CHAPTER 6

ORIGIN AND OCCURRENCE OF SEDIMENTS AND SEDIMENTARY ROCKS

6.1 FORMATION OF SEDIMENTARY ROCKS
6.2 TYPES OF SEDIMENTARY ROCK
6.3 FEATURES OF SEDIMENTARY ROCKS
6.4 PLATE TECTONICS AND SEDIMENTATION

"Clay and silt, sand and gravel, testimonials to the wreckage of the continents."
Anonymous
OVERVIEW

Most of us have dug our toes into a sandy beach, or picked our way over the gravels of a rushing stream, or perhaps slogged through the mud of a swamp. All of these—sand, gravel, and mud—are products of weathering, the subject of the last chapter, and none of them immediately suggests hard, solid rock. Yet deposits of this sort, or very similar ones, are the stuff from which 75 percent of the rocks exposed at the Earth's surface are made.

Sedimentary rocks form both from deposits of rock fragments and minerals created by weathering—detrital sedimentary rocks—and from chemical deposition—chemical sedimentary rocks. The process by which these materials are laid down is called sedimentation.

In this chapter we examine the mineral composition of the sedimentary deposits as well as their texture. These deposits will undergo changes in a process called diagenesis, and one of these changes is lithification, the process of converting an unconsolidated deposit to a firm rock.

As we seek a method of classification, we will find it useful to use detrital and chemical groups, composition, and texture. Once we have acquired a basis for recognizing the various sedimentary rocks, we will look at some of their special features, including bedding, the single most characteristic feature of the family.

We conclude the chapter with a discussion of geosynclines, and the relation of plate boundaries to sedimentation.

6.1 FORMATION OF SEDIMENTARY ROCKS

We found in Chapter 3 that igneous rocks harden from molten material that originates beneath the surface, under the high temperatures and pressures that prevail there. In contrast, sedimentary rocks form at the much lower temperatures and pressures that prevail at or near the Earth's surface (Figure 6.1).

ORIGIN OF MATERIAL

Sedimentary rocks or metamorphic rocks derived from them make up about 75 percent of the rocks exposed at the surface of the earth (see Box 6.1).

The material from which sedimentary rocks are fashioned originates in two ways. First, the deposits may be accumulations of minerals and rock fragments derived either from the erosion of existing rock or from the weathering products of these rocks. Deposits of this type are called

6.1 Grand Canyon from south rim showing flat-lying sedimentary rocks into which the Colorado River has cut its valley.
The graphs show relative abundance of sedimentary rocks and crystalline (igneous and metamorphic) rocks. (a) The great bulk (95 percent) of the outer 10 km of the Earth is made up of crystalline rocks. Only a small proportion (5 percent) is sedimentary. (b) In contrast, the areal extent of sedimentary rocks at the surface of the Earth's continents is three times that of crystalline rocks.

**Detrital** (from the Latin for "worn down"), and sedimentary rocks formed from them are called detrital sedimentary rocks. Second, the deposits may be produced by chemical processes. We refer to these deposits as nondenrital or chemical deposits and to the rocks formed from them as chemical sedimentary rocks.

Gravel, sand, silt, and clay derived from the weathering and erosion of a land area are examples of detrital sediments. Let us take a specific example. The quartz grains freed by the weathering of a granite may be winnowed out by the flowing water of a stream and swept into the ocean. There they settle out as beds of sand, a detrital deposit. Later, when this deposit is cemented to form a hard rock, we have a sandstone, a detrital rock.

Chemically formed deposits usually develop from the chemical precipitation of material dissolved in water. This process may take place either directly, through inorganic processes, or indirectly, through the intervention of plants and animals. The salt left behind after a salty body of water has evaporated is an example of a deposit laid down by inorganic chemical precipitation. On the other hand, certain organisms, such as the corals, extract calcium carbonate from the seawater and use it to build up skeletons of calcite. When the animals die, their skeletons collect as a biochemical (from the Greek for "life") deposit, and the rock that subsequently forms is called a biochemical rock—in this case, limestone.

Although we distinguish between the two general groups of sedimentary rocks—detrital and chemical—many sedimentary rocks are mixtures of the two. We commonly find that a chemically formed rock contains a certain amount of detrital material. In similar fashion, predominantly detrital rocks include some material that has been chemically deposited.

Each year the streams of the world deliver staggering amounts of material to the oceans. An estimate of this volume is given in Box 6.2.

Geologists use various terms to describe the environment in which a sediment originally accumulated. For example, if a limestone contains fossils of an animal that is known to have lived only in the sea, the rock is known as a marine limestone. Fluvial, from the Latin for "river," is applied to rocks formed by deposits laid down by a river. (Deltaic deposits occur near the mouth of major streams, whose overall shape appears like the Greek letter Δ.) Eolian, derived from Aeolus, the Greek wind god, describes rock made up of wind-deposited material. Rocks formed from lake deposits are termed lacustrine, from the Latin word for "lake." Glacial deposits include both those deposited directly by the ice and those deposited by meltwater from the glaciers (glaciofluvial).

Detrital and chemical, however, are the main divisions of sedimentary rocks based on the origin of material, and, as we shall see later, they form the two major divisions in the classification of sedimentary rocks.
**BOX 6.2 Mass of Material Delivered to Oceans**

We have some estimates of the total mass of sedimentary rocks on the Earth. Probably the best available estimate was made some time ago by Arie Poldervaart of Columbia University, who calculated their weight as $1.702 \times 10^{19}$ t. Of this total, he estimated that $480 \times 10^{13}$ t are presently on the continents and that the rest are in the oceans.

We have seen that sedimentary rocks are formed from the materials weathered from preexisting rocks. Eventually this material reaches the deep ocean basins. What estimate can we make about the amount of material delivered each year to the oceans? The figures presented in this box are order-of-magnitude estimates. Therefore a statement of the approximate amount of material delivered annually to the oceans—where most sedimentary rocks form—is approximately $10^{10}$ t. As these figures suggest, and as we implied in earlier chapters, the great bulk of this material is carried by rivers. That contributed by wind, by ice, or by extraterrestrial sources does not appreciably change the total amount of material deposited in the oceans.

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**Order-of-Magnitude Estimates by Source of Materials Delivered Annually to the Oceans**

<table>
<thead>
<tr>
<th>Source</th>
<th>t/yr</th>
</tr>
</thead>
<tbody>
<tr>
<td>Rivers</td>
<td>10 billion (10^9)</td>
</tr>
<tr>
<td>Glaciers</td>
<td>100 million to 1 billion (10^8 - 10^9)</td>
</tr>
<tr>
<td>Wind</td>
<td>100 million (10^8)</td>
</tr>
<tr>
<td>Extraterrestrial</td>
<td>0.03 to 0.3 (3 x 10^-2 to 3 x 10^-1)</td>
</tr>
</tbody>
</table>

*Estimated from various sources.*

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**SEDIMENTATION**

The general process by which rock-forming material is laid down is called *sedimentation*, or deposition. The factors controlling sedimentation are easy to visualize: To have any deposition at all, there must obviously be something to deposit—which is another way of saying that a source of sediments must exist. We also need some process to transport this sediment. Finally, there must be some place and some process for the deposition of the sedimentary material.

**Methods of Transportation**

Water—in streams and below glaciers, underground, and in ocean currents—is the principal means of transporting material from one place to another. In a stream the coarsest material is carried along the bed of the stream by rolling and sliding; the medium-sized material is carried partially within the flowing water, at times falling to the bottom only to bounce back up into the current; and the finer material is carried suspended within the flowing water. The material dissolved from the weathering of minerals is carried by the stream in solution and ultimately adds to the salinity of the ocean into which it finally empties.

Landslides and other movements induced by gravity also play a role, as do wind and ice. We shall look more closely at these processes in Chapters 12 through 17.

**Processes of Sedimentation**

Detrital material is deposited when its agent of transportation no longer has sufficient energy to move it farther. For example, a stream flowing along at a certain velocity possesses energy to move particles up to a certain maximum size. If the stream loses velocity, it also loses energy, and it is no longer able to transport all the material that it has been carrying at the higher velocity. The solid particles, beginning with the heaviest, start to settle to the bottom. The effect is much the same when a wind that has been driving sand across a desert suddenly dies—a loss of energy accompanies the loss in velocity.

Material that has been carried in solution is deposited in a different way, that is, by precipitation, a chemical process by which dissolved material is converted into a solid and separated from the liquid solvent. As already noted, precipitation may be caused by chemical interaction and evaporation concentration.

Although at first glance the whole process of sedimentation seems quite simple, it is actually as complex as nature itself. Many factors are involved, and they can interact in a variety of ways. Consequently the manner in which sedimentation takes place and the sediments that result from it differ greatly from one situation to another (see Figures 6.2 and 6.3). Think, for instance, of the different ways in which materials settle out of water. A swift, narrow mountain stream may deposit coarse to medium particles along its bed, but farther downstream, as the valley widens, the same stream may overflow its banks and spread fine particles, including mud, over the surrounding country. A lake provides a different environment, varying from the delta of the inflowing stream to the deep lake bottom and the shallow, sandy shore zones. In the oceans, too, environment and sedimentation vary, from the brackish tidal lagoon to the zone of plunging surf and out to the broad, submerged shelves of the continents and to the ocean depths beyond.
MINERAL COMPOSITION OF SEDIMENTARY ROCKS

As we discovered in the chapter on weathering, an igneous rock made up of (1) quartz, (2) feldspar, and (3) ferromagnesian silicates will break down to those specific minerals that are stable at the Earth's surface—namely, quartz, some feldspars, and clay minerals. Detrital sedimentary rocks are accumulations of these minerals combined with precipitated mineral matter that serves to cement the grains together. The grains of many sandstones are predominantly quartz, with some feldspar and other minor accessory minerals. The cementing material may be calcite, dolomite, silica, or iron oxide. Nondetrital sedimentary rocks are dominated by limestones and dolomites. Most sedimentary rocks are mixtures of more than one mineral, although one may dominate. Limestone, for example, is composed mostly of calcite, but even the purest limestone contains small amounts of other minerals, such as clay or quartz.

Clay In an earlier chapter we described how clay minerals develop from the weathering of the silicates, particularly the feldspars. These clays may be subsequently incorporated into sedimentary rocks; they may, for example, form an important constituent of mudstone and shale. Examination of recent and ancient marine deposits shows that the kaolinite and illite clays (illites are increasingly abundant in older rocks) are the most common clays in sedimentary rocks and that the smectite clays are relatively rare.

Quartz An important component of sedimentary rocks is silica, including the very common mineral quartz. The mechanical and chemical weathering of an igneous rock such as granite sets free individual grains of quartz that eventually may be incorporated into sediments. These quartz grains produce the detrital forms of silica and account for most of the volume of the sedimentary rock sandstone. But silica in solution is also produced by the weathering of an igneous rock. This silica may be precipitated or deposited in the form of quartz, particularly as a cementing agent in coarse-grained sedimentary rocks.

Calcite The chief constituent of the sedimentary rock limestone, calcite, CaCO₃, is also the most common cementing material in the coarse-grained sedimentary rocks. The calcium is derived from igneous rocks that contain calcium-bearing minerals, such as calcium plagioclase and some of the ferromagnesian minerals. It also comes from weathering of carbonate sedimentary rocks. Calcium is carried from the zone of weathering as calcium bicarbonate, Ca(HCO₃)₂, and is eventually precipitated as CaCO₃ through the intervention of plants, animals, or inorganic processes. The carbonate is ultimately derived from water and carbon dioxide.

Other Materials in Sedimentary Rocks Accumulations of clay, quartz, and calcite, either alone or in combination, account for all but a very small percentage of the sedimentary rocks, but certain other materials occur in quantities large enough to form distinct strata. The mineral dolomite, CaMg(CO₃)₂, for example, is usually intimately associated with calcite, although it is far less abundant. (Dolomite is named after an eighteenth-century French geologist, Déodat de Dolomieu.) When the mineral is present in large amounts in a rock, the rock itself is also known as dolomite or dolostone. The mineral dolomite is easily confused with calcite, and because they often occur together, distinguishing them is important. Calcite effervesces freely in dilute hydrochloric acid; dolomite effervescences very slowly or not at all unless it is finely ground or powdered. The more rapid chemical activity results from the increase in surface area, an example of the general principle discussed in Section 5.2.
The feldspars and micas are abundant in some sedimentary rocks. We have found that chemical weathering converts these minerals into new minerals at a relatively rapid rate. Therefore when we find mica and feldspar in a sedimentary rock, chances are that it was predominantly mechanical, rather than chemical, weathering that originally made them available for incorporation in the rock.

Iron produced by chemical weathering of the ferromagnesian minerals in igneous rocks may be caught up again in new minerals and be incorporated into sedimentary deposits. The iron-bearing minerals that occur most frequently in sedimentary rocks are hematite, goethite, and limonite. In some deposits these minerals predominate, but more commonly they act simply as coloring matter or as a cementing material.

Halite (NaCl) and gypsum (CaSO₄·2H₂O) are minerals precipitated from solution by evaporation of the water in which they were dissolved. The salinity of the water—that is, the proportion of the dissolved material to the water—determines the type of mineral that will precipitate out. The gypsum begins to separate from seawater when the salinity (at 30°C) reaches a little over three times its normal value. Then, when the salinity of the seawater has increased to about ten times its normal value, halite begins to precipitate.

Pyroclastic rocks, mentioned in Chapter 4, are sedimentary rocks composed mostly of fragments blown from volcanoes. The fragments may be large pieces that have fallen close to the volcano or extremely fine ash that has been carried by the wind and deposited up to many hundreds of kilometers from the volcanic eruption.

Finally, organic matter may be present in sedimentary rocks. In the sedimentary rock known as coal, plant materials are almost the only components. More commonly, however, organic matter is very sparsely disseminated throughout sedimentary deposits and the resulting rocks.

TEXTURE

Texture refers to the size, shape, and arrangement of the particles that make up a rock. There are two major types of texture in sedimentary rocks: clastic (or detrital) and non-clastic (crystalline).

Clastic (or Detrital) Texture  The term clastic is derived from the Greek for “broken” or “fragmental,” and rocks that have been formed from detritus of mineral and rock fragments are said to have clastic or detrital texture. The size and shape of the original particles have a direct influence on the nature of the resulting texture. A rock formed from a bed of gravel and sand has a coarse, rubble-like texture that is very different from the sugary texture of a rock developed from a deposit of rounded, uniform sand grains. Furthermore, the process by which a sediment is deposited also affects the texture of the sedimentary rock that develops from it. Thus the debris dumped by a glacier is composed of a jumbled assortment of rock material ranging from particles of clay size to large boulders. A rock that develops from such a deposit has a very different texture from one that develops from a deposit of windblown sand, for instance, in which all the particles are approximately 0.15 to 0.30 mm in diameter.

Chemical sedimentary rocks may also show a clastic texture. A rock made up predominantly of shell fragments from a biochemical deposit has a clastic texture that is just as recognizable as the texture of a rock formed from sand deposits (see Figure 6.4).

One of the most useful factors in classifying sedimentary rocks is the size of the individual particles. In practice, we usually express the size of a particle in terms of its diameter rather than in terms of its volume, weight, or surface area. When we speak of “diameter,” we may seem to imply that the particle is a sphere; but it is very unlikely that any fragment in a sedimentary rock is a true sphere. In geological measurements the term simply means the diameter that an irregularly shaped particle would have if it were a sphere of equivalent volume. Obviously, it would be a time-consuming, if not impossible, task to determine the volume of each sand grain or pebble in a rock and then to convert these measurements into appropriate diameters. So the diameters we use for particles are commonly determined by such rapid techniques as sieving. We can think of an irregularly shaped grain as having three mutually perpendicular axes or diameters; often these will be a long, an intermediate, and a short diameter. When we examine the “size” of large quantities of sand grains, we commonly pass the samples through nested sets of sieves, with the largest openings

6.4 Large number of oyster, clam, and snail shells make up this Miocene conglomerate of shell beds near Plum Point, Maryland.
in the sieve at the top of the stack, progressing downward to sieves with smaller and smaller openings at the bottom of the stack. The sand grains are bounced back and forth and up and down until they fit through the sieve with openings just a bit larger than the intermediate diameter and come to rest on the next sieve with openings just a bit smaller than that size grain. (Note that such sieve openings measure the intermediate diameter—not the long and not the short diameters!) By weighing the contents of each sieve we can demonstrate the distribution of the sand in various size classes and in this way compare and contrast one sample with others.

Several scales have been proposed to describe particles ranging in size from large boulders to minerals of microscopic dimensions. The Wentworth scale, presented in Table 6.1, is used widely, though not universally, by American and Canadian geologists. Notice that although the term clay is used in the table to designate all particles below 1/4500 (0.004) mm in diameter, the same term is also used to describe certain minerals. To avoid confusion, therefore, we must always refer specifically to either “clay size” or “clay mineral” unless the context makes the meaning clear.

Because determining the size of particles calls for the use of special equipment, the procedure is normally carried out only in the laboratory. In the examination of specimens in the field, therefore, comparisons are made to samples of known size or to scales prepared for that purpose.

Nonclastic Texture Some—but not all—sedimentary rocks formed by chemical processes have nonclastic texture, in which the grains are interlocked. These rocks have somewhat the same appearance as igneous rocks with crystalline texture. Actually, most of the sedimentary rocks with nonclastic texture have crystalline structure, although a few of them, such as opal, do not exhibit this structure.

The mineral crystals that precipitate from an aqueous solution are usually small. Because the fluid in which they form has a very low density, they usually settle out rapidly and accumulate on the bottom as fine sediment. Eventually, under the weight of additional sediments, this material is compacted more and more. Now the size of the individual crystals may begin to increase. Their growth may be induced by added pressure, which causes the favorably oriented grains to grow at the expense of less favorably oriented neighboring grains. Or crystals may grow as more and more mineral matter is added to them from the saturated solutions trapped in the original mud. In any event, the resulting rock is made up of interlocking crystals and has a texture similar to that of crystalline igneous rocks. Depending on the size of the crystals, we refer to these nonclastic textures as fine-grained (or finely crystalline), medium-grained, or coarse-grained. A coarse-grained texture has grains larger than 5 mm in diameter, and a fine-grained texture has grains less than 1 mm in diameter.

Sorting Clastic or detrital rocks commonly have a variety of different-sized materials mixed together. We refer to the range of different sizes that are present as the sorting of the sediment or sedimentary rock. A rock with a narrow range of sizes is called “well-sorted”; one with very wide range of sizes is considered to be poorly sorted. The degree of sorting can have significant ramifications in the search for fluids such as water, gas, and oil. A well-sorted sandstone has little or no fine silt- or clay-sized material filling the pores found among the sand grains, and therefore fluids may fill those pores. A poorly sorted sandstone, on the other hand, has fine-sized particles in those potential voids, resulting in a low porosity and less chance for water, gas, or oil to find pore spaces to occupy.

The degree of sorting varies with the depositional environment. Where currents are strong, for example, finer particles are swept away, leaving only the coarsest clasts, and a well-sorted sediment. In quiet environments only fine particles may be carried into the basin. Again, the sorting could be good. Where different transporting media provide different sizes of clasts, the sorting is poorer. Glacial environments produce poorly sorted sediments, for example. Wind-blown sediments tend to be very well sorted.

<table>
<thead>
<tr>
<th>Wentworth Scale of Particle Sizes for Clastic Sediments*</th>
</tr>
</thead>
<tbody>
<tr>
<td>Size, mm</td>
</tr>
<tr>
<td>---------</td>
</tr>
<tr>
<td>256</td>
</tr>
<tr>
<td>64</td>
</tr>
<tr>
<td>4</td>
</tr>
<tr>
<td>2</td>
</tr>
<tr>
<td>0.06</td>
</tr>
<tr>
<td>0.004</td>
</tr>
</tbody>
</table>


Shape of Particles Another component of texture is the shape of individual particles making up the sedimentary deposit. We refer to the roundness and sphericity of particles when describing their shape. Roundness is the degree to which the edges and corners have been ground off. Sphericity is the degree to which the shape of a fragment approaches the form of a sphere. Notice that a particle can have a high degree of roundness but a very low sphericity if it is bladelike or pencil-shaped, as long as the corners and edges have been rounded. It is somewhat less likely for a pebble or any other particle to be shaped almost like a sphere without having a concurrent high degree of roundness. (Exceptions include equidimensional minerals like garnet, magnetite, and ilmenite, which are very angular but would have a high sphericity.) Particles are shaped by the method and extent of transport to which they have been

6.1 Formation of Sedimentary Rocks 101
subjected. Stream-transported particles of sand size and larger tend to get better rounded with higher sphericity the longer and farther they are transported. Fine windblown silt tends to remain fairly angular even with long-distance transport because of the lack of grain contact, unlike bedload or saltation carpet. Clay-sized particles, because they commonly are clay minerals of sheetlike shapes, do not tend to become either rounded or spherical.

DIAGENESIS

After deposition a sediment or sedimentary rock may undergo a series of physical, chemical, and biologic changes that we call diagenesis. These changes may be in the recombination or rearrangement of a mineral, resulting in the formation of a new mineral. Such changes are specifically limited to those that occur after the sediment has been deposited, during and after its lithification, but before any significant increases in temperature and pressure associated with metamorphism. This process is also limited to changes occurring before the sediment or sedimentary rock has been lifted up to come into contact with the atmosphere. Changes that occur under these conditions have already been discussed in the chapter on weathering. Some examples of diagenesis include dewatering and decrease in porosity associated with compaction of sands, silts, and clay-muds; bacterial decomposition of organic matter; the distortion and destruction of bedding laminations by burrowing organisms; the removal of soluble materials such as calcium carbonate by dissolution in deep-sea sediments; the formation of concretions; the formation of a variety of authigenic minerals (those born or formed in place), such as orthoclase feldspar, illite, and other clay materials, and even quartz. During diagenesis many sedimentary rocks may undergo (1) dissolution, producing voids within the rock; (2) cementation, with the growth of new crystals into void space; or (3) replacement, involving the nearly simultaneous dissolution of existing minerals and the precipitation of new minerals.

LITHIFICATION

The process of lithification converts unconsolidated rock-forming materials into consolidated, coherent rock. The term is derived from the Greek lithos, meaning “rock,” and the Latin facere, “to make.” In the following subsections we shall discuss the various ways in which sedimentary deposits are lithified.

Compaction and Desiccation As sediments accumulate in any basin of deposition (such as the ocean, a lake, a pond, along the floodplain of a river, or on an open plain), they undergo compaction. In compaction the pore space between adjacent grains is gradually reduced by the pressure of overlying sediments or by pressures resulting from Earth movement. Coarse deposits of sand and gravel undergo some compaction, but fine-grained deposits of silt and clay respond much more readily because they contain a large volume of pore fluids. As the individual particles are pressed closer and closer together, the thickness of the deposit is reduced by expulsion of the fluids and its coherence is increased. It has been estimated that deposits of clay-sized particles buried to depths of 1 km have been compacted to about 60 percent of their original volume. This fluid expulsion forces out much of the water, gas, or oil that may have been present.

In desiccation the fluid that originally filled the pore spaces of water-laid clay and silt deposits is forced out (or drawn out by evaporation). Sometimes this is the direct result of compaction, but desiccation also takes place when a deposit is simply exposed to the air and the water evaporates.

Cementation Pressure alone does not produce an indurated rock. Fluids in pores of sediment contain varying amounts of dissolved matter in solutions and some of this material is dissolved by pressure solution at contacts between adjacent minerals. In cementation this dissolved material is precipitated in the spaces between individual particles binding them together. Of the many minerals that serve as cementing agents, the most common are calcite, dolomite, and quartz. Others include iron oxide, opal, chalcedony, anhydrite, pyrite, and especially the clay minerals. Apparently the cementing material is carried in solution by water that percolates through the open spaces among the particles of the deposit. Then some factor in the new environment causes the mineral to be deposited, and the former unconsolidated deposit is cemented into a sedimentary rock.

In coarse-grained deposits there are relatively large interconnecting spaces among the particles. As we should expect, these deposits are very susceptible to cementation because the percolating water can move through them with great ease. Deposits of sand and gravel are transformed by cementation into the sedimentary rocks sandstone and conglomerate.

Crystallization The crystallization of certain chemical deposits is in itself a form of lithification. Crystallization also serves to harden deposits that have been laid down by mechanical processes of sedimentation. For example, new minerals may crystallize within a deposit, or the crystals of existing minerals may increase in size. New minerals are sometimes produced by chemical reactions among amorphous, colloidal materials in fine-grained muds. Exactly how and when these reactions occur is not yet generally understood, but that new crystals have formed after the deposit was initially laid down becomes increasingly apparent as we make more and more detailed studies of sedimentary rocks. Furthermore, it seems clear that this crystallization promotes the process of lithification, particularly in the finer sediments.
6.2 TYPES OF SEDIMENTARY ROCK

CLASSIFICATION

Having examined some of the factors involved in the formation of sedimentary rocks, we are in a position to consider a classification for this rock family. The classification presented in Table 6.2 represents only one of several possible schemes, but it will serve our purposes very adequately. Notice there are two main groups—detrital and chemical—based on the origin of the sediment, and that the chemical category is further split into inorganic and biochemical. All the detrital rocks have clastic texture, whereas the chemical rocks have either clastic or nonclastic texture. We use particle size to subdivide the detrital rocks, and composition to subdivide the chemical rocks.

![Image of rounded fragments in consolidated conglomerate.](image)

6.5 Semirounded fragments in consolidated “conglomerate”; Precambrian from Great Slave Lake district of Canada.

6.6 When the fragments in a conglomerate are more angular than rounded, the rock is called a breccia. When the rock is made up of volcanic materials, as is this one in Iceland, it is called a volcanic breccia. [Roland Hellmann.]

**Conglomerate** A conglomerate, referred to by the common name *puddingstone*, is a detrital rock made up of more or less rounded fragments (Figure 6.5), an appreciable percentage of which are of gravel size (greater than 2 mm). If the fragments are more angular than rounded, the rock is called a *breccia* (Figure 6.6). An example of a conglomerate is *tillite*, a rock formed by the lithification of deposits laid down directly by glacier ice (see Chapter 15). The large particles in a conglomerate are usually rock fragments, and the finer particles are usually minerals derived from the weathering and erosion of preexisting rocks.

**Sandstone** A sandstone is formed by the consolidation of grains of sand size between $\frac{1}{64}$ (0.06) and 2 mm in diame-

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**TABLE 6.2**

<table>
<thead>
<tr>
<th>Origin</th>
<th>Texture</th>
<th>Particle size or composition</th>
<th>Rock name</th>
</tr>
</thead>
<tbody>
<tr>
<td>Detrital</td>
<td>Clastic</td>
<td>Granule or larger</td>
<td>Conglomerate (round grains) breccia (angular grains)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Sand</td>
<td>Sandstone</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Silt</td>
<td>Siltstone</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Clay</td>
<td>Mudstone and shale</td>
</tr>
<tr>
<td>Chemical:</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Inorganic</td>
<td>Clastic or nonclastic</td>
<td>Calcite, CaCO₃</td>
<td>Limestone</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Dolomite, CaMg(CO₃)₂</td>
<td>Dolostone</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Halite, NaCl</td>
<td>Salt</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Gypsum, CaSO₄·2H₂O</td>
<td>Gypsum</td>
</tr>
<tr>
<td></td>
<td></td>
<td>CaCO₃ shells</td>
<td>Limestone, chalk, coquina</td>
</tr>
<tr>
<td></td>
<td></td>
<td>SiO₂ diatoms</td>
<td>Diatomite</td>
</tr>
<tr>
<td></td>
<td></td>
<td>Plant remains</td>
<td>Coal</td>
</tr>
<tr>
<td>Biochemical</td>
<td>Clastic or nonclastic</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

6.2 Types of Sedimentary Rock 103
ter. Sandstone is thus intermediate between coarse-grained conglomerate and fine-grained mudrocks. The size of the grains varies sufficiently from one sandstone to another to allow us to speak of coarse-grained, medium-grained, and fine-grained sandstone.

Commonly the grains of a sandstone are almost all quartz. When this is the case, the rock is called a quartz arenite (from the Latin for “sand”). The name quartzose sandstone is also used. If the minerals are predominantly quartz and feldspar, the sandstone is called an arkose, a French word for the rock formed by the consolidation of debris derived from the mechanical weathering of a granite. Another variety of sandstone, called lithic sandstone or graywacke, is characterized by dark color and by angular grains of quartz, feldspar, and small fragments of rock set in a matrix of clay-sized particles. Arkosic and lithic sandstones result from more rapid sedimentation and burial than quartz-rich sandstones. These “immature” sediments have not normally been transported any great distance and hence have not been subjected to long periods of weathering, which would have decomposed the feldspars and lithic fragments.

Siltstone Particles smaller than \( \frac{1}{64} \) (.06) mm but larger than \( \frac{1}{256} \) (.004) mm are termed silt; when lithified, they form siltstones. The grains may be of any composition, but commonly consist of quartz and feldspar.

Mudrocks Fine-grained detrital rocks composed of clay-sized particles (less than 0.004 mm in diameter) are termed mudrocks, or pelites. Mudstones are fine-grained rocks with a massive or blocky aspect; shales are fine-grained rocks that split into small pieces more or less parallel to the bedding; the term shale is sometimes used for any very fine grained detrital sedimentary rocks. The particles in these rocks are so small that it is difficult to examine them visually; to determine the precise mineral composition, we need to use X-ray diffraction and related techniques. We do know that they contain not only clay minerals but also clay-sized particles of quartz, feldspar, calcite, and dolomite, to mention but a few.

### CHEMICAL SEDIMENTARY ROCKS OF INORGANIC ORIGIN

**Limestone** Limestone is a sedimentary rock that is made up chiefly of the mineral calcite that has been deposited by either inorganic or organic chemical processes. Many limestones have a clastic texture, but nonclastic, particularly crystalline, textures are common.

Inorganically formed limestone is made up of calcite that has been precipitated from solution by inorganic processes. Some calcite is precipitated from the fresh water of streams, springs, and caves, although the total amount of rock formed in this way is negligible. When calcium-bearing rocks undergo chemical weathering, calcium bicarbonate, \( \text{Ca} \left( \text{HCO}_3 \right)_2 \), is produced in solution. If enough of the water evaporates, or if the temperature rises, or if the \( \text{CO}_2 \) pressure falls, or if the water is agitated, calcite is precipitated from this solution. For example, most dripstone, or travertine, is formed in caves by the evaporation of water that is carrying calcium bicarbonate in solution. And tufa (from the Italian for “soft rock”) is a spongy, porous limestone formed by the precipitation of calcite from the water of streams and springs (see Figure 6.7).

Although geologists understand the inorganic processes by which limestone is formed by precipitation from fresh water, they are not quite sure how important these processes are in precipitation from seawater. Some observers have questioned whether they operate at all. On the floors of modern oceans and in rocks formed in ancient oceans, however, we find small spheroidal grains called

![Figure 6.7](image-url)
oolites, the size of sand and often composed of calcite; these grains are thought to be formed by the inorganic precipitation of calcium carbonate from seawater. (The term comes from the Greek for "egg" because an accumulation of oolites resembles a cluster of fish roe.) Cross sections show that many oolites, though not all, have grown around a mineral grain or around a small fragment of shell that acts as a nucleus. Some limestones are made up largely of oolites. One, widely used for building, is the so-called Indiana or Spergen Limestone.

A much larger percentage of limestone is of biochemical origin and will be discussed under that heading below.

Dolostone In discussing the mineral dolomite, CaMg(CO$_3$)$_2$, we mentioned that when it occurs in large concentrations, it forms a rock that is called dolostone, or dolomite. Extensive deposits of dolostone, the origin of which is not fully understood, appear to have been formed by replacement of preexisting deposits of calcite. There is now increasing agreement that most dolostone is in some way related to the local increase of the amount of magnesium in solution. Previously deposited calcite is modified by the movement through it of these magnesium-rich solutions. Field and laboratory observations show that in shallow-water intertidal zones evaporation of seawater may cause precipitation of calcium-bearing deposits. The waters may be increased by an order of magnitude in their content of magnesium relative to calcite. Such high-magnesium-content waters may then circulate through underlying calcite deposits, replacing some of the calcium with magnesium and thus converting limestone to dolostone. In supratidal zones, just above high tide, thin crusts of primary dolomite crystals form in tropical and subtropical regions today. Such dolostones have been found in the Bahamas and in the Florida Keys.

Evaporites An evaporite is a sedimentary rock composed of minerals that were precipitated from solution with the evaporation of the liquid in which they were dissolved. Rock salt (composed of the mineral halite) and gypsum are the most abundant evaporites. Anhydrite (from the Greek anhydros, "waterless") is an evaporite composed of the mineral of the same name, which is simply gypsum without its crystalization water, CaSO$_4$. Most evaporite deposits seem to have been precipitated from seawater according to a definite sequence. The less highly soluble minerals are the first to drop out of solution. Thus gypsum and anhydrite, both less soluble than halite, are deposited first. Then, as evaporation progresses, the more soluble halite is precipitated.

In the United States the most extensive deposits of evaporites are found in Texas, Louisiana, and New Mexico. Here gypsum, anhydrite, and rock salt make up over 90 percent of the Castile Formation, which has a maximum thickness of nearly 1,200 m. In central New York State there are thick deposits of rock salt; and in central Michigan there are layers of rock salt and gypsum. Some evaporite deposits are mined for their mineral content, and in certain areas, particularly in the Gulf Coast states, deposits of rock salt have pushed upward toward the surface to form salt domes producing commercially important reservoirs of petroleum (see Chapter 18).

Silica Silica occurs in sedimentary rocks in a variety of forms other than the typical clastic grains of quartz. It is a chemical constituent of many rocks, precipitated as opal, a hydrous silica, SiO$_2$·nH$_2$O. Opal is slightly softer than common quartz and has no specific crystal structure.

Silica also occurs in sedimentary rocks in a form called cryptocrystalline. This term (from the Greek kryptos, "hidden," and "crystalline") indicates crystalline structure so fine that it cannot be seen under most ordinary microscopes. The microscope does reveal, however, that some cryptocrystalline silica has a granular pattern and that some has a fibrous pattern. To the naked eye the surface of the granular form is somewhat duller than that of the fibrous form. Among the dull-surfaced, or granular, varieties is flint, usually dark in color. Flint is commonly found in certain limestone beds—the chalk beds of southern England, for example. Chert is the general term for these rocks. Jasper is a red variety of granular cryptocrystalline silica. The general term chalcedony is often applied to the fibrous types of cryptocrystalline silica, which have a higher, more waxy luster than the granular varieties. Sometimes the term is also used to describe a specific variety of brown translucent cryptocrystalline silica. Agate is a variegated form of silica, its bands of chalcedony alternating with bands of either opal or some variety of granular cryptocrystalline silica, such as jasper (Figure 6.8).

6.8 Agate showing banded character of rock. Maximum diameter is 15 cm. (John Simpson.)

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BIOCHEMICAL ROCKS

A significant quantity of sedimentary rocks is the direct result of the accumulation of shells, plant fragments, and other remains of organisms. Together these are called biochemical rocks. For example, diatoms are microscopic, single-celled plants that grow in marine or fresh water and secrete siliceous shells. When these shells accumulate in great numbers on the bottom of a basin, they may ultimately form a sedimentary rock called diatomite.

Limestone Biochemically formed limestones are created by the action of plants and animals that extract calcium carbonate from the water in which they live. The calcium carbonate may be either incorporated into the skeleton of the organism or precipitated directly. In any event, when the organism dies, it leaves behind a quantity of calcium carbonate, and over a long period of time thick deposits of this material may be built up. Reefs, ancient and modern, are well-known examples of such accumulations. The most important builders of modern reefs are algae, mollusks, corals, and one-celled animals, the same animals whose ancestors built up the reefs of ancient seas—the reefs, now old and deeply buried, that are often valuable reservoirs of petroleum.

There are virtual limestone “factories” in tropical regions like the Bahamas, the Florida Keys and Bay, Shark Bay in Western Australia, parts of the East and West Indies—all shallow-water continental shelf regions dominated by lime muds and lacking in terrigenous (land-derived) detritus. All these shelves have in common a warm climate and warm-surface ocean water that appears to be supersaturated with CaCO₃. In these regions organisms live and thrive, especially those that are particularly efficient in fixing large amounts of CaCO₃ in shells or other skeletal parts.

Carbonate-utilizing algae grow in abundance on shallow shelves, contributing enormous quantities of sand- and silt-sized carbonate sediments when they die and their stalks and branching arms disaggregate. It has been estimated that several centimeters of carbonate sediments can build up over a few hundred years. No wonder such regions have been called carbonate factories.

Deep-sea oozes commonly contain microscopic calcareous shells of pelagic organisms (free-swimmers and floaters). These calcareous oozes occur over wide regions of the oceans, except in the North Pacific, Arctic, and Antarctic. They accumulate slowly in water with maximum depths of 3 to 5 km. In deeper, colder waters these carbonate shells dissolve and are not found as deposits on the sea floor. (The depth at which carbonates disappear from bottom sediments is called the carbonate compensation depth.)

Freshwater carbonates occur in lakes of temperate regions. Here again, organisms play a role in the accumulation, concentration, and deposition of calcium carbonate. Freshwater limestones commonly have fairly large admixtures of fine clastic detritus, causing them to be calcareous muds (marls) or muddy limestones.

Chalk is made up in part of biochemically derived calcite in the form of the skeletal fragments of microscopic oceanic plants and animals. These organic remains are found mixed with very fine-grained calcite deposits of either biochemical or inorganic chemical origin. A much coarser type of limestone composed of organic remains is known as coquina (from the Spanish for “shellfish” or “ockle”) and is characterized by the accumulation of many large fragments of shells.

Coal Coal is a rock composed of combustible matter derived from the partial decomposition of plants. We shall consider coal as a biochemically formed sedimentary rock, although some geologists prefer to think of it as a metamorphic rock because it passes through various stages.

The process of coal formation begins with an accumulation of plant remains in a swamp. This accumulation is known as peat, a soft, spongy, brownish deposit in which plant structures are easily recognizable. Time, coupled with the pressure produced by deep burial and sometimes by Earth movement, gradually transforms the organic matter into coal. During this process the percentage of carbon increases as the volatile hydrocarbons and water are forced out of the deposit. Coals are ranked according to the percentage of carbon they contain. Peat, with the least amount of carbon, is the lowest ranking; then come lignite, or brown coal; bituminous, or soft coal; and finally anthracite, or hard coal, which has the highest percentage of carbon of all the coals (see Section 18.2).

RELATIVE ABUNDANCE OF SEDIMENTARY ROCKS

Sandstone, mudstone and shale, and limestone constitute about 99 percent of all sedimentary rocks. Of these, mudstone and shale are the most abundant. On the basis of extrapolation from measurements made in the field, the estimates of the percentages of mudstone and shale approximate 50 percent of all sedimentary rocks. Similar calculations for limestone and sandstone suggest that the limestone forms about 22 percent of these rock types and that sandstone accounts for the remaining 28 percent. These percentages, however, do not agree with theoretical determinations of relative abundances. They are based on the determination of the products to be expected from the weathering of an average igneous rock. If these weathering products are assigned to the three major sedimentary rock types, then we find that shale should be considerably more important volumetrically than it appears to be on the basis of field measurements. On such theoretical grounds, mudstone and shale constitute approximately 75 percent of the three major sedimentary rock types. Sandstone and limestone are approximately of equivalent volume and together constitute the other 25 percent. The discrepancy between the two estimates has not yet been resolved, though it is partly due to the loss of mud to the deep sea, where it is only rarely observed again in the sedimentary rock record.
Estimates have been made of the average chemical composition of the world's sediments. One of these estimates is presented in Table 6.3.

### TABLE 6.3
Average Composition of All Sediments (as Oxides)

<table>
<thead>
<tr>
<th>Oxide</th>
<th>Wt. %</th>
</tr>
</thead>
<tbody>
<tr>
<td>SiO₂</td>
<td>44.5</td>
</tr>
<tr>
<td>TiO₂</td>
<td>0.6</td>
</tr>
<tr>
<td>Al₂O₃</td>
<td>10.9</td>
</tr>
<tr>
<td>Fe₂O₃</td>
<td>4.0</td>
</tr>
<tr>
<td>FeO</td>
<td>0.9</td>
</tr>
<tr>
<td>MnO</td>
<td>0.3</td>
</tr>
<tr>
<td>MgO</td>
<td>2.6</td>
</tr>
<tr>
<td>CaO</td>
<td>19.7</td>
</tr>
<tr>
<td>Na₂O</td>
<td>1.1</td>
</tr>
<tr>
<td>K₂O</td>
<td>1.9</td>
</tr>
<tr>
<td>P₂O₅</td>
<td>0.1</td>
</tr>
<tr>
<td>CO₂</td>
<td>13.4</td>
</tr>
<tr>
<td>Total</td>
<td>100.0</td>
</tr>
</tbody>
</table>


### 6.3 FEATURES OF SEDIMENTARY ROCKS

We have mentioned that the stratification, or bedding, of sedimentary rocks is their single most characteristic feature. Now we shall look more closely at this feature, along with certain other characteristics of sedimentary rocks, including mud cracks and ripple marks, nodules, concretions, geodes, fossils, and color.

**BEDDING**

The beds, or layers, of sedimentary rocks are separated by bedding planes, along which the rocks tend to separate or break (see Figure 6.9). The varying thickness of the layers in a given sedimentary rock reflects the changing conditions that prevailed when each deposit was laid down. In general, each bedding plane marks the termination of one deposit and the beginning of another. Within each bed the texture and composition tend to be approximately uniform. Across bedding planes, however, something changes — the texture decreases from sand size to mud size or the composition goes from quartzose sandstone to limestone, for example.

As an illustration, let us imagine a bay of an ocean into which rivers normally carry fine silt from the nearby land. This silt settles out from the seawater to form a layer or bed of mud. Now heavy rains or melting snows may cause the river suddenly to flood and thereby pick up coarser material, such as sand, from the river bed. This material will be carried along and dumped into the bay. There it settles to the bottom and blankets the silt that was deposited earlier. The plane of contact between the silt and the sand represents a bedding plane. If, later on, the silt and the sand are lithified into siltstone and sandstone, the bedding plane persists in the sedimentary rock. In fact, it marks a plane of weakness along which the rock tends to break.

Bedding planes are usually horizontal, and, when parallel, the bedding is called parallel bedding. Closely spaced parallel bedding planes form laminated bedding. But some beds are laid down at an angle to the horizontal, and such bedding is variously called cross bedding, or false bedding (see Figure 6.10). Such nonhorizontal bedding can occur in several situations. For example, the bedding in sand dunes may have high angles on the leeward side of the dune (see Chapter 17). The deposits laid down at the growing edge of a delta may be inclined from 5° to 30°, and such beds are usually given a special name, foreset beds (see Chapter 13).