SHALES (MUDROCKS)

Shales are fine-grained, siliciclastic sedimentary rocks, that is, rocks that contain more than 50 percent siliciclastic grain less than 0.062 (1/256) mm. Thus, they are made up dominantly of silt-size (1/16–1/256 mm) and clay-size (<1/256 mm) particles. Shale is an historically accepted class name for this group of rocks (Tourtelot, 1960), equivalent to the class name sandstone, a usage accepted by Potter, Maynard, and Pryor (1980, p. 12–15). These authors use the term shale as the class name for all fine-grained siliciclastic sedimentary rocks, but they divide shales into several kinds, such as mudstones and mudshales, depending upon the percentage of clay-size constituents and the presence or absence of lamination (discussed subsequently under classification).

On the other hand, some authors prefer to use the class name **mudrock**, rather than shale, for all fine-grained rocks (e.g., Blatt, Middleton, and Murray, 1980, p. 382). They divide mudrocks into **shales** (if laminated) or **mudstones** (if nonlaminated). Thus, they restrict the usage of shale to fine-grained rocks, such as those in Figure 5.10A, that display lamination or fissility (the ability to split easily into thin layers). Fine-grained, nonlaminated rocks such as those shown in Figure 5.10B are, according to this usage, mudstones. In this book, I follow the usage of Potter, Maynard, and Pryor and apply the general class name shale to all fine-grained siliciclastic sedimentary rocks. Clearly, however, the usage of shale or mudrock as a class name for fine-grained siliciclastic rocks is a matter of personal preference.

Shales are abundant in sedimentary successions, making up roughly 50 percent of all the sedimentary rocks in the geologic record. Historically, shales have been an understudied group of rocks, mainly because their fine grain size makes them difficult to study with an ordinary petrographic microscope. This perspective is changing, however, as instruments are developed such as the scanning electron microscope and electron probe microanalyzer that allow study of fine-size grains at high magnification (e.g., Fig. 5.11).

Composition

Mineralogy

Shales are composed primarily of clay minerals and fine-size quartz and feldspars (Table 5.5). They also contain various amounts of other minerals, including carbonate minerals (calcite, dolomite, siderite), sulfides (pyrite, marcasite), iron oxides (goethite), and heavy minerals, as well as a small amount of organic carbon. Figure 5.11 is a high-magnification photograph, taken by use of backscattered scanning electron microscopy, which allows both the mineralogy and texture of this laminated shale to be examined. [See Krinsley et al., 1998, for discussion of the use of backscattered electron microscopy in the study of sedimentary rocks.] The

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data in Table 5.5 show mineral composition as a function of age. No discernible trend of mineralogy vs. age is evident from this table, except possibly a slight trend of decreasing feldspar with increasing age. Many factors affect the composition of shales, including tectonic setting and provenance (source), depositional environments, grain size, and burial diagenesis. Some minerals, such as carbonate minerals and sulfides, form in the shales during burial as cements or replacement minerals. Quartz, feldspars, and clay minerals are mainly detrital (terrigenous)

Table 5.5 Average percent mineral composition of shales of different ages

Age	Number of analyses	Clay minerals	Quartz	Potassium feldspar	Plagioclase feldspar	Calcite	Dolomite	Siderite	Pyrite	Other minerals	Organic carbon
Quaternary	5	29.9	42.3	12.4	—	6.6	2.4	—	5.6	—	0.9
Pliocene	4	56.5	14.6	5.7	11.9	3.2	—	2.9	1.8	<1.0	2.6
Miocene	9	25.3	34.1	7.4	11.7	14.6	1.2	—	1.9	2.4	1.4
Oligocene	4	33.7	53.5	3.0	—	5.5	—	—	—	4.0	0.4
Eocene	11	40.2	34.6	2.0	8.1	3.8	4.6	1.7	1.6	—	3.5
Cretaceous	9	27.4	52.9	3.6	1.6	2.9	7.9	0.1	1.6	—	2.0
Jurassic	10	34.7	21.9	0.6	4.4	14.6	1.6	0.4	10.9	—	10.9
Triassic	9	29.4	45.9	10.7	0.7	3.7	4.1	5.1	—	—	0.3
Permian	1	17.0	28.0	4.0	8.0	—	1.0	—	—	42.0	0.2
Pennsylvanian	7	48.9	32.6	0.8	6.2	1.4	2.1	3.4	3.5	—	1.0
Mississippian	3	57.2	29.1	0.4	2.9	—	—	0.6	5.1	—	4.7
Devonian	22	41.8	47.1	0.6	—	2.0	1.3	0.3	3.3	—	3.7
Ordovician	2	44.9	32.2	<1.0	6.3	9.8	0.5	0.5	3.4	—	1.5
Misc. ages	29	47.8	33.1	<1.0	5.5	5.2	2.3	0.8	3.1	—	4.5

Source: O'Brien, N. R., and R. M. Slatt, 1990, *Argillaceous rock atlas*, Springer-Verlag, New York, Table I, p. 124–125.
Note: Values adjusted to 100% for shales of each age

minerals, although some fraction of these minerals may also form during burial diagenesis. In particular, clay minerals appear to be strongly affected by diagenetic processes. As shown in Table 5.1, the principal clay mineral groups are kaolinite, illite, smectite, and chlorite. The relative proportions of these clay-mineral groups have been reported to change systematically with age (Fig. 5.12). With time, particularly in rocks older than the Mesozoic, the proportion of illite and chlorite increases at the expense of kaolinite and smectite. These trends are attributed to the diagenetic alteration of kaolinite and smectite to form illite and chlorite.

Chemical Composition

The chemical composition of shales is a direct function of their mineral composition. Compositions of some average North American and Russian shales are shown in Table 5.6. SiO_2 is the most abundant chemical constituent in these shales (57–68 percent), followed by Al_2O_3 (16–19 percent). The SiO_2 content of shales is affected by all silicate minerals present but particularly by quartz. Thus, shales tend to contain less SiO_2 than do sandstones, which commonly are enriched in quartz. Al_2O_3 is derived mainly from clay minerals and feldspars. It is more abundant in shales than in sandstones because of the greater clay mineral content of

Table 5.6 Average chemical composition of selected shales reported in the literature

	1	2	3	4	5	6	7	8	9	10	11	12	13
SiO_2	60.65	64.80	59.75	56.78	67.78	64.09	66.90	63.04	62.13	65.47	64.21	64.10	63.31
Al_2O_3	17.53	16.90	17.79	16.89	16.59	16.65	16.67	18.63	18.11	16.11	17.02	17.70	17.22
Fe_2O_3	7.11	—	—	—	—	—	—	--	—	—	—	2.70	0.82
FeO	—	5.66	5.59	6.56	4.11	6.03	5.87	7.66	7.33	5.85	6.71	4.05	5.45
MgO	2.04	2.86	4.02	4.56	3.38	2.54	2.59	2.60	3.57	2.50	2.70	2.65	3.00
CaO	0.52	3.63	6.10	8.91	3.91	5.65	0.53	1.31	2.22	4.10	3.44	1.88	3.52
Na_2O	1.47	1.14	0.72	0.77	0.98	1.27	1.50	1.02	2.68	2.80	1.44	1.91	1.48
K_2O	3.28	3.97	4.82	4.38	2.44	2.73	4.97	4.57	2.92	2.37	3.58	3.60	3.64
TiO_2	0.97	0.70	0.98	0.92	0.70	0.82	0.78	0.94	0.78	0.49	0.72	0.86	0.81
P_2O_5	0.13	0.13	0.12	0.13	0.10	0.12	0.14	0.10	0.17	—	—	—	0.10
MnO	0.10	0.06		0.08		0.07	0.06	0.12	1.10	0.07	0.05	—	0.06

Source:

(1) Moore, 1978 (Pennsylvanian shale, Illinois Basin)
(2) Gromet et al., 1984 (North American shale composite)
(3) Ronov and Migdisov, 1971 (average North American Paleozoic shale)
(4) Ronov and Migdisov, 1971 (average Russian Paleozoic shale)
(5) Ronov and Migdisov, 1971 (average North American Mesozoic shale)
(6) Ronov and Migdisov, 1971 (average Russian Mesozoic shale)
(7) Cameron and Garrels, 1980 (average Canadian Proterozoic shale)
(8) Ronov and Migdisov, 1971 (average Russian Proterozoic shale)
(9) Cameron and Garrels, 1980 (average Canadian Archean shale)
(10) Ronov and Migdisov, 1971 (average Archean shale)
(11) Clarke, 1924 (average shale)
(12) Shaw, 1956 (compilation of 155 analyses of shale)
(13) Average of values in columns 1 through 12

shales. Fe in shales is supplied by iron oxide minerals (hematite, goethite), biotite, and a few other minerals such as siderite, ankerite, and smectite clay minerals. K_2O and MgO abundance is related mainly to clay mineral abundance, although some Mg may be supplied by dolomite and K is present in some feldspars. Na abundance is related to the presence of clay minerals (e.g., smectites) and sodium plagioclase. Ca is supplied by calcium-rich plagioclase and carbonate minerals (calcite, dolomite).

Classification

Because special analytical techniques are required to determine the mineral composition of shales, and because such techniques are time consuming and expensive, many geologists do not routinely determine the mineral composition of shales (mudrocks). Therefore, most classifications that have been proposed for shales have not been based on mineral composition, or at least not entirely on mineral composition. These classifications, none of which has been widely accepted, commonly emphasize the relative amounts of silt and clay, the hardness or degree of induration of the rocks, and the presence or absence of fissile lamination. (Fissility is defined as the property of a rock to split easily along thin, closely spaced, approximately parallel layers.) Exceptions to this general practice of classification are the classifications of Picard (1971), which emphasizes mineral composition of the silt-size grains in mudrocks (shales), and the classification of Lewan (1978), which requires semiquantitative X-ray diffraction analysis to determine mineralogy.
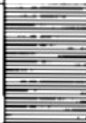
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The classification of Potter, Maynard, and Pryor (1980), shown in Table 5.7, is based on grain size, lamination, and degree of induration. It is similar to the field classification of shales proposed by Lundegard and Samuels (1980). This classification emphasizes the importance of clay-size constituents and bedding thickness, that is, whether bedded or laminated. Depending upon these variables, shales can be divided into **mudstone** (33–65% clay-size constituents and bedded) or **mudshale** (33–65% clay-size constituents and laminated), and **claystone** (66–100% clay-size constituents and bedded) or **clayshale** (66–100% clay-size constituents and laminated). Fine-grained siliciclastic rocks that contain less than 33 percent clay-size constituents are **siltstones**.

Additional informal terms can be used with this classification to provide further information about the properties of the shales. These may include terms that express color, type of cementation (calcareous, or limy; ferruginous, or iron-rich; siliceous); degree of induration (hard, soft); mineralogy if known (e.g., quartzose, feldspathic, micaceous); fossil content (e.g., fossiliferous, foram-rich); organic matter content (e.g., carbonaceous, kerogen-rich, coaly); type of fracturing (conchoidal, hackly, blocky); or nature of bedding (e.g., wavy, lenticular, parallel).

Origin and Occurrence of Shales (Mudrocks)

Shales form under any environmental conditions in which fine sediment is abundant and water energy is sufficiently low to allow settling of suspended fine silt

Table 5.7 Classification of shales and siltstone (>50% grains <0.062 mm)

	Percentage clay-size constituents		0–32	33–65	66–100
	Field adjective		Gritty	Loamy	Fat or slick
NONINDURATED	Beds >10 mm		Bedded silt	Bedded mud	Bedded claymud
NONINDURATED	Laminae <10 mm		Laminated silt	Laminated mud	Laminated claymud
INDURATED	Beds >10 mm		Bedded siltstone	Mudstone	Claystone
INDURATED	Laminae <10 mm		Laminated siltstone	Mudshale	Clayshale
METAMORPHOSED	Degree of metamorphism Low		Quartz Argillite	Argillite	
METAMORPHOSED			Quartz Slate	Slate	
METAMORPHOSED	High		Phyllite and/or Mica Schist		

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and clay. Shales are particularly characteristic of marine environments adjacent to major continents where the seafloor lies below the storm wave base, but they can form also in lakes and quiet-water parts of rivers, and in lagoonal, tidal-flat, and deltaic environments. The fine-grained siliciclastic products of weathering greatly exceed coarser particles; thus, fine sediment is abundant in many sedimentary systems. Because fine sediment is so abundant and can be deposited in a variety of quiet-water environments, shales are by far the most abundant type of sedimentary rock. They make up roughly 50 percent of the total sedimentary rock record. They commonly occur interbedded with sandstones or limestones in units ranging in thickness from a few millimeters to several meters or tens of meters. Nearly pure shale units hundreds of meters thick also occur. Shale units in marine successions tend to be laterally extensive.

A few shales that are particularly well known owing to their thickness, widespread areal extent, stratigraphic position, or fossil content include the Cambrian Burgess Shale of western Canada, which is famous for its well-preserved imprints of soft-bodied animals, the Eocene Green River (oil) Shale of Colorado; the Cretaceous Mancos Shale of western North America, which forms a thick, eastward-thinning wedge stretching from New Mexico to Saskatchewan and Alberta; the Devonian-Mississippian Chattanooga Shale and equivalent formations that cover much of North America and whose widespread extent is still poorly explained; the Silurian Gothlandian shales of western Europe, northern Africa, and the Persian Gulf region, which contain a pelecypod and graptolite faunal association, and the Precambrian Figtree Formation of South Africa, well known for its early fossils. The origin and occurrence of shales are discussed in detail by Potter, Maynard, and Pryor (1980) and Schieber, Zimmerle, and Sethi (1998).

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DIAGENESIS OF SILICICLASTIC SEDIMENTARY ROCKS

Siliciclastic sedimentary rocks form initially as unconsolidated deposits of gravels, sand, or mud. The mineral and chemical compositions of these deposits are functions of a complex system of conditions and processes, including source-rock lithology, sediment transport, and environmental conditions (e.g., Johnsson, 1993). Newly deposited sediments are characterized by loosely packed, uncemented fabrics; high porosities; and high interstitial water content. As sedimentation continues in subsiding basins, older sediments are progressively buried by younger sediments to depths that may reach tens of kilometers. Sediment burial is accompanied by physical and chemical changes that take place in the sediments in response to increase in pressure from the weight of overlying sediment, downward increase in temperature, and changes in pore-water composition. These changes act in concert to bring about compaction and **lithification** of sediment, ultimately converting it into consolidated sedimentary rock. Thus, unconsolidated gravel is eventually lithified to conglomerate, sand is lithified to sandstone, and siliciclastic mud is hardened into shale (mudrock).

The process of lithification is accompanied by physical, mineralogical, and chemical changes. Loose grain packing gives way with burial to more tightly packed fabrics having greatly reduced porosity. Porosity may be further reduced by precipitation of cements into pore spaces. Minerals that were chemically stable at low surface temperatures and in the presence of environmental pore waters become altered at higher burial temperatures and changed pore-water compositions. Minerals may be completely dissolved or may be partially or completely replaced by other minerals.

Thus, porosity, mineralogy, and chemical composition may all be changed to various degrees during burial diagenesis. Diagenesis, the final stage in the process of forming conglomerates, sandstones, and shales, is a process that begins with

weathering of source rocks and continues through sediment transport, deposition, and burial. To properly interpret the provenance, transport, and depositional history of sedimentary rocks, we must recognize and distinguish among features of sediment that were present at the time of deposition and those features of sedimentary rocks that resulted from burial alteration. Diagenesis also has economic significance because it can adversely affect the ability of siliciclastic rocks to store and transmit fluids, a subject of considerable interest to petroleum and groundwater geologists (e.g., Stonecipher, 2000). A short description of diagenetic processes and the physical and chemical effects of these processes is included here.

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Stages and Realms of Diagenesis

Diagenesis takes place at temperatures and pressures higher than those of the weathering environment but below those that produce metamorphism. There is no clear boundary between the realms of diagenesis and metamorphism; however, we commonly consider diagenesis to occur at temperatures below about 250°C (Fig. 5.13). Diagenesis can begin almost immediately after deposition, while sediment is still on the ocean or other basin floor, and may continue through deep burial and eventual uplift. Burial subjects sediments to conditions of pressure and temperature markedly different from those that exist in the depositional environment. Increases in geostatic (rock) pressure, hydrostatic (fluid) pressure, and temperature as a function of depth are shown in Figure 5.13. Pore-fluid composition changes also. There is both a general increase in salinity of pore waters with increasing burial depth and a change in pore-water chemistry (e.g., Heydari, 1997). Changes in pore-water chemistry are difficult to generalize, and they differ from basin to basin, but they include variations in abundance of such important mineral-forming ions as
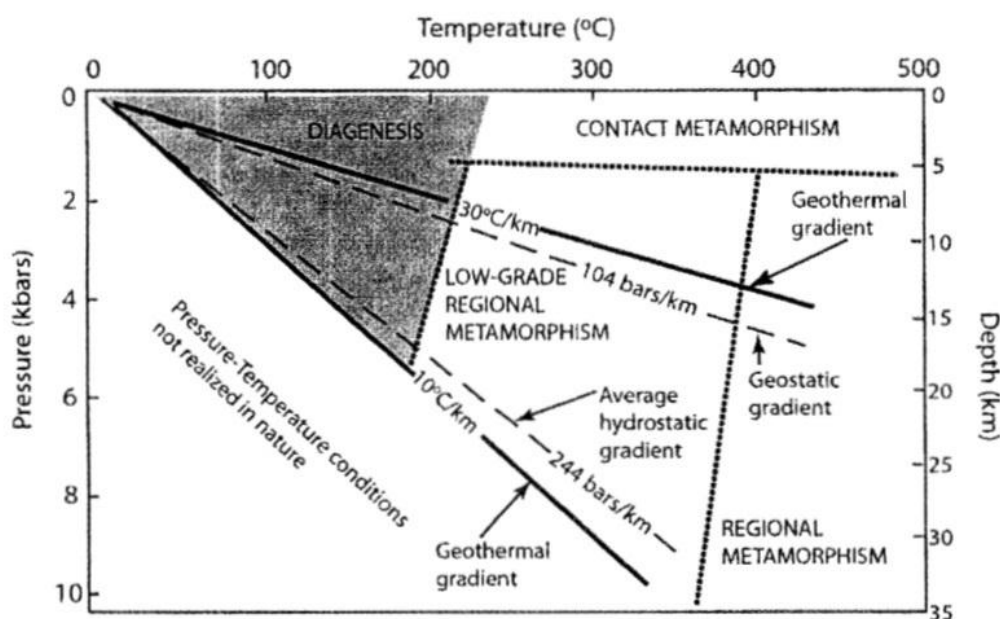
Figure 5.13

Pressure-temperature diagram relating diagenesis to metamorphic regimes and typical pressure-temperature, geostatic, and hydrostatic gradients in Earth's crust. The 10°C/km geothermal gradient is typical of stable cratons; the 30°C/km gradient is typical of rifted sedimentary basins. [Modified from Worden, R. H., and S. D. Burley, Sandstone diagenesis: The evolution of sand to stone, in Burley, S. D., and R. H. Worden, 2003, Sandstone diagenesis: Recent and ancient: Blackwell Pub., Malden, Mass. Fig. 1, p. 3. Reproduced by permission.]

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Si^{4+}, Al^{3+}, Ca^{2+}, K^+, Mg^{2+}, Na^+, and HCO_3^- (bicarbonate). Many of these ions increase in abundance with increasing burial depth, concomitant with increase in salinity. For a recent look at fluids in depositional basins and their role in diagenesis, see Kyser (2000).

Various authors have suggested that sediments go through three to six stages of diagenesis. Perhaps the most widely accepted stages of diagenesis are those proposed by Choquette and Pray (1970). **Eodiagenesis** refers to the earliest stage of diagenesis, which takes place at very shallow depths (a few meters to tens of meters) largely under the conditions of the depositional environment. **Mesodiagenesis** is diagenesis that takes place during deeper burial, under conditions of increasing temperature and pressure and changed pore-water compositions. **Telodiagenesis** refers to late-stage diagenesis that accompanies or follows uplift of previously buried sediments into the regime of meteoric waters. Sedimentary rocks that are still deeply buried in depositional basins have not, of course, undergone telodiagenesis. Some authors now refer to these stages simply as **eogenesis, mesogenesis,** and **telogenesis** (e.g., Worden and Burley, 2003; Fig. 5.14). The most important diagenetic processes that take place in each of these diagenetic regimes, and the effects of these processes, are summarized in Table 5.8. These processes and effects are discussed in greater detail below.

Major Diagenetic Processes and Effects

Shallow Burial (Eogenesis)

The principal diagenetic changes that take place in the eodiagenetic regime include reworking of sediments by organisms (bioturbation), minor compaction and grain repacking, and mineralogical changes. Organisms rework sediment at or near the depositional interface through various crawling, burrowing, and sediment-ingesting activities. Bioturbation can destroy primary sedimentary structures such
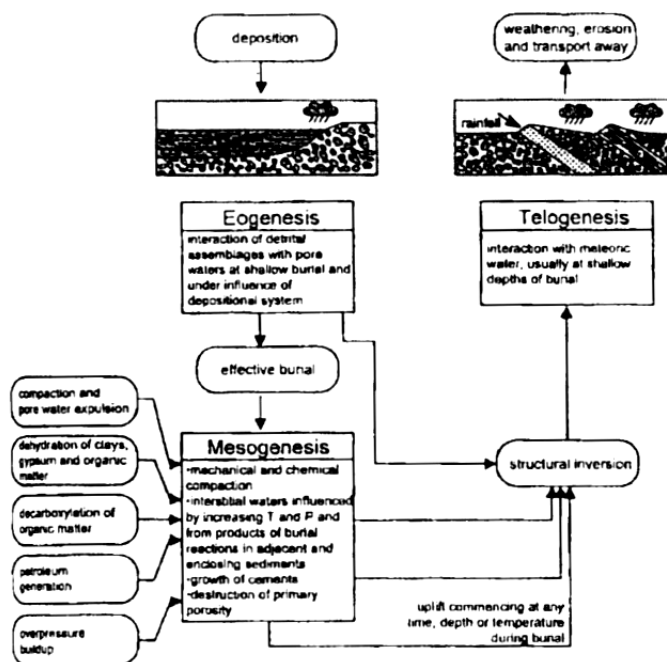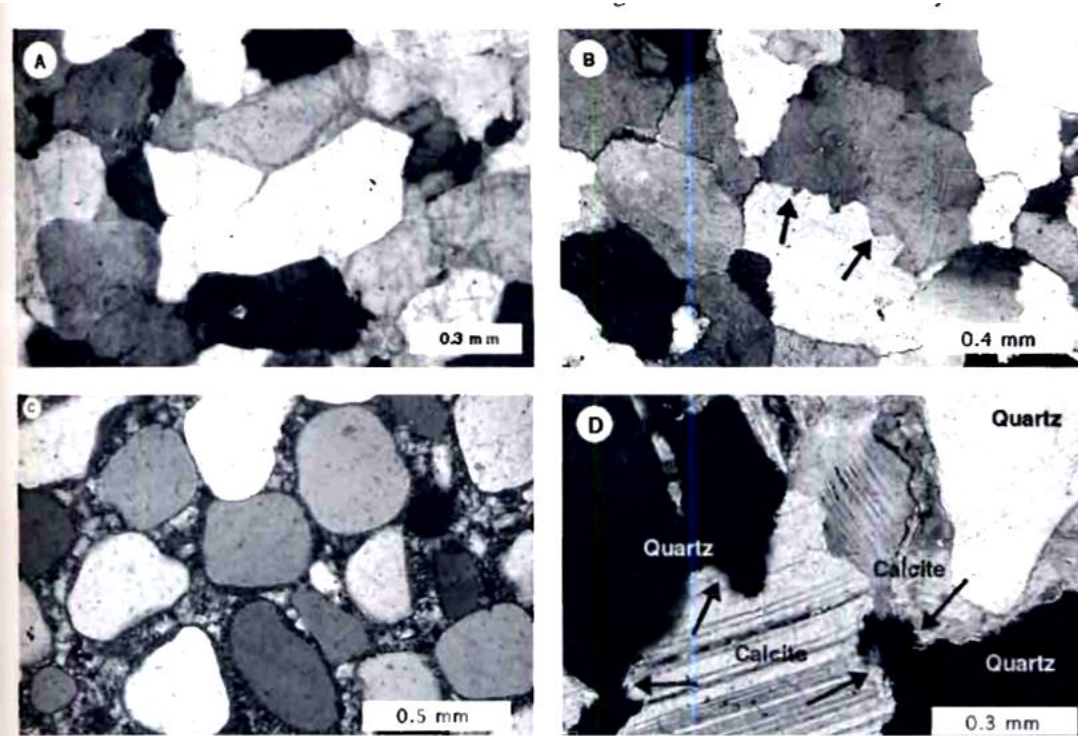

Figure 5.14
Flow chart illustrating the links between the regimes of diagenesis. Structural inversion refers to uplift. [From Worden, R. H., and S. D. Burley, Sandstone diagenesis: The evolution of sand to stone, in Burley, S. D., and R. H. Worden, 2003, Sandstone diagenesis: Recent and ancient: Blackwell Pub., Malden, Mass. Fig. 4, p. 7. Reproduced by permission.]

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Deep Burial (Mesogenesis)

Compaction.   The load pressures caused by deeper burial significantly increase the tightness of grain packing with concomitant loss of porosity (e.g., Fig. 5.15A) and thinning of beds. Increased pressure at the contact point between grains also
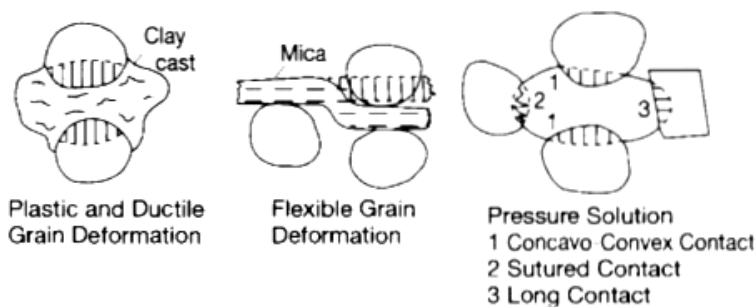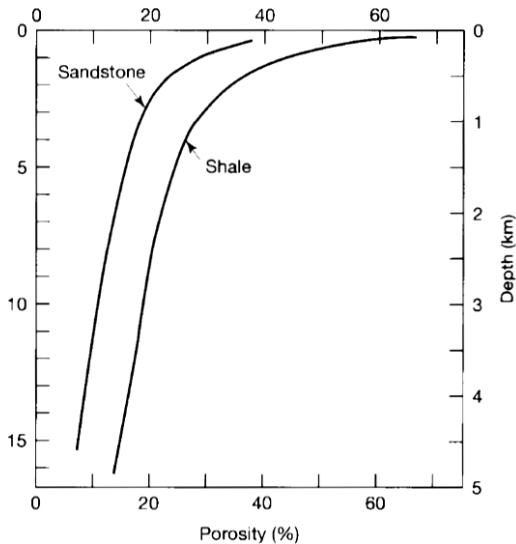

Figure 5.15
Fabrics in sandstones created by diagenetic processes: A. Physical compaction (note prevalence of concavo-convex and long contacts), Tuscarora Sandstone (Silurian), Pennsylvania. B. Chemical compaction owing to pressure solution (note irregular sutured contact indicated by arrows), Oriskany Quartzite (Devonian), Pennsylvania. C. Cementation by microquartz (chert), Jefferson City Fm. (Ordovician), Missouri. D. Replacement of quartz by calcite, creating "nibbled" contacts (arrows), Mauch Chunk Group (Mississippian), Pennsylvania. Crossed nicol photomicrographs.

increases the solubility of the grains at the contact, leading to partial dissolution of the grains. This process is referred to as **pressure solution** or **chemical compaction** (e.g., Fig. 5.15B). Chemical compaction further reduces porosity and increases bed thinning. Thus, under the influence of physical and chemical compaction, aided by cementation (below), the primary porosity of both sands and muds is reduced dramatically during deep burial (Fig. 5.16). Compaction also causes bending of flexible grains such as micas and squeezing of soft grains such as rock fragments (Fig. 5.17). Stone and Siever (1996) report that mechanical compaction and pressure solution cause porosity loss in quartzose sandstones mainly at burial depths less than about 2 km (Fig. 5.18) because the combined effects of compaction, pressure solution, and a small amount of quartz cement produce stable grain-packing arrangements. According to these authors, porosity loss at greater depths is primarily the result of quartz cementation. Worden and Burley (2003) suggest that some porosity loss owing to compaction can continue to depths of at least 5 km.

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Chemical Processes and Changes.   An increase in temperature of 10°C during burial can cause chemical reaction rates to double or triple. Thus, mineral phases that were stable in the depositional environment may become unstable during deep

Figure 5.16
Approximate best-fit curves showing changes in porosity of sediments related to burial compaction and cementation in some California (sandstone) and Louisiana (shale) basins. [Sandstone curve based on Wilson, J. C., and E. F. McBride, 1988, Compaction and porosity evolution of Pliocene sandstones, Ventura Basin, California: Am. Assoc. Petroleum Geologists Bull., v. 72, Fig. 4, p. 669; shale curve based on Dzevanshir, R. D., et al., 1986, Sed. Geology, v. 46, Fig. 1, p. 170.]

Plastic and Ductile
Grain Deformation

Flexible Grain
Deformation

Pressure Solution
1 Concavo Convex Contact
2 Sutured Contact
3 Long Contact

Figure 5.17
Schematic representation of textural criteria used to estimate volume loss in sandstones owing to compaction. The hachured areas indicate rock volume lost by grain deformation and pressure solution. [From Wilson, J. C., and E. F. McBride, 1988, Compaction and porosity evolution of Pliocene sandstones, Ventura Basin, California: Am. Assoc. Petroleum Geologists Bull., v. 72, Fig. 10, p. 679, reprinted by permission of AAPG, Tulsa, Okla.]

burial. Increasing temperature favors the formation of denser, less hydrous minerals and also causes an increase in solubility of most common minerals except the carbonate minerals. Thus, silicate minerals show an increasing tendency to dissolve with greater burial depths (and temperatures), whereas carbonate minerals such as calcite are more likely to precipitate. On the other hand, decrease in pH (increase in acidity) of pore waters with depth may bring about dissolution of carbonates. For example, organic materials may decompose during deep burial diagenesis to release CO_2. Increase in the CO_2 content of pore waters results in a decrease in pH (increase in acidity) that can bring about dissolution of carbonate minerals. As discussed above, increased pressure during deep burial causes an increase in solubility of minerals at point contacts, resulting in partial dissolution of

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the minerals. This process, which releases silica into pore waters, is an important mechanism for furnishing silica that can later precipitate as new silicate minerals. Several kinds of chemical/mineralogical diagenetic processes take place in siliciclastic sedimentary rocks during deep burial. The most important of these processes are cementation, dissolution, replacement, and clay-mineral authigenesis.
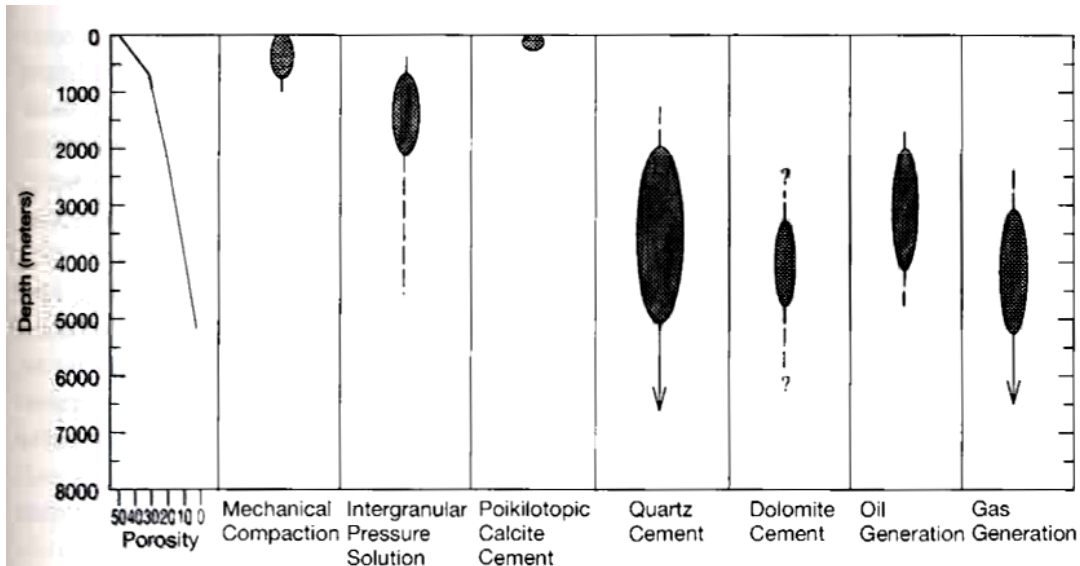
Figure 5.18
Summary diagram showing depth ranges at which mechanical compaction, pressure solution, and cementation reduce porosity in quartzose sandstones. Note that porosity is reduced from approximately 50 percent at the surface to virtually zero at a burial depth of about 5000 m. This diagram also shows the approximate depths at which oil and gas are generated in the subsurface (Chapter 7). [From Stone, W. N., and R. Siever, 1996, Quantifying compaction, pressure solution and quartz cementation in moderately- and deeply-buried quartzose sandstones from the greater Green River Basin, Wyoming, *in* Crossey, L. J., R. Loucks, and M. W. Totten, eds., 1996, Siliciclastic diagenesis and fluid flow: SEPM Special Publication No. 55, Fig. 5, p. 134.]

    Cementation refers to the precipitation of minerals into the pore space of sediment, thereby reducing porosity and bringing about lithification of the sediment. Carbonate and silica cements are most common; however, feldspars, iron oxides, pyrite, anhydrite, zeolites, and many other minerals can also form as cements. As mentioned in the discussion of sandstones, calcite is the dominant carbonate

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cement (e.g., Fig. 5.3C); aragonite, dolomite, siderite, and ankerite are less common. Carbonate cementation is favored by increasing concentration of calcium carbonate in pore waters and increasing burial temperature. Precipitation is inhibited by increased levels of CO_2 in pore waters, which may result from decomposition of organic matter in sediments during burial. Increased CO_2 levels (partial pressure) cause pore waters to become acidic and corrosive to carbonate minerals.

Figure 5.3C shows calcite cement that is restricted to a relatively small area within a sandstone. Cementation can be much more extensive, and cement eventually can fill most of the pore space in the sandstone. In other cases, cement may be concentrated around some object, such as a fossil or fossil fragment, which apparently acts as a nucleus for cementation. Cement can build up around this object to create a globular mass called a **concretion** (Fig. 5.19). In rare cases, calcite, as well as barite and gypsum, can crystallize (precipitate) as large crystals that envelop numerous sand grains, forming so-called **sand crystals** (Fig. 5.20).


Figure 5.19
Large concretions weathering out on the surface of a laminated sandstone bed, Coaledo Formation (Eocene), southern Oregon coast. Calcite, precipitated around some kind of nucleus, filled pore spaces in the sandstone, gradually building up the globular masses. Note the sandstone lamination preserved in the concretions. (Photograph courtesy of Robert Q. Oaks, Jr.)

Figure 5.20
Sand crystals, Miocene sandstone, Badlands, South Dakota. The length of the specimen is about 16 cm. [From Pettijohn, F. J., 1975, Sedimentary rocks, 3rd ed., Harper and Row, Publishers, Inc., New York, Fig. 1.2, p. 467.]

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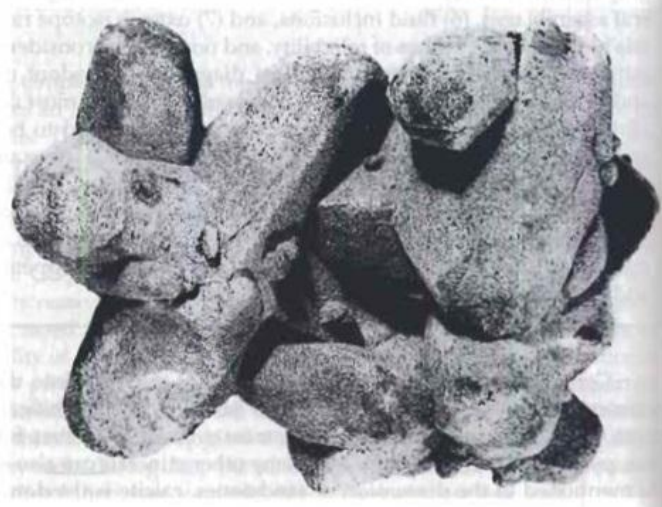
Quartz precipitated as overgrowths around existing detrital quartz grains (e.g., Fig. 5.3A) is the most common kind of silica cement. Quartz overgrowth cements are particularly abundant in many quartz arenites. Less commonly, silica precipitates as microcrystalline quartz (chert) cement (e.g., Fig. 5.15C) or opal. Quartz cementation is favored by high concentrations of silica in pore waters and by low temperatures. Some silica is supplied locally by pressure solution or by dissolution of the siliceous skeletons of fossil organisms such as diatoms and radiolarians. Silica may also be imported from other areas of a basin during episodes of fluid flow related to deep-basin mineral dehydration or tectonic activity (Stone and Siever, 1996). Quartz cementation is particularly likely to occur in sedimentary basins where waters that circulated downward deeply into the basin, and dissolved silica at higher temperatures, rise upward and cool along basin edges.

Dissolution of framework silicate grains and previously formed carbonate cements may occur during deep burial under conditions that are essentially the opposite of those required for cementation. For example, carbonate minerals are dissolved in cooler pore waters with high carbon dioxide partial pressures. Rock fragments and low-stability silicate minerals, such as plagioclase feldspars, pyroxenes, and amphiboles, may dissolve as a result of increasing burial temperatures and the presence of organic acids in pore waters. The selective dissolution of less stable framework grains or parts of grain during diagenesis is called **intrastratal solution**. Dissolution of framework grains and cements increases porosity, particularly in sandstones. Petroleum geologists, who are especially interested in the porosity of sandstones, now believe that much of the porosity that exists in sandstones below a burial depth of about 3 km is secondary porosity, created by dissolution processes.

Mineral **replacement** refers to the process whereby one mineral dissolves and another is precipitated in its place essentially simultaneously. Replacement appears to take place without any volume change between the replaced and replacing mineral. Thus, delicate textures present in the original mineral may, in some cases, be faithfully preserved in the replacement mineral. Well-known examples of such preserved textures can be found in petrified wood and carbonate fossils replaced by chert.

Common replacement events include replacement of carbonate minerals by microcrystalline quartz (chert), replacement of chert by carbonate minerals, replacement of feldspars and quartz by carbonate minerals (e.g., Fig. 5.15D), replacement of feldspars by clay minerals, replacement of clay matrix by carbonate minerals, replacement of calcium-rich plagioclase by sodium-rich plagioclase (albitization), and replacement of feldspars and volcanic rock fragments by clay or zeolite minerals. Replacement may be partial or complete. Complete replacement destroys the identity of the original minerals or rock fragments and thereby gives a biased view of the original mineralogy of a rock. Porosity may also be affected by replacement, particularly replacement of framework grains by clay minerals, which tend to plug pore space and reduce porosity. Much of the clay matrix in sandstones may be produced diagenetically by alteration of unstable framework grains to clay minerals.

In addition to these common replacement processes, one kind of clay mineral may alter to another during diagenesis. For example, smectite clays may alter to illite at temperatures ranging from about 55–200°C, with concomitant release of water. This process is particularly common in shales and is referred to as shale **dewatering**. Smectite may also alter to chlorite within about the same temperature range, and kaolinite typically alters to illite at temperatures between about 120 and 150°C. It is these diagenetic processes that are believed to account for the trend of changing clay-mineral relative abundance with age shown in Fig. 5.12.

Telogenesis

Sedimentary rocks that have undergone deep burial diagenesis may subsequently be uplifted by mountain-building activities and unroofed by erosion. These processes bring mineral assemblages, including new minerals formed during mesogenesis, into an environment of lower temperature and pressure and in which mesogenetic pore waters are flushed and replaced by oxygen-rich, acidic meteoric (rain) waters of low salinity. Under these changed conditions, previously formed cements and framework grains may dissolve (creating secondary porosity) or framework grains may alter to clay minerals, e.g., potassium feldspar to kaolinite (reducing porosity). Alternatively, depending upon the nature of the pore waters, silica or carbonate cements can be precipitated. Other changes may include oxidation of iron carbonate minerals and other iron-bearing minerals to form iron oxides (goethite and hematite), oxidation of sulfides (pyrite) to form sulfate minerals (gypsum) if calcium is present in pore waters, and dissolution of less stable minerals such as pyroxenes and amphiboles. The processes of telogenesis grade into those of subaerial weathering as sedimentary rocks are exposed at Earth's surface.

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