Sandstones

Introduction:

Sandstones make up nearly one-quarter of the sedimentary rocks in the geologic record. Sand and coarse silt-size particles (ranging in size from 1/16 to 2 mm.) in sandstones constitute the framework fraction of the sandstones. Sandstones may also contain various amounts of matrix (material <~0.03mm) and cement, which are present within interstitial pore space among the framework grains as well as empty pore space (porosity).

Particle composition

Sandstones are composed of detrital minerals, rock fragments and accessory minerals (Table 4.1). Detrital constituents are defined as those derived by mechanical–chemical disintegration of a parent rock. Most detrital constituents in sandstones are terrigenous siliciclastic particles that are generated through the process of weathering, explosive volcanism, and sediment transport from parent rocks located outside the depositional basin. A few detrital constituents in sandstones may be nonsiliciclastic particles, such as skeletal fragments or carbonate clasts, formed within the depositional basin by mechanical disruption of reef masses or other consolidated or semiconsolidated carbonate bodies.

The framework grains of most sandstones are composed predominantly of **quartz**, **feldspars**, and **rock fragments**. **Clay minerals** may be abundant in some sandstones as matrix constituents. **Coarse micas**, especially muscovite, make up a few percent of the framework grains of many sandstones. Finally, **heavy minerals** may constitute a small percentage of the detrital constituents of sandstones, such as zircon, tourmaline, and rutile.

Major Minerals

Quartz

The silica minerals are the most abundant minerals in most sandstones. **Quartz** is the dominant silica mineral. **Chalcedony** (fibrous quartz) and **opal** (amorphous and crystalline silica) may be present in chert grains a common rock fragment in sandstones, and opal may be present as a cement. Cristobalite and tridymite are uncommon in most sandstones.

Table 4.1 *Common minerals and rock fragments in siliciclastic sedimentary rocks*

Major minerals (abundance > ~ 1–2%)
- Stable minerals (greatest resistance to chemical decomposition)
  - Quartz – makes up approximately 65% of average sandstone, 30% of average shale; 5% of average carbonate rock
- Less stable minerals
  - Feldspars – include K-feldspars (orthoclase, microcline, sanidine, anorthoclase) and plagioclase feldspars (albite, oligoclase, andesine, labradorite, bytownite, anorthite); make up about 10–15% of average sandstone, 5% of average shale, < 1% of average carbonate rock
  - Clay minerals and fine micas – clay minerals include the kaolinite group, illite group, smectite group (montmorillonite a principal variety), and chlorite group; fine micas are principally muscovite (sericite) and biotite; make up approximately 25–35% of total siliciclastic minerals, but may comprise > 60% of the minerals in shales

Accessory minerals (abundances < ~ 1–2%)
- Coarse micas – principally muscovite and biotite
- Heavy minerals (specific gravity > ~ 2.9)
  - Stable nonopaque minerals – zircon, tourmaline, rutile, anatase
  - Metastable nonopaque minerals – amphiboles, pyroxenes, chlorite, garnet, apatite, staurolite, epidote, olivine, sphene, zoisite, clinozoisite, topaz, monazite, plus about 100 others of minor importance volumetrically
  - Stable opaque minerals – hematite, limonite
  - Metastable opaque minerals – magnetite, ilmenite, leucoxene

Rock fragments (make up about 10–15% of the siliciclastic grains in average sandstone and most of the gravel-size particles in conglomerates; shales contain few rock fragments)
- Igneous rock fragments – may include clasts of any igneous rock, but fragments of fine-crystalline volcanic rock and volcanic glass are most common in sandstones
- Metamorphic rock fragments – include metaquartzite, schist, phyllite, slate, argillite, and less commonly gneiss clasts
- Sedimentary rock fragments – any type of sedimentary rock possible in conglomerates; clasts of fine sandstone, siltstone, shale, and chert are most common in sandstones; limestone clasts are comparatively rare in sandstones

Chemical cements (abundance variable)
- Silicate minerals – predominantly quartz; others may include chalcedony, opal, feldspars, and zeolites
- Carbonate minerals – principally calcite; less commonly aragonite, dolomite, siderite
- Iron oxide minerals – hematite, limonite, goethite
- Sulfate minerals – anhydrite, gypsum, barite

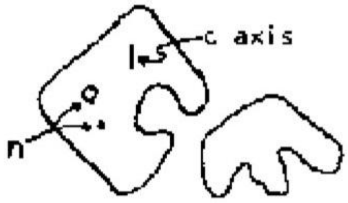
Quartz (SiO2) is the dominant mineral in sandstones and it is a stable mineral. Quartz may range from less than 5 percent (rare) to more than 95 percent. Its relative abundance in most source rocks, coupled with its chemical resistance to weathering and mechanical resistance to abrasion during transport, so it is abundance in sandstones.

There are three types of Quartz, **Igneous**, **Metamorphic** and **Sedimentary** in origin. The **Igneous Quartz** may be (1) **Plutonic** characterized by straight to slightly undulatory extinction. Usually contain no inclusions other than a small amount of scattered vacuoles. (2) **Volcanic** Quartz, the recognition of this quartz is based on a shape, whole or fragmental bi-pyramidal crystals with straight edges and rounded corners. They often have large rounded corrosion embayments. Extinction always straight and usually contain no inclusions. There are rarely any bubble inclusions but bits of volcanic glass.(3) **vein** quartz, many grains have straight extinction, and some

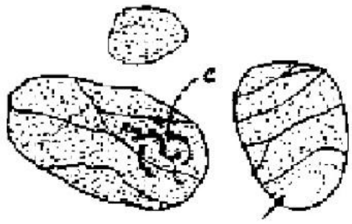
Dr. Zaid A. Malak (Dept. of Geology—Mousl University)
Email (zaid_692006@yahoo.com)

times the extinction slightly undulate. Vacuoles are very common. Vermicular Chlorite is very diagnostic of hydrothermal quartz.

Plutonic Quartz

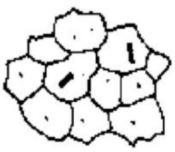


Volcanic Quartz

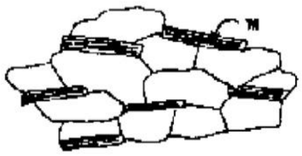


Vein Quartz

Metamorphic quartz may be (1) **recrystallised** if it have straight boundaries between equant interlacing grains forming mosaic, straight with slightly undulose extinctions, microlites and vacuoles are present. Or (2) **Schistose**, elongate ,composite with straight borders, Mica inclusions are present, with straight with slightly undulose extinctions. (3) **Stretched**, strong undulose extinctions, borders may be smooth, crenulated or granuled. Elongate, lenticular shape of crystal units. microlites and vacuoles are present.



Recrystallized Metamorphic Quartz



Schistose Metamorphic Quartz

Stretched Metamorphic Quartz

Reworked **sedimentary** quartz, characterized by the secondary silica cementation called (Overgrowth).

Detrital quartz grains, particularly small grains, tend to be subangular; however, grains that have undergone an episode of intensive eolian transport, or polycyclic grains that have undergone several episodes of transport and deposition, may be well rounded. Such rounded quartz grains that undergo secondary silica cementation during diagenesis may assume crystalline outlines owing to precipitation of silica over growths onto the rounded grains in the presence of open pore space (Fig. 4.1). The original rounded grain outline is commonly revealed by the presence of small specks of hematite, clay, or other material.

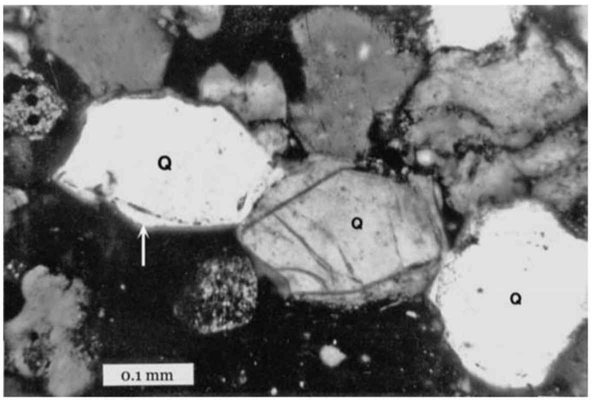


Figure 4.1 Detrital quartz grains (Q) that developed crystal faces owing to formation of overgrowths (arrow). Deadwood Formation (Cambrian–Ordovician). South Dakota. Crossed nicols.

Quartz occurs in sandstones as sand-size crystals (**monocrystalline quartz**) and **(polycrystalline quartz)**. Polycrystalline quartz, also called **composite quartz**, is quartz made up of aggregates of two or more crystals (Fig. 4.2). The individual crystals within a polycrystalline grain may be equant or elongate in shape, fine grained or coarse grained, all about the same size or variable in size, and have crystal boundaries that are relatively straight or sutured to various degrees.

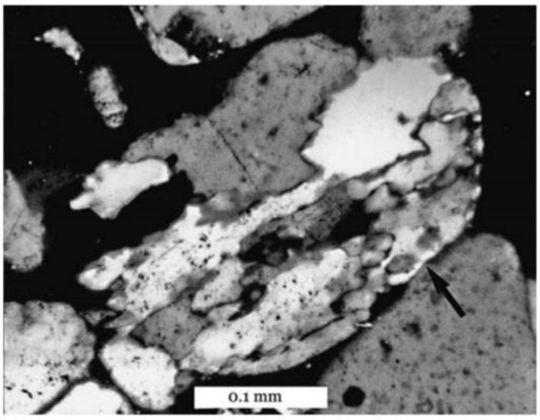


Figure 4.2 Large, well-rounded polycrystalline quartz grain (arrow) that displays sutured internal boundaries between composite crystals, indicating a probable early stage in development of metamorophic polycrystalline quartz. Unknown formation. Crossed nicols.

polycrystalline quartz can develop from monocrystalline quartz during metamorphism. Under the influence of increasing pressure and temperature, nonundulatory monocrystalline quartz changes progressively to undulatory quartz, polygonized quartz (quartz that shows distinct zones of extinction with sharp boundaries), and finally to polycrystalline quartz.

Polycrystalline quartz grains can also form in plutonic igneous rocks. The nature of the polycrystalline grains depends upon the type of igneous or metamorphic rock from which the grains were derived. In addition to polycrystalline quartz grains that originate as single grains in metamorphic or plutonic igneous rocks, clasts of quartzite, chert, and quartz-rich sandstone are considered by some authors to be polycrystalline quartz grains. Quartz may be derived from igneous rocks, especially acid plutonic rocks, various kinds of metamorphic rocks, and sedimentary rocks.

Feldspars

Feldspars are the most common framework mineral in sands and sandstones after quartz. Feldspars make up about 10–15 percent of the average ancient sandstone. Sandstones containing more than 25 percent feldspar are considered feldspar-rich. Feldspars are commonly divided into two main groups: alkali feldspars (K-Na feldspars) and plagioclase feldspars. Both groups are well represented in detrital sediments. alkali feldspars (K-Na feldspars) are generally regarded to be more abundant than plagioclase feldspars in the average sandstone. Sodic plagioclase (Na-Albite) tends to be abundant in sandstones, ((however, sandstones derived from source areas rich in volcanic rocks (Basalt-Basic Rocks) may contain more plagioclase than potassium feldspar)). Some important characteristics of detrital feldspars are summarized in Table 4.3.

Feldspars are chemically less stable than quartz and are more susceptible to chemical destruction during weathering and diagenesis. Because they are also softer than quartz, feldspars become more readily rounded during transport. They also appear to be somewhat more prone to mechanical shattering and breakup owing to their cleavage. Feldspars are less likely than quartz to survive several episodes of recycling, although they can survive more than one cycle if weathering occurs in a moderately arid or cold climate. Owing to this possibility of recycling, the presence of a few feldspar grains in a sedimentary rock does not necessarily mean that the rock is composed of first-cycle sediments derived directly from crystalline igneous or metamorphic rocks. On the other hand, a high content of feldspars, particularly on the order of 25 percent or more, probably indicates derivation directly from crystalline source rocks.

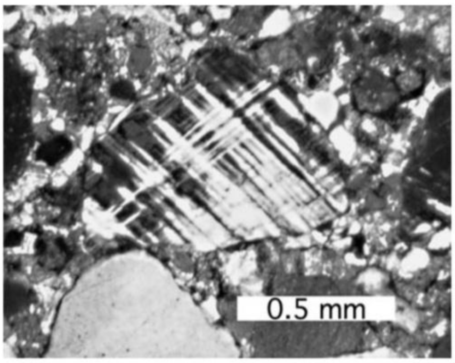


Figure 4.4 Microcline grain (center of photograph) with well-developed grid twinning. Fountain Formation (Pennsylvanian), Colorado. Crossed nicols.

Dr. Zaid A. Malak (Dept. of Geology—Mousl University)
Email (zaid_692006@yahoo.com)

Table 4.3 *Characteristics of detrital feldspars*

Alkali (potassium–sodium) feldspars

Complete solid-solution series from $K(AlSi_3O_8)$ (orthoclase, sanidine, microcline) to $Na(AlSi_3O_8)$ (anorthoclase); all biaxial (−) with low negative relief

Orthoclase – a common detrital feldspar characterized by birefringence lower than quartz and 2V ranging from ~40° to 75°; extinction angle ~13°–21°; twinned (Carlsbad twins most common) or untwinned; may appear cloudy owing to alteration products; distinguished from quartz most easily by staining

Sanidine – a high-temperature feldspar, with similar appearance to orthoclase, derived mainly from volcanic rocks; distinguished from orthoclase by smaller 2V (0°–47°); extinction angle ~5°–21°; commonly has fewer alternation products than orthoclase

Microcline – a common detrital feldspar distinguished particularly by distinctive cross-hatch twinning with twin lamellae approximately at right angles; 2V commonly larger than 65° (range 50°–85°); extinction angle ~5°–15°; may be cloudy owing to the presence of alteration products

Anorthoclase – comparatively rare in sandstones; extinction angle ~5°–20°; distinguished from microcline by finer-scale cross-hatch twinning and smaller 2V (~43°–54°)

Perthite – alkali feldspars characterized by platy intergrowths of albite

Plagioclase feldspars

Complete solid-solution series ranging from $NaAlSi_3O_8$ (albite) to $CaAl_2Si_2O_8$ (anorthite); composition commonly expressed as percent anorthite (An) molecule. Large 2V, which ranges with composition from about 50° to 105°; biaxial (+) or (−); extinction angles vary as a function of composition and temperature of crystallization and may range from ~3° to 60°; indices of refraction also vary with composition. Twinned or untwinned; if albite twinning present, easily distinguished from alkali feldspars. If untwinned, distinguished by large 2V, extinction angles, or by staining. Plagioclase from igneous rocks may display compositional zoning. Very common in sandstones derived from volcanic and metamorphic rocks but may also be derived from plutonic igneous rocks. [Check published references such as Nesse (1986) for details of extinction angles, 2Vs, and indices of refraction.] Members of the series are

Albite (An_0–An_{10})
Oligoclase (An_{10}–An_{30})
Andesine (An_{30}–An_{50})
Labradorite (An_{50}–An_{70})
Bytownite (An_{70}–An_{90})
Anorthite (An_{90}–An_{100})

Note: Data on extinction angles and 2Vs mainly from Nesse, 1986.

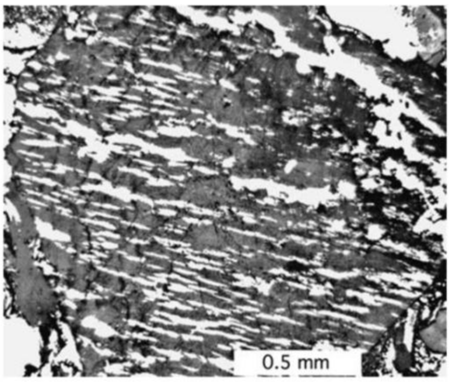


Figure 4.5 Large K-feldspar grain with abundant perthitic lamellae. Unknown formation. Crossed nicols.

Dr. Zaid A. Malak (Dept. of Geology—Mousl University)
Email (zaid_692006@yahoo.com)

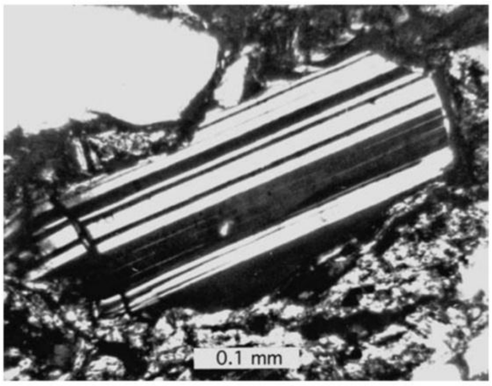


Figure 4.6 Large, twined plagioclase grain from Miocene deep-sea sandstone, Ocean Drilling Program Leg 127, Site 796, Japan Sea, 326 m below seafloor. Crossed nicols.

Coarse Mica

The principal coarse micas that occur as detrital grains in sandstones are muscovite and biotite. Chlorite, which occurs as detrital grains in some cases. The micas are distinguished from other minerals by their platy or flaky habit. (Fig. 4.7).

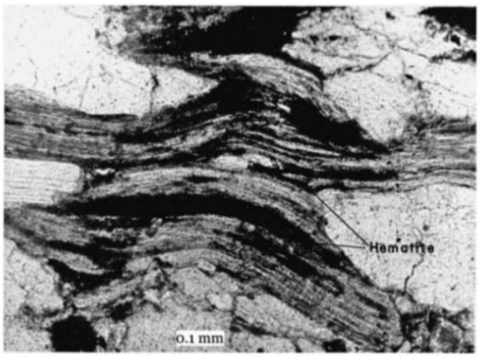


Figure 4.7 Large mica (biotite) grain cut normal to basal section. Hematite replaces biotite along some cleavage planes (black). Minturn Formation (Pennsylvanian), Colorado. Plane polarized light.

Muscovite is colorless in thin section and very weak. Biotite is normally yellow, brown, or green, but may be leached pale yellow or almost colorless. Chlorite is generally green. Micas and chlorite are derived primarily from metamorphic rocks, but biotite occurs also in volcanic rocks and in granites. Muscovite occurs also in granites and pegmatites. Coarse micas rarely form more than about two percent of the framework grains of sandstones and commonly much less. Muscovite is chemically more stable than biotite and is much more abundant in sandstones on the average than biotite. Coarse detrital chlorite is less abundant than biotite, possibly as a result of the greater tendency of chlorite to degrade mechanically to finer-size grains. Detrital micas are rarely rounded, and they are commonly deposited with their flattened dimension parallel to bedding. Owing to their sheetlike shape and consequent low settling velocity, they tend to be hydraulic equivalents of finer grains and thus are commonly deposited in very fine sands and silts rather than in coarser sands .

Clay minerals

Clay minerals are common in sandstones as matrix constituents. They occur also within argillaceous rock fragments. Because of their fine grain size, clay minerals cannot easily be identified under the petrographic microscope. Accurate identification requires X-ray dif- fraction methods or use of the scanning electron microscope or the electron probe micro- analyzer. In most cases, they are simply lumped together with fine-size (<0.03mm) quartz, feldspars, and micas as “matrix.” Furthermore, there is growing evidence that much of the matrix of sandstones may be authigenic, derived during diagenesis by chemical precipitation of clay minerals into pore space or alteration of framework grains to clays.

Heavy minerals

Most sandstones contain small quantities of sand-size accessory minerals. Most of these minerals have specific gravities that exceed 2.85 and are thus called heavy minerals. heavy minerals are separating by using heavy liquids. After washing and drying, a grain mount of the heavy minerals is prepared for microscopic study.

Heavy minerals are commonly divided into two groups on the basis of optical properties: opaque and nonopaque. Opaque heavy minerals include magnetite, ilmenite, hematite and limonite, pyrite, and leucoxene. The non - opaque heavy minerals encompass a very large group of more than 100 minerals, of which olivine, clinopyroxenes, orthopyroxenes, amphiboles, garnet, epidote, kyanite, sillimanite, andalusite, tourmaline, and zircon are particularly common. The average heavy-mineral content of sandstone is commonly reported to be about 1 percent or less. Heavy minerals are still thought to be useful indicators of sediment provenance.

Rock fragments

Rock fragments are detrital particles made up of two or more mineral grains. Depending upon source-rock composition, almost any kind of rock fragment can be present in a sandstone; however, clasts of fine-crystalline or fine-grained parent rocks are generally most abundant. The most common igneous rock fragments are volcanic rock fragments and glass. Metamorphic clasts include schist, phyllite, slate, and quartzite. Common sedimentary rock fragments are shale, fine sandstone and siltstone, and chert. Clasts of limestone and coarse plutonic and metamorphic rock areless common.

Rock fragments are common constituents of both modern sediments and ancient sandstones and conglomerates. They make up about 10–20 percent of the framework grains in average sandstones. Their range in abundances in ancient sandstones is very broad, however, and extends from less than l percent in some quartzose sandstones to well above 50 percent in some lithic sandstones. The abundance of rock fragments in ancient sandstones is a function of several factors, which include the lithology of their source rocks, their chemical and mechanical durability, and their transport and diagenetic history. On the whole, we find fewer rock fragments in compositionally mature (quartz-rich) sandstones than in immature sandstones. Chert and quartzite clasts are the most chemically stable and mechanically durable clasts. Shale and schist fragments may be less durable mechanically, and limestone fragments are probably the least stable chemically.

Mineral Cements

The framework grains in most siliciclastic sedimentary rocks are bound together by some type of mineral cement. These cementing materials may be either silicate minerals such as quartz and opal or nonsilicate minerals such as calcite and dolomite. Quartz is the most common silicate mineral that acts as a cement. In most sandstones, the quartz cement is chemically attached to the crystal lattice of existing quartz grains, forming rims of cement called overgrowths (Fig. 5.3A). Such overgrowths that retain crystallographic continuity of a grain are said to be syntaxial. Because syntaxial overgrowths are optically continuous with the original grain, they go to extinction in the same position as the original grain when rotated on the stage of a polarizing microscope. Overgrowths can be recognized by a line of impurities or bubbles that mark the surface of the original grain. Quartz overgrowths are particularly common in quartz-rich sandstones. Less commonly, quartz cement is present as microcrystalline quartz, which has a fine-grained, crystalline texture similar to that of chert. When silica cement is deposited as microcrystalline quartz, it forms a mosaic of very tiny quartz crystals that fill the interstitial spaces among framework silicate grains (Fig. 5.3B). Not uncommonly, the crystals next to the framework grains are small, slightly elongated, and are oriented normal to the surfaces of the framework grains. More rarely, opal (an isotropic mineral) occurs as a cement in sandstones, particularly in sandstones rich in volcanogenic materials. Like quartz and microcrystalline quartz (chert), opal is also composed of SiO_2, but, unlike these minerals, opal contains some water and lacks a definite crystal structure. Thus, it is said to be amorphous. Opal is metastable and crystallizes in time to microcrystalline quartz.

Carbonate minerals are the most abundant nonsilicate mineral cements in siliciclastic sedimentary rocks. Calcite is a particularly common carbonate cement. It is precipitated in the pore spaces among framework grains, typically forming a mosaic of smaller crystals (Fig. 5.3C). These crystals adhere to the larger framework grains and bind them together. Less common carbonate cements are dolomite and siderite (iron carbonate). Other minerals that act as cements in sandstones include the iron oxide minerals hematite and limonite, feldspars, anhydrite, gypsum, barite, clay minerals (Fig. 5.3D), and zeolite minerals. Zeolites are hydrous aluminosilicate minerals that occur as cements primarily in volcaniclastic sedimentary rocks (discussed in a subsequent section).

All cements are secondary minerals that form in sandstones after deposition and during burial.

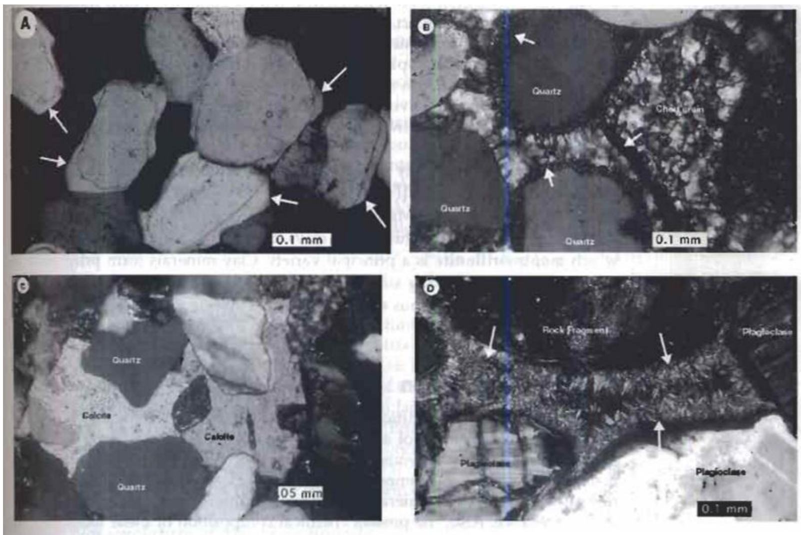


Figure 5.3
Common cements in sandstones. A. Quartz overgrowths (arrows), Lamotte Sandstone (Cambrian), Missouri. B. Microquartz (chert; arrows) cementing quartz and chert grains, Jefferson City Fm. (Ordovician), Missouri. C. Calcite, Kayenta Fm. (Triassic), Utah. D. Clay mineral (chlorite; arrows), Miocene sandstone, Japan Sea. Crossed nicol photomicrographs.

Matrix Minerals

Grains in sandstones smaller than about 0.03 mm, which fill interstitial spaces among framework grains, are referred to as matrix minerals. Matrix minerals may include fine-size micas, quartz, and feldspars; however, clay minerals make up the bulk of matrix grains. Because of their small size, clay minerals are difficult to identify by routine petrographic microscopy. They must be identified by X-ray diffraction techniques, electron microscopy, or other nonoptical methods. Clay minerals are compositionally diverse. They belong to the phyllosilicate mineral group, which is characterized by two-dimensional layer structures arranged in indefinitely extending sheets.

The most common clay mineral groups are **illite** [$K_2(Si_6Al_2)Al_4O_{20}(OH)_4$], **smectite** (montmorillonite) [$(Al, Mg)_8(Si_4O_{10})_3(OH)_{10} \cdot 12H_2O$], **kaolinite** [$Al_2Si_2O_5(OH)_4$], and **chlorite** [$(Mg, Fe)_5(Al, Fe^{3+})_2Si_3O_{10}(OH)_8$]. Kaolinite is a two-layer clay; the others are three-layer clays. Smectite is a clay-mineral group, of which **montmorillonite** is a principal variety. Clay minerals form principally as secondary minerals during subaerial weathering and hydrolysis, although they can also form by subaqueous weathering in the marine environment and during burial diagenesis.

Sandstone Maturity

The term maturity is applied to sandstones in two different ways. **Compositional maturity** refers to the relative abundance of stable and unstable framework grains in a sandstone. A sandstone composed mainly of quartz is considered compositionally mature, whereas a sandstone that contains abundant unstable minerals (e.g., feldspars) or unstable rock fragments is compositionally immature. **Textural maturity** is determined by the relative abundance of matrix and the degree of rounding and sorting of framework grains, as illustrated in Figure 5.6. Textural maturity can range from immature (much clay, framework grains poorly sorted and poorly rounded) to supermature (little or no clay, framework grains well sorted and well rounded). Textural maturity allegedly reflects the degree of sediment transport and reworking; however, it may also be affected by diagenetic processes (i.e., clay minerals may form in pore spaces during burial diagenesis).

Dr. Zaid A. Malak (Dept. of Geology—Mousl University)
Email (zaid_692006@yahoo.com)

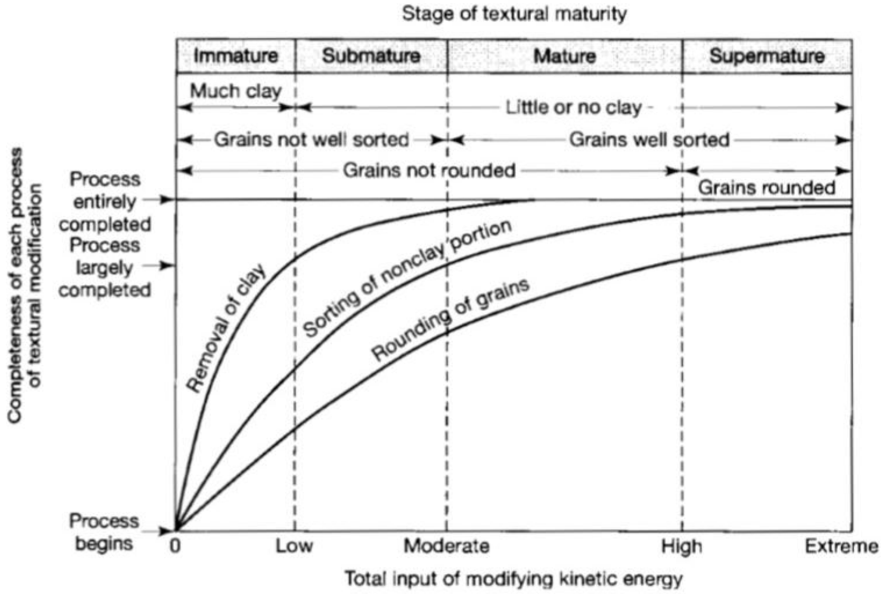


Figure 5.6

Classification of sandstones

Sandstones can be separated into two groups: epiclastic and volcani-clastic. Epiclastic deposits are formed from fragments of pre-existing rocks derived by weathering and erosion. Thus, they are composed mainly of silicate minerals and various kinds of igneous, metamorphic, and sedimentary rock fragments. Volcaniclastic deposits are those especially rich in volcanic debris, including glass.

Epiclastic and volcaniclastic deposits can be further classified on the basis of their composition. Unfortunately, there is little agreement among geologists about sandstone classification, particularly classification of epiclastic sandstones.

Classification of epiclastic sandstone

The framework grains in most sandstones are dominated by quartz, feldspars, and rock fragments. Many other minerals may be present in a given sandstone, but the abundances of these other minerals are so low in most sandstones that they can be ignored for the purpose of sandstone classification. Some sandstones contain matrix in addition to sand-size framework grains. As mentioned, matrix is defined as material less than about 0.03mm (30microns) in size. Thus, it is not a framework constituent. Rather, it occupies the interstitial spaces among sand-size grains. The matrix content of sandstones may range from zero to several tens of percent.

According to Friedman and Sanders (1978, p. 190), more than 50 classifications for sandstones have been published since the late 1940s in ten countries and seven languages.

Examples of classifications

Two types of classifications have emerged: those based upon both the composition of framework grains and the abundance of matrix, and those based entirely upon the composition of framework grains.

Dott (1964) classification, shown in Fig. 4.12, is an example of the first type of classification. Sandstones are divided into two broad groups: **arenites**, containing matrix (<15 percent), and **wackes**, containing matrix (15-75%) and ***Mudstone*** if the matrix more than 75 percent. Combining the matrix parameter with composition (QFL, see Fig. 4.12 for explanation) yields seven kinds of sandstones: quartz arenite and quartz wacke; subarkose and sublitharenite; arkosic arenite and feldspathic graywackes; lithic aenite and lithic graywacke. Gilbert (1982) classification sets the boundary between arenites and wackes at 5 percent matrix.

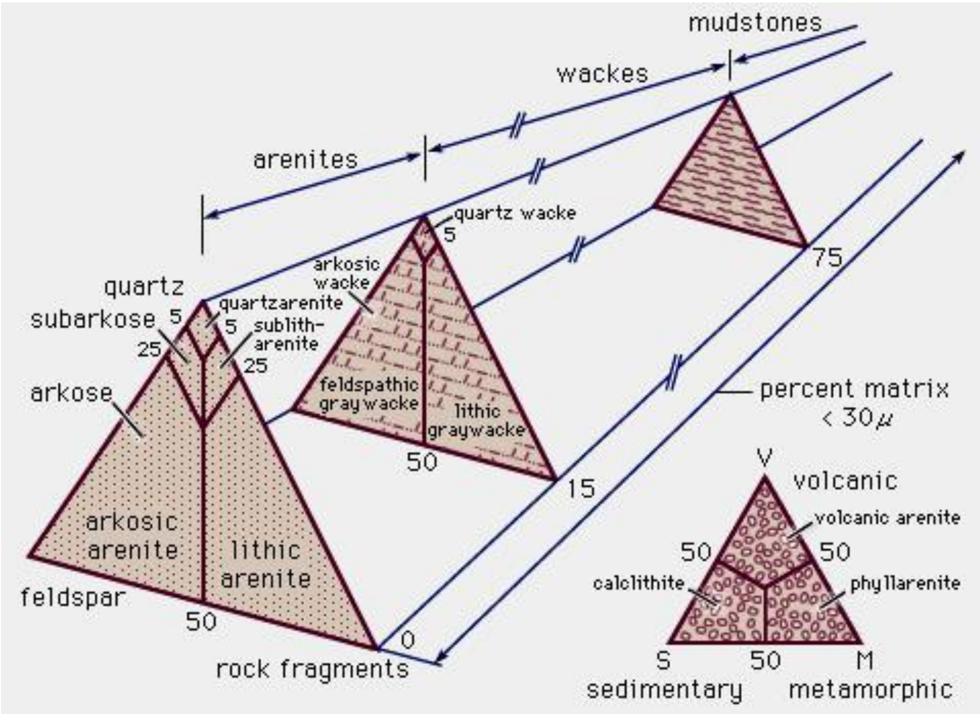


Figure 4.12 (Dott classification).

The two classifications shown in Fig. 4.13 are examples of the second type of classification. These classifications divide sandstones into seven or eight compositional types; however, the classifications do not use matrix as a classification parameter. Note that Fig. 4.13B allows even finer subdivision of sandstones by erecting "daughter" compositional triangles on the main composition diagram.

Two common sandstone names that do not appear in many formal classifications require some additional explanation. The term **arkose** is used in some sandstone classifications and is in general use by many geologists. It has been applied to any sandstone containing obvious amounts of feldspars, as seen in hand specimens, to sandstones containing more than 20 percent, more than 25 percent, or more than 30 percent feldspars, and to sandstones derived from granitic source rocks. Probably the most widely accepted definition for an arkose is **a feldspathic sandstone** containing more than 25 percent feldspars. Presumably, an arkose also has a lower rock-fragment content than feldspar content. The second term **greywacke** is still used by many geologists. In general, the name **graywacke** is applied to dark-gray, greenish-gray, or black, matrix-rich, well- indurated sandstones. The name was originally given to rocks on the basis of their field characteristics.

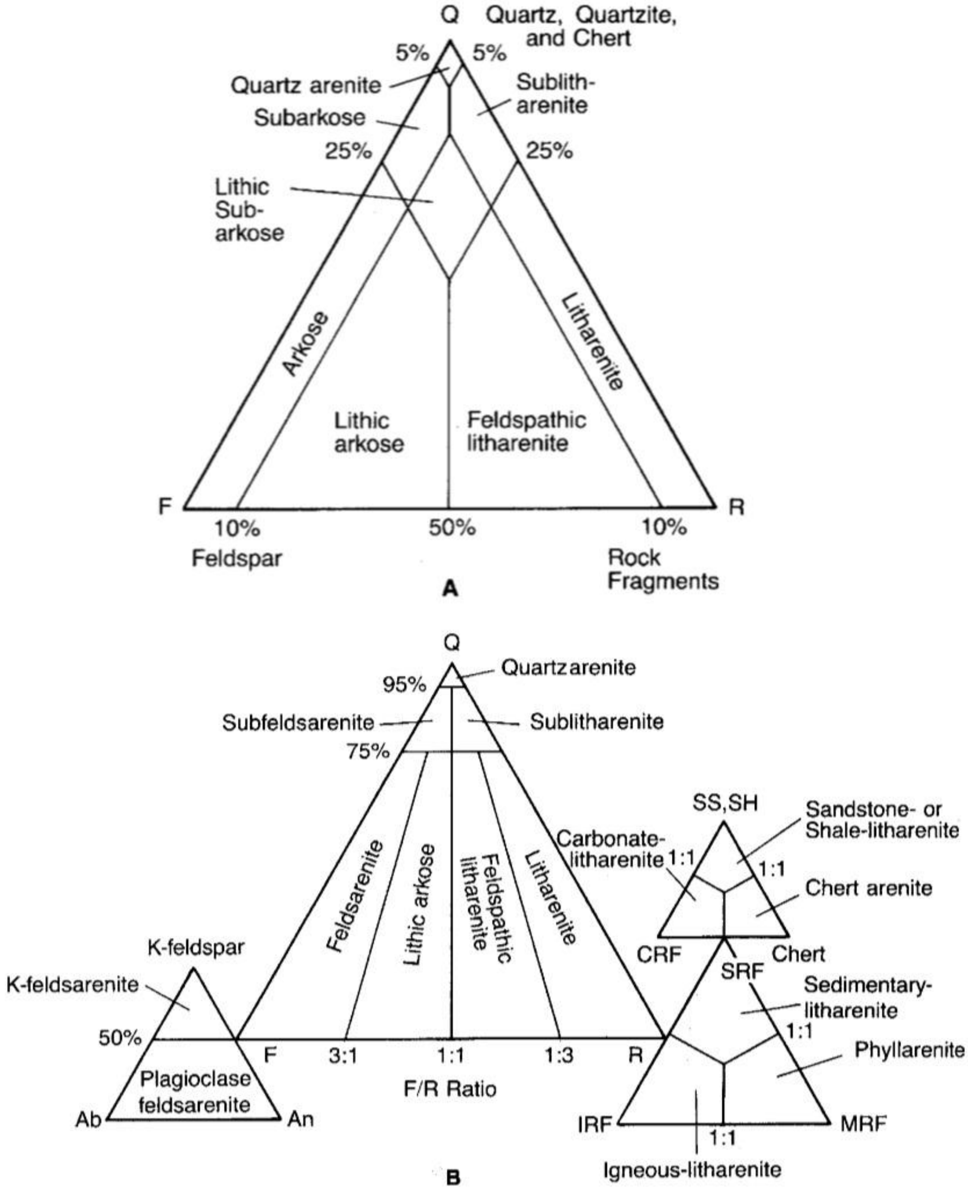


Figure 4.13 Classification of sandstones according to (A) McBride (1963) and (B) Folk *et al.* (1970). In Folk's classification , chert is included with rock fragments at the R pole, and granite and gneiss fragments are included with feldspars at the F pole. SS = sandstone, SH = shale, CRF = carbonate rock fragments, SRF = sedimentary rock fragments, IRF = igneous rock fragments, MRF = metamorphic rock fragments. (A, from McBride, E. F., 1963, A classification of common sandstones: *J. Sediment. Petrology*, **34**, Fig. 1, p. 667, reprinted by permission of SEPM. B, from Folk, R. L., P. B. Andrews, and D. W. Lewis, 1970, Detrital sedimentary rock classification and nomenclature for use in New Zealand: *New Zealand J. Geol. Geophys.*, **13**, Fig. 8, p. 955, and Fig. 9, p. 959, British Crown copyright, reprinted by permission.)

Petrography and chemistry of sandstones

Major petrographic divisions

The preceding discussion indicates that epiclastic sandstones can be divided by particle composition into three major groups: quartz-rich sandstones (quartz arenites and wackes), feldspar-rich sandstones (feldspathic arenites and wackes), and rock-fragment-rich sand- stones (lithic arenites and wackes). These major sandstone clans may differ in more than just particle composition. They may show also detectible variations in texture, types of cements, matrix content, and bulk chemistry.

Quartz arenites

Quartz arenites are composed of >90–95 percent siliceous grains (quartz, chert, quartzose rock fragments). They are commonly white or light gray, but may be stained red, pink, yellow, or brown by iron oxides. They are generally well lithified and well cemented with silica or carbonate cement; however, some are porous and friable. Quartz arenites typically occur in association with assemblages of rocks deposited in stable cratonic environments such as eolian, beach, and shelf environments. Most quartz arenites are texturally mature to supermature according to Folk's(1951) textural maturity classification, but quartz arenites with low maturity exist. Cross-bedding is particularly characteristic of these rocks, and ripple marks are moderately common. Fossils are rarely abundant in these sandstones, possibly owing to poor preservation or to the eolian origin of some quartz arenites, but both fossils and carbonate grains may be present. Also, trace fossils such as burrows of the Skolithos facies may be locally abundant in some shallow-marine quartz arenites. Quartz arenites are common in the geologic record. Pettijohn (1963) estimates that they make up about one-third of all sandstones.

The extreme compositional maturity of quartz arenites requires that these sandstones originate under rather specialized conditions. If quartz arenites are first-cycle deposits, they must form under weathering, transport, and depositional conditions so vigorous that most grains chemically and mechanically less stable than quartz are eliminated. Conceivably, extreme chemical leaching under hot, humid, low-relief weathering conditions, prolonged transport by wind, intensive reworking in the surf zone or tidal zone (reworking by reversing tides), or a combination of these factors might be adequate to generate a first-cycle quartz arenite.

Stream transport is produce little rounding of sand-size quartz grains, whereas wind transport is an effective rounding agent. Therefore, the well-rounded and well-sorted character of many quartz arenites, plus the common presence of large-scale sets of cross-beds with high-angle foresets, suggests that many quartz arenites are **eolian** deposits.

On the other hand, many other quartz arenites are clearly of **marine** origin. These marine quartz arenites may have been deposited in beach or barrier-island settings.

Ferree et al.(1988) conclude that both modern beach sand and ancient beach sandstones have frameworks that are mineralogically more mature than their fluvial counterparts.

Johnsson et al.(1988) report that definite first-cycle quartz arenites are forming in the Orinoco River basin in Venezuela and Colombia. They cite two conditions that are necessary to produce first-cycle quartz arenites: an environment of intense chemical weathering and a mechanism to provide extended time over which weathering can operate (e.g. temporary storage on extensive alluvial plains).

Feldspathic arenites

Feldspathic arenites contain less than 90 percent quartz grains, more feldspar than unstable rock fragments, and minor amounts of other minerals such as micas and heavy minerals. They may contain as little as 10 percent feldspar grains, but most feldspathic arenites show greater feldspar enrichment. Sandstones that contain more than about 25 percent feldspar are commonly called **arkoses**. Some feldspathic arenites are colored pink or red owing to the presence of K-feldspars or iron oxides. They are typically medium to coarse grained and may contain high percentages of subangular to angular grains. Matrix content may range from trace amounts to more than 15 percent, and sorting of framework grains can range from moderately well sorted to poorly sorted. Thus, feldspathic sandstones are commonly texturally immature or submature.

Feldspathic arenites are not especially characterized by any particular kinds of sedimentary structures. Bedding may range from essentially structureless to parallel laminated or cross-laminated. Fossils may be present in marine examples. Feldspathic arenites typically occur in cratonic or stable shelf settings, where they are associated with conglomerates, shallow-water quartz arenites or lithic arenites , carbonate rocks ,and evaporates .Less typically, they occur in sedimentary successions that were deposited in unstable basins or other deeper-water, mobile-belt settings. Feldspathic arenites of the latter type, which are matrix rich and well indurated owing to deep burial, are often called feldspathic graywackes.

Pettijohn (1963) estimates that arkoses make up about 15 percent of all sandstones. Feldspathic arenites in total are probably more abundant than 15 percent, especially if feldspathic graywackes are included.

Feldspathic arenites originate mainly by weathering of feldspar-rich crystalline rocks, either plutonic igneous rocks or feldspar-rich metamorphic rocks. Therefore, most feldspathic arenites are probably first-cycle deposits. Most feldspathic arenites contain considerably more K-feldspar than plagioclase; however, several plagioclase feldspathic arenites are known. These plagioclase-rich feldspathic arenites are derived mainly from volcanic sources.

The preservation of large quantities of feldspars during the process of weathering appears to require that feldspathic arenites originate either (1) under very cold or very arid climatic conditions where chemical weathering processes are inhibited or (2) in warmer, more humid climates where marked relief of local uplifts allows rapid erosion of feldspars before they can be decomposed.

Some arkoses originate essentially *in situ* when granite and related rocks disintegrate to produce a granular sediment called **grus**. These residual arkosic materials may be shifted a short distance downslope and deposited as fans or aprons of waste material, commonly referred to as clastic wedges. These fans may extend into basins and become intercalated or interbedded with better stratified and better sorted sediments. Other feldspathic arenites undergo considerable transport and reworking by rivers or the sea before they are deposited. These reworked sandstones commonly contain less feldspar than do residual arkoses, and they are better sorted and grains are better rounded.

Lithic arenites

Lithic arenites are an extremely diverse group of rocks that are characterized by generally high content of unstable rock fragments. Classified according to Fig. 4.12, any sandstone that contains less than 90 percent quartz (plus chert and quartzite) and unstable rock fragments in excess of feldspars is a lithic arenite. Colors may range from light gray ("salt and pepper") to uniform, medium to dark gray. Many lithic arenites are poorly sorted; however, sorting ranges from well sorted to very poorly sorted. Quartz and many other framework grains are generally poorly rounded. Lithic arenites tend to contain substantial amounts of matrix, most of which may be of secondary origin. They may range from irregularly bedded, laterally restricted, cross-stratified fluvial units to evenly bedded, laterally extensive, graded, marine turbidite units. They occur in association with fluvial conglomerates and other fluvial deposits and in association with generally deeper-water, marine conglomerates, pelagic shales, cherts, and submarine basalts. Lithic arenites include sandstones that many geologists continue to refer to as graywackes. Graywackes differ from "normal" lithic arenites in that they are dark gray to dark green, are well indurated or lithified, and commonly have a matrix consisting of secondary chlorite. The term graywacke is used so loosely, however, that it might be best to simply drop it, as mentioned. Pettijohn (1963) estimates that lithic arenites and graywackes together make up nearly one-half of all sandstones.

Lithic arenites are typically compositionally immature sandstones that originate under conditions favoring the production and deposition of large volumes of relatively unstable materials. The mechanically weak character of many of the lithic fragments in these sandstones suggests that they are probably derived from rugged, high-relief source areas. Lithic arenites may be deposited in nonmarine settings in proximal alluvial fans or other fluvial environments. Alternatively, they may be deposited in marine foreland basins ( ) adjacent to fold-thrust belts, or they may be transported by large rivers off the continent into deltaic or shallow shelf environments. Lithic sediments deposited in coastal areas may be retransported into deeper water by turbidity currents or by other sediment gravity-flow mechanisms. These deeper-water sediments are particularly likely to undergo deep burial and incipient metamorphism, leading to development of characteristics generally ascribed to graywackes.

Common examples of lithic sandstones include the Paleozoic sandstone successions of the central Appalachians in the eastern United States (e.g., Ordovician Juniata Formation, Mississippian Pocono Formation, Pennsylvanian Pottsville Formation); many sandstones associated with the Coal Measures throughout the world; many Jurassic and Cretaceous sandstones of the U.S. and Canadian Rocky Mountains and the U.S. West Coast (e.g., Cretaceous Belly River Sandstone of Canada, Jurassic Franciscan Formation of California); and Tertiary sandstones of the Gulf Coast, the West Coast, and the Alps.

Volcaniclastic sandstones are a special kind of lithic arenite composed primarily of volcanic detritus (Fig. 5.7D). Volcaniclastic sandstones may be made up largely of pyroclastic materials that have been transported and reworked, or they may contain volcanic detritus derived by weathering of older volcanic rocks. They are especially characterized by the presence of euhedral feldspars, pumice fragments, glass shards, and volcanic rock fragments, and they generally have a very low quartz content (e.g., Boggs, 1992, p. 197–209).

Other Sandstones

The sandstones discussed above are composed of constituents derived primarily by weathering of preexisting rocks or by explosive volcanism. A few less abundant types of "sandstones" are known whose constituents formed largely within the depositional basin by chemical or biochemical processes. These rocks, called hybrid sandstones by some authors, include such uncommon varieties of sandstones as **greensands** (glauconitic sands), phosphatic sandstones, and **calcarenaceous sandstones** (composed of sand-size carbonate grains). These rocks are not true sandstones (siliciclastic rocks) but rather are chemical/biochemical sedimentary rocks

Dr. Zaid A. Malak (Dept. of Geology—Mousl University)
Email (zaid_692006@yahoo.com)