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GENESIS AND EVOLUTION OF STRONTIUM DEPOSITS OF THE GRANADA BASIN (SOUTHEASTERN SPAIN): EVIDENCE OF DIAGENETIC REPLACEMENT OF A STROMATOLITE BELT

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ABSTRACT

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There are important strontium deposits in the Granada Basin, the most noteable of which (Montevives) has an annual production of approximately 50,000 tons of almost pure celestite. Two types of mineralization can be recognized: a primary variety consisting of stromatolitic carbonate that has been partially replaced by celestite, and a secondary variety consisting of celestite–pebble karst deposits. Both are included in an evaporitic Messinian succession. The primary variety of mineralization is within carbonates that interfinger with, and prograde across, gypsum deposits.

The development of these deposits can be interpreted in the context of the general evolution of the Granada Basin during Late Tortonian and Messinian times. Open marine conditions prevailed during the late Tortonian. During the transition from Tortonian to Messinian a restriction of the basin resulted in evaporite sedimentation, with stromatolites thriving at the basin's margin. The stromatolites were distributed along a coastal belt that was limited on the east by the tectonically active Sierra Nevada with its local alluvial fans. Runoff from the Sierra Nevada produced a freshwater lens and surface salinities that permitted the development of stromatolites, rather than the accumulation of gypsum.

The replacement of stromatolitic carbonate by celestite occurred within the mixing zone of the coastal aquifers during sedimentation and/or early diagenesis. An essentially marine origin is considered for the strontium. Supplementary influxes from continental weathering are also thought to have been produced.

Further restriction of the Granada Basin led to complete desiccation and the deposition of a 20 m thick halite layer. Later, gypsum deposits were exposed, and resulting cavities ("dolinas") were filled with celestite pebbles. The return of sediment accumulation within lakes buried and preserved these deposits.

INTRODUCTION

The Granada Depression is a postorogenic intramountainous basin situated at the boundary between the internal and external zones of the Betic Cordillera. It is partially filled with Neogene and Quaternary materials (González Donoso, 1970).

The oldest of the Tertiary deposits consists of Upper Serravalian (Middle Miocene)–Lower Tortonian (lowermost Upper Miocene) conglomerates, sandstones, claystones, limestones and, locally, evaporites of continental and/or lacustrine origin (Rodríguez Fernández, 1982). Unconformably overlying these materials are Tortonian coastal and shallow marine sandstones and conglomerates (Gallego Guarnido, 1978), algal limestones, coral reefs and marls with planktonic forams. This sequence is, in turn, unconformably overlain by Upper Tortonian marine conglomerates ("delta fans") that exhibit local patch reefs and which change laterally to turbidite sandstones and claystones with planktonic forams. The Miocene succession is capped by alluvial-fan conglomerates, on top of which fluvial sediments occur unconformably (Dabrio et al., 1978a, b). In the centre of the basin, overlying marine lutites, Messinian evaporites occur that are represented in two tectonosedi-

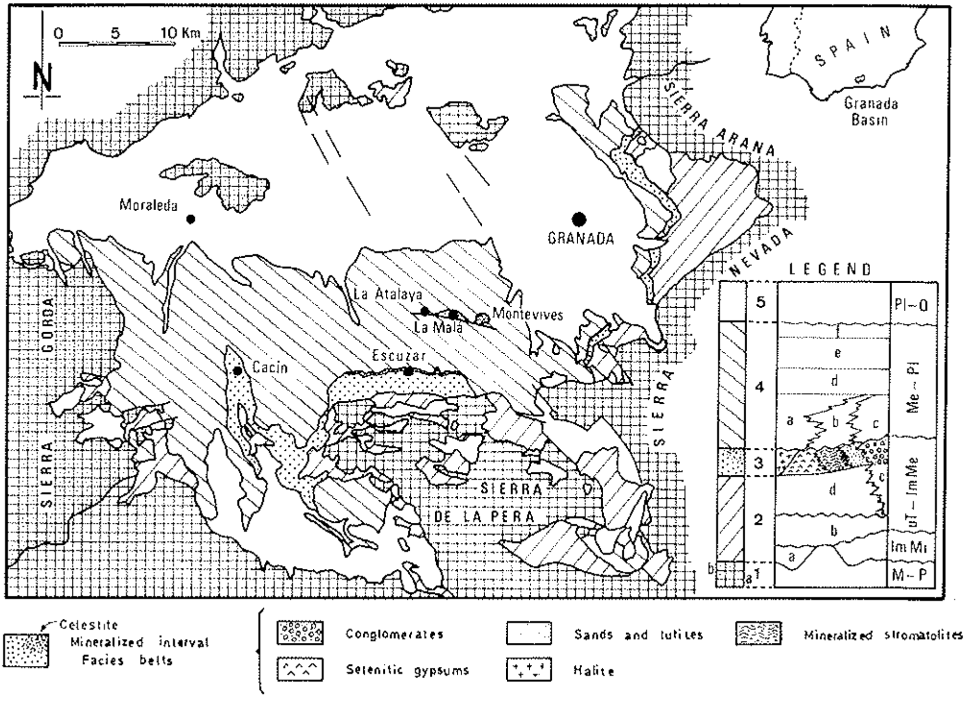

Fig. 1. Geological map and schematic stratigraphic section of the Granada basin, showing the distribution of the main celestite outcrops. Modified after Dabrio et al., 1982. *1a* = Nevado-Filábride and Alpujárride basements; *1b* = Subbetic basement; *2a* = conglomerates, sandstones, clays, limestones and evaporites; *2b* = conglomerates, sandstones and limestones; *2c* = delta fan conglomerates; *2d* = turbiditic sandstones and marine lutites with planktonic forams; *3* = mineralized interval; *4a* = lacustrine lutites of Cacin; *4b* = lacustrine turbidites of La Malá; *4c* = fluviatile sediments; *4d* = gypsiferous turbidites; *4e* = lutites, sandstones and lignites; *4f* = lacustrine limestones; *5* = recentmost conglomeratic formations.

M-P = Mesozoic–Paleozoic; *L-mMi* = Lower–Middle Miocene; *uT-Lm Me* = Upper Tortonian–lowermost Messinian; *Me-Pl* = Messinian–Pliocene; *Pl-Q* = Pliocene–Quaternary.

mentary units separated by an unconformity (Dabrio et al., 1982). The evaporites are, again, overlain by Messinian–Pliocene lutites, fine-grained sandstones, lignites and lacustrine limestones (Bone et al., 1978). The Upper Pliocene–Quaternary materials comprise several conglomeratic continental formations that are separated one from another by unconformities (Fig. 1).

The occurrence of celestite within Neogene deposits of the Granada Basin has been known for a long time. The older mines date back to the beginning of this century. Nevertheless, extensive mining at the most important of the deposits (Montevives) only began in the early fifties. The rate of extraction has increased notably in the last few years to the present level of 50,000 tons yr^{-1} with a celestite content higher than 95%.

There are works that make direct reference to these deposits. Some earlier papers deal mainly with the mineralogy (Arana, 1973; Ortega Huertas, 1973; Ortega Huertas et al., 1973, 1974). Other reports refer to particular mineralogical occurrences, such as wulfenite (Arana et al., 1978), and to late mobilization by thermal waters (Fernández Rubio et al., 1975). In still others a single stratigraphic position for the celestite has been noted (Sanz de Galdeano et al., 1976). As reported by Dabrio et al. (1982), there are two Messinian evaporite units present within the Granada Basin: the lower (marine?) division consists of selenite gypsum, halite, carbonates and celestite. The upper, lacustrine division, consists of detrital gypsum derived from the lower gypsum. A widespread detrital succession (turbidites of La Malá; Dabrio et al., 1972) is placed in between.

The present study shows for the first time that the celestite was formed by synsedimentary and/or early diagenetic replacement of freshwater stromatolitic carbonates. On this basis, and taking the current state of knowledge of the evolution of the Granada Basin into account, we suggest a genetic model for the formation of the celestite deposits.

DESCRIPTION OF CELESTITE OUTCROPS

Celestite outcrops of the Granada Basin occur in two parallel east–west belts approximately 5 km apart. The more northerly belt, which is relatively short, includes the Montevives and La Atalaya deposits. The Escúzar belt to the south is approximately 10 km long and appears to be prospective, especially near the village of Escúzar. The most significant features of each of these three areas are described separately in the following paragraphs.

Montevives

Montevives exposures occur on a small hill to the east of, and very close to, the village of La Malá (Fig. 2). Here there occurs the only deposit currently mined. The average concentration of celestite ranges from 70 to 80%. The mineralized mass, 2

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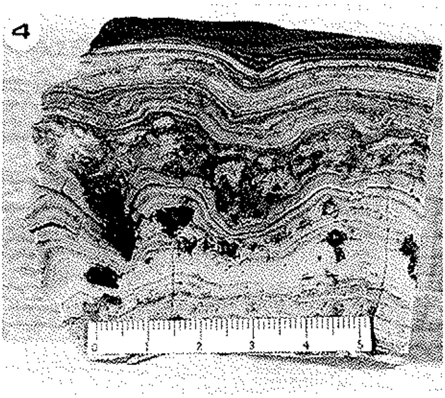

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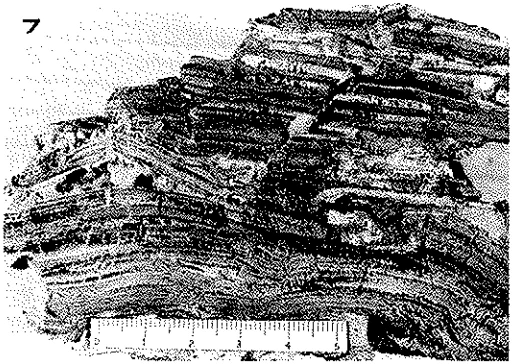

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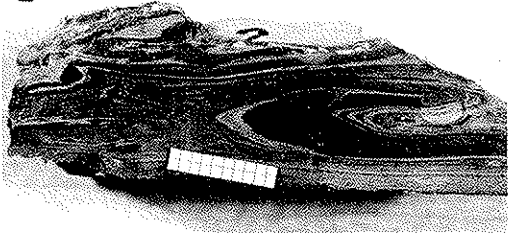


Figs. 2–8. *2*: General view of the Montevives deposit. *3*: Mineralized stromatolite. Close-up showing its internal fine lamination (bar equal to 2 cm). *4* and *5*: Mineralized stromatolites. The degree of replacement by celestite differs in the two samples. In 4 the grade of substitution is approximately 60%. Sample of 5 is completely replaced by celestite. Note that in the last one the internal lamination is partly obliterated. *6*: Mineralized stromatolite. Horizon with abundant evaporite molds (former gypsum crystals). Coin diameter = 2 cm. *7*: Mineralized stromatolite. Horizon with desiccation structures such as "tepees", "mud-cracks", and "flat-pebbles". *8*: Mineralized stromatolite with slump folds affecting its internal lamination (bar equal to 2 cm).

which is a carbonate body partially replaced by celestite, has a thickness of about 40 m and is visible for nearly 1 km. It dips 20°–50° toward the northwest. The southern limit of the outcrop is marked by an east–west fault that is down-to-the-south.

The mineralized carbonate body is well stratified in layers ranging from 20 to 50 cm in thickness. Seams of detrital clay are specially rich in montmorillonite–illite and also contain notable quantities of kaolinite, chlorite, and small amounts of paragonite (Ortega Huertas et al., 1973, 1974). It appears to be finely laminated (Fig. 3). These laminae tend to be obliterated where the replacement of carbonate by celestite is more thorough. It is marked by the existence of alternating clear and dark laminae a millimetre thick or less (Fig. 3). Although planar disposition appears to be the rule, there are undulating or dome-shaped forms (Figs. 4 and 5). Planar forms clearly correspond to algal mat structures ("planar stromatolites" or "cryptalgal laminites" according to the terminology of Hoffman, 1976, and Monty, 1976); undulating forms to stromatolites s.s. Microscopic study corroborates these observations. There are random centimetric bands of a spongy-pustular texture without any internal lamination, similar to that described by Buchbinder (1981) in some fossil stromatolitic structures from the Pleistocene of the Dead Sea. Hand specimens of these laminites and stromatolites exhibit moulds of lenticular gypsum crystals (Fig. 6) that disrupt the laminations, and desiccation structures such as "mud cracks", "flat-pebble breccias", "skrinkage pores", "tepees", etc. (Fig. 7). Small slump folds also occur, selectively concentrated at some levels (Fig. 8).

It is important to note the absence of primary gypsum in exposures at Montevives, which only appears to have been detected in drillholes (Ortega Huertas, 1973).

La Atalaya

Atalaya outcrops are situated at the base of a hill west of La Malá village beside the old "baths" where there are some abandoned galleries.

Mineralization, like that at Montevives, has affected laminated stromatolitic carbonate that has been irregularly replaced by celestite. The thin (10–30 cm) mineralized layers, appear to be intercalated in the upper part of the selenite gypsum unit, largely transformed into alabaster in this area (Dabrio and Martín, 1981). The deposit exhibits little lateral continuity and appears not to be prospective.

The Escúzar Belt

Escúzar outcrops extend east–west from the vicinity of "Cortijo de Santapudia" to "Cerro de Doña María", with a lateral continuity of 10 km. Escúzar village is centrally located within the belt and marks one the better exposures.

The mineralized carbonate, which is also stromatolitic, appears to overlie and prograde onto selenite gypsum of the lower evaporitic unit. The celestite seems to be in layers 20–50 cm thick dipping 20°–30° towards the north. The maximum thickness of the deposit is approximately 20 m. It thins laterally both east and west and is locally absent. The average celestite content is somewhat lower than that of Montevives (50–55%). Although the prevailing structure is laminated, a spongy-pustular texture does occur especially where the gypsum and mineralized zones interfinger.

One of the most significant features of the Escúzar Belt, especially in its eastern half, is the existence of a Miocene karst developed on top of the gypsum of the lower evaporite unit, whose cavities (circular "dolinas" of 1–5 m in diameter and up to 2–3 m deep), and depressed areas, appear to be filled with breccias containing pebbles of celestite and stromatolitic carbonate partly replaced by celestite (Fig. 9). This karst is the product of a period of exposure, during which most of the celestite outcrops were eroded. This erosion also explains the irregular thickness of the mineralized belt. The average celestite concentration within this breccia is low (about 15–20%). The existence of Miocene breccia with celestite pebbles argues in favour of an early origin for the primary mineralization, inasmuch as it is overlain by Upper Miocene lacustrine materials.

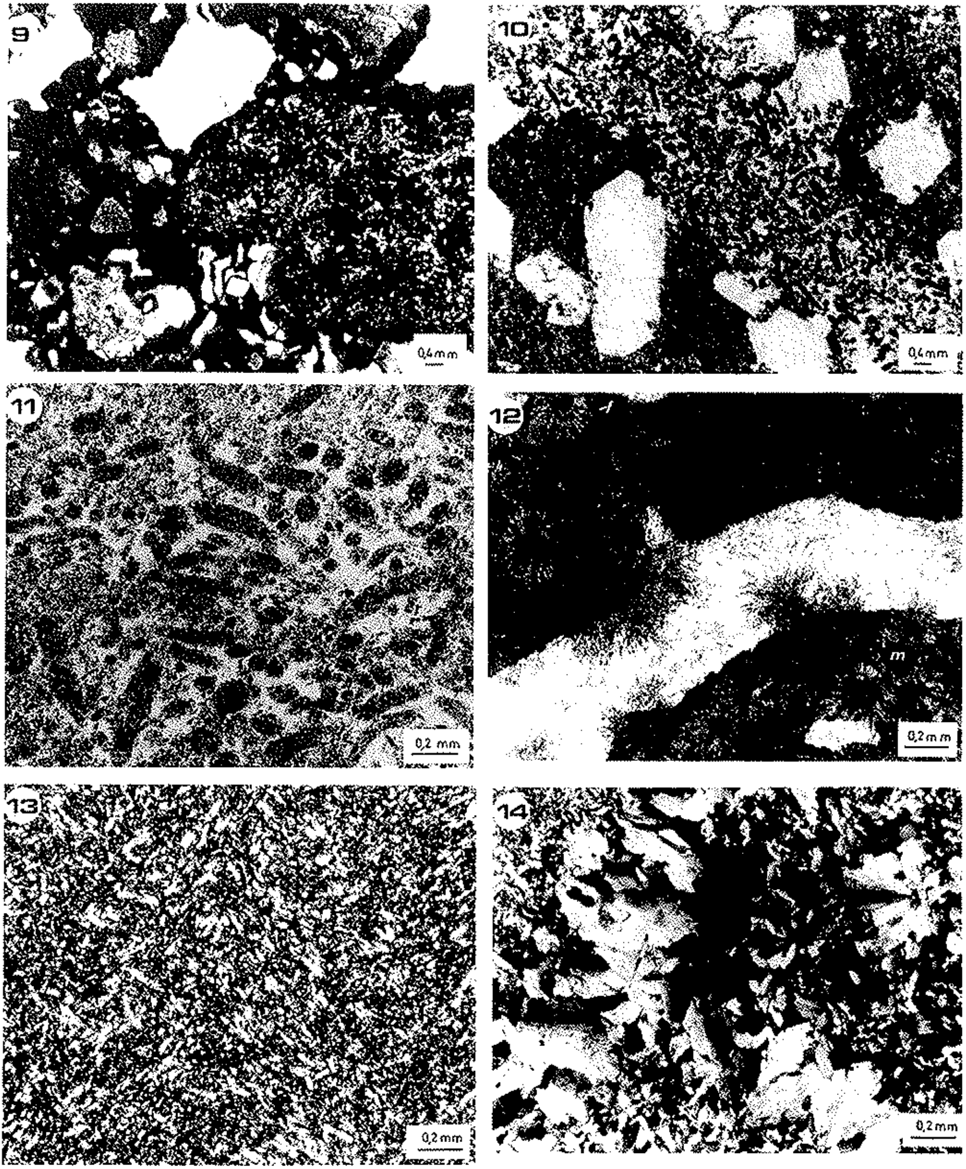
From the study of surface exposures we may infer two types of mineralization: a primary variety that corresponds to a stromatolitic carbonate partly replaced by celestite, and a secondary corresponding to a karst-filled Miocene breccia with celestite pebbles. The primary mineralization, which is quantitatively the more important, occurs marginal to selenite gypsum of the lower evaporite unit, interfingering laterally with it and prograding partly on it.

CHARACTERISTICS OF THE CELESTITE MINERALIZATION

Here we shall consider only particular aspects of the mineralization, most of them not described until now, which are important in the interpretation of the genesis and evolution of the celestite deposits. Complementary mineralogical data have been reported by earlier authors (Arana, 1973; Ortega Huertas, 1973; Ortega Huertas et al., 1973, 1974; Fernández Rubio et al., 1975; Sanz de Galdeano et al., 1976; Arana et al., 1978).

Microstructure of the mineralized stromatolites

Microscopic study corroborates the stromatolitic character of the mineralized carbonate. Laminar stromatolites appear to consist of an alternation of clear and dark laminae ranging in thickness from 0.05 to 1 mm. Both types of laminae are rich in calcified filaments of blue-green algae (Figs. 10 and 11). The thinner dark bands mark dense layers of filaments piled one of top of another. The thicker clear bands have fewer and more irregularly distributed filaments. They usually appear isolated and fragmented within microsparite carbonate matrix, and, for this reason, are more obvious than those within the dark bands. They consist of micrite impregnated by



Figs. 9–14. *9*: Karst-filled breccia with stromatolite pebbles partly replaced by celestite. Nx. *10*: Microesparite level partly replaced by celestite with abundant *Scytonema*-like "filaments" between micritic levels with evaporite molds. Plane polarized light. *11*: Partly leached *Scytonema*-like "filaments" included in a celestite matriz. Plane polarized light. *12*: Microscopic appearance of the pustular stromatolite structures. Note the existence of micrite nucleus (*m*) and fibrous-radiating growths (*f*). The latter also partly fill synsedimentary voids. Plane polarized light. *13*: Microcrystaline acicular celestite replacing stromatolitic carbonate. Nx. *14*: Fibrous-radiating celestite filling synsedimentary open spaces. Nx. (Nx = crossed polars.)

organic matter and are extraordinarily large (up to 0.1 mm in diameter and several millimetres in length).

These filaments are identical to those described by Rouchy and Monty (1981) from cryptalgal laminite levels in gypsum crystals of Messinian sequences in Cyprus. Those authors pointed out their remarkable resemblance to examples of *Scytonema*. This blue-green alga is abundantly represented in supratidal fresh and/or brackish-water ponds and marshes in the Bahamas (Monty and Hardie, 1976) and Florida, where it also builds stromatolitic structures. The "filaments" of this alga also cluster together in bundles, calcifying as if they were a single macrofilament (Monty, 1965), which explains their huge size compared with other blue-green alga filaments. In fact they only calcify in fresh water (Monty, 1967).

The internal laminate structure of these stromatolites tends to be obliterated in samples highly replaced by celestite. Nevertheless, more or less leached remains of filaments are recognizable in most of them (Fig. 11) and, in some cases, moulds of filaments with an internal filling of celestite are identifiable. In general, the microsparite of the clear bands tends to have been completely replaced, whereas the dark bands and filaments are less altered.

The pustular stromatolites exhibit an internal structure identical to that described by Buchbinder (1981) from similar stromatolite structures from the ancestral Dead Sea (Lake Lisan) (Fig. 12). This author interpreted such structures as the result of the calcification of mats of coccoid unicellular algae. The salinity of the environment at the time of the development of these structures is estimated to have been about 20–25%, i.e. that tolerated by present-day coccoid algal mats of Great Salt Lake, Utah (Halley, 1976).

Types of celestite

There are two types of celestite ascribed to primary mineralization: one variety consists of acicular crystals (0.15 mm long and 0.03 mm wide) that are generally restricted to the clear bands (Fig. 13). The other variety fills open spaces such as gypsum moulds and synsedimentary fractures (mud cracks and shrinkage pores). They occur as either tabular-prismatic crystals up to 0.6 mm long, or, less commonly, as fibrous-radiating crystals (Fig. 14). This void-filling celestite appears to be restricted to margins of earlier voids as a first-generation cement. The rest of the open space is filled with later sparry calcite crystals up to several centimetres in size.

Other syngenetic primary minerals

The lenticular gypsum crystals, whose moulds now appear to be partly filled by celestite and/or calcite, are interpreted as early (synsedimentary) diagenetic growths inside algal mats. Similar growths have been described from algal mats covering tidal flats in the Persian Gulf, where precipitation of gypsum occurs in the high intertidal zone from sulfate-enriched water within the more superficial layers of algal mats (Kinsman, 1966).

The presence of dolomite has been detected by X-ray analysis of samples from different outcrops (Ortega Huertas et al., 1974; Sanz de Galdeano et al., 1976). It appears, however, to occur in small quantities, usually less than 10%.

Late remobilization

As the result of the circulation of thermal waters through the deposits, some ions (mainly CO_3^{2-}, SO_4^{2-}, Ca^{2+} and Sr^{2+}) have been selectively dissolved and later precipitated as strontium and calcium minerals within cavities and cracks, the best known being that of "Las Fumarolas" (Fernández Rubio et al., 1975). The most abundant secondary mineral is strontianite, which is followed, in order of importance, by calcite and gypsum. Small quantities of barite, celestite, dolomite and quartz, as well as patinas of goethite, hematite and iron sulfates also occur (Fernández Rubio et al., 1975; Sanz de Galdeano et al., 1976).

ORIGIN AND EVOLUTION OF THE STRONTIUM DEPOSITS

The origin and evolution of these strontium deposits must be understood within the geodynamic context of the Granada Basin during the Late Miocene. The history of the basin during this time may be summarized in several stages (Fig. 15):

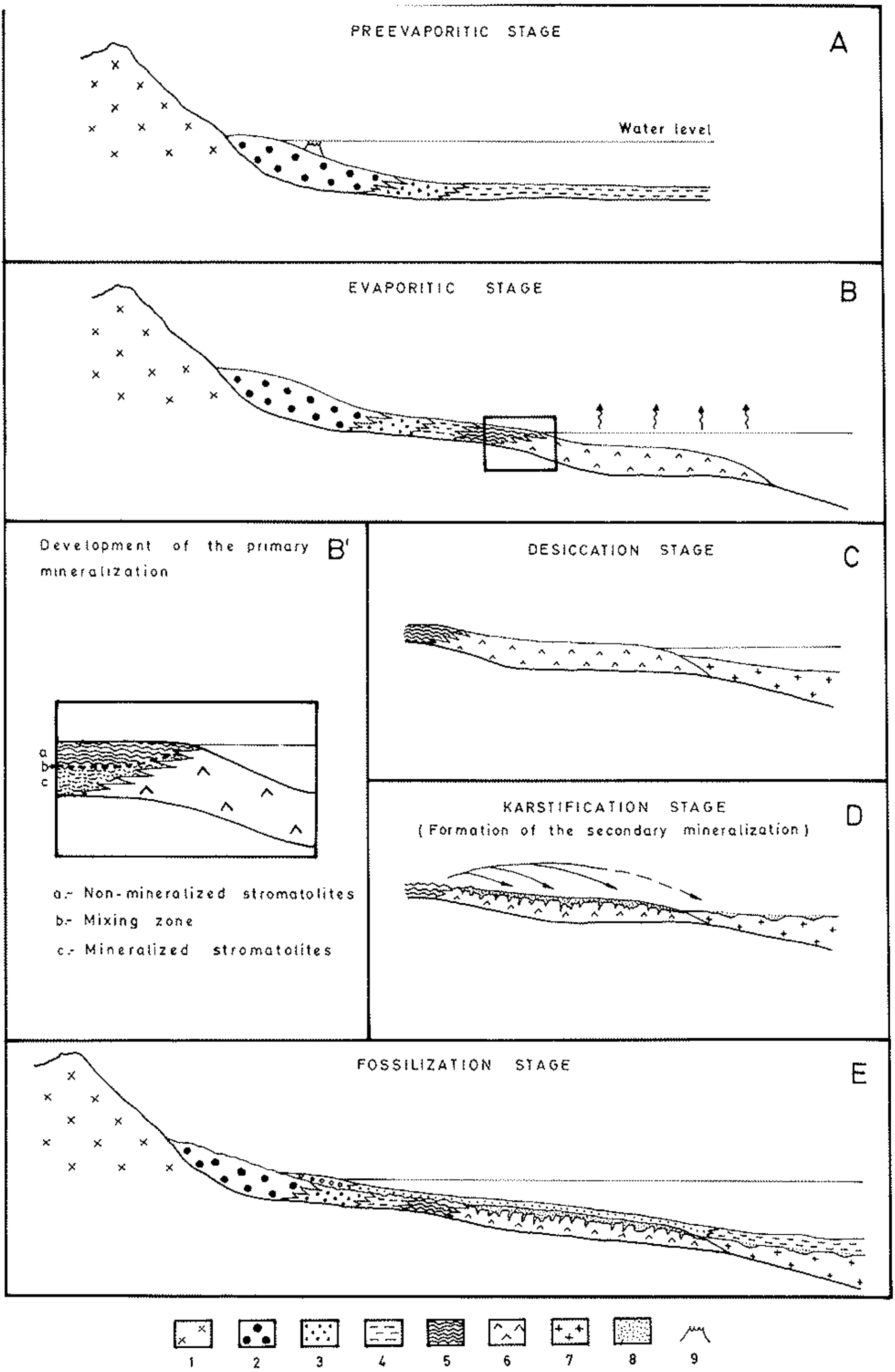
Pre-evaporitic stage

The Granada Basin already had its present size during the Late Tortonian (Rodríguez Fernández, 1982). There already existed tectonic reliefs at its margins, the most important of which were effected by ancestral Sierra Nevada and Sierra Arana, which maintained fans penetrating the sea. These delta fans were locally colonized by coral reefs in epochs of inactivity, which were destroyed later, as soon as the conglomeratic sedimentation in the area was renewed (Dabrio et al., 1978a,b). Sedimentation in the central, and usually deeper, parts of the basin resulted in marine lutites with abundant planktonic forams and turbidite sandstones.

Evaporite stage

Approximately at the time of the Tortonian–Messinian transition, uplifting of areas marginal to the Granada Basin produced the primitive Sierra Nevada and Sierra Arana. Alluvial fans derived from these highlands were permanently subaerially exposed (Dabrio et al., 1978a,b). The basin was thereby transformed into a depression where evaporitic sedimentation occurred (Dabrio et al., 1982).

The lower evaporitic sequences consist of "balatino" and selenite gypsum layers



PREEVAPORITIC STAGE
A
Water level
EVAPORITIC STAGE
B
Development of the primary mineralization
B'
a
b
c
a.- Non-mineralized stromatolites
b.- Mixing zone
c.- Mineralized stromatolites
DESICCATION STAGE
C
KARSTIFICATION STAGE
(Formation of the secondary mineralization)
D
FOSSILIZATION STAGE
E
1
2
3
4
5
6
7
8
9

alternating with lutites (Dabrio et al., 1982). Higher up, selenite gypsum dominates. This upper selenite gypsum changes laterally, at the basin margins, into mineralized stromatolitic carbonate.

The existence of Upper Miocene planktonic forams in the lutite intervals intercalated at the base of the gypsum indicates a marine origin. Sequences in which they appear intercalated with lutite-"balatino"-selenite are interpreted as the result of a progressive increase in the salinity of the environment (Rouchy, 1976; Dabrio et al., 1982). The predominance of selenite gypsum at the top of the unit indicates that the saturation conditions of the bottom became more permanent as time went on.

A shallow-water and shallow-basin evaporitic sedimentation model (Hardie and Eugster, 1971; Kendall, 1979) seems to be the most applicable. According to this model, the deposition of gypsum takes place mainly in the shallow marginal areas of the basin, which are also those which reach a greater concentration because of the considerable evaporation to which they are subjected. As we shall see, this also explains the absence of gypsum in the centre of the basin and the presence of halite higher within the section (Dabrio et al., 1982).

According to Lucia (1972), the precipitation of gypsum in marginal basins of this type requires the previous complete isolation of the water body. In such conditions, the influx of ions into the basin by runoff or groundwater may be very important (Kendall, 1979), which would explain Rouchy and Pierre's data (1979) on the isotopic composition of selenite gypsum of the Granada Basin. These authors found that the isotopic composition of this gypsum is closer to that of Keuper evaporites (outcropping in the northern part of the basin) than to Messinian evaporites of the centre of the Mediterranean Basin.

What is of interest here, however, is the fact that, by the time mineralization of stromatolitic carbonate occurred, the basin was occupied by a hypersaline water body, and that these carbonates accumulated in a peripheral position, marginal to the gypsum.

Fig. 15. Origin and evolution of the strontium deposits of the Granada basin. Stage *A*, preevaporitic, is characterized by the development of delta fans at the bottom of the paleoreliefs, on which locally small coral reefs appear. In the centre of the basin, turbiditic sandstones and lutites with planktonic forams were deposited. During stage *B* (evaporitic) the basin changed to a relict depression where evaporitic sedimentation occurred. Stromatolites developed at the margins. The outermost facies belts correspond to the sand- and mud-flats and to the fringe of alluvial fans at the foot of the paleorelief. As is shown in detail in B, the coastal aquifers were, at one time, responsible for the superficial development of the stromatolites, and their subsuperficial mineralization at the fresh and salt groundwater mixing zone (see text for details). Stage *C* corresponds to the moment when the basin was desiccated with the deposition of halite in its centre. In stage *D*, as a result of subaerial exposition, gypsum was karstified and the resulting cavities (dolinas, etc.) were partly filled by celestite. Stage *E* corresponds to the renewal of the sedimentation in lacustrine conditions and the fossilization of the primary (replacement type) and secondary (karstic type) celestite mineralization.

Legend: *1* = Basement rocks; *2* = conglomerates; *3* = sands; *4* = lutites; *5* = stromatolites; *6* = selenite gypsum; *7* = halite; *8* = karst-filled breccia; *9* = coral reefs.

As has been pointed out, the predominant laminated stromatolites consist of calcified "filaments" of blue-green algae very similar to *Scytonema*, which only calcify in freshwater. There is an apparent contradiction here, inasmuch as these stromatolites change laterally to selenite gypsum precipitated from hypersaline water. It should be remembered that there existed, at that time, at the eastern border of the basin, two tectonically important regions, the Sierra Nevada and the Sierra Arana, which were subject to intense erosion in semi-arid climatic conditions (Dabrio et al., 1978a,b; Ortega Huertas, 1978), at the base of which, conglomerates and sands were deposited in alluvial fans forming a fringe around them.

Infiltration (and/or runoff) waters coming from those reliefs accumulated in the coastal aquifers by slow lateral migration through the sand and mud flats facing the fans. Freshwater of the coastal lens produced a groundwater level that intersected the surface, so that fresh- (and/or brackish) water ponds and swamps, where *Scytonema* stromatolites developed, proliferated all along this coastal margin. Meanwhile selenite gypsum was precipitated in the immediately adjoining areas of the hypersaline basin.

Dry periods produced reductions in the groundwater level that resulted in the desiccation of superficial stromatolite structures and the appearance of "tepees" and mud cracks. At the same time, strong evaporation raised hypersaline waters by capilary action, from which small lenticular gypsum crystals in the top few centrimetres of the sediments precipitated, accounting for local dolomite crusts. Moreover, oscillations of the water level of the basin caused temporary flooding of the coastal border by hypersaline water, resulting in changes in the salinity of the pond water favourable to the development of pustular stromatolite structures. Inasmuch as *Scytonema* structures are clearly abundant, it is possible to infer that the normal situation was that of a stable freshwater coastal lens with a groundwater level cropping out at or near the surface. In these conditions, the subsuperficial evolution was of extreme importance in the genesis of the deposits, since a series of reactions would occur in the aquifers involving all the elements carried in solution or incorporated in the structure of previously formed carbonate minerals.

Genesis of the primary celestite

Examples of celestite in / or associated with evaporitic environments

It is well known that restricted basins in arid regions subject to intense evaporation favour the formation of celestite. In them, most of the celestite concentrations occur preferentially in marginal zones. A good number of such deposits are known, but relatively few, such as Hemmelle West Field and Zechstein in West Germany, those in Central Tunisia and Algeria, northeastern Mexico and Neuquen in Argentina, are of economic importance.

The Hemmelle West deposits are of upper Jurassic age. Müller (1962) considers them to be of synsedimentary syngenetic origin, formed by precipitation resulting from evaporation in saline basins. Puchett and Müller (1964) suggest a similar origin for the upper Permian Zechstein deposits. These authors, arguing from physicochemical considerations and the stratigraphic and palaeogeographic situation of the deposits, suggest that celestite began to be precipitated in the transitional region between the calcium carbonate and calcium sulphate deposits when the solution was reduced to between 1/3 and 1/5 of its original volume.

Brodtkorb et al. (1982) have suggested a similar origin for primary celestite in the penesaline supratidal Mesozoic facies of the Neuquen deposits.

On the other hand, Braitsch (1971) and Usdovski (1973), also basing their arguments on physicochemical considerations relating to the solubility of $SrSO_4$, place the beginning of celestite precipitation by evaporation of seawater at the start of the halite deposition phase. Similarly, Braitsch (1971) suggests the negative temperature coefficient of $SrSO_4$ solubility as a possible contributary cause of its preferential precipitation in shallow nearshore regions.

A different genesis has also been proposed for celestite deposits located in the transitional zones between freshwater coastal and evaporitic basinal facies. Chabou-Mostefai et al. (1978) suggest for the Eocene of Central Tunisia and Algeria a model, which, in its essentials has been applied to other celestite deposits (Samama, 1969; Fuchs, 1978, 1980, etc.). In this, the palaeogeographic position of celestite deposits is attributed to the creation of a geochemical barrier in the basin's marginal zone as the result of the mixture of strontium-bearing continental waters and sulfated connate waters of evaporitic facies. Celestite formation is considered to be of synsedimentary and/or early diagenetic origin.

Celestite deposits in similar contexts have been described by various authors: in the Upper Tortonian sediments of the Carpathian region: Srebrodol'ski, 1966; Lower Permian in the Volgograd Oblast area: Mokiyenko, 1966; Oslo area, Norway: Olaussen, 1981; Yate area near Britol, U.K.: Wood and Shaw, 1976; in the northeast of Mexico: Kesler and Jones, 1981; etc.

The existence of synsedimentary celestite of modern and/or recent formation has been reported in "sabkhas" of the Persian Gulf. Its origin seems to be closely related to early diagenetic dolomitization processes. The strontium, which was previously incorporated in aragonite structure (mollusc shells, etc.), is released during aragonite–dolomite transformation and combines with the sulfate ion of interstitial waters to form celestite (Kinsman, 1966). Small quantities of celestite have also been reported by Skinner (1963) and Wopfner and Twidale (1973) in recent lagoonal sediments in Southeastern Australia, and by Skocek and Saadallah (1972) and Khalaf et al. (1982) in recent alluvial fans and aeolian dunes in the deserts of Kuwait and Iraq.

Primary celestite formation in the Granada Basin

The main geological factors to be borne in mind in establishing a genetic model are:

(1) The deposits are palaeogeographically situated marginally to the basin and only in its south-southeast sector.

(2) Celestite has not been formed by simple precipitation as a result of seawater evaporation, but replaces earlier stromatolitic carbonates of largely freshwater origin.

Although the two bands of celestite outcrops are now separate, their characteristics suggest that they originally formed part of a single coastal fringe or strip. This, as the directions of progradation of the celestite indicate, traced an arc limited on the south by the Sierra de la Pera, which has been almost inactive since Late Tortonian (Gallego Guarnido, 1978) and to the east by the primitive Sierra Nevada. It is worth pointing out that an important part of the Sierra Nevada consists of Permo-Triassic materials, mainly carbonate rocks. A significant portion of such rocks accumulated in very restricted lagoonal and tidal-flats environments with preevaporitic characteristics (Martín and Torres Ruiz, 1982; etc.), and a generally high strontium content (Arana, 1973; Ortega Huertas et al., 1974; Gómez Pugnaire et al., 1981).

In order to explain the formation of celestite in the Granada Basin, we propose a genetic model based on a mechanism similar to that which gave rise to dolomite formation in transitional environments such as tidal flats, fringing reefs, etc., described by Hanshaw et al. (1971), Land (1972, 1973) and Badiozamani (1973).

Essentially we are dealing with the creation of a geochemical barrier to $SrSO_4$ in the coastal border zone caused by the mixing of subterranean freshwater with Sr-rich sulfated waters of evaporitic facies, which must have occurred at the beginning of and/or during precipitation of calcium sulphate, as we may deduce from the spatial relationship between these rocks and the celestite deposits and from the fact that the latter prograde on selenitic gypsums. In our model Sr derives principally, though not exclusively, from the seawater in the basin.

A detailed description of the physico-chemical conditions in which the creation of such a geochemical barrier took place is difficult to make, since data on the solubility of $SrSO_4$ are sometimes contradictory, especially concerning the influence of other ions and/or salts, above all of NaCl (Braitsch, 1971).

We see three determining factors:

—A relative dilution of the NaCl in connate waters from evaporitic brines resulting from their mixture with freshwater. According to various authors' data quoted by Braitsch (1971), dilution of concentrations of Na_2Cl_2 in sea water at the time of the precipitation of $CaSO_4$ (15 mol Na_2Cl_2/1000 mol H_2O) would reduce the solubility of $SrSO_4$.

—The rise in the temperature of the solutions in the coastal margin, since, according to Braitsch (1971) $SrSO_4$ shows a negative coefficient of solubility with respect to temperature, which he cites as a possible cause of the preferential concentration of celestite on the borders of evaporitic basins.

—Supplementary influxes of Sr in freshwater from the Sierra Nevada highlands, which would partially compensate the effect of dilution of Sr in evaporitic connate waters and explain why celestite deposits are restricted to the south-southeast of the coastal margin.

The combination of these factors would give rise to physico-chemical conditions ideal for the development of dissolution–precipitation reactions, which produce the replacement of calcium carbonate by celestite.

Later evolution

Desiccation of the basin. Formation of karst-filling celestite

The progressive restriction and isolation of the basin led finally to its complete desiccation and to the deposition, in its centre, of a salt layer 20 m thick (Dabrio et al., 1982), completing the final distribution of facies. These are arranged in a series of parallel bands in the following order, from the margin to the centre: an alluvial fan–conglomerate belt, a sand-flat sandy belt, a mud-flat clay belt, a coastal fringe mineralized stromatolitic carbonate (celestite) belt, a hypersaline platform selenite gypsum belt, and a central halite belt.

Following desiccation, selenite gypsum was subject to weathering, and the resulting cavities (dolinas, etc.) were filled with breccias derived from erosion of the surrounding highlands. These include abundant celestite pebbles, such as those of the Escúzar Band. This karst-filling celestite has already been described in detail.

The evolution of the Granada Basin during the Messinian is a local example of the Messinian evolution of the entire western Mediterranean, which ended with its complete desiccation and the deposition of up to 1000 m of halite in the centre of the abyssal Balearic plain (Hsü et al., 1977). Evaporitic sedimentation occurred earlier in marginal basins of the peripheral belt of the Mediterranean (e.g. the Granada Depression) than in those at its very border (Rouchy, 1981).

Deposition of the upper lacustrine unit. Preservation of the celestite deposits

Lacustrine materials of different lithologies were deposited on top of the erosion surface and the karst-filling breccia, thereby preserving the primary mineralization that had escaped erosion. The lowest deposits consist of terrigenous turbidites (la Malá turbidites, Dabrio et al., 1972) and fine terrigenous sediments (Cacín lutites, Dabrio et al., 1982) which pass laterally into fluviatile Messinian sediments (Dabrio and Ruiz Bustos, 1979).

The Sierra Nevada was the source of both the fluviatile materials and the turbidites. Turbidite gypsum of the upper evaporite unit, resulting from erosion and reworking of the lower gypsum (Dabrio et al., 1982), occurs on top of these deposits. The upper lacustrine sequence is capped by lutites, sands, lignites and limestones (Bone et al., 1978).

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