

Authigenic minerals

Minerals that are formed in sediment or a sedimentary rock. Their in-place origin distinguishes them from minerals that are formed elsewhere and transported to the site of deposition (detrital minerals). Authigenic minerals form at the Earth's surface as well as during subsequent burial. The postdepositional processes are referred to as diagenesis, and the resulting minerals are important clues to post-depositional physical and chemical changes in the rock. *See* DIAGENESIS; SEDIMENTARY ROCKS.

Mode of formation. Authigenic minerals precipitate from the overlying water column, pore fluids in the sediment, recrystallization or alteration of preexisting minerals, and structural transformation of one mineral to another. The minerals change in an attempt to equilibrate to the physical and chemical conditions present at any given time. Critical factors in their formation are initial mineral assemblage, temperature, pressure, ionic concentration, pH, electron availability, and the fluid flux through the rock.

In sedimentary rocks it is common to find a record of multiple diagenetic events based on the authigenic minerals. For example, in sediments near the surface, meteoric water may displace original marine pore water, resulting in distinct types of cements. Iron oxide can result from oxidizing fluid. Depletion of oxygen by bacteria may result in the formation of iron sulfides, such as pyrite. During burial, the sediments respond to increasing temperature (up to 200°C; 390°F), pressure (up to 2.5 kilobars; 250 megapascals), and fluid movement from compaction-driven waters or influx of water from the basin flanks. As a result, the sedimentary rock may contain authigenic minerals that record a sequence of events ranging from processes occurring near the sediment-water interface to those forming during deep burial. Unlike metamorphic rocks, the preexisting (detrital) mineral assemblage is at least partially retained, in part due to the sluggish reaction rates at diagenetic conditions. Early cementation processes often seal up the rock, preventing subsequent diagenetic reactions and preserving the original detrital mineral assemblage.

Examples. Authigenic minerals occur in all sedimentary rock and can vary from trace amounts to virtually the total rock (see **table**). The carbonate minerals calcite, dolomite, and siderite are some of the most common types. They form in a wide range of depositional environments and at varying burial depths. Calcite and dolomite form the principal minerals in limestones and dolostones, respectively, as well as cements in sandstones or shales. Carbonate cements result from recrystallization of detrital carbonates and from dissolution of other calcium, iron, and magnesium minerals with carbon dioxide from organic reactions. Much of the calcite in limestones initially consisted of aragonite or magnesium-rich calcite, whereas most dolomite has been formed by the chemical alteration of calcite. Recrystallization may change aragonite to calcite. Aragonite (orthorhombic) is a naturally unstable form of calcium carbon-

Common authigenic minerals

Mineral	Formula
Albite	$\text{NaAlSi}_3\text{O}_8$
Anatase	TiO_2
Anhydrite	CaSO_4
Apatite*	$\text{Ca}_5(\text{PO}_4)_3(\text{F}, \text{Cl}, \text{OH})$
Aragonite (orthorhombic)	CaCO_3
Barite	BaSO_4
Boehmite	$\text{AlO}(\text{OH})$
Calcite (hexagonal)	CaCO_3
Celestite	SrSO_4
Clay minerals	
Chlorites*	$(\text{Mg}, \text{Fe}^{2+}, \text{Fe}^{3+})_8 - (\text{Al}, \text{Si}_3)\text{O}_{10}(\text{OH})_8$
Illites*	$\text{K}(\text{Al})_2(\text{AlSi}_3)\text{O}_{10}(\text{OH})_2$
Kaolinite	$\text{Al}_2\text{Si}_2\text{O}_5(\text{OH})_4$
Smectites*	$(\text{Na}, 0.5\text{Ca})_{0.5}(\text{Al}, \text{Mg}, \text{Fe})_2 - (\text{Al}, \text{Si}_3)\text{O}_{10}(\text{OH})_2 \cdot n\text{H}_2\text{O}$
Dolomite	$\text{CaMg}(\text{CO}_3)_2$
Gibbsite	$\text{Al}(\text{OH})_3$
Glauconite*	$\text{K}(\text{Al}, \text{Mg}, \text{Fe}^{2+}, \text{Fe}^{3+})_2 - (\text{Al}, \text{Si}_3)\text{O}_{10}(\text{OH})_2$
Goethite	$\text{Fe}_2\text{O}_3 \cdot n(\text{H}_2\text{O})$
Gypsum	$\text{CaSO}_4 \cdot 2(\text{H}_2\text{O})$
Halite	NaCl
Hematite	Fe_2O_3
Leucosene	TiO_2
Limonite	$\text{FeO}(\text{OH}) \cdot n(\text{H}_2\text{O})$
Opal (amorphous)	$\text{SiO}_2 \cdot n(\text{H}_2\text{O})$
Orthoclase	KAlSi_3O_8
Pyrite (isometric)	FeS_2
Pyrolusite	MnO_2
Quartz	SiO_2
Siderite	FeCO_3
Zeolites*	$\text{X}^{1+}, 2^{+} \text{Al}_x\text{Si}_{1-x}\text{O}_2 \cdot n\text{H}_2\text{O}$
Clinoptilolite*	$(\text{Na}, 0.5\text{Ca}, \text{K})_{3.5} \text{Al}_{3.5}\text{Si}_{14.5}\text{O}_{36} \cdot n\text{H}_2\text{O}$
Analcime	$\text{NaAlSi}_2\text{O}_6 \cdot \text{H}_2\text{O}$
Laumontite	$\text{CaAl}_2\text{Si}_4\text{O}_{12} \cdot 4\text{H}_2\text{O}$

* Group of minerals characterized by considerable chemical variation.

ate. With the passage of geologic time, aragonite normally inverts to the more stable calcite (hexagonal). The substitution of magnesium for calcium is responsible for the conversion of calcite or aragonite to dolomite, and it has been shown that dedolomitization (replacement of magnesium by calcium) is also possible. *See* ARAGONITE; CALCITE; CARBONATE MINERALS; CEMENT; DOLOMITE.



Fig. 1. Calcite concretion formed in Paleocene marine siltstone, Moeraki beach, New Zealand. Uncemented siltstone has been eroded away by wave action. Concretion is about 2 m (7 ft) in diameter, which is unusually large.

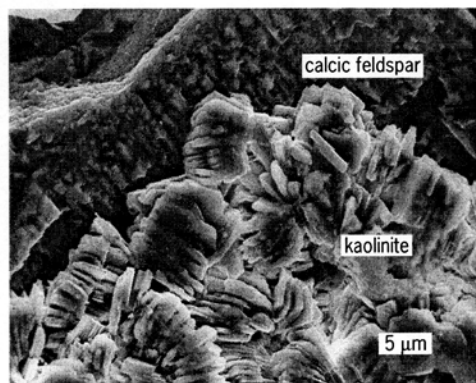


Fig. 2. Scanning electron micrograph of authigenic kaolinite. The clay mineral formed from dissolution of calcic feldspar in a Miocene sandstone, San Joaquin basin, California.

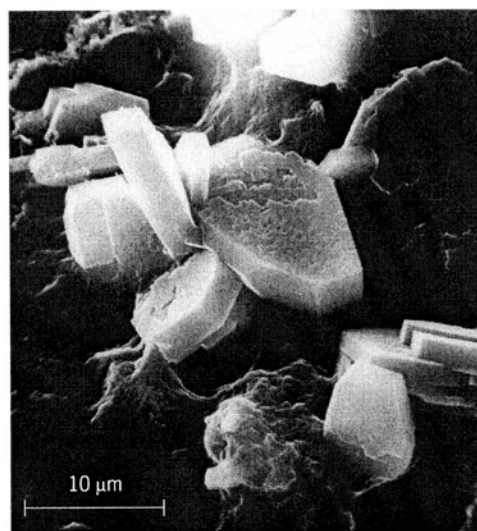


Fig. 3. Scanning electron micrograph of clinoptilolite zeolite (white crystals) on smectite-coated (dark background) sand grains. Sample is from Paleocene volcanic sandstone, Baja California, Mexico.

Sandstones and shales are often cemented near the sediment-water interface by carbonate, which forms subrounded shapes called concretions (Fig. 1). Early-formed iron oxides of hematite or goethite may coat grains when sediments are exposed to oxidizing waters or iron sulfides (such as pyrite) if the water is low in dissolved oxygen. Quartz cements in sandstones form overgrowths on detrital quartz grains at advanced burial conditions. Potential reactions producing silica include quartz dissolution at quartz-clay grain contacts and breakdown of aluminosilicates such as clay minerals and feldspars. Overgrowth of authigenic feldspar on detrital feldspar is very common. At higher temperatures, calcic plagioclase is replaced by albite.

Authigenic clay minerals—chlorite, illite, and smectite—occur as grain coatings and cements. Clay

minerals in sandstones commonly result from alteration of feldspars (Fig. 2). Commonly the dissolution within the feldspar will result in porosity, and the clay forms a pore-filling cement between the sand grains. In shales, mixed-layer smectite/illite clays are particularly common, and during progressive burial the increases in temperature and time cause the smectite to be converted to illite. Volcanic glass in sediments is often altered to zeolite and smectitic clay minerals (Fig. 3). With increasing geologic time the zeolite mineral assemblage becomes simpler due to the breakdown of relatively unstable early-formed zeolites. Pure fine-grain quartz rocks called chert (flint, jasper) crystallize from biogenic opaline silica during burial. See CHERT; CLAY; FELDSPAR; ZEOLITE.

Commonly, authigenic mineral assemblages are a result of a series of dehydration and desilicification reactions during progressive burial. Examples are reaction of gypsum to anhydrite, goethite to hematite, opaline silica to quartz, smectite to illite, and transformation of zeolites to feldspars. To balance these reactions, other processes and reactions consume silica and water. Recrystallization in authigenic mineral formation leads to an enlargement of crystal size through the process of dissolution followed by precipitation. Examples are large halite, gypsum, and anhydrite crystals in evaporite deposits and calcite crystals in some limestones. See GYPSUM; HALITE.

Identification. Since many of the authigenic minerals shown in the table may be deposited in sediments as detrital minerals, it is important that their authigenic origin be clearly established, especially when these minerals are utilized to interpret diagenetic chemical, temperature, and pressure conditions. When grown into pore space, authigenic minerals often exhibit crystalline shapes. In contrast, detrital mineral surfaces are irregular, rough, or rounded because they were leached in a weathering environment, were abraded during transportation, or have been leached further by pore fluids after deposition. Authigenic minerals also form obvious cements that coat grains and completely fill the original pore space in the sediment. In many cases, the authigenic minerals use preexisting crystal structures as templates for their growth. Examples are feldspars and zeolites forming on detrital feldspar, quartz overgrowths on quartz, and carbonates growing on shell tests. Although the authigenic origin of minerals can often be established from the study of thin sections, the scanning electron microscope and electron microprobe are particularly useful for their identification and chemical study. The identification of many fine-grain minerals such as clays and zeolites is possible only with x-ray diffraction and scanning electron microscope techniques (Fig. 2). The high magnification and resolution attainable with the scanning electron microscope make it possible to differentiate these minerals from like species of detrital origin by their crystal form. See MICROSCOPE; SCANNING ELECTRON MICROSCOPE; X-RAY DIFFRACTION.

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