

# Chapter 6

## OTHER SEDIMENTARY ROCKS: CHERT AND EVAPORITES

### 1. CHERT

#### 1.1 Introduction

**1.1.1** *Chert* is the general term for *very fine-grained and nonporous sedimentary rocks that consist mostly or entirely of silica*, in the form of either amorphous silica or microcrystalline quartz presumably derived from recrystallization of amorphous silica. The non-scientific equivalent term is *flint*.

**1.1.2** Cherts are fairly easy to identify, even though you can't see the constituents with your unaided eye, because of the way they fracture and the way they look on fracture surfaces. Cherts are widespread (they are the last of the rocks we'll consider in this course that you are likely to see commonly), but their total contribution to the sedimentary record is probably not much more than a percent or two.

**1.1.3** The crystal size of quartz in recrystallized chert is usually in the range 5–20  $\mu\text{m}$ . Electron microscopy of fractured surfaces shows the quartz to be *polyhedral, equant to elongate, and closely fitted to surrounding grains*. Cryptocrystalline geometries in the transition from amorphous silica to recrystallized quartz are complex. Thin-section studies don't help much because the quartz is too fine.

**1.1.4** Chert comes in two distinct varieties, *nodular chert* and *bedded chert*, whose sedimentological occurrence is rather different. The problems with chert come in its origin; it's another one of those rocks whose origin we have a hard time observing directly. The relative importance of nodular chert and bedded chert has changed through geologic time: *bedded chert is much more common in the Precambrian, and nodular chert is more common in the Phanerozoic*.

#### 1.2 Terminology

**1.2.1** Here are a few terms for kinds of amorphous silica and chert:

**amorphous silica:** material composed of relatively pure  $\text{SiO}_2$  but with only very local crystallographic order. Amorphous silica includes various kinds of *hydrated and dehydrated silica gels, silica glass, siliceous sinter formed in hot springs*, and (certainly of greatest geological importance) *the skeletal materials of silica-secreting organisms* (see below, under *opal*).

**opal** (or **opaline silica**) is a solid form of amorphous silica with some included water. It's abundant in young cherts, back into the Mesozoic. Its geological occurrence is by alteration of volcanic ash, precipitation from hot springs, and, by far most importantly, *precipitation as skeletal material by certain silica-secreting organisms* (see a later section). Opal starts out as what is called **opal-A**, which shows only a very weak x-ray diffraction pattern, indicating that any crystallographic order is very local. With burial, during the initial stage of diagenesis, opal-A is transformed into **opal-CT**, which shows a weak x-ray diffraction pattern characteristic of cristobalite another silica mineral; see below). Upon further diagenesis, opal-CT is transformed into *crystalline quartz*, resulting in chert that consists of an equant mosaic of microquartz crystals. By that stage, most or all of the fossil evidence of origin is obliterated.

**chalcedony** is a very finely crystalline form of silica consisting of radiating needles or fibers, often spherulitic, of quartz. There's probably amorphous silica in among the needles, and a variable water content. This stuff is metastable with respect to ordinarily crystalline quartz, but it hangs around a long time; it's found even in Paleozoic cherts.

**flint** is the general-language equivalent of *chert*, usually applied to dark gray chert in nodules or as beds.

**jasper**: *chert that's red because it contains hematite* (often more than a few percent).

**porcelanite** (also spelled porcellanite): *a minutely porous form of chert* with a dull appearance on the fresh surface.

### 1.3 Silica Geochemistry

**1.3.1** *The solubility of quartz in pure water is very small*, several parts per million. Figure 6-1 shows a graph of quartz solubility as a function of pH. You can see that the solubility of quartz is very low for pH values up to about 8 (slightly alkaline) but then rises sharply with increasing pH. It's hard to measure, but it can be calculated. Moreover, attainment of equilibrium is very slow. The solubility of amorphous silica is an order of magnitude higher, and attainment of equilibrium is also slow, but much faster than for quartz; you can do it in the laboratory. The dominant species of silica in solution in natural waters in the usual range of pH is silicic acid,  $\text{H}_4\text{SiO}_4$ , a weak acid.

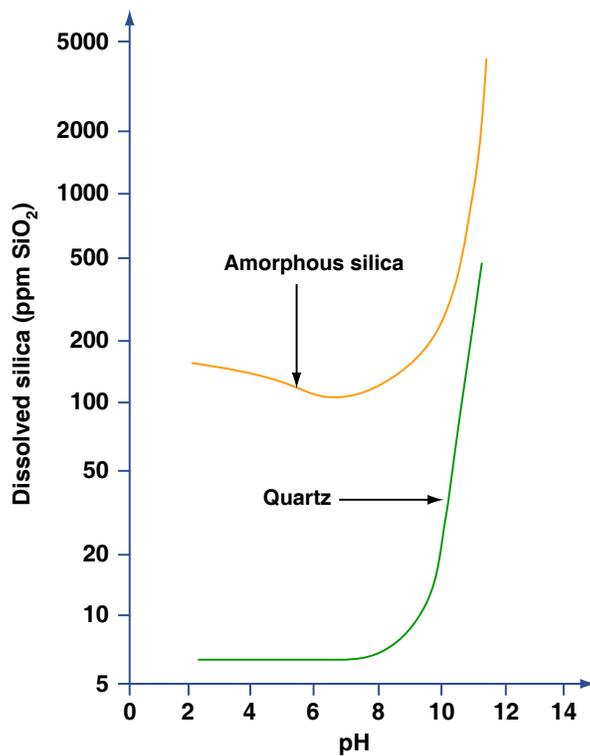


Figure by MIT OCW.

Figure 6-1: The solubility of quartz and amorphous silica at 25°C

**1.3.2** At room temperature and pressure, *amorphous silica is metastable with respect to quartz, but precipitates formed from supersaturated solutions are always amorphous silica.* This is because the building of the quartz crystal structure at low temperatures takes a very long time. Solutions unsaturated with respect to amorphous silica but supersaturated with respect to quartz remain stable for many years if not longer. Clear crystalline quartz can't be precipitated in the laboratory at the low temperatures and pressure of sedimentary environments. Presumably nature can do it because of the longer times involved.

**1.3.3** *The concentrations of dissolved silica in rivers, streams, and lakes are a few tens of parts per million.* Concentrations are also in this range in groundwater; the deeper the groundwater, the higher the silica concentration.

**1.3.4** *What are the sources of silica in solution?*

**weathering of silicate minerals:** this is *the ultimate source* of most of the silica in solution in the Earth's surface waters.

**thermal springs:** the concentration of dissolved silica concentration is very high but the absolute volume is small.

**dissolution of amorphous silica:** this must be important in areas underlain by silica-containing sedimentary rocks, but it is certainly not as important overall as weathering of silicates

**dissolution of quartz:** most natural fresh waters are in the pH range for which the solubility of quartz is very low, so dissolution of quartz is not an important source of silica in solution.

Here's the bottom line: it's generally agreed that *by far the greater part of dissolved silica comes from weathering of silicate minerals in source rocks.*

**1.3.6** *The concentration of dissolved silica in the oceans is surprisingly and extremely small:* only a few tenths of a part per million. In the present oceans, there's no possibility of precipitating silica inorganically. The reason for this very low silica concentration in the oceans is that *several kinds of organisms are very effective in extracting silica from sea water and fixing it in the form of opaline silica in their skeletons.* They do this out of equilibrium, by metabolic concentration processes.

## **1.4 Silica-Secreting Organisms**

### **1.4.1 Introduction**

**1.4.1.1** Several kinds of important marine organisms secrete amorphous silica to form part or all of their skeletons: *diatoms*, *radiolarians*, and *sponges*. These organisms play an important part, directly or indirectly, in the deposition of chert.

### **1.4.2 Diatoms**

**1.4.2.1** Diatoms are unicellular photosynthesizing organisms, a kind of algae, mostly from 20 to 200  $\mu\text{m}$  in size. The cell wall of the organism is silicified with opaline silica to form what's called a *frustule*, consisting of two valves that overlap one another like the lid of a shallow dish.

**1.4.2.2** Some diatoms are benthic (living on the sea floor), others (by far the most important sedimentologically) are *planktonic*, living in the photic zone of the oceans (the uppermost layer in the oceans, in which there is abundant sunlight), in less than 200 m water depth. They solve the problem of buoyancy by being fitted with tiny fat droplets. They like cool water best, apparently because of greater nutrient supply, higher water density, and upwelling. At certain times of the year diatom blooms produce concentrations of up to a billion organisms per cubic meter.

**1.4.2.3** Diatoms first evolved in the Cretaceous and have been expanding in importance since. A fairly high percentage of the total number of known diatom species are living today.

**1.4.2.4** Like foraminifera, *diatoms rain onto the floor of the deep ocean over large areas.* The resulting sediment is called **diatom ooze**. Fossil diatomaceous deposits are called **diatomite**, or **diatomaceous earth** if friable. It's mined as a filtration medium for water filters. Diagenesis of diatomite leads to **diatomaceous chert**, and if diagenesis is sufficiently intense, most or even all the evidence of the originally diatomaceous nature of the rock is obliterated.

**1.4.2.5** Figure 6-2: Brasier, M.D., 1980, *Microfossils*: George Allen & Unwin, 193 p. (Figure 7.3, p. 42) shows what planktonic diatoms look like.

### **1.4.3 Radiolarians**

**1.4.3.1** Radiolarians are planktonic protozoans with approximately spherical skeletons generally from 100 to 2000  $\mu\text{m}$  in diameter. The protoplasm of the cell is divided into endoplasm and ectoplasm, separated by what's called a central capsule. The ectoplasm has frothy gelatinous material in which are embedded swarms of symbiotic algae called zooxanthellae, in the shallow-water species. Radiating outward from the central capsule through the ectoplasm are threadlike pseudopods. The skeleton, situated within the cell, consists of radial and tangential elements of highly varied geometry: spicules, spines, bars, spheres, spindles, and cones. The composition of the skeletal material is either celestite,  $\text{SrSO}_4$  (a strontium sulfate mineral, which is unimportant geologically because the celestite dissolves too fast), or opaline silica. Radiolarians have been around since at least the Cambrian.

**1.4.3.2** Radiolarian species live in a wide range of water depths. They like warmer water. They trap and paralyze passing small plankton with their sticky pseudopods. They maintain buoyancy by fat globules or gas-filled vacuoles.

**1.4.3.3** Like diatoms, *radiolarians sink to the bottom and can accumulate to form radiolarian ooze if they are not dissolved before burial*. **Radiolarian ooze** tends to be found at greater depths than calcareous ooze, and at low latitudes. Fossil deposits of radiolarian ooze are called **radiolarite**, and chert containing radiolarians is called **radiolarian chert**. The effects of diagenesis are similar to those on diatomaceous cherts.

**1.4.3.4** Figure 6-3: Brasier, M.D., 1980, *Microfossils*: George Allen & Unwin, 193 p. (Figure 12.1, p. 81, p. 48) shows what radiolarians look like.

### **1.4.4 Sponges**

**1.4.4.1** Sponges are filter-feeding organisms that live attached to the bottom in shallow and deep marine environments. They are multicellular but are not considered to be true metazoans because they are not organized into tissues and they have no nervous systems; in degree of organization they fall between protozoans and metazoans.

**1.4.4.2** Here's how sponges work: they have an upright body shaped like a bag or vase with a hole at the top. The outer surface of the sponge has tiny holes leading inward to small chambers in the wall of the sponge that are lined with cells with flagella that whirl continually to create a current flowing inward into the central cavity. Plankton sticks to the surface of these cells, which then ingest the material. Sponges are thus low-pressure water pumps, which can pump a volume equal to their body volume in a minute.

**1.4.4.3** Sponges have skeletons composed of spongin (an insoluble and chemically inert protein, what you buy as a “natural sponge”), calcareous spicules, or siliceous spicules. It’s the siliceous spicules that we’re interested in here. They can be straight, radiating, or irregular in shape, and they range in size from microscopic to quite large. Silica-secreting sponges have been around since the Cambrian.

**1.4.4.4** Figure 6-4: (upper left) Clarkson, E.N.K., 1979, *Invertebrate Palaeontology and Evolution*: George Allen & Unwin, 323 p. (Figure 4.1, p. 49). (lower left) Clarkson, E.N.K., 1979, *Invertebrate Palaeontology and Evolution*: George Allen & Unwin, 323 p. (Figure 4.4, p. 51) (right) Shrock, R.R., and Twenhofel, W.H., 1953, *Principles of Invertebrate Paleontology*: McGraw-Hill, 816 p. (Figure 3-2, p. 76) shows what some sponge spicules look like.

## ***1.5 Nodular Chert***

**1.5.1** Chert is widespread as *nodules in limestone*. Chert nodules are varied in shape, from more or less regular discoidal or egg-shaped bodies (that’s the common shape for relatively small chert nodules) to highly irregular knobby and warty bodies (the common shape of relatively large chert nodules). Their size ranges from a few centimeters to a few tens of centimeters. They tend to be concentrated along certain bedding planes. Where abundant, they often form a two-dimensionally or three-dimensionally interconnected network.

**1.5.2** Chert nodules are usually structureless, but some show faint traces of stratification coincident with that in the enclosing limestone. If the limestone is fossiliferous, the fossils in the chert can be either calcareous or silicified. And the fossils in the rest of the limestone may or may not be silicified themselves. Sometimes the boundary between the nodule and the limestone passes right through fossils—suggesting that the nodules are a diagenetic feature, not a primary feature.

**1.5.3** Although in the past some geologists believed that nodules form by direct precipitation of silica gel on the ocean bottom, today the evidence for a *replacement origin* is considered to be overwhelming:

- irregular shape and interconnectedness of many nodules;
- presence of irregular patches of limestone in nodules;
- association of chert and silicified fossils in many limestones;
- presence of replaced fossils in some nodules;
- traces of bedding passing through nodules;
- contacts of nodules passing through fossils.

**1.5.4** But the source of the silica is a problem. Did it come from within the bed or from somewhere else? The general feeling seems to be that, in most cases at least, the nodules can be explained by *the presence of abundant biogenic amorphous silica in the original sediment and then diagenetic reorganization*. By diagenetic reorganization is meant the process by which the disseminated bodies of opaline silica (sponge spicules, diatoms, radiolarians) are dissolved, whereupon the silica in solution migrates to certain places in the sediment where it is reprecipitated in the form of opal-CT to form the nodules. This happens because the pore fluids are undersaturated with respect to the original biogenic silica, which consists of opal-A, but are supersaturated with respect to opal-CT, which has lower solubility than opal-A. Frustratingly (for me, at least), the literature on chert nodules never seems to provide the classic “straight answer” for what the controls are on the origin of the nodules: Why do they form where they do? Why are the size and spacing the way they commonly are, rather than much smaller or much larger?

**1.5.5** Where the chert nodules form only a small part of the bulk volume of the rock, a good case can be made that the silica that forms the nodules was present in the sediment from the time of deposition. But how about when the chert forms the greater part of the rock? Then a stronger case could be made for *introduction of silica in solution after deposition*, by circulating pore solutions.

**1.5.6** The chert in shallow marine limestones probably comes mostly from sponge spicules; that in deep marine cherts probably comes mostly from diatoms and radiolarians. In either case, the limestones that contain chert nodules are commonly fine-grained wackestones and mudstones, because the remains of the organisms responsible for the silica were deposited in quiet-water carbonate environments conducive to the various silica-secreting organisms.

## ***1.6 Bedded Chert***

**1.6.1** Chert is also found as continuous beds, from centimeters up to as much as a few meters thick, often, but not always, interbedded with shale. Bedded cherts are also often interbedded with turbidite sandstones and submarine volcanics. Most such bedded cherts show abundant evidence of having been deposited in the deep ocean.

**1.6.2** Two scenarios seem most attractive for explaining these cherts:

- Open-ocean siliceous ooze is conveyor-belted to subduction margins and incorporated into a subduction mélange;
- Siliceous sediment is deposited near a subduction margin, interbedded with subduction-zone volcanics and turbidites shed from the island arc, and incorporated into a subduction mélange.

**1.6.3** Here we need to distinguish between Phanerozoic bedded cherts and Precambrian bedded cherts.

- **Phanerozoic bedded cherts** usually contain radiolarians, and they can be explained by lithification and diagenesis of radiolarian-rich bottom sediments, although some may have been inorganically precipitated with radiolarians as a nonessential constituent.

- **Precambrian bedded cherts**, on the other hand, show no convincing evidence of having started out as biogenic sediment, inasmuch as no silica-secreting marine organisms are known from the Precambrian. (Some of the oldest and most exotic unicellular organisms are found preserved in Precambrian cherts, going way back into the Archean—but that does not mean that the cherts were formed biogenically from silica-secreting organisms). Most Precambrian cherts seem to have been inorganically precipitated, although the processes involved are not entirely clear. The concentrations of dissolved silica in the Precambrian oceans, before the appearance of the silica-secreting organisms characteristic of the Phanerozoic, were probably in the range of tens of parts per million, almost an order of magnitude greater than today. It's common for such cherts to be interbedded with chemically precipitated ironstones. Some of the very oldest sedimentary rocks, in greenstone terranes, are bedded cherts. Chert is characteristically interbedded, down to centimeter scale, with Precambrian iron formations.

## 2. EVAPORITES

### 2.1 Introduction

**2.1.1** Enormous volumes of sediment in the ancient sedimentary record are composed mainly or entirely of what are called *evaporite minerals*: *minerals precipitated from a saline solution that has been concentrated by evaporation*. Most evaporites in the sedimentary record are marine evaporites (solutions derived from normal sea water by evaporation are said to be *hypersaline*), but there are also nonmarine evaporites. There are a great many evaporite minerals, but only a few are common. It's been estimated that something like one percent of Phanerozoic rocks are evaporites, and that a quarter to a third of the area of the continents is underlain by evaporite deposits in the subsurface, with thicknesses ranging from a few meters to hundreds of meters. (Think "salt mines": there really are such things.) Figure 6-5 shows areas of the conterminous United States that are underlain by evaporites, and Figure 6-6 shows the location of the major evaporite deposits of the world.



Figure by MIT OCW.

Figure 6-5: Areas of the conterminous US that are underlain by evaporite deposits

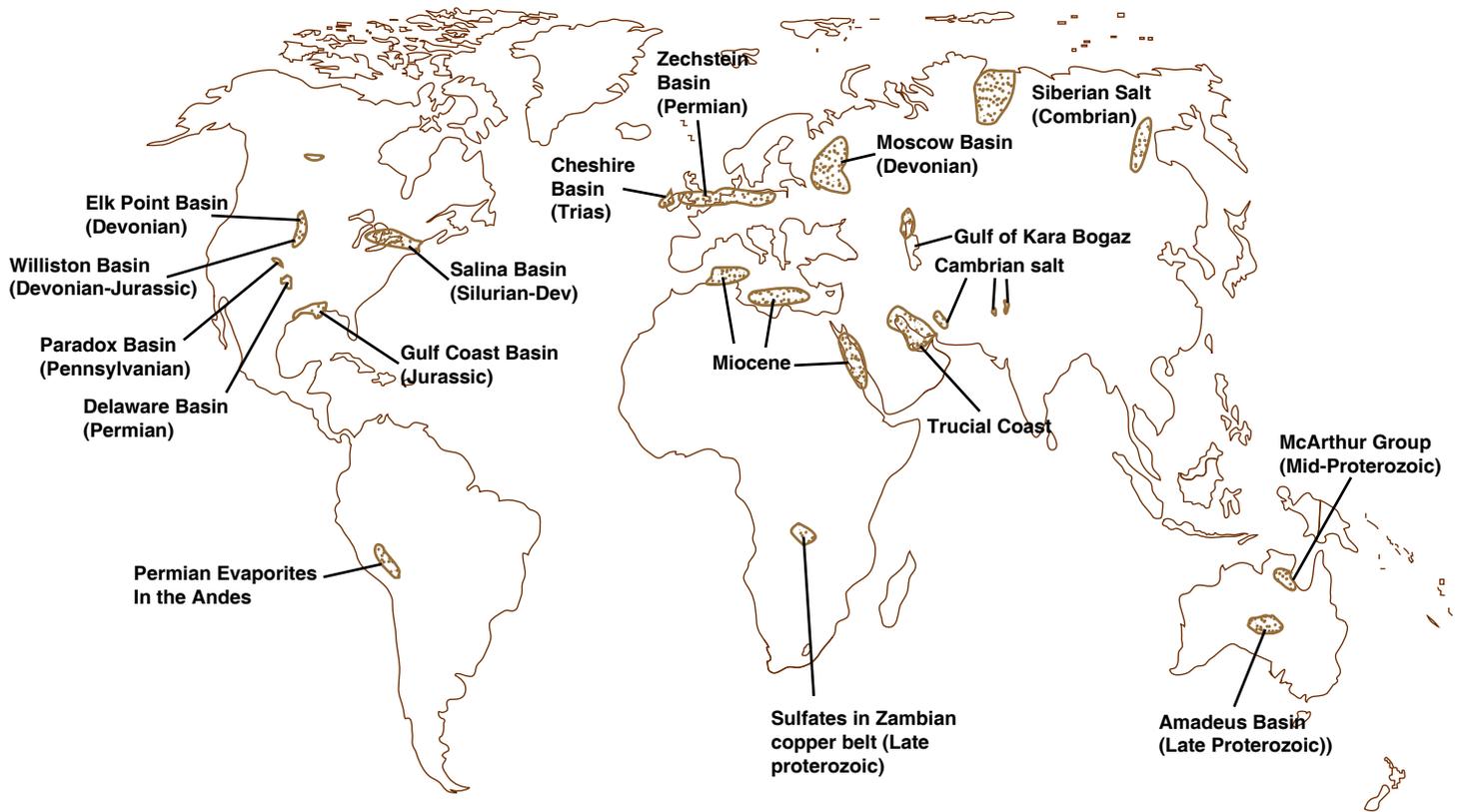


Figure by MIT OCW.

Figure 6-6: Major evaporite deposits of the world

**2.1.2** The Permian Period was the time of greatest evaporite deposition—presumably because much of the great supercontinent of Pangea lay in a low-latitude belt of continental aridity in what we call, in the present time, the “horse latitudes”, where the troposphere experiences widespread subsidence in the downgoing leg of the low-latitude Hadley cell of the general circulation of the atmosphere.

**2.1.3** If you leave dolomite out of consideration (and some nearly primary dolomite must have been associated with hypersaline conditions), the three most common evaporite minerals are **gypsum** (a hydrated calcium sulfate mineral), **anhydrite** (an anhydrous calcium sulfate mineral), and **halite**, NaCl. But *the only evaporite you as field geologists are likely to see is gypsum*, because halite dissolves so readily in surface waters except in the most arid regions, and anhydrite, being the high-pressure anhydrous form of calcium sulfate, is mostly restricted to depths greater than tens to hundreds of meters.

**2.1.4** *Are you likely to see evaporites in outcrop?* The answer is yes, for gypsum, but no, for halite (the two most voluminous evaporite deposits). Only in the driest of climates—in the interior of Iran, for example—is rock salt exposed at the Earth’s surface.

**2.1.5** Evaporites are *excellent indicators of paleoclimate*: it takes a hot and arid climate for major evaporite deposits to form. As you will see in a later chapter on economic sedimentology, evaporites can form seals for the accumulation of petroleum hydrocarbons in reservoir rocks. On a more prosaic note, gypsum is the mineral that’s used to make plaster and “drywall” (gypsum wallboard).

**2.1.6** Evaporite deposits are known from all the continents, with ages ranging from Precambrian to Late Cenozoic (although Precambrian evaporites are scarce, either because they were not deposited or because they have been dissolved away during diagenesis through geologic time). Unfortunately, however, *there are no modern evaporite depositional environments representative of the extensive areas of evaporite deposition that existed at various times and places in the geologic past*. As with dolostones, this has led to great difficulties and a diversity of competing theories in interpreting the depositional environments of evaporites.

## **2.2 Evaporite Minerals**

**2.2.1** More than eighty naturally occurring evaporite minerals have been identified. The intricate equilibrium relationships among these minerals have been the subject of many studies over the years. But there are only about a dozen major evaporite minerals. Here’s a list of the most important evaporite minerals:

anhydrite  $\text{CaSO}_4$   
calcite  $\text{CaCO}_3$   
carnallite  $\text{KMgCl}_3 \cdot 6\text{H}_2\text{O}$   
dolomite  $\text{CaMg}(\text{CO}_3)_2$   
gypsum  $\text{CaSO}_4 \cdot 2\text{H}_2\text{O}$   
halite  $\text{NaCl}$   
kainite  $\text{KMg}(\text{SO}_4)_2 \cdot \text{H}_2\text{O}$   
kieserite  $\text{MgSO}_4 \cdot \text{H}_2\text{O}$   
langbeinite  $\text{K}_2\text{Mg}_2(\text{SO}_4)_3$   
magnesite  $\text{MgCO}_3$   
polyhalite  $\text{K}_2\text{Ca}_2\text{Mg}(\text{SO}_4)_4 \cdot 2\text{H}_2\text{O}$   
sylvite  $\text{KCl}$

**2.2.2** But of the above minerals, only three are of paramount importance: gypsum, anhydrite, and halite.

**Gypsum** is a calcium sulfate mineral containing water of hydration. It's monoclinic with one perfect cleavage, and usually white or colorless but sometimes gray or reddish. It's one of the softest of common minerals: you can scratch it with your fingernail. Its specific gravity is relatively low, 2.32. Although phase equilibria in the  $\text{CaSO}_4\text{--H}_2\text{O}$  system are far from clear, because of the difficulty of studying equilibrium precipitation and dissolution in the laboratory, it seems clear that gypsum is the mineral that is stable at surface conditions and is virtually always the primary evaporite precipitation product.

**Anhydrite** is an anhydrous calcium sulfate mineral. It's orthorhombic, has one perfect cleavage and two other good cleavages, and is usually white to colorless but sometimes gray. Its specific gravity is relatively high, 2.93. It is usually seen as massive granular aggregates rather than well formed crystals. It's stable at higher pressures and temperatures than is gypsum, so gypsum is converted to anhydrite when the evaporite deposit is buried. The deepest reported occurrence of gypsum is about 1300 m.

**Halite** is a cubic mineral with a perfect cubic cleavage. It's colorless when pure but can be reddish from iron and various other colors as well. It's easy to identify by its salty taste. It forms surface outcrops only in the most arid regions. It's also a relatively low-density mineral: the specific gravity is 2.16.

**2.2.3** Not all evaporites are marine deposits. Nonmarine evaporites are common, although the volume of nonmarine evaporites is not nearly as great as that of marine evaporites. Nonmarine evaporites are common in continental extensional basins. There is a great variety of nonmarine evaporite minerals.

## 2.3 Chemistry of Evaporites

**2.3.1** You know, of course, that the oceans are salty. The salinity of sea water is close to 35 parts per thousand, ppt (parts dissolved materials, by mass, per thousand parts of bulk sea water, by mass). Generally the salinity varies only slightly from place to place in the oceans, because of dilution and evaporation, although these slight differences are of extreme importance in controlling the circulation and also the biology of the oceans. But *in certain unusual situations, the salinity can rise to much higher values, giving rise to evaporite precipitation.*

**2.3.2** *Almost all of the salt in sea water is accounted for by seven ions.* Here's a table that lists these ions and their concentration in percent by weight. These seven ions account for 99.7% by weight of all the dissolved salts in sea water.

Cl <sup>-</sup>	55
Na <sup>+</sup>	31
SO <sub>4</sub> <sup>2-</sup>	7.7
Mg <sup>2+</sup>	3.7
Ca <sup>2+</sup>	1.16
K <sup>+</sup>	1.10
HCO <sub>3</sub> <sup>-</sup>	0.41

Of these, the last is controlled by the equilibrium between sea water and the CO<sub>2</sub> of the atmosphere. The dominant control on the rest is *the dissolved-salt content of river runoff.*

**2.3.3** Back in 1849 an Italian chemist named Usiglio made a classic, widely cited, but somewhat misleading experiment on evaporite deposition. He took a volume of normal sea water and slowly evaporated it, and kept track of the composition and mass of precipitated salts as a function of extent of evaporation. He found that when the water is reduced in volume to about 50% of the original volume, a little iron oxide and some aragonite are precipitated, but precipitation of the major evaporite minerals doesn't begin until the volume is reduced to about 20%, when *gypsum* is precipitated. Then when the volume is reduced to about 10%, *halite* begins to be precipitated. Further reduction in volume then leads to the precipitation of various *magnesium sulfates and chlorides*, and finally to *NaBr and KCl*.

**2.3.4** In terms of volumes of precipitated salts, experiments like that of Usiglio show that if a column of sea water 1000 m thick is evaporated to dryness, the precipitated salt deposit would be about 15 m thick. Of this, 0.5 m would be gypsum, 11.8 m would be halite, and the rest, 2.7 m, would be mainly salts of potassium and magnesium. But is this how most evaporite deposits are formed? No; read on.

## ***2.4 Kinds of Evaporite Deposits***

**2.4.1** Gypsum rock is usually finely granular but sometimes coarsely crystalline. The mode of occurrence of gypsum is rather variable, because, as you will see presently, diagenetic changes are usually far-reaching. Details of bedding are usually not well preserved. Gypsum is also common as nodules in a matrix of carbonate or shale.

**2.4.2** Anhydrite rock is usually bedded, often with delicate lamination that's continuous for long distances. It's usually fine grained. You seldom see anhydrite in outcrop, however, for reasons that will become apparent soon.

**2.4.3** Rock salt is a massive, coarsely crystalline, and nonjointed rock. It sometimes shows bedding, especially by alternation of beds with greater or lesser concentration of nonhalite impurities (gypsum, anhydrite, carbonates, and siliciclastics).

## ***2.5 Diagenesis of Evaporites***

**2.5.1** Changes in evaporites during burial and unroofing are usually substantial, for three reasons:

- Evaporite minerals in general are very susceptible to solution and reprecipitation.
- Gypsum is converted to anhydrite upon burial, and then the anhydrite is usually reconverted to gypsum as it makes its way to the surface again. When gypsum is replaced by anhydrite, there is a 38% loss in volume, because of the great difference in density between the two minerals. So there's great extra compaction on the way down (which, incidentally, is conducive to nontectonic deformation, because if the pore water, preexisting and newly released, can't drain away fast enough, the reduction in solids volume tends to cause a liquefied state in which the pore-water pressure can approach the lithostatic pressure). Conversely, on the way up the anhydrite is reconverted to gypsum, provided that extra water is available. This expansion tends to cause even more local deformation of the deposit. The change from anhydrite to gypsum on the way up is usually delayed until the near-surface zone of influence of surface groundwater is reached. This gypsum–anhydrite diagenetic cycle is shown in Figure 6-7.

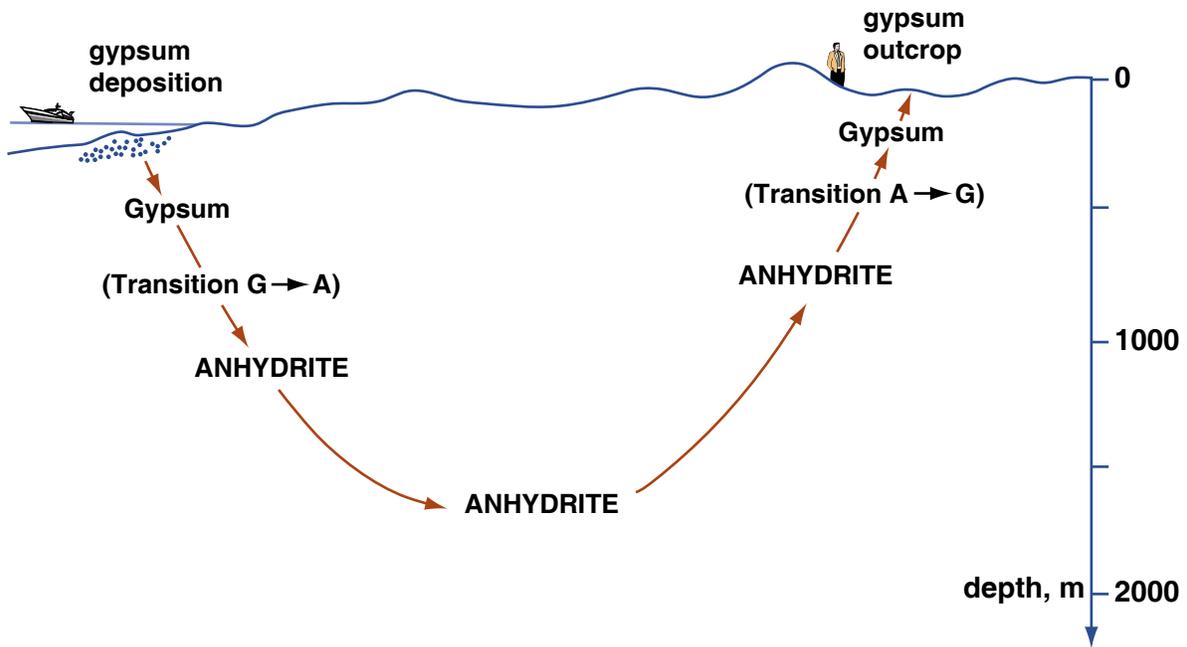


Figure by MIT OCW.

Figure 6-7: The gypsum-anhydrite diagenetic cycle

- *Halite tends to flow plastically on a large scale*, because it's relatively weak and because its deposits have a substantially lower bulk density than the “normal” rocks (carbonates and siliciclastics) with which it's interstratified on a large scale. A horizontal layer of halite lying between layers of denser rock represents a gravitationally unstable situation, and even slight preexisting nonplanarity of the layer leads to uprise of columns or domes of salt, called *diapirs*, typically on scales of several kilometers. Hence the term *salt tectonics*.

## 2.6 Sedimentary Structures in Evaporites

**2.6.1** Do evaporites precipitate as *bottom growths* or as *discrete mineral particles*? The former is demonstrable in some cases, but the widespread existence of primary sedimentary structures, like lamination, cross stratification, ripple marks, and graded bedding—just as in siliciclastic sediments and rocks—gives good evidence that the latter is the more important mode of deposition. It's clear that once evaporite crystals are precipitated they are susceptible to reworking and transportation before final deposition.

**2.6.2** Evaporites are highly susceptible to early and significant diagenesis, owing to the extreme solubility of the evaporite minerals. Most of the sedimentary structures in evaporites are diagenetic. Various kinds of postdepositional folds and veins are common. Veins of gypsum cutting fine siliciclastic rocks (mudrocks and fine sandstones), both parallel and oblique to stratification, are common;

presumably the largely siliciclastic sediment contained an admixture of gypsum, which was remobilized upon shallow burial.

**2.6.3** A striking and common diagenetic structure in calcium sulfate evaporites is *nodular anhydrite* (Figure 6-8: Blatt, H., 1992, *Sedimentary Petrology*, Second Edition: W.H. Freeman, 514 p. (p. 327, Figure 10-7)), also called, picturesquely, *chicken-wire anhydrite*.

## 2.7 Theories of Evaporite Deposition

**2.7.1** The great thickness of some evaporite deposits, as well as the common dominance of gypsum in the deposit, argue against simple evaporation of a body of sea water to dryness. One has to appeal to a continuous evaporation process whereby *a water body partially isolated from the open ocean is subjected to net evaporation* (evaporation is greater than fresh-water runoff into the basin).

**2.7.2** If in such a continuous-evaporation basin there is no outflow of water, only inflow, then all of the dissolved salts must be precipitated somehow, whatever the ratio of sea-water inflow and fresh-water runoff. The common dominance of one or another evaporite mineral suggests, however, that *some percentage of the concentrated brine flows out of the basin, somehow, during the evaporation process*. Presumably the outflow would contain a disproportionate concentration of the harder-to-precipitate salts. The outflow could be as an outgoing underflow of denser, more saline water beneath an inflowing surface current of normal sea water (Figure 6-9A). Or it could be as a gravity-driven diffuse flow through a permeable barrier (Figure 6-9B).

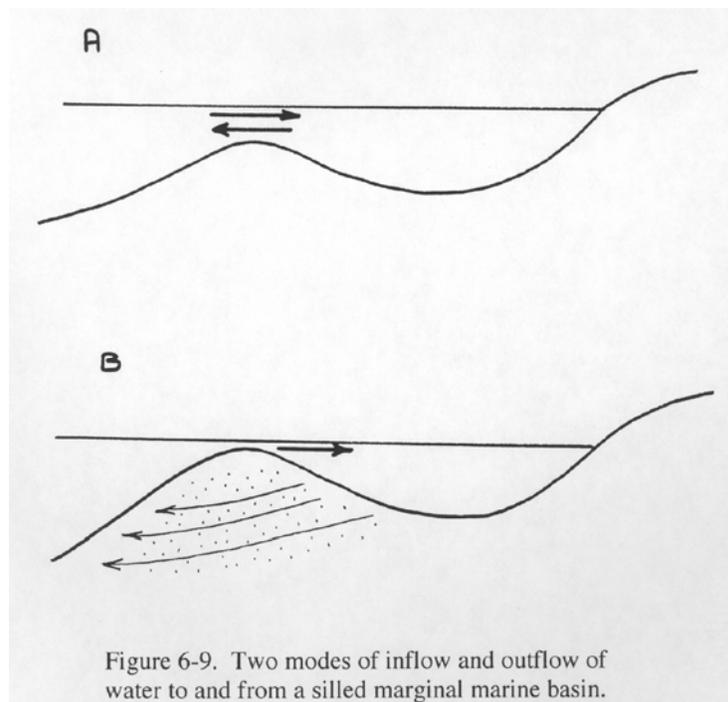


Figure 6-9. Two modes of inflow and outflow of water to and from a silled marginal marine basin.

**2.7.3** If a greater percentage of the water is returned to the open ocean by reflux, then gypsum would be the dominant evaporite, whereas if a lesser percentage of the water is returned by reflux, then halite would be the dominant evaporite. In the least common situation, where an even smaller percentage of the basin water is returned by reflux, the more exotic evaporites would be formed.

**2.7.4** Shallow ocean-margin continuous-evaporation basins of the kind described above are a natural way of accounting for the thick and extensive deposits of evaporites seen in the sedimentary record, but keep in mind that *no good examples are known from the modern*, so it's all a matter of deduction rather than of observation.

**2.7.5** Many evaporite deposits are relatively thin and intimately associated with sediments of other compositions, and show excellent evidence of very shallow-water conditions, like desiccation cracks. An appealing way of accounting for such evaporites is to assume that they are sabkha deposits. A *sabkha* is a low-lying but supratidal surface along a coast with net evaporating conditions and little supply of siliciclastic sediment. (The adjective *supratidal* refers to elevations just above the high-tide level.) Most of the time the sabkha surface is emergent, but occasional storms cover the surface with sea water. That water tends to seep down into the sabkha surface, and during later evaporation the water rises again by capillary action, and the dissolved salt is precipitated as evaporite minerals. If the shoreline is prograding, a layer of evaporites some meters thick may be deposited in that way.