1998), these contradictions are possibly easy to resolve assuming that the basin was of a salinity type and the brine was neither exclusively marine nor exclusively non-marine but was a mixture of marine and various nonmarine waters.

THE BADENIAN SALINITY CRISIS IN THE NORTHERN CARPATHIAN FOREDEEP (CENTRAL PARATETHYS)

The Badenian salinity crisis was a crucial event in the history of the Central Paratethys (RÖGL 1999). In Badenian times the seas occupying the area of the emerging Carpathian orogen lost their open connection with the Mediterranean and were transformed into evaporite basins with an impoverished and/or largely absent marine fauna (Text-fig. 2). These basins were re-flooded with marine water and colonized with new fauna at the end of the crisis. Later they became brackish Sarmatian Paratethys lakes. The Badenian salinity crisis occurred at ca. 12.5 Ma (OSZCZYPKO 1999), which preceded, by 6-7 Ma years, the well-known Messinian crisis in the Mediterranean. The Badenian events took place during the global cooling and general trend of sea-level fall following the Middle Miocene climatic optimum, at ca. 16 Ma (LOURENS & HILGEN 1997, BICCHI & *al.* 2003). The Badenian evaporite basins occupied depressed areas both within and surrounding the emerging Carpathians. The ensuing orogeny, overthrust and uplift of the mountain chain largely destroyed the original sedimentary record of these basins (Text-fig. 3). This is why the Badenian palaeogeography and connections between particular evaporite basins and, more generally, with the Mediterranean, remain controversial (compare KRACH 1962; PAUCA 1968, fig. 1; PISHVANOV & TKACHENKO 1971; KWIATKOWSKI 1972; SONNENFELD 1974, fig. 7; RÖGL & MÜLLER 1978, fig. 5; STEININGER & *al.* 1978, fig. 25; GARLICKI 1979; KARPENCHUK 1979, fig. 3; KHRUSHCHOV & PETRICHENKO 1979, fig. 2; KHRUSHCHOV 1980; PANOV 1983, fig. 26; TRASHLIEV 1984a; POPESCU & GHEȚA 1984; RÖGL & STEININGER 1984, fig. 10.9; KITYK & PANOV 1985, figs 2-3; PETRICHENKO 1988, fig. 18; OSZCZYPKO & ŚLĄCZKA 1989, fig. 5; MARINESCU & MĂRUNȚEANU 1990; POŁTOWICZ 1993, figs 2, 9; PERYT 1996, 2001; OSZCZYPKO 1998, fig. 14; RÖGL 1998, plate 8; 1999, fig. 9; ANDREYEVA-GRIGOROVICH & *al.* 2003). Basically, the hydrological and depositional system of the Badenian evaporite basins and

their sedimentary history are not well understood and this and the companion studies to this paper (BABEL 2004a in press, 2004b in press) are an attempt to clarify some of the confusion.

Of the Carpathian Foredeep its part stretching between the Czech Republic and Poland and further east to the Ukraine, is the largest and best elaborated (Text-fig. 4). The evaporites are present as laterally continuous layers of gypsum, anhydrite, halite and carbonate deposits which locally exceed 60 m in thickness (Tab. 2). The evaporites are underlain by marine Badenian clastics and carbonates, up to several tens of metres, resting transgressively on an eroded, mostly Mesozoic and Palaeozoic substrate, and are overlain by marine to brackish Badenian-Sarmatian clastics reaching up to 4 km. The Ca sulphate evaporites predominate to the north, while the halite evaporites occur in the south. The evaporites contain economically significant native sulphur deposits associated with the widespread gypsum and carbonate facies in the north (KLIMCHOUK 1997; GASIEWICZ 2000, with references). Undisturbed primary gypsum deposits with spectacular selenite facies crop out along the northern margin of the basin and are well documented (LILL DE LILIENBACH 1833, ATLAS GEOLOGICZNY GALICYI 1885-1914, NOWAK 1938, KUDRIN 1955, KWIATKOWSKI 1972, KUBICA 1992), especially in the last decade (KASPRZYK 1993a, 2003; PERYT 1996, 2000, 2001; PETRICHENKO & *al.* 1997; KASPRZYK & ORTÍ 1998; ROSELL & *al.* 1998; BABEL 1999a, b). To the south, they pass into anhydrite and halite deposits which are deeply buried in the axial area of the Foredeep (up to 4.5 km, BUROV 1963). Originally up to ca. 50 km wide, at the southern margin of the basin (OSZCZYPKO 1981, p. 69; OSZCZYPKO & ŚLĄCZKA 1989; POŁTOWICZ 1993), this part of the record is both tectonically and erosionally destroyed, and the evaporites occur both in situ below the flysch nappes, within folds, and resting on tectonically displaced nappes along the Carpathian overthrust. These folded evaporites are clearly seen in several salt mines, including the old Wieliczka mine, which has been open since the thirteenth century (GAWEL 1962, GARLICKI 1979, KOLASA & ŚLĄCZKA 1985). Sparse radiometric data from tuffites under- and overlying halite strata near the environs of Bochnia suggest a limited time span from ca. 13.4 to 11.1 Ma for possible evaporite deposition (VAN COUVERING & *al.* 1981, BUKOWSKI 1999, WIESER & *al.* 2000). The biostratigraphic investigation indicates that the Badenian evaporites roughly represent nannoplankton zone NN6 (GAŹDZICKA 1994; PERYT & PERYT 1994a, b; PERYT & *al.* 1997b; PRISYAZHNIUK & *al.* 1997; PERYT 1999a, b; MĂRUNȚEANU 1999; RÖGL 1999; ANDREYEVA-GRIGOROVICH & *al.* 2003, with references therein), which extends between 13.6-11.8 Ma (BERGGREN & *al.*

Name of the model	Ancient or recent example	References
Marine salina	Lake Larnaca, Cyprus	GRABAU 1920
Seepage influx model	Lake Assal, Republic of Djibouti	DEGOUTIN 1922, DRONKERT 1985, GASSE & FONTES 1989
Desiccated deep basin or shallow-water deep-basin or deep dry basin	Messinian (Upper Miocene) of the Mediterranean	HSÜ 1972, SONNENFELD 1984, BENSON & RAKIC-EL BIED 1991, KENDALL & HARWOOD 1996
Seepage-inflow and surface-inflow basins	MacLeod salina, Australia	LOGAN 1987, WARREN 1991, KENDALL 1992
Sequence-stratigraphic model of carbonate-evaporite basins with incomplete and complete drawdown	Zechstein (Upper Permian), Central Europe	TUCKER 1991, GOODALL & <i>al.</i> 1992, references in WARREN 1999
Drawdown basin	Paradox Basin (Upper Carboniferous), USA	WILLIAMS-STROUD 1994
Closed-basin, deep-marine-brine model	Castile Formation (Upper Permian), USA	KIRKLAND & <i>al.</i> 2000, KIRKLAND 2003
Seawater-fed terminal evaporitic lake	Dead Sea Rift valley (Neogene), Israel	KLEIN-BENDAVID & <i>al.</i> 2004

Tab. 1. Selected examples of salina-type evaporite basin models

1995) or 15-11.7 Ma (see BÁLDI & *al.* 2002, BICCHI & *al.* 2003). The Badenian evaporites were tentatively correlated with the global sea-level lowstand related to HAQ's & *al.* (1987) cycle TB2.6 at ca. 12.5 Ma (OSZCZYPKO 1998, 1999; MILLER & *al.* 1996).

The regional stratigraphy, palaeogeography, and sources of brine for the Badenian evaporites in the Carpathian Foredeep remain controversial (see KWIATKOWSKI 1972; GARLICKI 1979; LISZKOWSKI 1989; POBEREZHSKYI 1991, 2000; POŁTOWICZ 1993; SMIRNOW & *al.* 1994; PERYT 1996, 2001; PETRICHENKO & *al.* 1997; KASPRZYK & ORTÍ 1998; PRYSJAZHNJUK 1998). Recently it is accepted that evaporite deposition in the Polish Carpathian Foredeep took place in several subbasins; halite-dominated and containing laminated anhydrites and clay intercalations to the south, and gypsum-dominated and displaying widespread selenite facies in the north (KRACH 1956, GAWEL 1962, CZAPOWSKI 1994, GARLICKI 1994, KASPRZYK 2003). The gypsum subbasins were very shallow – 0 to several metres deep; the halite basins are considered to have been deeper (KRACH 1962, GARLICKI 1979, POŁTOWICZ 1993, SMIRNOW & *al.* 1995, KASPRZYK & ORTÍ 1998, KASPRZYK 2003). Unlike the marginal gypsum basins, which have some prominent and recognisable emersion surfaces, no emersion surfaces were recognised in these buried halite-bearing evaporites (GARLICKI 1979, 1994; POŁTOWICZ 1993; PERYT 2000; KASPRZYK 2003). Nevertheless, the presence of primary growth zonation in some halite crystals suggests deposition in brine shallow enough for it to have been subjected to temperature fluctuations (BUKOWSKI 1994, GALAMAY 1997). Scarce oscillation

ripples (CHARYSZ & WIEWIÓRKA 1977) suggest a depth below wave base. The depth of the halite subbasins was estimated as less than 30-40 m (KHRUSHCHOV & PETRICHENKO 1979, BUKOWSKI 1994), or even less than 15 m (PETRICHENKO 1988, tab. 13; SHAIDETSKA 1997).

Because the study area belongs to the northernmost and most landward part of the Paratethys, it could have been supplied indirectly with saline water, not directly from the Mediterranean Sea, but via intermediate evaporite subbasins (*cf.* ALEKSENKO 1961, 1967, BOBROVNIK 1962). The Carpathian Foredeep evaporite basin could have had a complicated pattern of marine brine influx. Brine could flow into the basin only from the east (KRACH 1962, KWIATKOWSKI 1972, KITYK & PANOV 1985), both from the west and from the east (GARLICKI 1979; KHRUSHCHOV 1980, PETRICHENKO 1988, PRYSJAZHNJUK 1998), or additionally also from the south across the orogenic arc (see VENGLYNSKYI & KOPYSTIANSKAYA 1979; RÖGL & STEININGER 1984, fig. 10.9; POŁTOWICZ 1993, figs 2, 5).

Palaeobotanical data from the northern Carpathian Foredeep indicate that the climate during the Badenian salinity crisis was of the Mediterranean type: warm and relatively humid (RANIECKA-BOBROWSKA 1957, ILJINSKAJA 1962, KITA 1963, OSZAST 1967, KOLASA & ŚLĄCZKA 1985, PRYSJAZHNJUK & *al.* 1997, SŁODKOWSKA 2004; see also KWIATKOWSKI 1972, LISZKOWSKI 1989). The climate was not extremely dry, but this does not exclude an evaporite origin of the salts (SCHMALZ 1971, WALTON 1978). Mediterranean climate is characterised by wet and cool winters and dry hot summers, and annual evaporation can be higher than annual precipitation – a factor which is fundamental for evaporite deposition.

MODEL FOR THE BADENIAN EVAPORITE BASIN

Most previous authors accepted the classic barrier model of evaporite sedimentation, and believed that the evaporites were deposited in drying out marine lagoons developed in front of the Carpathian orogen during regression of the Badenian sea (e.g. TEISSEYRE 1921, GAWEL 1955, KOWALEWSKI 1958, KRACH 1962, PRYSJAZHNJUK 1998, cf. DROMASHKO 1955, SMIRNOV & *al.* 1995, OSZCZYPKO 1998, KASPRZYK 2003). Using this model, it was commonly assumed that fluctuations in sea and/or meteoric water influx to the evaporite lagoons were responsible for differentiated or cyclothemetic deposition of the salt layers (e.g. TKACHUK & *al.* 1960, GAWEL 1962; VENGLYNSKY & GORETSKYI 1966). Other authors believed that the Badenian evaporites were deposited in salt lakes (e.g. ŁASZKIEWICZ 1936), in a Kara Bogaz-type 'lagoon' without return brine outflow to the sea (TIEZE 1882 in SIDORENKO 1904; KUDRIN 1955; STRACHOV 1962, p. 315; cf. TRASHLIEV 1969a, GONERA 2001), or in a salina-like depression nearly totally separated from the sea by barriers (PERYT & *al.* 1995, PERYT 2001).

The above two groups of views are essentially different and represent two opposite end-member models of evaporite basin development: a salt lagoon and a salina model. These models were briefly compared at the beginning of this paper. The sedimentological features of the Badenian gypsum deposits better fit salina depression than marine saline lagoon sedimentation (PERYT & *al.* 1995, PERYT 2001). Badenian gypsum deposits show primary depositional structures that are the same as those of modern salina deposits (KASPRZYK 1993a; BABEL 1999a, with references). The salina model is a subject of more detailed analysis in the companion paper (BABEL 2004), which introduces an integrated group of hydrological and sedimentological models for gypsum deposition in a salina basin environment.

FEATURES OF THE SALINA MODEL IN BADENIAN EVAPORITES

Drawdown and extent of Badenian evaporites

The areal extent of the evaporites in Upper Silesia (MICHAEL 1914), the Miechów Upland (KRACH 1947), the northernmost Carpathian Foredeep (KUBICA 1992), Bulgaria (TRASHLIEV & *al.* 1963, TRASHLIEV 1988), and Ukraine (ATLAS GEOLOGICZNY GALICYI 1885-1914, RYCHLICKI 1913, NOWAK 1938) is significantly smaller than that of the underlying marine Badenian deposits. In Ukraine such a limited area cannot be attributed exclusively to subsequent erosion because the gypsum deposits

apparently occur in topographically lower areas than the pre-evaporite sediments (TEISSEYRE 1900, RYCHLICKI 1913). The smaller areas of the Badenian evaporites are interpreted here as an effect of evaporite drawdown, although the areal extent of the evaporites in some parts of the basin is also modified by tectonics (mainly on the southern margin of the basin). In Ukraine, the marine pre-evaporite sediments, commonly represented by subaqueous sand bar facies, are found at topographically the same (or higher) levels as gypsum deposits in the immediate vicinity. This was observed by the author between Chortkiv and Mlynki, and by KUDRIN (1960, 1966) in many other areas. Such spatial relations reflect the palaeotopography of the regressive and drying out basin (SIDORENKO 1904, RYCHLICKI 1913, NOWAK 1938). According to FRIEDBERG (1912), the most significant palaeouplift was situated between Przemyślany and Czernelica. Gypsum was deposited mainly in deeper areas south of this uplift, but only locally on shoals to the north. The Podniestrze area, between Mykolaiv, Bibrka, Rohatyn, Buchach and Dnister, supposedly represents the original depression in which the gypsum deposits accumulated (TEISSEYRE 1896). Gypsum deposition in palaeodepressions is well known in the Miechów Upland and in some parts of the Nida area (KRACH 1947; RADWAŃSKI 1968, 1969), and locally in some areas containing native sulphur deposits (see ALEKSENKO 1967, GAŚIEWICZ 2000).

The view that Badenian evaporite sedimentation was directly related to sea-level fall and regression, and was followed by transgression during, or at the end of chemical deposition, is well established (FRIEDBERG 1912; MICHAÏLOV 1951; OLEKSYSHYN 1953; KRACH 1962; TRASHLIEV & *al.* 1963; PISHVANOVA 1963; KWIATKOWSKI 1972, with references; VENGLYNSKYI 1985; SENEŠ 1989; OSZCZYPKO 1998, 1999; DZIADZIO 2000; KASPRZYK 2003; PORĘBSKI & *al.* 2003). Many authors believed that all the marginal Badenian gypsum sediments were deposited during transgression (VENGLYNSKY & GORETSKYI 1966; DZIADZIO 2000; and reviews in: DROMASHKO 1955, KUDRIN 1955, KWIATKOWSKI 1972) on the eroded or washed out substrate (see DROMASHKO 1955, ALEKSENKO 1961, KWIATKOWSKI 1972). However, any reliable evidences of emersion and erosional unconformity below the gypsum strata are absent or scarce (SMIRNOW & *al.* 1994; PRYSJAZHNJUK 1998, p. 132). They are present only on original elevations like on the Miechów Upland (see BABEL 2004a in press) or the Rzeszów Island (DZIADZIO 2000), while in the deeper basinal areas they are not documented.

Supposed emersion surfaces below gypsum strata can be interpreted as hiatuses solely representing non-deposition. Direct sedimentation of gypsum on the Cretaceous substrate on the Miechów Upland and in the Nida area

was interpreted as deposition on subaqueous elevations lacking pre-evaporite deposits due to non-deposition related to action of wave currents (see RADWAŃSKI 1968, 1969; KWIATKOWSKI 1972, pp. 68-69, 92). However, thin clastic deposits underlying evaporites on the eastern peripheries of the Rzeszów Island, interpreted by DZIADZIO (2000, p. 1135) as a result of 'forced regression' and rapid fall of water level, can represent the drawdown phase of basin evolution. Hiatuses and/or erosional gaps were recognised both within and at the top of evaporite strata especially in Ukraine (e.g. PERYT & PERYT 1994b, PRYSIAZHNIUK 1998). The shallow-water, submerged, up to several metres deep, nature of the Badenian gypsum deposits is documented by many sedimentological studies (TRASHLIEV 1969b; KWIATKOWSKI 1970, 1972; KASPRZYK 1993a, b; PERYT 1996, 2001; BĄBEL 1999a). The evaporites are sandwiched within typical marine deposits that show features of relatively deeper environments (e.g. SMIRNOW & *al.* 1994, OSZCZYPKO 1999). Scarce marine fauna, mainly foraminifera, pteropods and nannoplankton, occur in clay intercalated within evaporites. These fossils are commonly interpreted as redeposited from older deposits (including pre-evaporite Badenian deposits) or as derived from seawater incursions into the evaporite basin (PISHVANOVA 1966, ALEKSENKO 1967, ODRZYWOLSKA-BIEŃKOWA 1975, VENGLYNSKYI & KOPYSTIANSKAYA 1979, VENGLYNSKYI 1985, GAŻDZICKA 1994, PERYT & *al.* 1997b, PRYSIAZHNIUK & *al.* 1997, PERYT 1999a, ANDREYEVA-GRIGOROVICH & *al.* 2003). Fish remains common in the Czech area and Upper Silesia could derive from brackish or fresh waters (KRACH 1939).

The evaporite drawdown predicted by the salina basin model fits well to all these data. Rapid drawdown and flat relief explain the lack of regressive sediments preceding gypsum deposition (but see DZIADZIO 2000). Emersions and shallow-water deposition are consistent with the model of a drying out salina depression. Scarce marine microfossils present in evaporites can have been washed out from pre-evaporite deposits emerged on the basin margins or may have been transported by marine incursions into a drawdown basin.

Range of drawdown

The range of evaporite drawdown can be roughly compared with the depth of the depression. It was suggested that the depth of ancient salina depressions at least equals the thickness of the evaporites (SONNENFELD 1984, p. 88). The thickness of shallowing-up gypsum units in salinas of southern Australia roughly coincides with the initial water depth and equals ca. 10 m (WARREN & KENDALL 1985).

On the northern margin of the Carpathian Foredeep, the onset of gypsum deposition could have more or less coincided with maximum drawdown when the brine depth was less than several m (BĄBEL 1999b). It took place in basins with flat bottoms (RADWAŃSKI 1969, OSZCZYPKO & TOMAŚ 1976). The maximum thickness of the Badenian marginal gypsum deposits is ca. 60 m but commonly is less due to subsequent erosion or non-deposition (Tab. 2). Taking into account the above-mentioned maximum

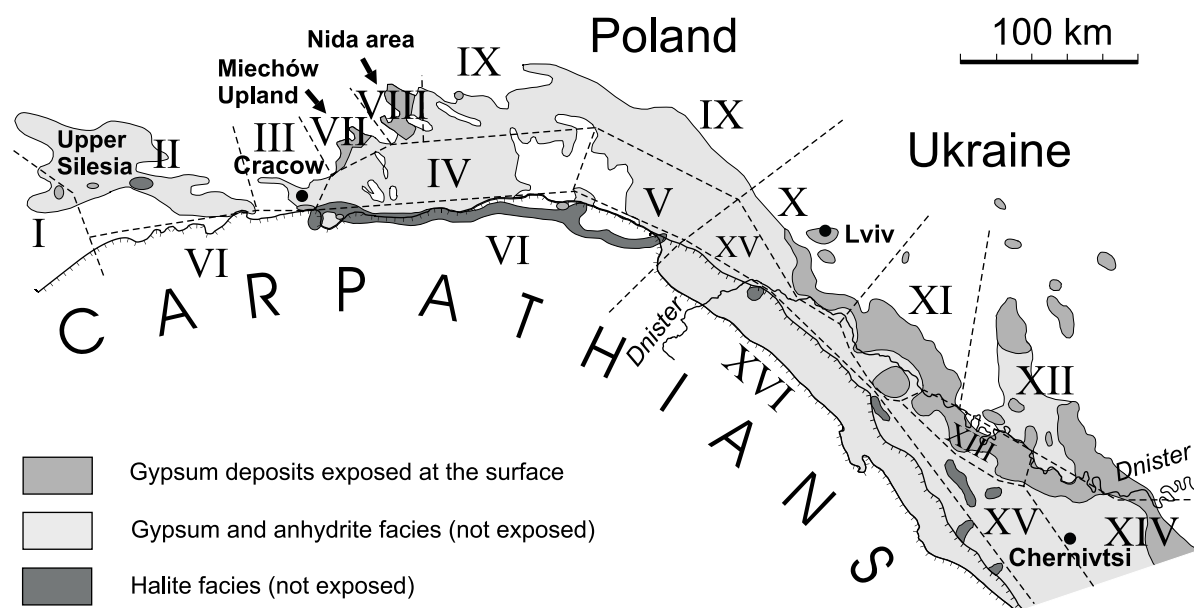


Fig. 4. Distribution of Badenian evaporites in northern Carpathian Foredeep (after various sources; references to areas I-XIV are given in Table 2)

Areas of evaporite basin	Dominated lithology and facies	Thickness of evaporites (some data from drill cores may be not corrected by dip)	Detailed references
I. Environs of Opava, Czech Republic	Gypsum (selenites), clay	30-40 m on average; maximum 61.1 (OV-4a Třebom borehole)	MÁTĽ 1973, 1980, 1981; ČIČHA & <i>al.</i> 1985; PERYT & <i>al.</i> 1997a, b
II. Upper Silesia	Gypsum (locally selenites), anhydrite, clay, locally halite, native sulphur deposits	20-30 m on average; Ca-sulphates with halite intercalation: maximum 112 m (Otto borehole, S of Palowice), 106 m (Pallowitz 10 borehole, SW of Palowice); Ca-sulphates: 57.3 m (Pilchowice 10 borehole), 54 m (Twardawa IG 1 borehole), ca. 60 m at Dzierżysław; thickest halite intercalation: 40 m (Pallowitz 7 borehole, SW of Palowice)	MICHAEL 1914; TREMBECKI 1952; KRACH 1956; RANIECKA-BOBROWSKA 1957; ALEXANDROWICZ 1961, 1963, 1964, 1965; KUBICA 1997; GONERA 2001; BICCHI & <i>al.</i> 2003
III. Environs of Cracow	Clay, gypsum	10 m on average; over 22 m at Wola Krzeszowicka borehole	BOGACZ 1965, GARLICKI 1968
IV. Central area between Cracow and Rzeszów	Ca-sulphates (gypsum-anhydrite), clay, halite	5-25 m on average; Ca-sulphates: 34 m (Kłaj 1 borehole)	OLTUSZYK 1967, OSZCZYPKO & TOMAŚ 1976, GARLICKI 1968, HOLLENDONNER 1995, POŁTOWICZ 1998, KARNKOWSKI & OZIMKOWSKI 2001, KRZYWIEC 2001, KASPRZYK 2003
V. Eastern part of the Polish Carpathian Foredeep	Anhydrite	5-10 m in average; 14.8 m (Ryszkowa Wola 7 borehole)	NEY 1963, PERYT & <i>al.</i> 1998b, DZIADZIO 2000, KRZYWIEC 2001
VI. Carpathian nappes and its substrate in Poland	Halite deposits with anhydrite and clay intercalations; gypsum deposits (locally with selenites)	ca. 70 m for restored, lower bedded part of halite deposits at Wieliczka and halite evaporites at Bochnia; over 70 m of gypsum in boreholes at environs of Broniszów	OLEWICZ 1968; NEY & <i>al.</i> 1974; GARLICKI 1979; URBANIAK 1985; NIEMCZYK 1986, 1995; POŁTOWICZ 1993, 1995, 1997; CZAPOWSKI 1994, with references; BUKOWSKI 1999, with references; NOWAK & POŁTOWICZ 2000; ANDREYEVA-GRIGOROVICH & <i>al.</i> 2003
VII. Miechów Upland	Gypsum (selenites), native sulphur deposits	10-15 m on average; maximum 27.8 m (Topola 5 borehole)	KRACH 1947; KWIATKOWSKI 1972, 1974; OSMÓLSKI 1972; KRYSIAK 1986; ROMAN 1998
VIII. Nida river valley area	Gypsum (selenites), native sulphur deposits	25-30 m on average; maximum ca. 52 m (Gartatowice-Stawiany, boreholes VI-4, V-5); over 37 m at Borków quarry	KWIATKOWSKI 1972, 1974 and BĄBEL 1999a, b; with references.
IX. Northern area of the basin between Chmielnik and Basznia	Gypsum (selenites), anhydrite, native sulphur deposits	20-40 m on average; maximum: 60.4 m at borehole NW of Baranów Sandomierski; 46.6 m at Cieszanów 1 borehole	KOWALEWSKI 1957, 1966; NEY 1969; KASPRZYK 1989a, 1993a, 1995; KUBICA 1992; ROSELL & <i>al.</i> 1998; GAŚIEWICZ 2000
X. Western Ukraine at environs of Lviv	Gypsum (selenites), anhydrite, native sulphur deposits	10-30 m on average; maximum 65 m at Bartatov; 45 m summary thickness at Pisky-Schytets'	ŁOMNICKI 1897, 1898; KUDRIN 1966; KITYK & <i>al.</i> 1979; BOGUCCI & WOŁOSZYN 1993
XI. Galicia at environs of Halich	Gypsum (selenites), anhydrite	30-40 m on average; maximum ca. 50 m at Hannusivka and Uzin	DUNIKOWSKI 1881, ŁOMNICKI 1881, BIENIASZ 1887, ŁOMNICKI 1905, MALICKI 1938, PAZDRO 1953, DROMASHKO 1955, KUDRIN 1955
XII. Podillia between Seret and Zbruch river valleys	Gypsum (selenites and gypsum microbialites)	10-15 m on average; 23 m at Zavallia	ALTH 1887, TEISSEYRE 1900, ŁOMNICKI 1901, OLEKSYSHYN 1953, DROMASHKO 1955, KUDRIN 1955, AGEANOVA & <i>al.</i> 1968, DUBLYANSKY & LOMAEV 1980, KLIMCHOUK & <i>al.</i> 1995
XIII. Podolian Pokutyá	Gypsum (selenites)	30-40 m on average; maximum: 55 m - summary thickness at Palahychi N3 and S1 (BABEL 2004a in press); ca. 50 m at Odaiv, Voroniv; over 40 m at borehole 6 km WNW of Verenchanka	ALTH 1887, MALICKI 1938, ZOLOTUCHYN 1954, DUBLYANSKY & SMOLNIKOV 1969, PRYSIAZHNIUK & <i>al.</i> 1997, PERYT 2001
XIV. Bukovyna and north Moldova	Gypsum (gypsum microbialites)	10-20 m on average; up to 35 m at Staľnivtsi; 29.5 m 3 km SE Hudești (borehole 81), Moldova	SIDORENKO 1904, KOLTUN & <i>al.</i> 1972, CEHLAROV & ŢIBULEAC 1996, PRYSIAZHNIUK 1998
XV. Area in front of Ukrainian Carpathians	Gypsum and anhydrite with clay intercalations, locally very thick and with halite intercalations, native sulphur deposits	Ca-sulphates: 5-10 m on average; 32.6 m at C-9 borehole near Drohobych; maximum thickness of NaCl-evaporites in tectonic grabens or paleovalleys: 360 m at Korshiv-Ispas-22 borehole; 347 m at Bratkovtsy 11 borehole	WYSZYŃSKI 1939, GOFŠTEIN & PISHVANOVA 1961, GURZHYI 1969, PETRICHENKO & <i>al.</i> 1974, KORENEVSKYI & <i>al.</i> 1977, KITYK & <i>al.</i> 1979, KROPACHEVA 1980, KHRUSHCHOV 1980, KASPRZYK 1995a, SMIRNOV & <i>al.</i> 1995, PANOV & PLOTNIKOV 1996, PANOV & PLOTNIKOV 1999
XVI. Carpathian folds in Ukraine and Romania	Ca-sulphates, halite, clay, supposed coeval K-salts, controversial stratigraphy	Controversial estimations of thickness	LADYZHENSII & ANTIPOV 1961, KORENEVSKYI & DONCHENKO 1963, STOICA & GHERASIE 1981, CEHLAROV 1999, HRYNIV & <i>al.</i> 1999, with references; WÓJTOWICZ 2001, with references; ANDREYEVA-GRIGOROVICH & <i>al.</i> 2003, WÓJTOWICZ & <i>al.</i> 2003

Tab. 2. Selected references to geology of Badenian Ca sulphate evaporite deposits in particular areas of the Carpathian Foredeep marked in Text-fig. 4; the extensive literature on halite and K-salt deposits is omitted

thickness, the range of evaporite drawdown in the northern margin of the Carpathian Foredeep can be estimated as less than 60 m. The thickest gypsum sections apparently occur in areas of increased subsidence or in a tectonic trough (KUBICA 1992, PERYT 2001). The upper allochthonous part of the gypsum deposits, in particular, seems to have been locally deposited in tectonic depressions (ROSELL & *al.* 1998, PERYT 2001, KASPRZYK 2003). Therefore, the range of evaporite drawdown is better comparable with the thickness of the underlying lower part, largely composed of selenite strata, which is ca. 25 m. Previously, OSZCZYPKO (1998, p. 424; 1999) assumed a 50-100 m water level drop during chemical sedimentation in the axial area of the Foredeep based on the palaeobathymetric significance of the foraminifera. TRASHLIEV (1984a) suggested a 50-60 m water level drop for the maximum 45 m thick Badenian gypsum deposits in Bulgaria (Text-fig. 3). Because of the relatively wet climate, the evaporite drawdown in the Badenian basin presumably was low.

The thickest evaporite sections occur in some halite subbasins that are known only from borehole cores (Tab. 2). In Ukraine, the thickest halite-bearing deposits appear to occur in pre-Badenian palaeovalleys (SHPAK & *al.* 1999, VYSHNIAKOV & *al.* 2000) or tectonic grabens (KUDRIN 1958, GOFSHTEIN 1962, TURCHINOV 1999). Similar structures occur in Poland and are commonly devoid of evaporites (see POŁTOWICZ 1998, KARNKOWSKI & OZIMKOWSKI 2001, KRZYWIEC 2001). An exception to this is provided by the halite deposits in Upper Silesia (ALEXANDROWICZ 1964). The geological history of these areas is controversial. One of the possibilities is that the supposed palaeovalleys or synsedimentary grabens experienced greater drawdown, which could have resulted in erosion on the slopes of these structures and sometimes in halite deposition in their deepest part. It is possible that the extent of drawdown was different in separate subbasins, which could reflect various water levels. This interpretation is in accordance with the salina basin model, but is not supported by any specific sedimentological studies.

Water level changes, accommodation and aggradation

Apparent fluctuations of water level, including both rapid emersions and floods, were recognised in gypsum sections in the northern Carpathian Foredeep (KASPRZYK 1991, 1993a; BABEL 1999b, 2004a in press) and, to some extent, in the southern halite subbasins (KASPRZYK 1994b, 1995b; GAŚIEWICZ & CZAPOWSKI 1995).

High accommodation is the reason for the presence and predominance of extremely shallow-water features in Badenian gypsum sections. The clearest evidence for high

accommodation, however, is provided by the 5-20 m thick, continuous sequences of gypsum microbialites recorded in Bulgaria (TRASHLIEV 1969b), Ukraine and Moldova (PERYT 1996, 2001; BABEL 2004a in press). Such extremely thick continuous microbialite sequences are unknown from high energy tidal flat environments controlled by sea level, which show low accommodation (see GERDES & KRUMBEIN 1994, p. 115). Such sequences are possible in some coastal salinas and salt lakes, similar to the Solar Lake and MacLeod salinas or lakes on Christmas Island (GERDES & KRUMBEIN 1994, TRICHET & *al.* 2001) but located in some depressed areas of the large salina basin. There the continuous accretion of gypsum microbialites is not limited by low accommodation (TRICHET & *al.* 2001).

During progressive filling in of the salina basin the accommodation decreases. This can be reflected by cyclic evaporite units thinning upward to the top of the evaporites (e.g. TUCKER 1999a). Such a feature is seen in the Badenian gypsum deposits in the highest selenite unit (unit F), which repeats features of the lower selenites, but is reduced in thickness (BABEL 2004a in press, 2004b in press).

The concept of layer-cake stratigraphy is well documented both in gypsum (e.g. WALA 1963, KUBICA 1992, PERYT & *al.* 1998a, BABEL 1999b) and halite strata (GARLICKI 1979, 1994) and strongly supports the aggradational mode of deposition. The layer-cake pattern is best recorded in selenite strata. Sequences of 5-30 cm thick-, grass-like selenite layers continue over distances of a few hundred km (BABEL & *al.* 1998, BABEL 2004a in press). The onlap of younger gypsum layers is recorded in Podillia and the Miechów Upland (PERYT 2001, BABEL 2004a in press, 2004b in press) and can be interpreted as autocyclic transgression (cf. SIDORENKO 1904, VENGLYNSKYI & KOPYSTIANSKAYA 1979, VENGLYNSKYI 1985).

Badenian deepening-upward sequences

Monotonous, up to ca. 20 m thick- Badenian sequences composed of gypsum microbialites are present near the Ukraine-Moldova boundary (PERYT 2001, BABEL 2004a in press). Such thick and continuous gypsum microbialite sequences are improbable in marine peritidal or lagoon environments, but fit to isolated salinas placed in depressions beyond direct sea-level control (GERDES & KRUMBEIN 1994, see also TRICHET & *al.* 2001). The sequences represent water-level rise keeping pace with gypsum microbialite deposition – a case which appear to be typical of a salina basin.

Both shallowing-upward and deepening-upward sequences were recognised in the Badenian gypsum deposits (KASPRZYK 1993a, BABEL 1999b), and occur-

rence of the latter type is the expected feature of the salina basin environment. Two prominent deepening-upward sequences enclose: an up to ca. 10 m-thick interval between layers c and h recorded throughout the NW margin of the basin (see relative water-level changes interpreted by KASPRZYK 1993a, 1994b; BABEL 1999b, 2004b in press), and a transition from 'basal' gypsum microbialites to selenite deposits below layer c in Ukraine (PERYT 1996, 2001; BABEL 2004a in press). The latter transition is strikingly similar to that described in Egypt (YOUSSEF 1988, AREF 2003). Both of these Badenian sequences can be also interpreted as brining-upward (see BABEL 1999b, 2004b in press), and this interpretation in the case of the c-h interval is supported by geochemical data, namely by the Sr content in the primary bottom-grown gypsum crystals (KASPRZYK 1994a, ROSELL & *al.* 1998). Additionally, the trend to falling $\delta^{34}\text{S}$ and $\delta^{18}\text{O}$ values (Text-fig. 5) recorded throughout the basin can be at least partly interpreted as the reservoir effect (see HAŁAS & *al.* 1996; PERYT & *al.* 1997a, 2002, with references) typical of more closed evaporite systems (compare ROUCHY & *al.* 1995), which is in accordance with the salina basin model.

These Badenian deepening-up and brining-up sequences seem not to be a result of increased subsidence, because they occur in the platform area far from the axis

of the foredeep. They are interpreted here as a simple effect of a higher rate of water-level rise than the rate of gypsum deposition (cf. BABEL 1999b). Because the rate of gypsum deposition is usually very high (up to a few cm a year), this rate of water-level rise was apparently much higher than any rate of sea-level rise.

Badenian brines

The marine origin of the Badenian evaporites has been accepted for a long time mainly because they are sandwiched by marine sediments and contain marine fauna. However widespread clay intercalations and common plant remains clearly indicate a significant input of continental water to the basin (e.g. KWIATKOWSKI 1972). Recent geochemical studies of Badenian evaporites led to some controversies concerning brine derivation.

Chemical analyses of fluid inclusions in sedimentary halite from the southern subbasins show that the halite crystallised from brine very similar to recent marine brine at the initial to middle stages of halite saturation – long before saturation with K-Mg salts (KHRUSHCHOV & PETRICHENKO 1979; GALAMAY 1997; GALAMAY & *al.* 1997, 2003; GALAMAY & KAROLI 1997; KOVALEVICH &

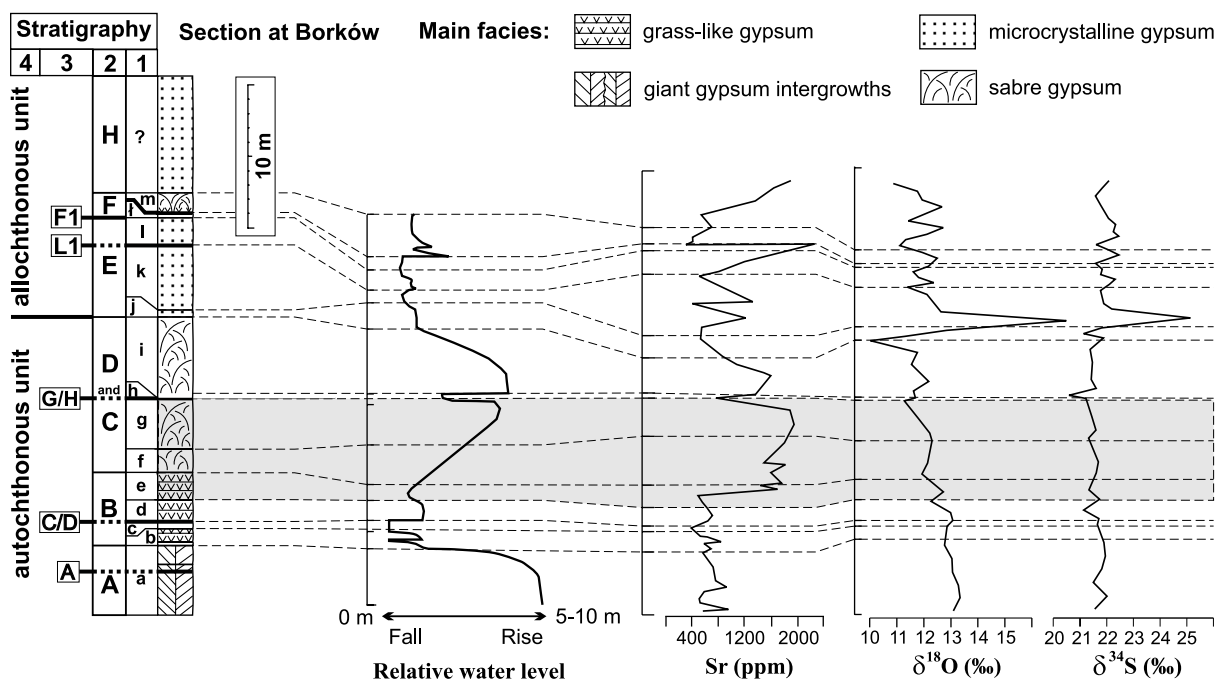


Fig. 5. Deepening-brining-upward sequence (shaded) in the Badenian gypsum; gypsum facies and section at Borków in the Nida area, Poland, after BABEL (1999b, with references). Stratigraphy: columns 1-2, 4 – lithostratigraphic units: 1 – layers lettered after WALA (1963), 2 – lithosomes lettered after KUBICA (1992), 4 – main lithostratigraphic units after ROSELL & *al.* (1998); column 3 – isochronous surfaces lettered after BABEL 1999b, 2004a in press. Water level curve after KASPRZYK 1993a, modified. Sr content in selenite crystals after ROSELL & *al.* (1998); S and O isotopic composition in sulphate ions in selenite crystals after HAŁAS & *al.* (1996); for more geochemical data from this interval of the section in other localities see KASPRZYK (1989b, 1994a, 1997) and PERYT & *al.* (1997a, 2002)

PETRICHENKO 1997; SHAIDETSKA 1997). Earlier results of trace element studies in halite rocks were interpreted in the same way, although the contents of Br were commonly lower than expected for recent marine brine saturated with halite (see GARLICKI & WIEWIÓRKA 1981, CZAPOWSKI 1994). This discrepancy was related to recycling of salt and brine dilution (e.g. BUKOWSKI 1997). There is a great consistency in the results of the above-mentioned fluid inclusion and Br analyses in the many separate halite subbasins. This supports the view that these subbasins were supplied with the same Badenian seawater (POBEREZHSKYI & KOVALEVYCH 2001). This interpretation is based on the widely accepted opinion that Cenozoic seawater did not significantly differ from recent seawater (see KOVALEVICH & PETRICHENKO 1997, ZIMMERMANN 2000). The halite subbasins were located very close to the source of seawaters south of the Carpathian orogen. Nevertheless GARCÍA VEIGAS & *al.* (1997, p. 185) commented: 'although the composition of the fluid inclusions studied is very similar to the composition of seawater at the beginning of halite precipitation, it could also be achieved by evaporation of a special type of continental water'. EASTOE & PERYT (1999) studied chloride isotopes in the same halite subbasins and recognised non-marine sources of Cl⁻ ions in many halite samples. GALAMAY & *al.* (1997), GARCÍA VEIGAS & *al.* (1997), CENDON & *al.* (1999) and BUKOWSKI & *al.* (2001) interpreted geochemical data from fluid inclusions in halite and found evidences of recycling and dissolution of earlier evaporites by continental water during halite deposition.

The S and O isotopic composition of Badenian sulphates (mostly in primary gypsum) coincides with the values expected for precipitation from Cenozoic seawater (TRASHLIEV & KANTOR 1978; PERYT & *al.* 2002, with references; KASPRZYK 2003). Deviations of up to $\pm 2\%$ were attributed to various processes typical of evaporite basin (sulphate reduction, reservoir effect, dissolution and reprecipitation). However, the S and O isotopes alone are not unequivocal indicators of marine brine. They are very insensitive to non-marine contributions because of the much higher SO₄²⁻ concentration in seawater brine than in most freshwaters (KIRKLAND & *al.* 1995, DENISON & *al.* 1998, LU & MEYERS 2003). The Badenian S and O isotopic compositions do not exclude the possibility of significant meteoric water influx to the basin.

Contrary to the O and S isotopes in the sulphates, the results of studies of fluid inclusions in primary gypsum crystals did not reflect a marine source of the brine (KULCHITSKAYA 1982). They showed that gypsum-precipitating waters in the northern landward subbasins were not related to sea brines but to low salinity waters very rich in Ca sulphate, similar to those of the modern Aral Sea. Both the salinity and the NaCl content were much lower than expected for marine brine at a stage of gypsum satu-

ration (KULCHITSKAYA 1982; PETRICHENKO & *al.* 1988, 1997; POBEREZHSKYI 1991, 2000; PETRYCZENKO & *al.* 1995; PERYT 2001).

Summarising, all the results seem to reflect both marine and mixed geochemically very complex brines for the Badenian evaporites. Non-marine water and a mixture of marine and non-marine water was important in some areas and times of evaporite deposition (GASIEWICZ 2000, PERYT 2001, PERYT & *al.* 2002). This is in accordance with the salina basin model.

The lack of K-Mg salts in the Badenian evaporites was explained either by a relatively low concentration of brine persisting at the beginning of the halite precipitation stage (e.g. GARLICKI & WIEWIÓRKA 1981; GALAMAY 1997, with references), caused by brine refreshment (GARLICKI 1968, GARLICKI & WIEWIÓRKA 1981), or by subsequent destruction of such salts (POŁTOWICZ 1993). One 0.3 m thick polyhalite layer was noted in halite deposits in the Sielec-Stupnica-Kotovane area, 15 km NW of Drohobych (PETRICHENKO & *al.* 1974). Many authors claimed that the strongly folded K-salts at Kalush in Ukraine are Badenian in age and are more or less coeval with the studied gypsum deposits (see DZHINORIDZE 1976; HRYNIV & *al.* 1999; PANOV & PLOTNIKOV 1999; WÓJTOWICZ 2001, with references). Recently this view was supported by biostratigraphic studies based on nannoplankton (ANDREYEVA-GRIGOROVICH & *al.* 2003), but questioned by radiometric data which suggested that the primary Kalush salts were much older (WÓJTOWICZ & *al.* 2003). These Kalush salts also appear to have been deposited in salt lakes or salina basins cut off from the sea (e.g. HRYNIV & PARAFINIUK 2001).

SUMMARY

The Badenian evaporite basin, and particularly the Badenian gypsum deposits in the northern Carpathian Foredeep, show features of a salina basin. Although the palaeogeography, stratigraphy, tectonic evolution and sedimentary history of the whole basin are not yet fully understood, many sedimentary features of the evaporites, appear to be clearly indicative of a salina basin environment.

Following the interpretation presented, it can be stated that during the Badenian salinity crisis the northern Carpathian Foredeep area passed through three main stages: a marine basin (represented by pre-evaporite deposits), passing to a salina basin (evaporite unit), and back to a marine basin (post-evaporite deposits). In terms of depositional sequences, the most important phases are: (i) deposition controlled by relative sea-level changes before the semi-isolation from Mediterranean

influx; (ii) a break in surface connection and rapid water-level fall during evaporite drawdown; (iii) aggradational shallow-water evaporite deposition controlled by water-level fluctuations within the basin, controlled mainly by regional climate, with common emersions and variable water levels between particular subbasins during low-stands; and (iv) marine flooding and rapid transition into marine deposition again controlled by relative sea-level changes. This scenario fits the geological data from the northern Carpathian Foredeep. The pre-evaporite sediments can be interpreted as deposited during rapid sea-level fall, and the deposition of the post-evaporite deposits as starting from rapid sea level rise; in both sequences additional fluctuations of sea level are recognisable (e.g. DZIADZIO 2000, PORĘBSKI & *al.* 2003). The evaporite, and particularly the gypsum deposits, apparently reflect 'internal' fluctuations of water level in the basin and aggradational deposition. The stratigraphy and facies development of the Badenian gypsum deposits are interpreted in the light of a salina basin model in separate papers (BĄBEL 2004a in press, 2004b in press).

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