Al-Karkh University for Sciences
College of Remote Sensing and Geophysics
Geophysics Department

Lecture ONE

Geologic Time:

Concepts and Principles

by

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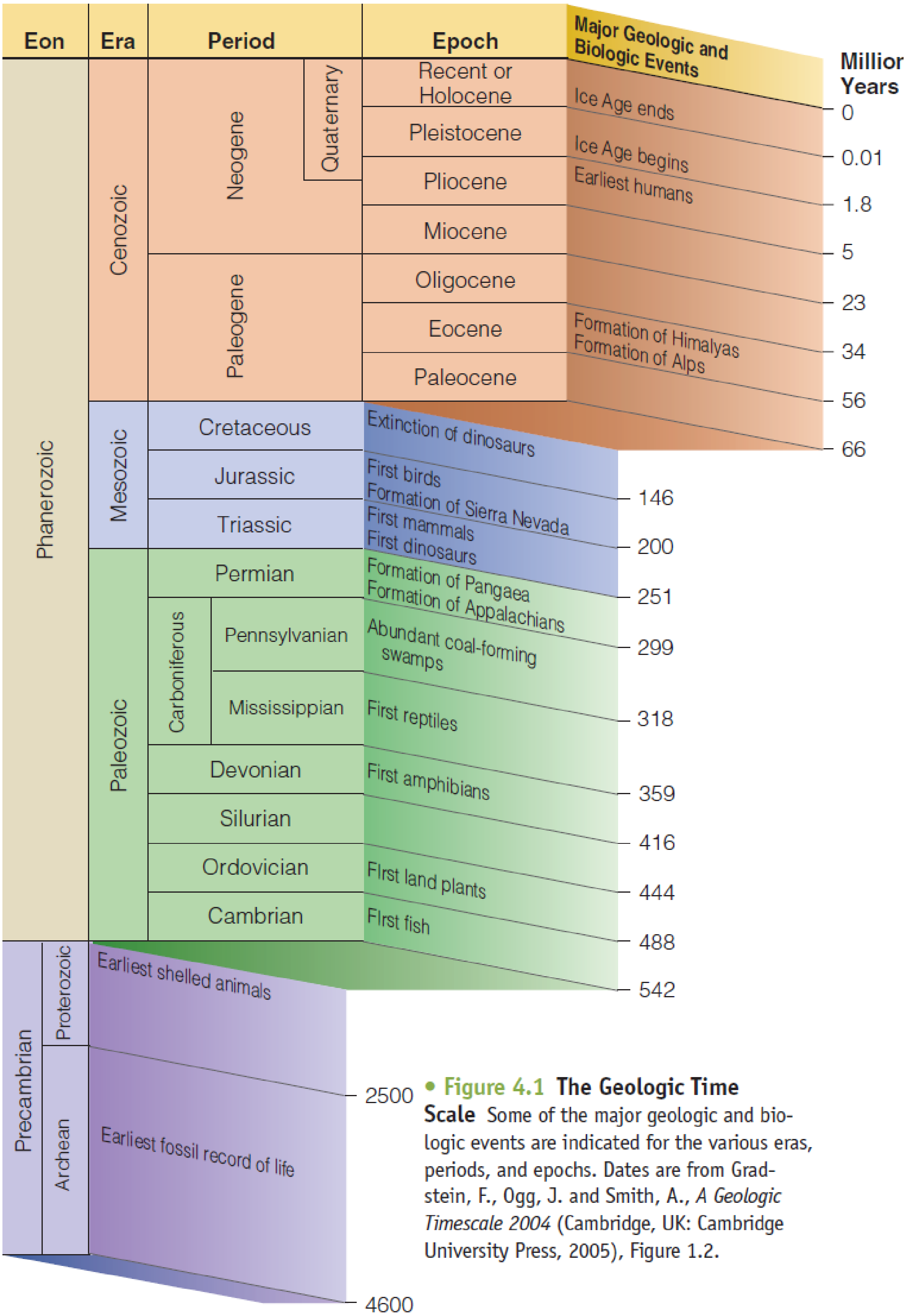
Introduction

We begin this chapter by asking the question "What is time?" We seem obsessed with time, and we organize our lives around it. Yet most of us feel we don't have enough of it—we are always running "behind" or "out of time." Whereas physicists deal with extremely short intervals of time, and geologists deal with incredibly long periods of time, most of us tend to view time from the perspective of our own existence; that is, we partition our lives into seconds, hours, days, weeks, months, and years. Ancient history is what occurred hundreds or even thousands of years ago. Yet when geologists talk of ancient geologic history, they are referring to events that happened millions or even billions of years ago!
Vast periods of time set geology apart from most of the other sciences, and an appreciation of the immensity of geologic time is fundamental to understanding the physical and biological history of our planet. In fact, understanding and accepting the magnitude of geologic time are major contributions geology has made to the sciences.

How is Geologic Time Measured?

In some respects, time is defined by the methods used to measure it. Geologists use two different frames of reference when discussing geologic time.
Relative dating is placing geologic events in a sequential order as determined from their positions in the geologic record. Relative dating will not tell us how long ago a particular event took place; only that one event preceded another. The various principles used to determine relative dating were discovered hundreds of years ago, and since then they have been used to construct the *relative geologic time scale* (Figure 4.1 in Wicander and Monroe, 2010). Furthermore, these principles are still widely used by geologists today, especially in reconstructing the geologic history of the terrestrial planets and their moons.
Absolute dating provides specific dates for rock units or events expressed in years before the present. In our analogy of the television guide, the time when the programs are actually shown would be the absolute dates. In this way, you not only can determine whether you have missed a show (relative dating), but also know how long it will be until a show you want to see will be shown (absolute dating).
Radiometric dating is the most common method of obtaining absolute ages. Dates are calculated from the natural rates of decay of various radioactive elements present in trace amounts in some rocks. It was not until the discovery of radioactivity near the end of the 19th century that absolute ages could be accurately applied to the relative geologic time scale. Today, the geologic time scale is really a dual scale—a relative

scale based on rock sequences with radiometric dates expressed as years before the present (Figure 4.1).



• **Figure 4.1 The Geologic Time Scale** Some of the major geologic and biologic events are indicated for the various eras, periods, and epochs. Dates are from Gradstein, F., Ogg, J. and Smith, A., *A Geologic Timescale 2004* (Cambridge, UK: Cambridge University Press, 2005), Figure 1.2.

Advances and refinements in absolute dating techniques during the 20th century have changed the way we view Earth in terms of when events occurred in the past and the rates of geologic change through time. The ability to accurately determine past climatic changes and their causes has important implications for the current debate on global warming and its effects on humans (see the Epilogue).

Early Concepts of Geologic Time and Earth's Age

During the 18th and 19th centuries, several attempts were made to determine Earth's age on the basis of scientific evidence rather than revelation. The French zoologist Georges Louis de Buff on (1707–1788) assumed ***Earth gradually cooled*** to its present condition from a molten beginning. To simulate this history, he melted iron balls of various diameters and allowed them to cool to the surrounding temperature. By extrapolating their cooling rate to a ball the size of Earth, he determined that Earth was at least 75,000 years old. Although this age was much older than that derived from Scripture, it was still vastly younger than we now know our planet to be.

Other scholars were equally ingenious in attempting to calculate Earth's age. For example, if ***deposition rates could be determined*** for various sediments, geologists reasoned that they could calculate how long it would take to deposit any rock layer. They could then extrapolate how old Earth was from the total thickness of sedimentary rock in its crust. Rates of deposition vary, however, even for the same type of rock. Furthermore, it is impossible to estimate how much rock has been removed by erosion, or how much a rock sequence has been reduced by compaction. As a result of these variables, estimates of Earth's age ranged from younger than 1 million years to older than 2 billion years. Another attempt to determine Earth's age involved ***ocean salinity***. Scholars assumed that Earth's ocean waters were originally fresh and that their present salinity was the result of dissolved salt being carried into the ocean basins by streams. Knowing the volume of ocean water and its salinity, John Joly, a 19th-century Irish geologist, measured the amount of ***salt currently in the world's streams***. He then calculated that it would have taken at least 90 million years for the oceans to reach their present salinity level. This was still much younger than the now accepted age of 4.6 billion years for Earth, mainly because Joly had no way to calculate either how much salt had been recycled or the amount of salt stored in continental salt deposits and seafloor clay deposits.

Relative Dating Methods:

Before the development of radiometric dating techniques, geologists had no reliable means of absolute dating and therefore depended solely on relative dating methods. Relative dating places events in sequential order but does not tell us how long ago an event took place. Although the

principles of relative dating may now seem self-evident, their discovery was an important scientific achievement because they provided geologists with a means to interpret geologic history and develop a relative geologic time scale.

Six fundamental geologic principles are used in relative dating:

- *Superposition.*
- *Original Horizontality.*
- *Lateral Continuity.*
- *Cross-Cutting Relationships.*
- *Inclusions.*
- *Fossil Succession.*

The 17th century was an important time in the development of geology as a science because of the widely circulated writings of the Danish anatomist Nicolas Steno (1638–1686). Steno observed that when streams flood, they spread out across their floodplains and deposit layers of sediment that bury organisms dwelling on the flood plain. Subsequent floods produce new layers of sediments that are deposited or superposed over previous deposits. When lithified, these layers of sediment become sedimentary rock.

Thus, in an undisturbed succession of sedimentary rock layers, the oldest layer is at the bottom and the youngest layer is at the top. This **principle of superposition** is the basis for relative-age determinations of strata and their contained fossils (Figure 4.2a).

• Figure 4.2 **The Principles of Original Horizontality, Superposition, and Lateral Continuity**

a The sedimentary rocks of Bryce Canyon National Park, Utah, illustrate three of the six fundamental principles of relative dating. These rocks were originally deposited horizontally in a variety of continental environments (principle of original horizontality). The oldest rocks are at the bottom of this highly dissected landscape, and the youngest rocks are at the top, forming the rims (principle of superposition). The exposed rock layers extend laterally in all directions for some distance (principle of lateral continuity).

b These shales and limestones of the Postolonnec Formation, at Postolonnec Beach, Crozon Peninsula, France, were originally deposited horizontally but have been significantly tilted since their formation.

Steno also observed that, because sedimentary particles settle from water under the influence of gravity, sediment is deposited in essentially horizontal layers, thus illustrating the **principle of original horizontality** (Figure 4.2a and chapter opening photograph). Therefore, a sequence of sedimentary rock layers that is steeply inclined from the horizontal must have been tilted after deposition and lithification (Figure 4.2b).

Steno's third principle, the **principle of lateral continuity,** states that sediment extends laterally in all directions until it thins and pinches out or terminates against the edge of the depositional basin (Figure 4.2a). The Scottish geologist James Hutton (1726–1797) is considered by many to be the founder of modern geology. His detailed studies and observation of rock exposures and present-day geologic processes were instrumental in establishing the ***principle of uniformitarianism*** (see Chapter 1). Furthermore, Hutton is also credited with discovering the **principle of cross-cutting relationships**, whereby he recognized that an igneous intrusion or a fault must be younger than the rocks it intrudes or displaces (Figure 4.3).

Figure 4.3 The Principle of Cross-Cutting Relationships

a A dark gabbro dike cuts across granite in Acadia National Park, Maine. The dike is younger than the granite it intrudes.

b A small fault (arrows show direction of movement) cuts across, and thus displaces, tilted sedimentary beds along Templin Highway in Castaic, California. The fault is therefore younger than the youngest beds that are displaced.

Establishment of Geology as a Science:

The Triumph of Uniformitarianism over Neptunism and Catastrophism, (for more details see Wicander and Monroe, 2010 p 69).

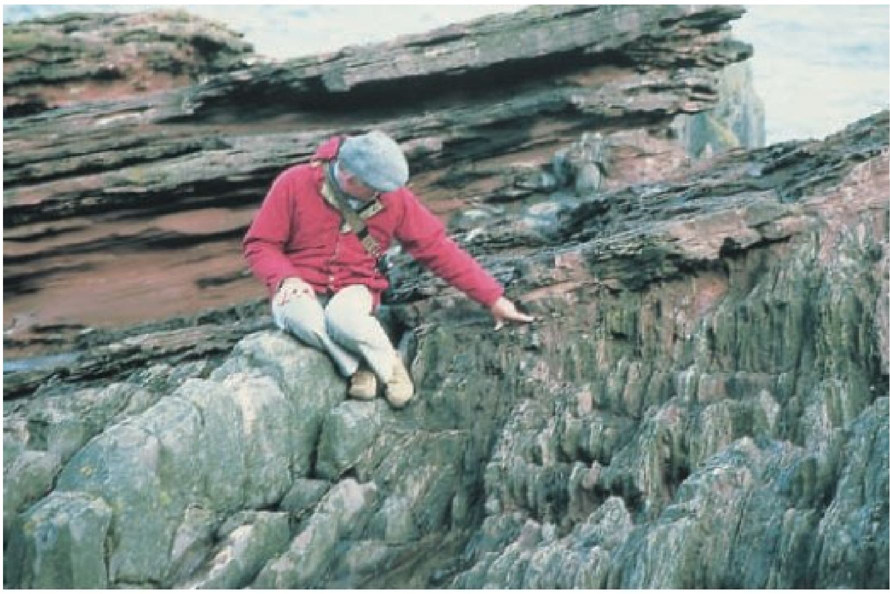
Uniformitarianism: The Scottish geologist James Hutton observed the processes of wave action, erosion by running water, and sediment transport, and concluded that, given enough time, these processes could account for the geologic features in his native Scotland. He reasoned that "the past history of our globe must be explained by what can be seen to be happening now." This assumption that present-day processes have operated throughout geologic time was the basis for the **principle of uniformitarianism.**

Although Hutton developed a comprehensive theory of uniformitarian geology, it was Charles Lyell (1797–1875) who became the principal advocate and interpreter of uniformitarianism. William Whewell, however, coined the term itself, in 1832.

Hutton viewed Earth history as cyclical—that is, continents are worn down by erosion, the eroded sediment is deposited in the sea, and uplift of the seafloor creates new continents, thus completing a cycle. He thought the mechanism for uplift was thermal expansion from Earth's hot interior. Hutton's field observations, and experiments performed by his contemporaries involving the melting of basalt samples, convinced him that igneous rocks were the result of cooling magma. This interpretation of the origin of igneous rocks, called *plutonism,* eventually displaced the neptunian view that igneous rocks precipitated from seawater.

Hutton also recognized the importance of unconformities in his cyclical view of Earth history. He observed steeply inclined metamorphic rocks that had been eroded and covered by flat-lying younger rocks (Figure 4.4). It was clear to him that severe upheavals had tilted the lower rocks and formed mountains. These were then worn away and covered by younger, flat-lying rocks. The erosion surface meant there was a gap in the geologic record, and the rocks above and below this surface provided evidence that both mountain building and erosion had occurred. Although Hutton did not use the word ***unconformity***, he was the first to understand and explain the significance of such gaps in the geologic record. Hutton was also instrumental in establishing the concept that geologic processes had vast amounts of time in which to operate. Because Hutton relied on known processes to account for Earth history, he concluded that Earth must be very old. However, he estimated neither how old Earth was, nor how long it took to complete a cycle of erosion, deposition, and uplift. He merely allowed that "we find no vestige of a beginning, and no prospect of an end," which was in keeping with a cyclical view of Earth history.

Unfortunately, Hutton was not a particularly good writer, and so his ideas were not widely disseminated or accepted. In fact, neptunism and catastrophism continued to be the dominant geologic concepts well into the 1800s. In 1830, however, Charles Lyell published a landmark book, ***Principles of Geology***, in which he championed Hutton's concept of uniformitarianism.



• *Active* Figure 4.4 **Angular Unconformity at Siccar Point, Scotland** It was at this location in 1788 that James Hutton first realized the significance of unconformities in interpreting Earth history. Visit the Geology Resource Center to view this and other active figures at www.cengage.com/sso.

Instead of relying on catastrophic events to explain various Earth features, Lyell recognized that imperceptible changes brought about by present-day processes could, over long periods of time, have tremendous cumulative effects. Not only did Lyell effectively reintroduce and establish the concept of unlimited geologic time, but he also discredited catastrophism as a viable explanation of geologic phenomena. Through his writings, Lyell firmly established uniformitarianism as the guiding principle of geology. Furthermore, the recognition of virtually limitless amounts of time was also necessary for, and instrumental in, the acceptance of Darwin's 1859 theory of evolution (see Chapter 7).

Perhaps because uniformitarianism is such a general concept, scientists have interpreted it in different ways. Lyell's concept of uniformitarianism embodied the idea of a steady-state Earth in which present-day processes have operated at the same rate in the past as they do today. For example, the frequency of earthquakes and volcanic eruptions for any given period

of time in the past must be the same as it is today. Or, if the climate in one part of the world became warmer, another area would have to become cooler so that overall the climate remains the same. By such reasoning, Lyell claimed that conditions for Earth as a whole had been constant and unchanging through time.

Modern View of Uniformitarianism

Geologists today assume that the principles, or laws, of nature are constant but that rates and intensities of change have varied through time. For example, volcanic activity was more intense in North America during the Miocene Epoch than today, whereas glaciation has been more prevalent during the last 1.8 million years than in the previous 300 million years. Because rates and intensities of geologic processes have varied through time, some geologists prefer to use the term *actualism* rather than *uniformitarianism* to remove the idea of "uniformity" from the concept. Most geologists, though, still use the term *uniformitarianism* because it indicates that, even though rates and intensities of change have varied in the past, laws of nature have remained the same.

Uniformitarianism is a powerful concept that allows us, through analogy and inductive reasoning, to use presentday processes as the basis for interpreting the past and for predicting potential future events. It does not eliminate occasional, sudden, short-term events such as volcanic eruptions, earthquakes, floods, or even meteorite impacts as forces that shape our modern world. In fact, some geologists view Earth history as a series of such short-term, or punctuated, events, and this view is certainly in keeping with the modern principle of uniformitarianism.

Earth is in a state of dynamic change and has been since it formed. Although rates of change may have varied in the past, natural laws governing the processes have not.

Absolute Dating Methods

Although most of the isotopes of the 92 naturally occurring elements are stable, some are radioactive and spontaneously decay to other, more stable isotopes of elements, releasing energy in the process. The discovery in 1903 by Pierre and Marie Curie that radioactive decay produces heat meant that geologists finally had a mechanism for explaining Earth's internal heat that did not rely on residual cooling from a molten origin. Furthermore, geologists now had a powerful tool to date geologic events accurately and to verify the long time periods postulated by Hutton and Lyell.

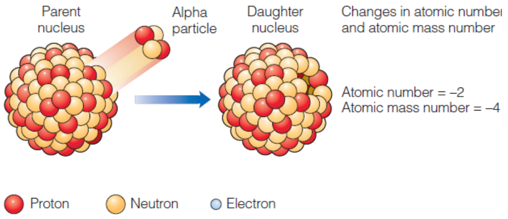
Atoms, Elements, and Isotopes As we discussed in Chapter 2, all matter is made up of chemical *elements*, each composed of extremely small particles called *atoms*. The *nucleus* of an atom is composed of *protons* (positively charged particles) and *neutrons* (neutral particles) with *electrons* (negatively charged particles) encircling it (see Figure 2.2).

The number of protons defines an element's ***atomic number*** and helps determine its properties and characteristics. The combined number of protons and neutrons in an atom is its ***atomic mass number***. However, not all atoms of the same element have the same number of neutrons in their nuclei. These variable forms of the same element are called ***isotopes*** (see Figure 2.3). Most isotopes are stable, but some are unstable and spontaneously decay to a more stable form. It is the decay rate of unstable isotopes that geologists measure to determine the absolute age of rocks.

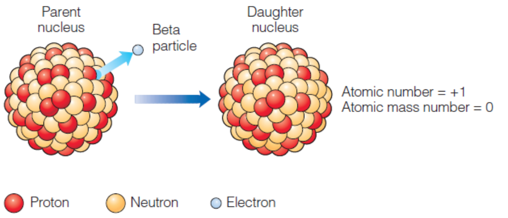
Radioactive Decay and Half-Lives

Radioactive decay is the process whereby an unstable atomic nucleus is spontaneously transformed into an atomic nucleus of a different element. Scientists recognize three types of radioactive decay, all of which result in a change of atomic structure (Figure 4.5). In *alpha decay,* two protons and two neutrons are emitted from the nucleus, resulting in the loss of two atomic numbers and four atomic mass numbers. In *beta decay*, a fast-moving electron is emitted from a neutron in the nucleus, changing that neutron to a proton and consequently increasing the atomic number by one, with no resultant atomic mass number change. *Electron capture decay* is when a proton captures an electron from an electron shell and thereby converts to a neutron, resulting in a loss of one atomic number but not changing the atomic mass number. Some elements undergo only one decay step in the conversion from an unstable form to a stable form. For example, rubidium 87 decays to strontium 87 by a single beta emission, and potassium 40 decays to argon 40 by a single electron capture. Other radioactive elements undergo several decay steps. Uranium 235 decays to lead 207 by seven alpha steps and six beta steps, whereas uranium 238 decays to lead 206 by eight alpha and six beta steps (Figure 4.6). When we discuss decay rates, it is convenient to refer to them in terms of half-lives. The **half-life** of a radioactive element is the time it takes for one-half of the atoms of the original unstable *parent element* to decay to atoms of a new, stable *daughter element*. The half-life of a given radioactive element is constant and can be precisely measured. Half-lives of various radioactive elements range from less than a billionth of a second to 106 billion years.

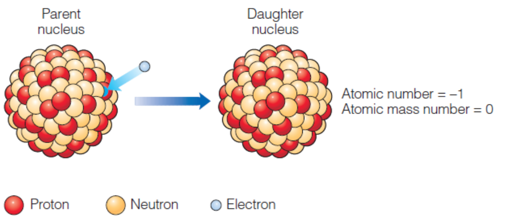
• **Figure 4.5** **Three Types of Radioactive Decay**



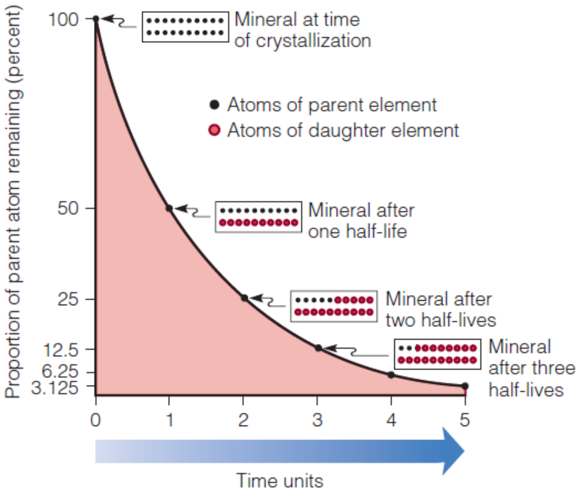
a Alpha decay, in which an unstable parent nucleus emits 2 protons and 2 neutrons.



b Beta decay, in which an electron is emitted from a neutron in the nucleus.

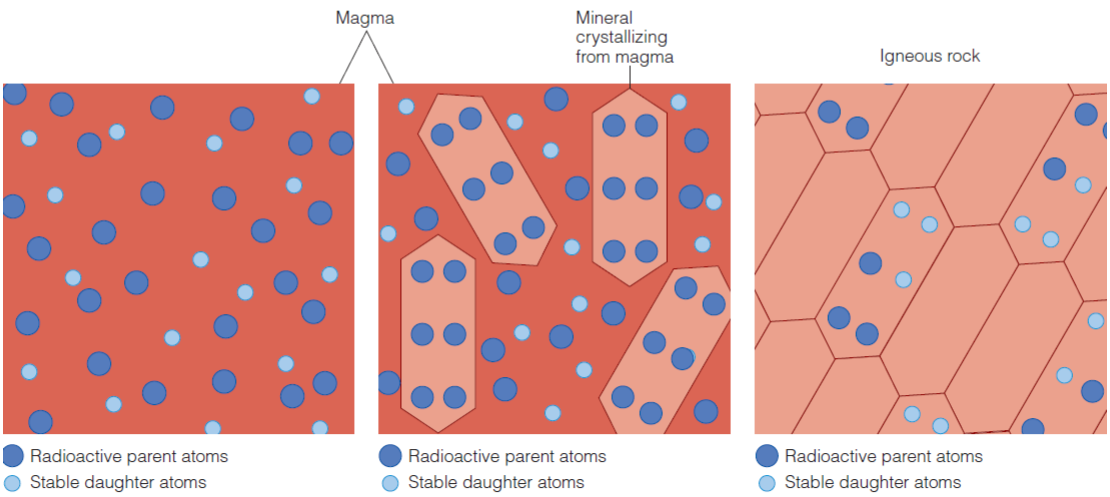


c Electron capture, in which a proton captures an electron and is thereby converted to a neutron.



b A geometric radioactive decay curve, in which each time unit represents one half-life, and each half-life is the time it takes for half of the parent element to decay to the daughter element.

• **Figure 4.8** **Crystallization of Magma Containing Radioactive Parent and Stable Daughter Atoms**



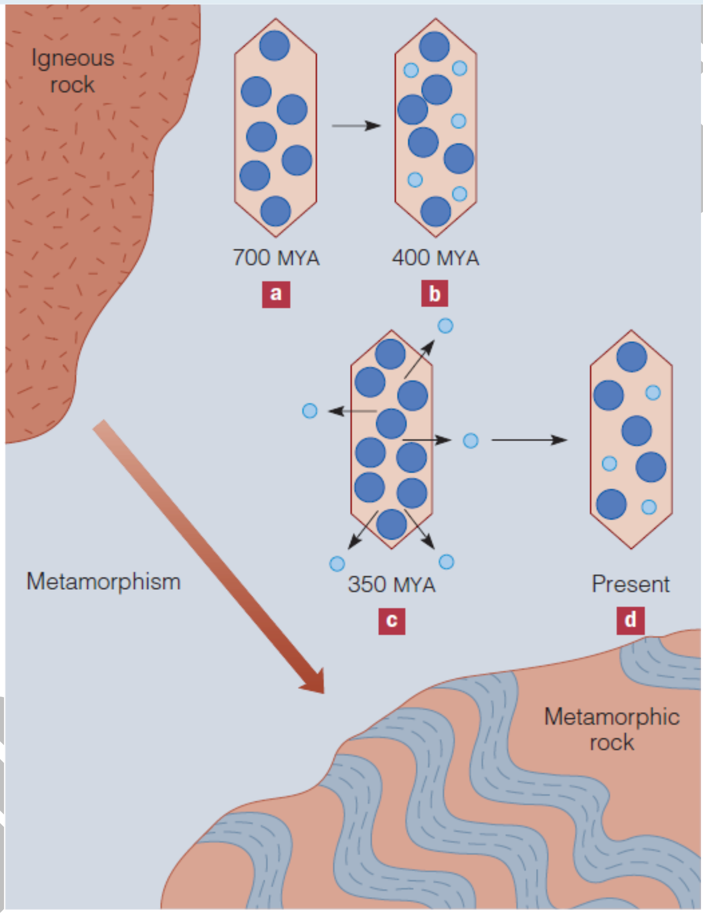
a Magma contains both radioactive parent atoms and stable daughter atoms. The radioactive parent atoms are larger than the stable daughter atoms.

b As magma cools and begins to crystallize, some of the radioactive parent atoms are incorporated into certain minerals because they are the right size and can fit into the crystal structure. In this example, only the larger radioactive parent atoms fit into the crystal structure. Therefore, at the time of crystallization, minerals in which the radioactive parent atoms can fit into the crystal structure will contain 100% radioactive parent atoms and 0% stable daughter atoms.

c After one half-life, 50% of the radioactive parent atoms will have decayed to stable daughter atoms, such that those minerals that had radioactive parent atoms in their crystal structure will now have 50% radioactive parent atoms and 50% stable daughter atoms.

TABLE 4.2 Five of the Principal Long-Lived Radioactive Isotope Pairs Used in Radiometric Dating

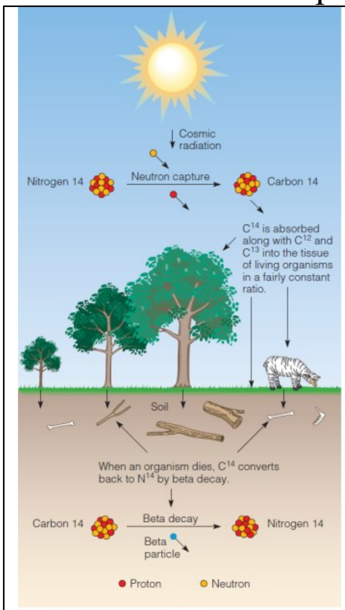
Isotopes		Half-Life of Parent (Years)	Effective Dating Range (Years)	Minerals and Rocks That Can Be Dated
Parent	**Daughter**			
Uranium 238	Lead 206	4.5 billion	10 million to 4.6 billion	Zircon Uraninite
Uranium 235	Lead 207	704 million		
Thorium 232	Lead 208	14 billion		
Rubidium 87	Strontium 87	48.8 billion	10 million to 4.6 billion	Muscovite Biotite Potassium feldspar Whole metamorphic or igneous rock
Potassium 40	Argon 40	1.3 billion	100,000 to 4.6 billion	Glauconite Muscovite Biotite Hornblende Whole volcanic rock



• **Figure 4.9 Effects of Metamorphism on Radiometric Dating** The effect of metamorphism in driving out daughter atoms from a mineral that crystallized 700 million years ago (MYA). The mineral is shown immediately after crystallization a, then at 400 MYA b, when some of the parent atoms had decayed to daughter atoms. Metamorphism at 350 MYA c drives the daughter atoms out of the mineral into the surrounding rock. If the rock has remained a closed chemical system throughout its history, dating the mineral today d, yields the time of metamorphism, whereas dating the whole rock provides the time of its crystallization, 700 MYA.

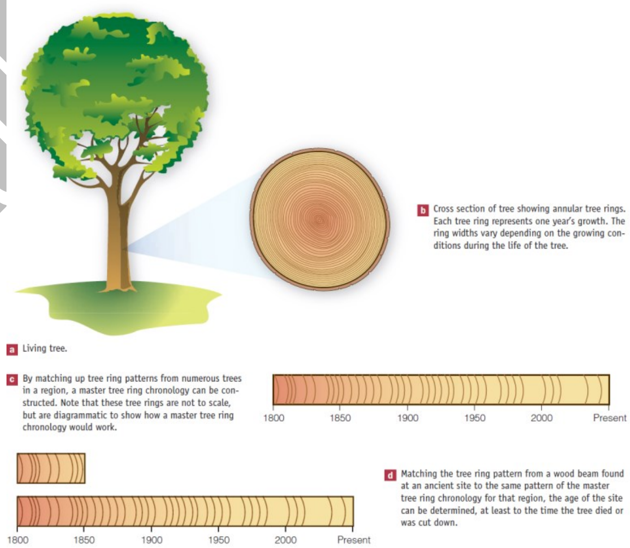
Fission-Track Dating is of particular interest to archaeologists and geologists because the technique can be used to date samples ranging from only a few hundred to hundreds of millions of years old. It is most useful for dating samples between about 40,000 and 1.5 million years ago, a period for which other dating techniques are not always particularly suitable. One of the problems in fission-track dating occurs when the rocks have later been subjected to high temperatures. If this happens, the damaged crystal structures are repaired by annealing, and consequently the tracks disappear. In such instances, the calculated age will be younger than the actual age.

Radiocarbon and Tree-Ring Dating Methods Carbon is an important element in nature and is one of the basic elements found in all forms of life. It has three isotopes; two of these, carbon 12 and 13, are stable, whereas carbon 14 is radioactive (see Figure 2.3). Carbon 14 has a half-life of 5730 years plus/ minus 30 years. The **carbon-14 dating** technique is based on the ratio of carbon 14 to carbon 12 and is generally used to date once-living material. The short half-life of carbon 14 makes this dating technique practical only for specimens younger than about 70,000 years. Consequently, the carbon-14 dating method is especially useful in archaeology and has greatly helped unravel the events of the latter portion of the Pleistocene Epoch.



• *Active* Figure 4.11 **Carbon-14 Dating Method** The carbon cycle involves the formation of carbon 14 in the upper atmosphere, its dispersal and incorporation into the tissue of all living organisms, and its decay back to nitrogen 14 by beta decay. Visit the Geology Resource Center to view this and other active figures at www.cengage.com/sso.

• Figure 4.12 **Tree-Ring Dating Method** In the cross-dating method, tree-ring patterns from different woods are compared to establish a ring-width chronology backward in time.



Reference : Wicander, R., and Monroe, J. S. 2010. Historical geology-Books-Cole.

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Lecture TWO

Rock, Fossils, and Time

Part (1)

by

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Introduction

In their efforts to decipher Earth history, geologists study all of the families of rocks, but they pay particular attention to sedimentary rocks. But how is it possible to interpret events that no one witnessed? Perhaps an analogy will help. Suppose you are walking in a forest and observe a shattered, charred tree. Considering your knowledge of trees and how they grow you immediately know that it did not always exist in its present form. So, perhaps your observations can be explained by an exploding bomb, but you quickly reject this hypothesis because you see no evidence of a bomb crater or shrapnel (metallic fragments) from a bomb. Finally, you conclude that the tree was hit by lightning, a hypothesis that you can check by comparing with trees known to have been hit by lightning.
In our hypothetical example, you did not witness the event but you determined what happened because evidence of the event having occurred was present. It does not matter whether the lightning struck the tree a few days before or many years ago; the only requirement is that the evidence still exists. Likewise, the geologic record provides evidence of past physical and biological events having occurred. For instance, when mud dries it shrinks and cracks and waves in shallow water deform sand into ripples (see Chapter 6), which if found in ancient rocks must have formed by the same processes responsible for them now.

Stratigraphy

The branch of geology called ***stratigraphy*** is concerned with the composition, origin, age relationships, and geographic extent of sedimentary rocks, but the principles of stratigraphy apply to any sequence of stratified rocks. Sedimentary rocks are, with few exceptions, stratified (• Figure 5.1), but volcanic rocks, including lava flows and ash beds, as well as many metamorphic rocks are also stratified. Where sedimentary rocks are well exposed, as in the walls of deep canyons, you can easily determine the vertical relationships among individual layers or strata (singular *stratum*). Lateral relationships are equally important in analysing the geologic record, but they usually must be determined from a number of separate rock exposures, or what geologists call ***outcrops***.

Vertical Stratigraphic Relationships In vertical successions of sedimentary rocks, surfaces known as ***bedding planes*** separate individual strata from one another (Figure 5.1), or the strata grade vertically from one rock type into another. The rocks below and above a bedding plane differ in composition, texture, colour, or a combination of these features, indicating a rapid change in sedimentation or perhaps a period of non-deposition and/or erosion followed by renewed deposition. In contrast, gradually changing conditions of sedimentation are account for those

rocks that show a vertical gradation. Regardless of the nature of the vertical relationships among strata, the correct order in which they were deposited, that is, their relative ages, must be determined.

• Figure 5.1 **Stratified Sedimentary Rocks**

a These rocks in South Dakota have been deeply eroded, but the stratification (layering) is still clearly visible.

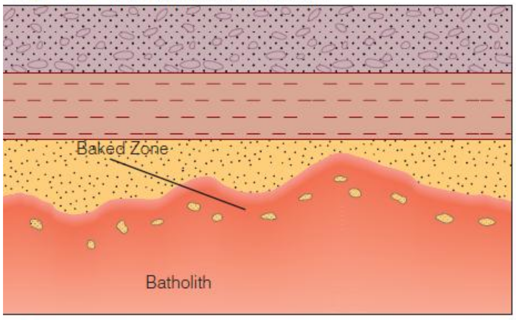
b Stratified rocks in California that have been deformed so that they no longer lie in their original position. It is not obvious from the image but the youngest layer is on the right.

Superposition in Chapter 4, we discussed the *principle of superposition* that resulted from the work of Nicolas Steno during the 17th century. According to this principle, you can determine the correct relative ages of under formed strata by their position in a sequence; the oldest layer is at the bottom of the sequence with successively younger layers upward in the sequence (see Figure 4.2). If strata are deformed by faulting, folding, or both, the task is more difficult (Figure 5.1b), but several sedimentary structures and some fossils allow geologists to resolve these kinds of problems (see Chapter 6).

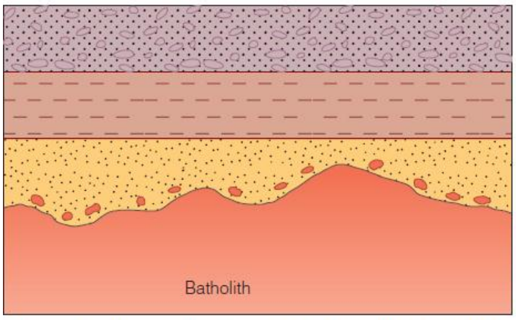
The **principle of inclusions** is yet another way to figure out relative ages, because inclusions, or fragments, in a body of rock must be older than the rock itself. Obviously the sand grains making up sandstone are older than the sandstone, but what about the relative ages of the granite and sandstone in • Figure 5.2? Two interpretations are possible: The granite was intruded into the sandstone and thus is the youngest; the granite was the source of the sand and the sandstone is oldest.

Suppose you encounter a sequence of mostly sedimentary rocks but one layer is made up of basalt (• Figure 5.3). If the basalt is a buried lava flow, you can determine its relative age by the principle of superposition, but if it is a sill-a sheet-like intrusive body—it is younger than the layers below it and younger than the layer immediately above it. Study Figure 5.3 closely and note that the principle of inclusions as well as contact metamorphic effects help resolve this problem.

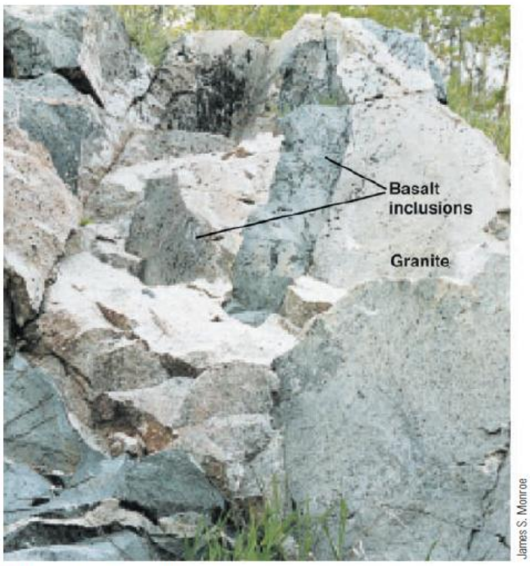
• Figure 5.2 **The Principle of Inclusions**



a The batholith is younger than sandstone because the sandstone has been baked at its contact with the granite and the granite has sandstone inclusions.

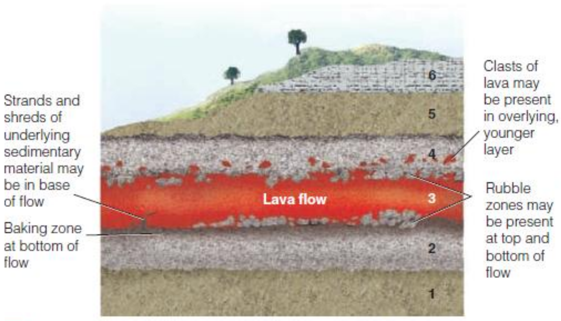


b Granite inclusions in the sandstone indicate that the batholith was the source of the sand and is therefore older.

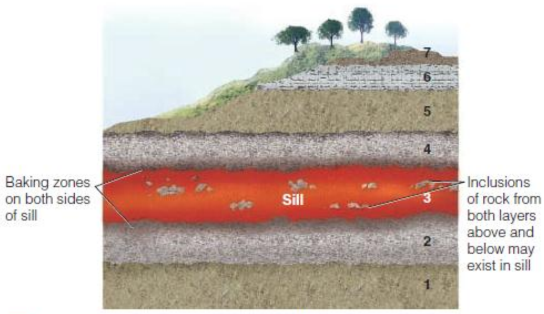


c Basalt inclusions (black) in granite in Wisconsin show that the basalt is oldest.

• Figure 5.3 **How to Determine the Relative Ages of Lava Flows, Sills, and Associated Sedimentary Rocks**



a A buried lava flow has baked underlying bed 2, and clasts of the lava were deposited along with other sediment in bed 4. The lava flow is older than beds 4, 5, and 6.



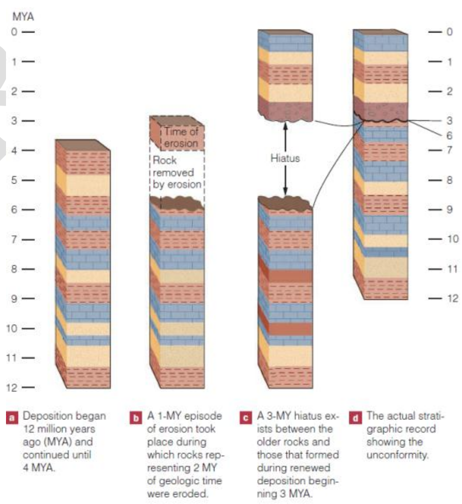
b The layers above and below the sill have been baked, so the sill is younger than layers 2 and 4. Can you determine its age relative to beds 5, 6, and 7?

Unconformities So far, we have discussed vertical relationships among **conformable** strata—that is, sequences of rocks in which deposition was more or less continuous. A bedding plane between strata may represent a depositional break of anywhere from minutes to tens or hundreds of years but is inconsequential in the context of geologic time. However, in many

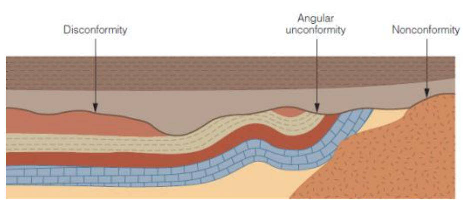
sequences of strata, surfaces known as **unconformities** are present that represent times of nondepositation and/or erosion. These uncomformities encompass long periods of geologic time, perhaps millions or tens of millions of years. Accordingly, the geologic record is incomplete at that particular location, just as a book with missing pages is incomplete, and the interval of geologic time not represented by strata is a *hiatus* (• Figure 5.4). Unconformity is a general term that encompasses three distinct types of surfaces called *disconformity*, *nonconformity*, and *angular unconformity*. Furthermore, unconformities of regional extent may change from one type to another, and they do not necessarily encompass equivalent amounts of geologic time everywhere (Figure 5.5a). Unconformities are common, so the geologic record is incomplete at those locations where unconformities are present. Nevertheless, the geologic time not recorded by rocks in one area is represented by rocks elsewhere. A **disconformity** is an erosion surface in sedimentary rocks that separates younger rocks from older rocks, both of which are parallel to each other (Figure 5.5b).

However, an erosion surface cut into plutonic rocks or metamorphic rocks that is overlain by sedimentary rocks is nonconformity (Figure 5.5c). And finally, an **angular unconformity** is present if the strata below an erosion surface are inclined at some angle to the strata above (Figure 5.5d). In this case, we can infer that the sequence of events included deposition, lithification, deformation, erosion, and, finally, renewed deposition.

• Figure 5.4 The Origin of an Unconformity and a Hiatus



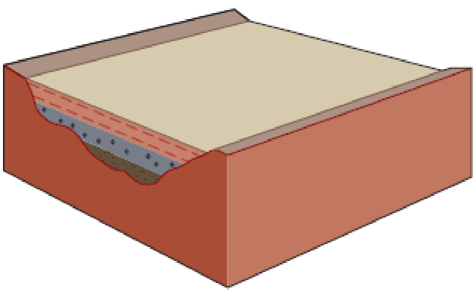
• Figure 5.5 Types of Unconformities



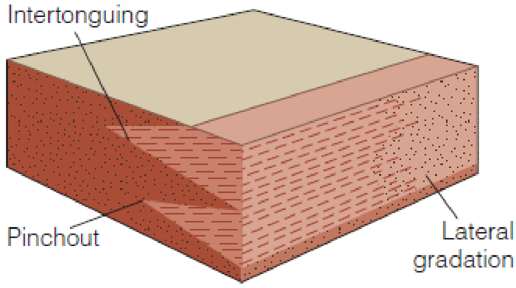
Lateral Relationships—Facies In 1669, Nicolas Steno proposed his principle of lateral continuity, meaning that layers of sediment extend outward in all directions until they terminate (see Chapter 4). Rock layers may terminate abruptly where they abut the edge of a depositional basin, where they are eroded, or where they are cut by faults. A rock unit may also become progressively thinner until it *pinches out*, or it splits laterally into thinner units each of which pinches out—a phenomenon known as *intertonguing*. And finally, a rock unit might change by *lateral gradation* as its composition and/or texture become increasingly different (Figure 5.6). Both intertonguing and lateral gradation indicate the simultaneous operation of different depositional processes in adjacent environments. For example, on the continental shelf sand may accumulate in the high-energy near shore environment, while at the same time mud and carbonate deposition takes place in offshore low-energy environments. Deposition in each of these laterally adjacent environments yields a **sedimentary facies,** a body of sediment with distinctive physical, chemical, and biological attributes.

Armanz Gressly, in 1838, was the first to use the term *facies* when he carefully traced sedimentary rocks in the Jura Mountains of Switzerland and noticed lateral changes such as sandstone grading into shale. He reasoned that these changes indicated deposition in different environments that lie next to one another. Any attribute of sedimentary rocks that makes them recognizably different from laterally adjacent rocks of about the same age is sufficient to establish a sedimentary facies.

• Figure 5.6 Lateral Termination of Rock Layers



a Lateral termination at the edge of a depositional basin.



b Lateral termination by intertonguing, pinchout, and lateral gradation.

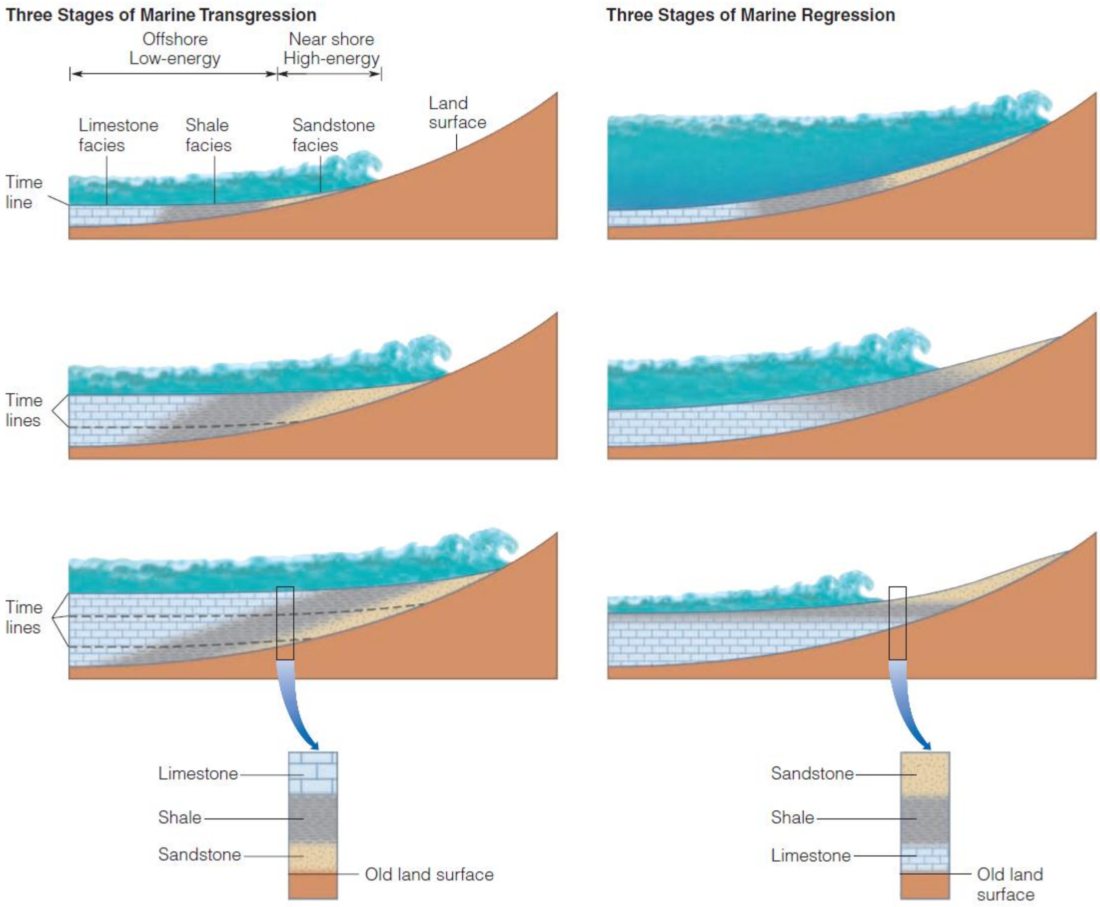
Marine Transgressions and Regressions

Three rock units exposed in the walls of the Grand Canyon of Arizona consist of sandstone followed upward by shale and finally limestone (Figure 5.7). These three facies, all with fossils of marine-dwelling trilobites and brachiopods, were deposited on one another. Superposition tells us their relative ages, but what accounts for their presence in Arizona

far from the sea, and how were these facies deposited in the order observed? These deposits formed during a time when sea level rose with respect to the land, giving rise to a **marine transgression.** During a marine transgression the shoreline migrates landward, as do the environments that parallel the shoreline as the sea progressively covers more and more of a continent (Figure 5.8). Remember that each laterally adjacent depositional environment is an area where a sedimentary facies develops. So during a marine transgression, the facies that formed in off shore environments become superposed on facies deposited in nearshore environments. Another aspect of a marine transgression is that the rocks making up each facies become younger in a landward direction. The sandstone, shale, and limestone facies in Figure 5.8 were deposited continuously as the shoreline moved landward, but none of the facies was deposited simultaneously over its entire geographic extent. In other words, the facies are ***time transgressive,*** meaning that their ages vary from place to place.

Obviously the sea is no longer present in the Grand Canyon area, so the transgression responsible for the rocks in Figure 5.7 must have ended. In fact, it did—during a **marine regression** when sea level fell with respect to the continent and the environments that paralleled the shoreline migrated seaward. In other words, a marine regression is the opposite of a transgression and it yields a vertical sequence with nearshore facies overlying off shore facies (Figure 5.8). In this case individual rock units become younger in a seaward direction. In our discussion of marine transgressions and regressions, we considered both vertical and lateral facies relationships, the significance of which was first recognized by Johannes Walther (1860–1937). When Walther traced rock units laterally, he reasoned, as Gressly had, that each sedimentary facies he encountered was deposited in laterally adjacent environments. In addition, Walther noticed that the same facies he found laterally were also present in a vertical sequence. His observations have since been formulated into **Walther's law,** which holds that the facies seen in a conformable vertical sequence will also replace one another laterally.

The application of Walther's law is well illustrated by the marine transgression and regression shown in Figure 5.8. In practice, it is usually difficult to follow rock units far enough laterally to demonstrate facies changes. It is much easier to observe vertical facies relationships and use Walther's law to work out the lateral relationships. Remember, though, that Walther's law applies only to conformable sequences of rocks; rocks above and below an unconformity are unrelated and Walther's law does not apply.



• Figure 5.8 Marine Transgressions and Regressions

Extent, Rates, and Causes of Marine Transgressions and Regressions

Geologists carefully analyse sedimentary rocks to determine the maximum extent of marine transgressions and the greatest withdrawal of the sea during regressions. Six major marine transgressions followed by regressions have taken place in North America since the Neoproterozoic, yielding unconformity-bounded rock sequences that provide the stratigraphic framework for our discussions of Paleozoic and Mesozoic geologic history. The paleogeographic maps in Chapters 10, 11, 14, and 16 show the extent to which North America was covered by the sea or exposed during these events. Shoreline movements during transgressions and regressions probably amount to no more than a few centimetres per year. Suppose that a shoreline moves landward 1000 km in 20 million years, giving 5 cm/yr as the average rate of transgression. Our average is reasonable, but we must point out that large-scale transgressions are not simply events during which the shoreline steadily moves landward. In

fact, they are characterized by a number of reversals in the overall transgressive trend, thus accounting for the intertonguing we see among some sedimentary rock units (Figure 5.6b).

Geologists agree that uplift and subsidence (downward movement) of the continents, the amount of water frozen in glaciers, and rates of seafloor spreading are responsible for marine transgressions and regressions. During uplift of a continent, the shoreline moves seaward, and just the opposite takes place during subsidence. Widespread glaciers expanded and contracted during the Pennsylvanian Period (see Chapter 11), which caused several sea level changes and resulted in transgressions and regressions. Indeed, if all of Earth's present-day glacial ice were to melt, sea level would rise by about 70 m.

Geologic evidence indicates that sea level may have been as much as 250 m higher during the Cretaceous Period, and as a result, widespread marine transgressions occurred during which large parts of the continents were invaded by the sea (see Figure 14.6). The probable cause of this event was comparatively rapid seafloor spreading during which heat beneath the mid-oceanic ridges caused them to expand and displace water onto the continents. When seafloor spreading slows, the mid-oceanic ridges subside, increasing the volume of the ocean basins, and the seas retreat from the continents.

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Lecture THREE

Rock, Fossils, and Time

Part (2)

by

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Fossils and Fossilization

You now know that superposition, cross-cutting relationships, and original horizontality are essential for interpreting geologic history, but consider the situation in Figure 5.9. For the rocks in either of the two columns, or what geologists call columnar sections, the relative ages are apparent from superposition, but what are the ages of the rocks in one column compared to those in the other column? With the data provided, you cannot resolve this problem. The solution to this problem involves using fossils, if present, and some physical events of short duration such as volcanic ash falls.

Fossils, the remains or traces of prehistoric organisms preserved in rocks, are most common in sedimentary rocks, but they may also be found in volcanic ash and volcanic mudflows. Geologists use fossils extensively to determine the relative ages of strata, but fossils also provide useful information for determining environments of deposition (see Chapter 6), and they constitute some of the evidence for the theory of evolution (see Chapter 7). In short, fossils are essential to fully decipher Earth and life history.

Today, it is apparent that bones, teeth, and shells in rocks are the remains of once-living creatures, yet this view is rather recent. Indeed, during most of historic time in the Western world, people variously believed fossils were inorganic objects formed within rocks by some kind of molding force, or even objects placed in rocks by the Creator to test our faith or by Satan to sow seeds of doubt. Some perceptive observers—such as Leonardo da Vinci in 1508, Robert Hook in 1665, and Nicolas Steno in 1667—recognized the true nature of fossils, but their views were largely ignored. By the 18th and 19th centuries, though, it was apparent that fossils were truly the remains of organisms, and it was also clear that many fossils were of organisms now extinct.

How Do Fossils Form? Our definition of fossil includes the term "remains," or what are called body fossils (Figure 5.10a and b), consisting mostly of skeletal parts such as shells, bones, and teeth. However, under some exceptional circumstances soft parts may be preserved by freezing or mummification. In contrast, **trace fossils** are not actual remains but an indication of organic activity such as tracks, trails, burrows, and nests (Figure 5.10c and d). The trace fossil known as a coprolite is fossilized feces that may provide information about the diet and size of the animal that produced it.

• Figure 5.10 **Body Fossils and Trace Fossils**

a Body fossils are actual remains of organisms, such as these dinosaur bones.

b Shells of invertebrate marine animals such as this ammonite known as *Parapuzosia seppenradensis* are also body fossils. This specimen measures about 1.0 m in diameter.

c These bird tracks are one kind of trace fossil. This slab of rock formed over the actual tracks, so it is a cast of the tracks.

d Fossilized feces (coprolite) of a carnivorous mammal. The specimen is about 5 cm long and it contains small fragments of bones and teeth.

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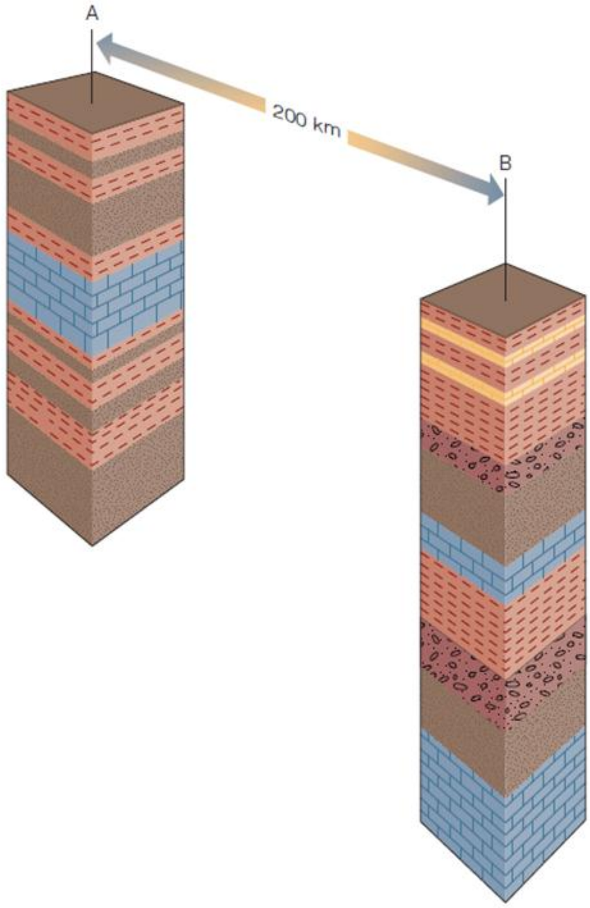

• Figure 5.9 **Relative Ages of Rocks** You can determine the relative ages of rocks in either column A or B, but you cannot tell the relative ages of rocks in A compared to B. The rocks in A may be younger than those in B, the same age as some in B, or older than all of the rocks in B.

The most favorable conditions for preserving body fossils are that an organism has a durable skeleton and lives where rapid burial is likely. Consequently, corals, clams, and brachiopods have good fossil records because they meet these criteria, whereas jellyfish with no skeletons have a poor record. Bats have skeletons but they are delicate and they live where burial is not so likely, so they too have a poor fossil record. Even given these qualifications and the fact that bacterial decay, scavenging, and metamorphism destroy organic remains, fossils of some organisms are common.

For any organism, death is certain but preservation as a fossil is rare. Nevertheless, fossils are more common than most people realize because so many billions of creatures have lived during so many millions of years. Indeed, at many localities you can collect hundreds of fossils of corals and brachiopods, and even the fragmented bones and teeth of dinosaurs are common in some areas. It is true that for fossilization to occur, shells

or bones must be buried rapidly in some protective medium such as sand or mud, but rapidly means only before remains are destroyed by physical and chemical processes, which might take many years in some environments.

Thousands of fossils have been recovered from the La Brea Tar Pits in Los Angeles, California; a bone bed in Canada has the remains of hundreds of horned dinosaurs; and hundreds of rhinoceroses, three-toed horses, saber-toothed deer and other animals are in Miocene-age ash deposits in Nebraska. Such phenomenal concentrations of fossils might seem to be unrelated to any everyday process with which we are familiar, but we know of many instances of the first stages of fossilization taking place now. For example, small animals today are trapped in the sticky reside at oil seeps at the La Brea Tar Pits, and wildebeests by the hundreds drown in river crossings, probably much like the horned dinosaurs in Canada did millions of years ago. Burial in volcanic ash should come as no surprise because that is exactly what has happened during recent eruptions. In short, just as with the record of past events preserved in rocks we use the principle of uniformitarianism to determine how fossils were buried and preserved.

Body fossils may be preserved as unaltered remains (Figure 5.11), meaning they retain their original composition and structure, or as altered remains (Figure 5.12), in which case some change has taken place in their composition and/or structure. (Table 5.1). In addition, molds and casts are common in the fossil record (Figure 5.13). A mold forms when buried organic remains dissolve, leaving a cavity shaped like a clam or bone, for example. If minerals or sediment should later fill the cavity, it forms a cast— that is, a replica of the original.

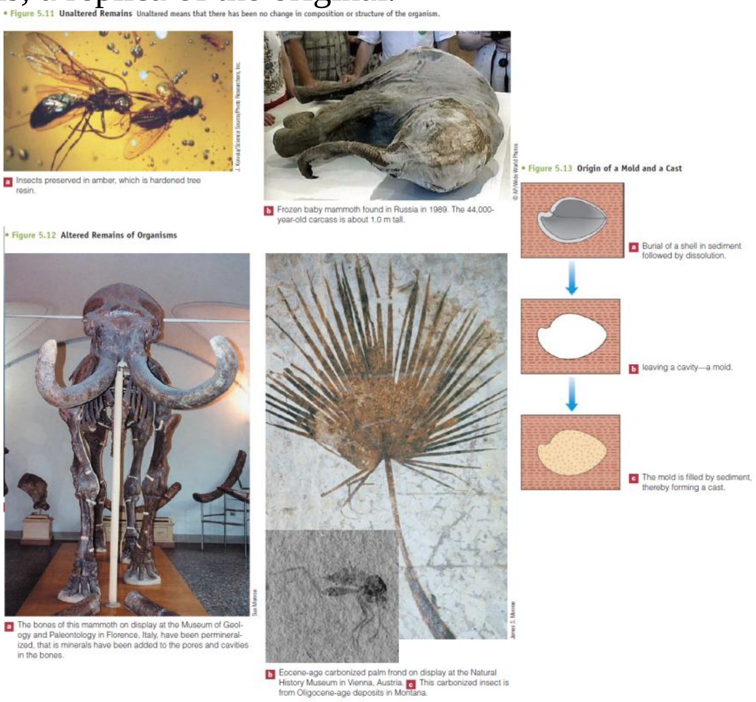

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TABLE 5.1	Types of Fossil Preservation
Body fossils—unaltered remains	Original composition and structure preserved
Freezing	Large Ice Age mammals frozen in sediment
Mummification	Air drying and shriveling of soft tissues
Preservation in amber	Leaves, insects, and small reptiles trapped and preserved in hardened tree resin
Preservation in tar	Bones, insects preserved in asphaltlike substance at oil seeps
Body fossils—altered remains	Change in composition and/or structure of original material
Permineralization	Addition of minerals to pores and cavities in shells and bones
Recrystallization	Change in the crystal structure—for example, aragonite in shells recrystallizes as calcite
Replacement	One chemical compound replaces another—for example, pyrite (FeS_2) replaces calcium carbonate ($CaCO_3$) of shells
Carbonization	Volatile elements lost from organic matter leaving a carbon film; most common for leaves and insects
Trace fossils	Any indication of organic activity such as tracks, trails, burrows, droppings (coprolites), and nests
Molds and casts	Mold—a cavity with the shape of a bone or shell; cast—a mold filled by minerals or sediment

The ***fossil record***, the record of ancient life, just as the geologic record of which it is a part, must be analyzed and interpreted. As a repository of prehistoric organisms, this record provides our only knowledge of such extinct animals as trilobites and dinosaurs. Furthermore, the study of fossils has several practical applications that we will discuss in the following sections and in Chapter 6.

Fossils and Telling Time We began the section on fossils by referring to the columnar sections in Figure 5.9 and asking how you might determine the relative ages of the strata in these geographically separate areas. Our answer was "fossils"—but exactly how do fossils resolve this problem? To fully understand the usefulness of fossils, we must examine the historical development of an important geologic principle.

The use of fossils in relative dating and geologic mapping was demonstrated during the early 1800s in England and France. William Smith (1769–1839), an English civil engineer who was surveying and building canals in southern England, independently discovered Steno's principle of superposition. He reasoned that in a sequence of strata, the oldest is at the bottom and the youngest is at the top, and he came to the same conclusion regarding any fossils the rocks contained. Smith made numerous observations at outcrops, mines, and quarries and discovered that the sequence of fossils, and especially groups of fossils, is consistent from area to area. In short, he discovered that the relative ages of sedimentary rocks at different locations could be determined by their fossil content (• Figure 5.14).

By recognizing the relationship between strata and fossils, Smith could predict the order in which fossils would appear in rocks at some locality

he had not previously visited. In addition, his knowledge of rocks and fossils allowed him to predict the best route for a canal and the best areas for bridge foundations. As a result, his services were in great demand. Smith gets much of the credit for the idea of using fossils to determine relative ages, but other geologists, such as Alexander Brongniart in France, also recognized this relationship. In any case, their observations served as the basis for what we now call the **principle of fossil succession** (also known as the ***principle* or *law of faunal succession***). This important principle holds that fossil assemblages (groups of fossils) succeed one another through time in a regular and determinable order. But why not simply match up similar rock types in Figure 5.9 and conclude they are the same age? Rock type will not work because the same kind of rock— sandstone, for instance—has formed repeatedly through time. Fossils have also formed continuously through time, but because different organisms existed at different times, fossil assemblages are unique. In short, an assemblage of fossils has a distinctive aspect compared with younger or older fossil assemblages. Accordingly, if we match up rocks containing similar fossils, we can assume they are of the same relative age.

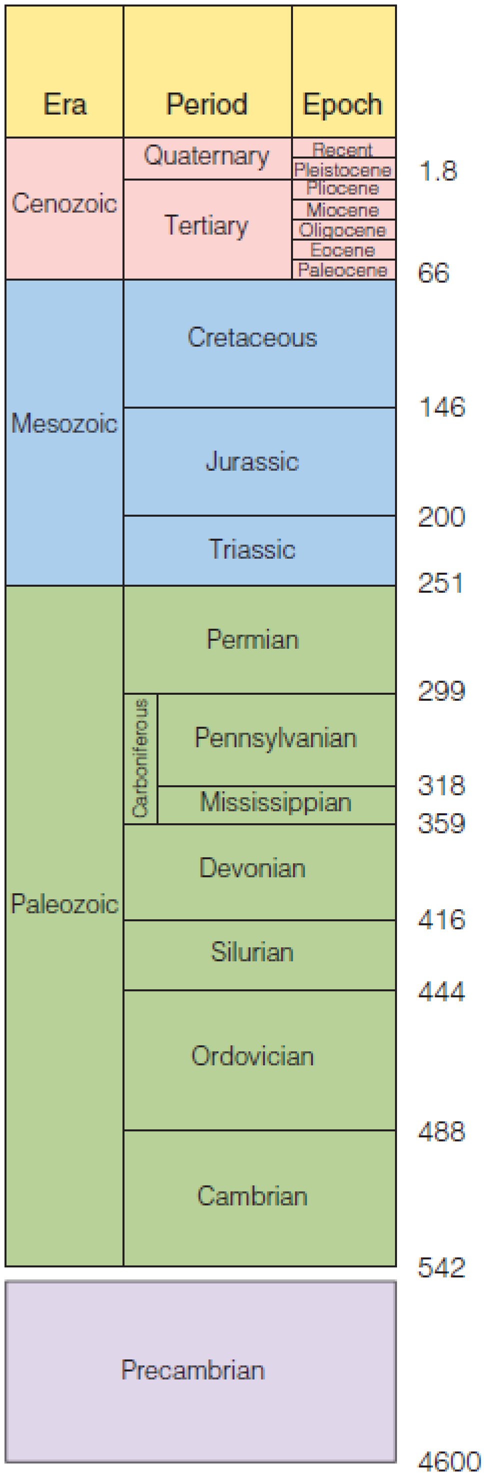
Stratigraphic Terminology

Establishing systems and a relative geologic time scale brought some order to stratigraphy, but problems remained. Because many sedimentary rock bodies are time transgressive (Figure 5.8), they may belong to one system in one area and to another system elsewhere, or they may simply straddle the boundary between systems. Accordingly, geologists now use terminology for two fundamentally different kinds of units: those defined by their content and those expressing or related to geologic time (Table 5.2).

TABLE 5.2 Classification of Stratigraphic Units

Units Defined by Content		Units Expressing or Related to Geologic Time	
Lithostratigraphic Units	Biostratigraphic Units	Time-Stratigraphic Units	Time Units
Supergroup	Biozone	Eonothem	Eon
Group		Erathem	Era
Formation		System	Period
Member		Series	Epoch
Bed		Stage	Age

Lith and *litho* mean “stone” or “stonelike.”
Chron and *chrono* are added to words to indicate time.



Lithostratigraphic Units and Biostratigraphic Units:

Rock type with no consideration of time of origin is the only criterion used to define a **lithostratigraphic1 unit**, which has unifying features that set it apart from other lithostratigraphic units (Table 5.2). The basic lithostratigraphic unit is the **formation,** which is a map-able body of rock with distinctive upper and lower boundaries. Formations may consist of a single rock type (the Redwall Limestone), a variety of related rock types

(the Morrison Formation), and the term *formation* may also apply to metamorphic and igneous rocks (the Lovejoy Basalt). Many formations are subdivided into *members* and *beds*, and they may be parts of more inclusive units such as *groups* and *supergroups* (Figure 5.16, Table 5.2).

Biostratigraphic units, in contrast, are defined solely on the basis of their fossil content with no regard to rock type or time of origin (Table 5.2), and their boundaries do not necessarily correspond with those of lithostratigraphic units. The fundamental biostratigraphic unit is the **biozone,** which is discussed more fully in the following section on correlation.

TABLE 5.2 Classification of Stratigraphic Units

Units Defined by Content		Units Expressing or Related to Geologic Time	
Lithostratigraphic Units	**Biostratigraphic Units**	**Time-Stratigraphic Units**	**Time Units**
Supergroup	Biozone	Eonothem	Eon
Group		Erathem	Era
Formation		System	Period
Member		Series	Epoch
Bed		Stage	Age

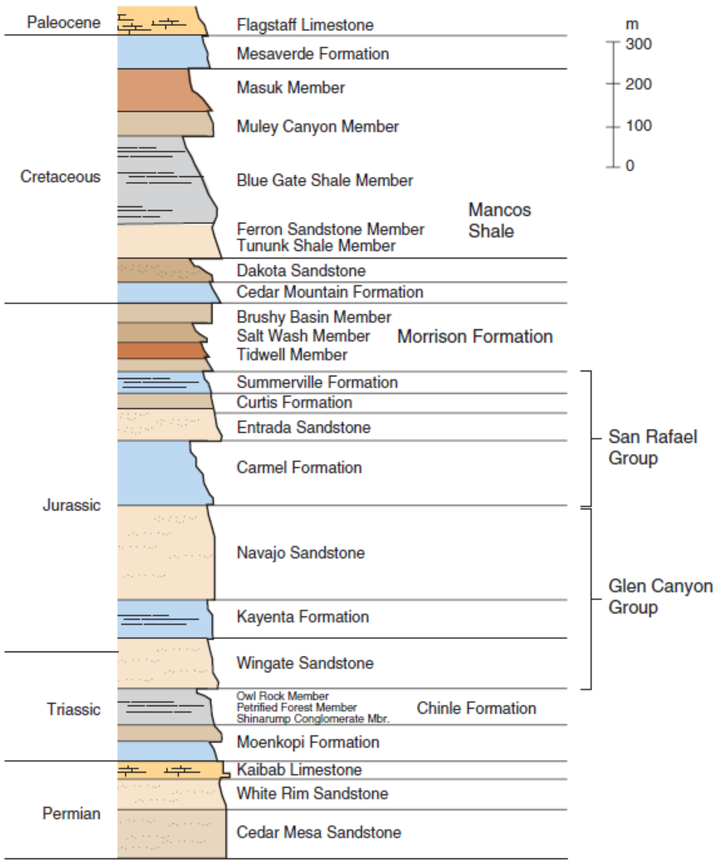
Time-Stratigraphic Units and Time Units

Time-stratigraphic units (also called chronostratigraphic units) and time units make up the category related to or expressing geologic time (Table 5.2). Geologists define a **time-stratigraphic unit** as the rock that formed during a particular interval of geologic time. The most commonly used time-stratigraphic unit is the **system,** which is based on a *stratotype* consisting of rocks in the area where the system was first described. Systems are recognized beyond their stratotype areas by their fossil content. Remember our discussion on how geologists defined and recognized the systems during the 1830s and 1840s, and how these systems served as the basis for the relative geologic time scale (Figure 5.15).

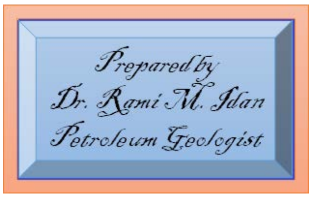
Time units are simply designations for certain intervals of geologic time. The **period** is the most commonly used time unit, but two or more periods may be designated as an *era*, and two or more eras make up an *eon*. Periods may also be subdivided into shorter intervals such as *epochs* and *ages*. All of these time units have corresponding time-stratigraphic units, although the terms *erathem* and *eonothem* are not oft en used (Table 5.2). These two types of units referring to time and their relationship are particularly confusing to beginning students.

Remember, though, that time-stratigraphic units are material bodies of rock that occupy a position in a sequence of strata (for example, the Devonian System), whereas time units refer only to time (the Devonian Period). Thus, a system is a body of rock with lower, medial, and upper parts. In contrast, a time unit such as the Devonian Period is the time during which the Devonian System was deposited, and we refer to early, middle, and late subdivisions.

Consider this example and perhaps the distinction between time-stratigraphic units and time units will be clearer. Think of a three-story building that was built one floor at a time, the lowest first and so on. The building—let us call it Brooks Hall—is the Brooks Hall System, a material body that was built during a specific interval of time, the Brook Hall Period. Lower Brooks Hall was built during the Early Brooks Hall Period and so on. And because system refers to a position in a sequence, we would refer to the third floor as Upper Brooks Hall but not Late Brooks Hall, and likewise we would call the first floor Lower Brooks Hall but not Early Brooks Hall.



• Figure 5.16 **Graphic Representation of the Lithostratigraphic Units in Capital Reef National Park in Utah** Notice that some of the formations are further divided into members, and some are parts of more inclusive groups.



Correlation

In geology, the term **correlation** refers to matching up geologic phenomena in two or more areas. For example, we may simply correlate the same rock units, with no regard to time, over an area in which they are no longer continuous, in which case we refer to ***lithostratigraphic*** *correlation*. In a previous section, we noted that systems were identified beyond their stratotype areas by applying the principle of fossil succession. In this case, we match up rocks of the same relative age, which is a ***time- stratigraphic correlation***.

Correlation of lithostratigraphic units involves demonstrating that a formation or group was once continuous over a particular area. If outcrops are adequate, lithostratigraphic units may be traced laterally even if occasional gaps are present (• Figure 5.17a). Other criteria for lithostratigraphic correlation include the presence of a distinctive key bed, position in a sequence, and composition (Figure 5.17b and c); keep in mind that composition indicates only the geographic extent of a rock unit (see Perspective). Rock cores and well cuttings from drilling operations as well as geophysical data are useful for correlating rock units below the surface.

Many sedimentary rock units, such as widespread formations, are time transgressive, so we cannot rely on lithostratigraphic correlation to demonstrate time equivalence. For example, geologists have accurately correlated sandstone in Arizona with similar rocks in Colorado and South Dakota, but the age of the rocks varies from Early Cambrian in the west to Middle Cambrian further east. The most effective way to demonstrate the time stratigraphic equivalence of sedimentary rocks in different areas involves using biozones, but several other methods are useful too.

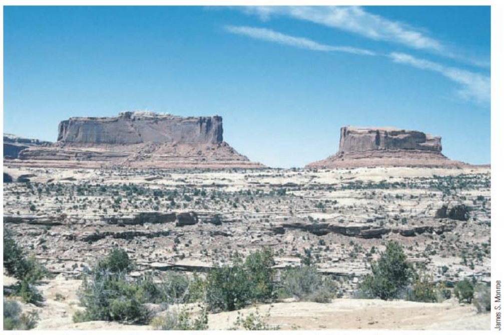
The existence of all organisms now extinct marks two points in time-time of origin and time of extinction. One type of biozone, known as the **range zone,** is defined as the geologic range (total time of existence) of a particular fossil group, such as a species or a group of related species called a genus. The most useful fossils are those that are easily identified, geographically widespread, and that had a rather short geologic range. *Lingula* is a type of brachiopod that meets the first two of these criteria, but its geologic range makes it of little use, whereas the brachiopod

Atrypa and the trilobite *Paradoxides* are **guide fossils** because they are well suited for time-stratigraphic correlation (Figure 5.18).

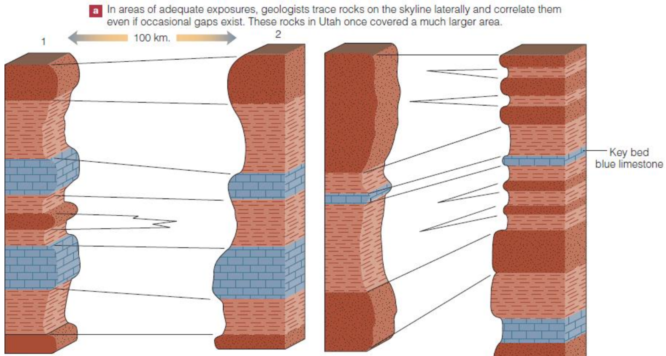
Geologists identify several types of *interval zones* which are defined by the first and last occurrence of a species or genus, but one of the most useful is the **concurrent range zone** that is established by plotting the overlapping ranges of two or more fossils with different ranges (Figure 5.19). Correlating concurrent range zones is probably the most accurate method to determine time equivalence between sedimentary rocks in widely separated areas. Some physical events are also used to demonstrate time equivalence. Remember that rock composition in most cases is useless in this endeavor, but a few rocks, particularly lava flows and ash falls, form rapidly over their entire extent. Accordingly, correlating an ash fall, for example, is time significant (Figure 5.20). Furthermore, ash falls are not restricted to a specific environment, so they may extend from continental to marine environments.

The methods of time-stratigraphic correlation work well for Phanerozoic-age rocks, but Precambrian rocks have few fossils and those present are not very useful for correlation. For these rocks absolute ages based on radioactive dating techniques are most useful.

• Figure 5.17 **Lithostratigraphic Correlation** This kind of correlation demonstrates the original geographic continuity of rock unit, such as a formation or group.

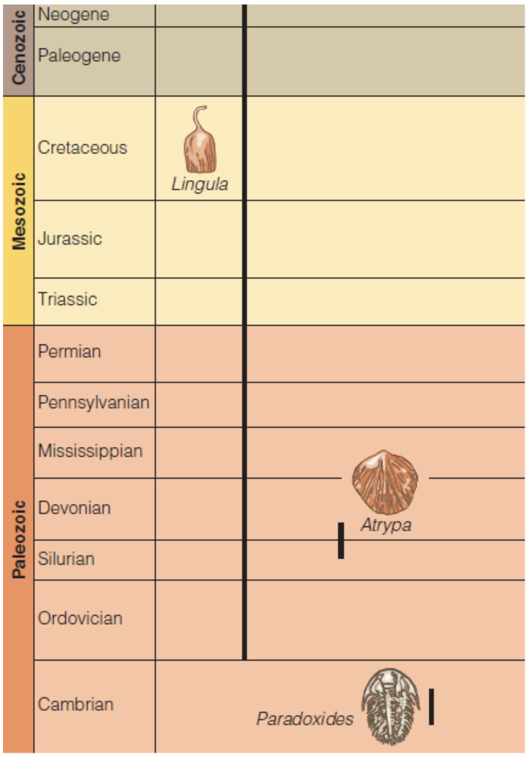


a In areas of adequate exposures, geologists trace rocks on the skyline laterally and correlate them even if occasional gaps exist. These rocks in Utah once covered a much larger area.

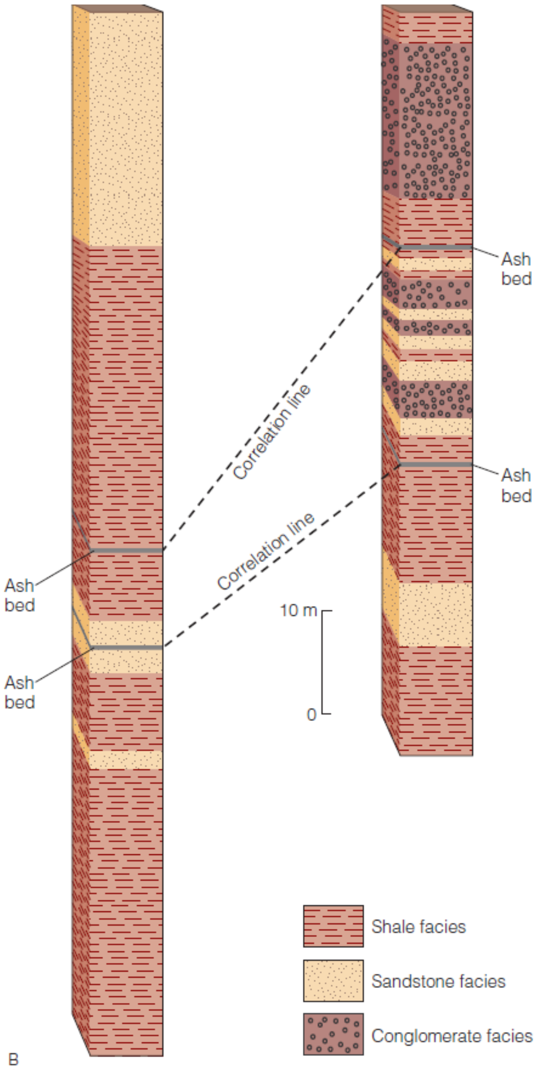


b Lithostratigraphic correlation by similarities in rock type and position in a sequence. We assume the middle sandstone in column 1 intertongues or grades laterally into the shale in column 2.

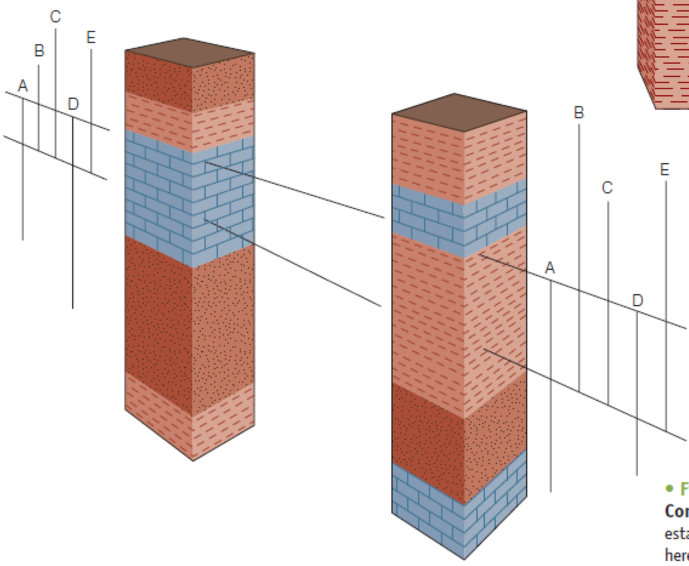
c Lithostratigraphic correlation using a key bed, a distinctive blue limestone in this case.



• **Figure 5.18 Comparison of the Geologic Ranges (Heavy Vertical Lines) of Three Marine Invertebrate Animals** *Lingula* is of no use in biostratigraphy because it has such a long range. *Atrypa* and *Paradoxides* are good guide fossils because both are widespread, easily identified, and have short geologic ranges.



• **Figure 5.20 Ash Beds Used in Time-Stratigraphic Correlation** Volcanic ash accumulates quickly over its entire geographic extent, so that correlating from outcrop to outcrop is time significant. What can you say about sedimentation rates in the two columns during the time encompassed by the correlation lines?



• **Figure 5.19 Time-Stratigraphic Correlation Using Concurrent Range Zones** This concurrent range zone was established by the overlapping ranges of fossils symbolized here by the letters A through E.

: References

Wicander, R., and Monroe, J. S. 2016. Historical geology-Books-Cole.

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Lecture FOUR

Sedimentary Rocks – The Archives of Earth History

Part (1)

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Introduction

We also mentioned previously that sedimentary rocks have a special place in deciphering Earth history, but we should also emphasize that all rocks are important in this endeavor. Certainly igneous rocks help geologists identify ancient convergent plate boundaries and areas of ancient volcanism, and most of our absolute dates for sedimentary rocks come from associated igneous rocks (see Chapter 5).

Metamorphic rocks tell us something of the temperatures and pressures that prevailed in Earth's crust, especially where plates collided and mountains formed.

Nevertheless, our main emphasis in this chapter is on sedimentary rocks because:

(1) they preserve evidence of surface processes responsible for deposition of sediment by running water, wind, waves, glaciers, and so on.

(2) Many contain fossils, our only record of prehistoric life.

So, we focus on areas or environments where sediment accumulates—that is, ***depositional environments***—as well as the criteria used to recognize deposits of specific depositional environments. Accordingly, geologists examine the distinctive properties of sedimentary rocks, keeping in mind that the processes responsible for these features are the same as those operating now (Figure 6.1). Sedimentary rocks are also important in any discussion of geology because many are resources—sand and gravel used in construction, for example—or they contain resources such as petroleum and natural gas. Coal is certainly an important resource, most of which is used at electrical-power generating plants, and phosphorus-rich sedimentary rocks are used in fertilizers, animal feed supplements, metallurgy, matches, ceramics, and preserved foods. Furthermore, a type of sedimentary rock known as banded iron formation is the main source of the world's iron ores.

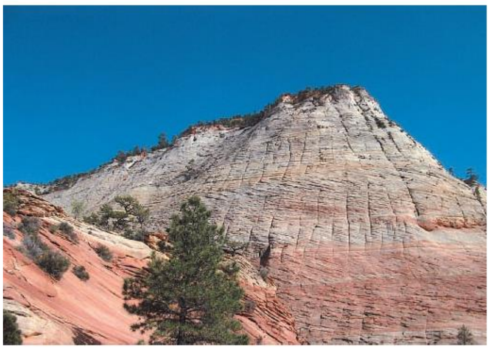


Figure 6.1 Jurassic-age Navajo Sandstone in Zion National Park, Utah Geologists conclude that the Navajo Sandstone was deposited as a vast blanket of sand dunes based on several features of these rocks as well as comparisons with present-day deposits. Much of this chapter covers the criteria geologists use to make such determinations. This rock exposure is called Checkerboard Mesa, so named because of the pattern formed where vertical fractures intersect the layering in the rocks.

Sedimentary Rock Properties:

The first step in investigating sedimentary rocks—or any other rock type, for that matter—is observation and data gathering by visiting rock exposures (outcrops) and carefully examining textures, composition, fossils (if present), thickness, and relationships to other rocks. During these initial field studies, geologists may make some preliminary interpretations. For example, color is a useful feature of some sedimentary rocks: Red rocks likely were deposited on land (see the likely were deposited in a marine environment. However, exceptions are numerous, so color must be used with caution. After completing fieldwork, geologists study the rocks more carefully by microscopic examination, chemical analyses, fossil identification, and construction of diagrams that show vertical and lateral facies relationships. A particularly important aspect of these studies is to compare the features of sedimentary rocks with those in present-day sediment accumulations that formed in known depositional environments. When all the data have been analyzed, geologists make an environmental interpretation.
Determining how sedimentary rocks were deposited and examining their fossils satisfies our curiosity about Earth and life history. If you live in Kansas you might be interested to know that a widespread shallow sea populated by shelled animals known as ammonites as well as sharks and huge marine reptiles were present there during the Cretaceous Period. In addition, there are economic reasons to determine environments of deposition in exploration for mineral deposits, petroleum, and natural gas.

Composition and Texture

Detrital (Clastic) sedimentary rocks contain more than 100 minerals, but only quartz, feldspars, and clay minerals are very common. Rock composition depends mostly on the composition of the rocks in the area from which the detrital sediment was derived (the source area), but it tells little about how deposition took place. Quartz, for example, may have been deposited in a stream channel, in desert dunes, or on a beach, so composition taken alone is not very useful for environmental determinations.

Among the chemical sedimentary rocks, limestone (composed of calcite [CaCO3] and dolostone (composed of dolomite [Ca Mg (CO3)2] are the most common, and both are usually deposited mostly in warm, shallow seas, but a small amount forms in lakes. Evaporites such as rock salt (composed of halite [Na Cl]) and rock gypsum (made up of gypsum

[CaSO4.2H2O]*) invariably indicate arid environments where evaporation rates are high.

Grain size in detrital sedimentary rocks tells us something of the conditions of transport and deposition. Conglomerate is made up of gravel, so we can be sure that it was transported by energetic processes such as swiftly flowing streams, by waves, or by glaciers. Sand also requires high energy transport and tends to be deposited in stream channels, desert dunes, and on beaches, but silt and clay are transported by weak currents and are deposited in low-energy environments such as lagoons, lakes, and river floodplains.

Texture refers to the size, size distribution, shape, and arrangement of clasts in detrital sedimentary rocks. Recall from Chapter 2 that gravel, sand, silt, and clay are simply size designations for detrital particles (see Figure 2.13). The degree to which detrital particles have had their sharp edges and corners smoothed off by abrasion is called **rounding** (Figure 6.2a). **Gravel tends to become rounded very quickly as particles collide with one another, and with considerable transport, sand is also rounded, but smaller particles are carried suspended in water and usually are not so well rounded**.

• Figure 6.2 Rounding and Sorting in Sediments

a Beginners often mistake rounding to mean spherical or ball-shaped. All of these stones are rounded—that is, their sharp corners and edges have been worn smooth—but only the one in the upper left is spherical.

b Sorting in this deposit is only moderate, but all of the stones are well rounded.

c Angular, poorly sorted gravel.

Another textural feature is **sorting,** which refers to the size variation in a sedimentary deposit or rock. If most of the particles are of about the same

size, the sediment or rock is **well sorted**, but if a wide range of sizes is present, the material is **poorly sorted** (Figure 6.2c). Wind has a limited capacity to transport and deposit sediment so its deposits tend to be well sorted, whereas glaciers transport anything supplied to them and their deposits are poorly sorted.

Sedimentary Structures:

Most sedimentary rocks have features known as ***sedimentary structures*** that formed during deposition or shortly thereafter. All are manifestations of the physical and biological processes that take place in depositional environments, so analyses of these structures is the single most important aspect of sedimentary rocks for determining how deposition took place.

The origin of sedimentary structures is well known because geologists see them forming today and many have been produced experimentally (Table 6.1). With very few exceptions, sedimentary rocks have a layered aspect called ***stratification or bedding*** (Figure 6.3); layers less than 1 cm thick are ***laminations***, whereas beds are thicker. Laminations are most common in the mudrocks and form as silt and clay settle from suspension, but they are also found in sandstones and limestones. However, *coarser grained rocks, sandstone and conglomerate, as well as many limestones are frequently bedded.* The surfaces separating one bed from another are ***bedding planes***, above and below which the rocks differ in composition, texture, color, or a combination of these features.

Some of the deposition that takes place in the seas marginal to continents results from ***turbidity currents***, which are sediment–water mixtures that move along the seafloor because they are denser than seawater. When the flow velocity in these currents diminishes, they deposit large particles followed by progressively smaller ones thus forming ***graded bedding*** in which the grain size decreases upward (Figure 6.4). Graded bedding is also found in some stream channels where flow velocity decreases rapidly as during flash floods.

Deposition that takes place on the downwind, sloping side of a sand dune or a similar feature in a stream channel produces ***cross-bedding,*** which accumulates at an angle to the surface on which deposition occurs (Table 6.1, Figure 6.5). Transport and deposition by either wind or water produces cross-bedding and individual cross beds are inclined downward

or dip in the direction of flow. Accordingly, cross-bedding indicates the direction of flow of ancient currents.

TABLE 6.1 Summary of Sedimentary Structures

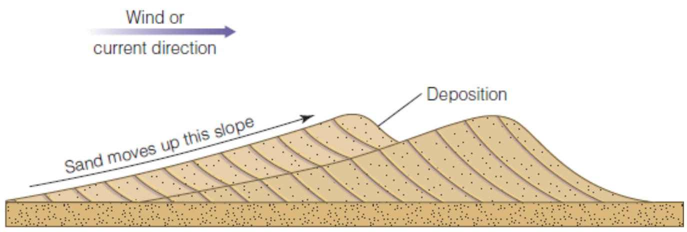
Physical Sedimentary Structures (produced by processes such as currents)	Characteristics	Origin
Laminations (or laminae)	Layers less than 1 cm thick	Form mostly as particles settle from suspension
Beds	Layers more than 1 cm thick	Form as particles settle from suspension and from moving sediment as sand in a stream channel
Graded-bedding	Individual layers with an upward decrease in grain size	Deposition by turbidity currents or during the waning stages of floods
Cross-bedding	Layers deposited at an angle to the surface on which they accumulated	Deposition on a sloping surface at the downwind side of a sand dune
Ripple marks	Small (<3 cm high) ridges and troughs on bedding planes	
Current ripple marks	Asymmetric ripple marks	Result from deposition by water or air currents flowing in one direction
Wave-formed ripple marks	Symmetric ripple marks; generally with a sharp crest and broad though	Formed by oscillating currents (waves)
Mud cracks	Intersecting cracks in clay-rich sediments	Drying and shrinkage of mud along a lakeshore, a floodplain, or on tidal flats
Biogenic Sedimentary Structures (produced by organisms)		
Trace fossils	Tracks, trails, tubes, and burrows	Indications of organic activity. Intense activity results in *bioturbation* involving disruption of sediment

• Figure 6.3 Stratification in Sedimentary Rock

a These rocks in Valley of the Gods in Utah are layered or stratified. They were originally continuous over this entire area but have been deeply eroded.

b Laminations, layers less than 1 cm thick, in the chemical sedimentary rock called banded iron formation in Michigan. The pencil measures about 14 cm long.

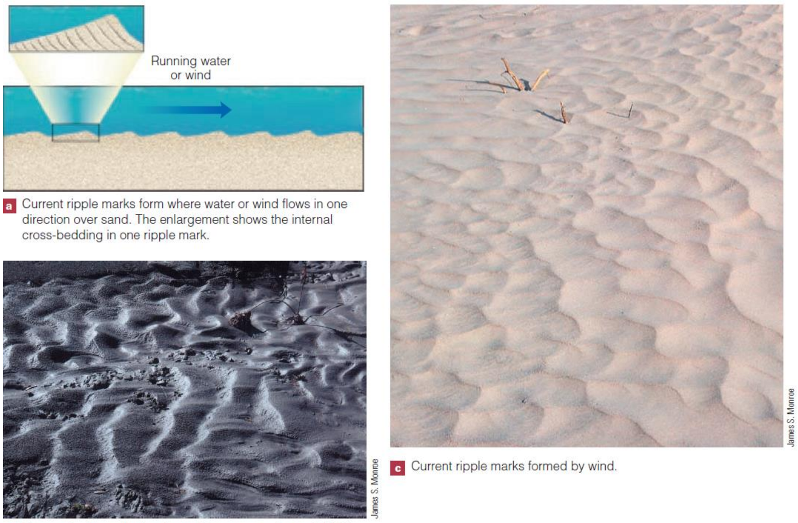
• Figure 6.5 Cross-Bedding Cross-bedding forms when individual beds or strata are deposited at an angle to the surface upon which they accumulate.



a Origin of cross-bedding by deposition on the sloping surface of a desert dune. Cross-bedding also forms in dune-like structures in stream and river channels and on the continental shelf.

Small-scale alternating ridges and troughs common on bedding planes, especially in sand, are **ripple marks** (Table 6.1). Ripple marks, or simply ripples, form either by unidirectional flow of wind or water, and they form by the to and fro motion of waves. ***Current* ripples, those formed by the flow of wind or water, have asymmetric profiles with a gentle upstream slope and a seep downstream slope (Figure 6.6).** Because of their asymmetry they indicate the original flow direction. ***Wave-formed ripple marks* tend to have symmetrical profiles (Figure 6.7).**

• Figure: 6.6 **Current Ripple Marks** Current ripple marks are small (<3 cm high) sedimentary structures with an asymmetric profile.



a Current ripple marks form where water or wind flows in one direction over sand. The enlargement shows the internal cross-bedding in one ripple mark.

b Current ripple marks that formed in a stream channel. Which way did the current flow?

c Current ripple marks formed by wind.

• Figure 6.7 **Wave-Formed Ripple Marks** Just as with current ripple marks, these are small-scale sedimentary structures, but they tend to have symmetric profiles.

a Wave-formed ripples form where waves move to and fro.

b Wave-formed ripples in shallow seawater.

Clay-rich sediments tend to shrink as they dry and crack into polygonal forms called **mud cracks** (Table 6.1, Figure 6.8). If you see mud cracks in ancient rocks you can be sure that deposition took place in an environment where alternating wetting and drying took place, such as along a lakeshore, on a river fl oodplain, or where mud is exposed at low tide along a seashore.

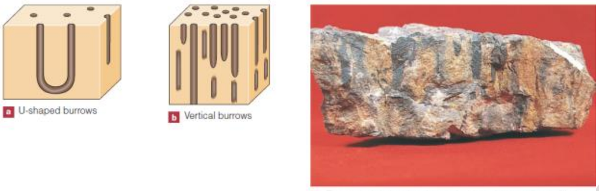
The sedimentary structures discussed so far all form by physical processes, but biological processes yield **biogenic sedimentary structures** and include tracks, trails, and burrows (Table 6.1, Figure 6.9). Remember from Chapter 5 that these biologically produced features are also called *trace fossils*. Extensive burrowing by organisms, or what is known as **bioturbation,** may so thoroughly disrupt sediments that other structures are destroyed.

• Figure 6.8 Mud Cracks Form in Clay-Rich Sediments When They Dry and Contract

a Mud cracks in a present-day environment.

b Ancient mud cracks in Glacier National Park, Montana. Notice that the cracks have been filled in by sediment.

• Figure 6.9 Bioturbation Results from Organisms Burrowing Through Sediments

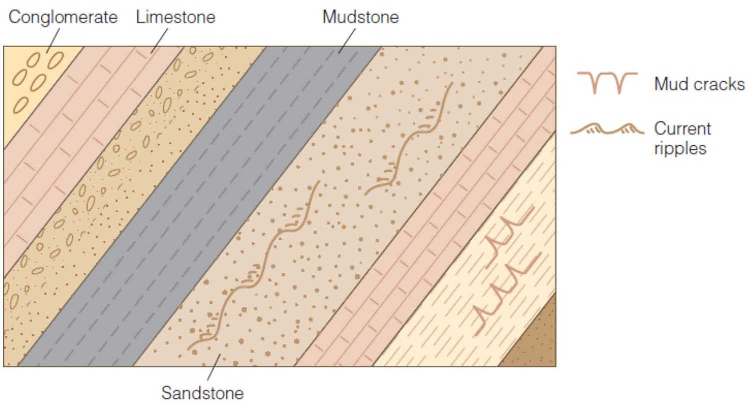


a U-shaped burrows

b Vertical burrows

c The vertical dark-colored areas in this rock are sediment-filled burrows

Sedimentary structures are important for environmental analyses, but be aware that no single structure is unique to a specific environment—current ripples can form in desert dunes, in stream channels, and in tidal channels in the sea, for example. However, associations of sedimentary structures taken along with other sedimentary rock properties allow geologists to make environmental interpretations with a high degree of confidence. We noted in Chapter 5 that to decipher the geologic history of any area geologists use the principle of superposition to determine the relative ages of sedimentary rocks, but if the rocks have been deformed the task is more difficult. Several sedimentary structures as well as some fossils allow geologists to resolve these kinds of problems (Figure 6.10).



• **Figure 6.10 Using Sedimentary Structures to Determine the Relative Ages of Deformed Sedimentary Rocks** According to the principle of original horizontality, these strata were not deposited in their present position but rather deposited more or less horizontally and then deformed. To determine which layer is oldest and youngest notice the sedimentary structures. The mud cracks form a "V" that opens up toward the younger strata, and the shape of the current ripple marks indicate that the youngest layer is the one in the lower right.

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Wicander, R., and Monroe, J. S. 2016. Historical geology-Books-Cole.

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Lecture FIVE

Sedimentary Rocks – The Archives of Earth History

Part (2)

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Geometry of Sedimentary Rocks

The three-dimensional shape, or geometry, of a sedimentary rock body may be helpful in environmental analyses, but it must be used with caution because the same geometry can be found in more than one environment. Moreover, geometry can be modified by sediment compaction during lithification, and by erosion and deformation. Nevertheless, it is useful when considered in conjunction with other features.

Some of the most extensive sedimentary rocks in the geologic record resulted from marine transgressions and regressions (see Figure 5.8). These rocks cover hundreds or thousands of square kilometres but are perhaps only a few tens to hundreds of meters thick. That is, they are not very thick compared to their dimensions of length and width and thus, have a *blanket* or *sheet geometry*.

Some sand deposits have an *elongate* or *shoestring geometry*, especially those deposited in stream channels or barrier islands. Delta deposits tend to be lens shaped when viewed in cross profile or long profile, but lobate when observed from above. Buried reefs are irregular, but many are long and narrow, although rather circular one also exist.

Fossils—the Biologic Content of Sedimentary Rocks

We defined fossils as the remains or traces of prehistoric organisms, and we discussed how geologists use fossils in some aspects of stratigraphy to establish biostratigraphic units.

Fossils are also important constituents of some rocks, especially limestones that may be composed largely of shells of marine-dwelling animals such as brachiopods, clams, and corals (see Figure 2.15a and c below), or even the droppings (pellets) of these organisms. Fossils are not present in all sedimentary rocks, but if they are, they are important for determining depositional environments.

We must consider two factors when using fossils in environmental analyses. First, did the organisms in question live where they were buried, or were their remains transported there? Fossil dinosaurs, for example, usually indicate deposition in some land environment such as a river floodplain, but if their bones are found in rocks with clams, corals, and

sea lilies, we must assume a carcass was washed out to sea. Second, what kind of habitat did the organisms originally occupy? Studies of a fossil's structure and its living relatives, if any, are helpful.

▸Figure 2.15 Varieties of Limestone

(a) Limestone with numerous fossil shells is called fossiliferous limestone.

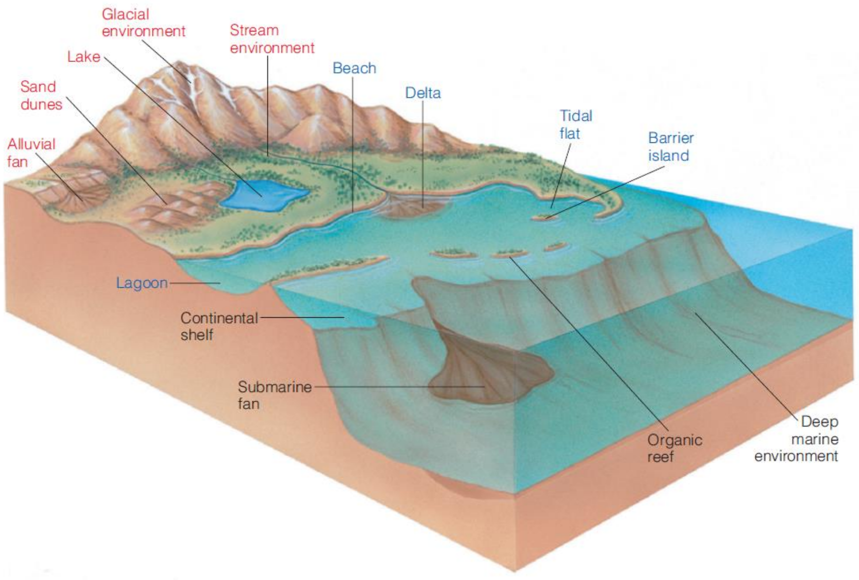
(c) Coquina is limestone made up entirely of broken shells. Compare with the fossiliferous limestone in (a).

Clams with heavy, thick shells typically live in shallow, turbulent water, whereas those with thin shells are found in low-energy environments. Most corals live in warm, clear, shallow marine environments where symbiotic bacteria can carry out photosynthesis.

Depositional Environments

We defined a ***depositional environment*** *as any area where sediment accumulates, but more specifically it entails a particular area where physical, chemical, and biological processes operate to yield a distinctive kind of deposit.* Geologists recognize three broad areas of deposition *continental*, *transitional*, and *marine*—each of which has several specific environments (Figure 6.9).

▸Figure 6.9 Depositional Environments Continental environments are shown in red type. The environments along the seashore, shown in blue type, are transitional from continental to marine. The others, shown in black type, are marine environments.



Continental Environments

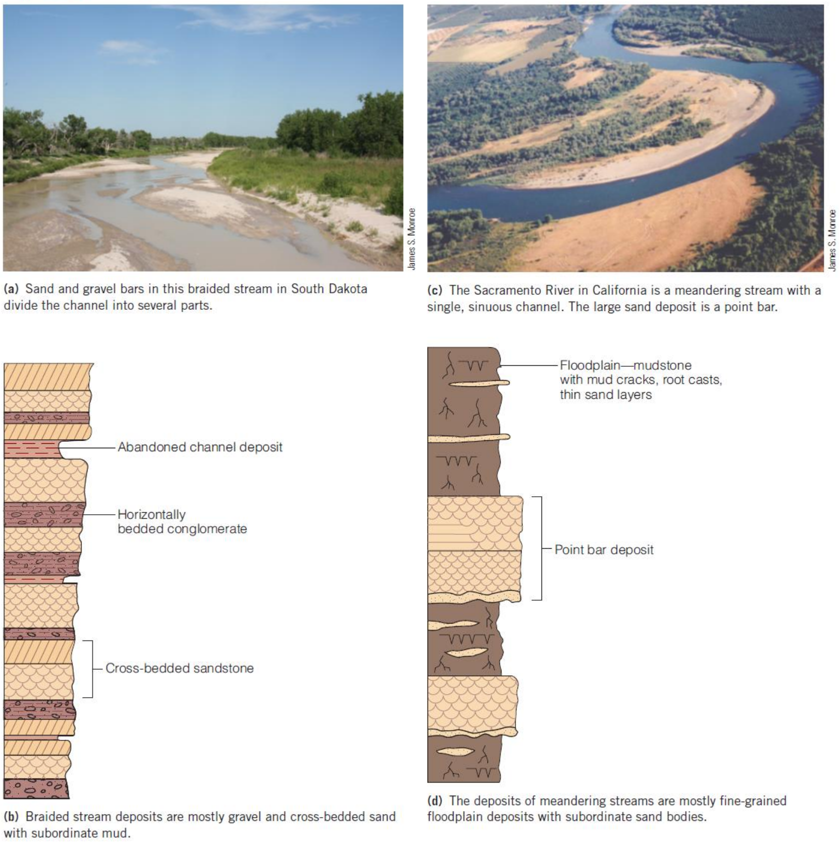
Deposition on the continents—that is, on land—takes place in fluvial systems (rivers and streams), lakes, deserts, and areas covered by or adjacent to glaciers (Table 6.2). The sedimentary deposits in each of these environments show combinations of features that allow us to determine that they were in fact deposited on land and to distinguish one from the other.

Fluvial refers to river and stream activity and to their deposits. These fluvial deposits accumulate in either braided stream systems or meandering stream systems.

Desert environments are commonly inferred from an association of features found in **sand dune, alluvial fan,** and **playa lake** deposits (Fig. below).

Varves represent yearly accumulations of sediment—the light layers form in spring and summer and consist of silt and clay, whereas the dark layers consist of clay and organic matter that settled when the lake froze over. Dropstones liberated from icebergs might also be present in glacial lake deposits

▸**Figure 6.10** **Fluvial Deposits of Braided and Meandering Streams**



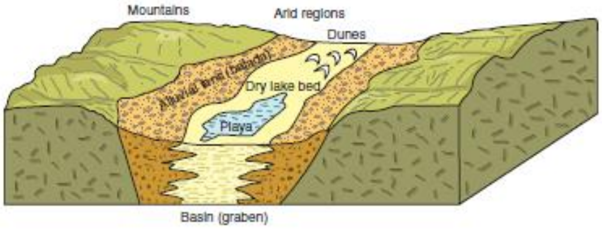
(a) Sand and gravel bars in this braided stream in South Dakota divide the channel into several parts.

(c) The Sacramento River in California is a meandering stream with a single, sinuous channel. The large sand deposit is a point bar.

(b) Braided stream deposits are mostly gravel and cross-bedded sand with subordinate mud.

(d) The deposits of meandering streams are mostly fine-grained floodplain deposits with subordinate sand bodies.

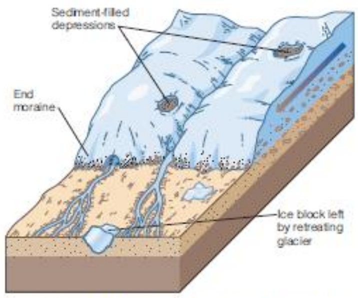
▸Figure 6.11 Deposits of Sediment in a Desert Environment



(a) Block diagram showing the association of deposits found in a desert environment.

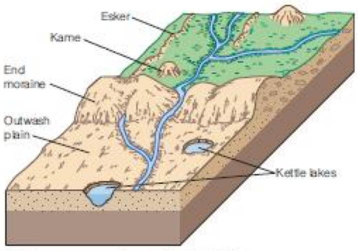
(b) A playa lake near Fallon, Nevada. Notice the deposits of rock salt that were forming when this image was taken in June 2007.

▸Figure 6.12 Glaciers and Their Deposits



(a) Glaciers transport huge quantities of sediment that is deposited as moraines and outwash.

(c) This ridge of gravel is a small end moraine that was deposited in 1995 by the Exit Glacier near Seward, Alaska. The gravel layers in the foreground are outwash.



(b) The same area as shown in (a), but with the ice gone.

(d) Dropstone in glacial lake deposits showing varves in Canada.

Transitional Environments

Transitional environments include those in which both marine processes and processes typical of continental environments operate (Table 6.2). For instance, deposition where a river or stream (fluvial system) enters the sea yields a body of sediment called a **delta,** but the deposit is modified by marine processes, especially waves and tides. Small deltas in lakes are also common, but the ones along seashores are much larger, far more complex, and many deltas are important economically because they form the reservoir rocks for petroleum and natural gas.

On broad continental margins with abundant sand, long **barrier islands** lie offshore separated from the mainland by a lagoon. Barrier islands are

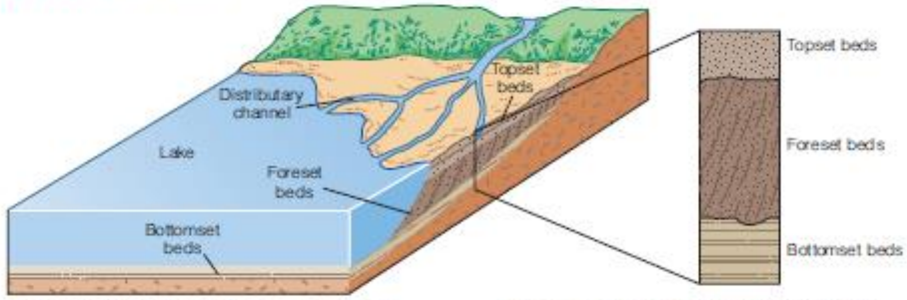
common along the Gulf Coast and much of the Atlantic Coast of the United States, and many ancient deposits in those areas formed in this environment. Notice in Figure 6.15a that a barrier island has several subenvironments: beach sands grade offshore into finer deposits of the shoreface, dune sands contain shell fragments and thus differ from dune sands of deserts, and fine-grained lagoon deposits (mostly mud) have marine fossils and bioturbation.

Tidal flats are present along many coastlines where part of the shoreline environment is periodically covered by seawater at high tide and then exposed at low tide. Many tidal flats build or prograde seaward and yield a sequence of rocks grading upward from sand to mud. One of their most distinctive features is *herringbone crossbedding* sets of cross beds that dip in opposite directions.

TABLE 6.2 Summary Chart of Rocks and Features in Depositional Environments

Environment	Dominant Type of Rocks
Continental Environments	
Fluvial	
Braided stream	Mostly horizontally bedded conglomerate and cross-bedded sandstone; mudrocks not common
Meandering stream	Mostly mudrocks deposited on floodplains; subordinate but distinctive lenticular sandstones deposited in point bars
Desert	
Alluvial fan	Poorly sorted conglomerate from debris flows and sandstone- conglomerate-filled channels
Sand dune	Well sorted, rounded sandstone with large-scale cross-beds
Playa lake	Laminated mudstone/siltstone; evaporites, rock salt, rock gypsum, and others
Glacial	
Outwash	Much like braided stream deposits
Moraines	Unsorted, nonstratified deposits of sand and gravel
Transitional Environments	
Delta (marine)	Mudrocks and sandstone in coarsening upward sequences; associated rocks of marine origin; fossils of marine and land-dwelling organisms
Beach	Rounded sandstone with variable soring, commonly with shells or shell fragments, wave-formed ripple marks, and small-scale cross-bedding
Barrier Island	
Beach	As above
Sand dunes	Much like desert dunes but with sand-sized shell fragments
Tidal flat	Mudstone and sandstone in fining-upward sequences; distinctive herringbone cross-bedding in sandstone
Marine Environments	
Continental shelf	
Inner Shelf	Mostly cross-bedded sandstone with wave-formed ripples, marine fossils, and bioturbation
Outer Shelf	Mostly mudrocks with subordinate sandstone; marine fossils and bioturbation
Continental slope and rise	Turbidite sequences in submarine fans with graded bedding in sandstone and mudrocks
Carbonate shelf	Limestone (dolostone). Limestone varies from coquina (made of shell fragments) to oolitic limestone to micrite (carbonate mud). Cross-beds, mud cracks, ripple marks common; marine fossils.
Deep-ocean basin	Pelagic clay and calcareous and siliceous oozes.
Evaporite environments	Rock salt and rock gypsum the most common, but others including potassium and magnesium salts may be present.

▸**Figure 6.13** **Origin of a Delta**



(a) Surface view and internal structure of the simplest type of prograding delta.

(b) Coarsening upward vertical sequence of deposits that result from a prograding delta.

▸**Figure 6.14** **Stream-, Wave-, and Tide-Dominated Deltas**

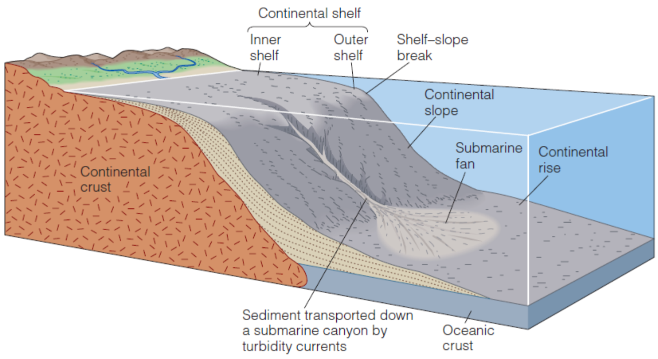
(a) The Mississippi River delta on the U.S. Gulf Coast is stream-dominated.

(b) The Nile delta of Egypt is wave-dominated.

(c) The Ganges-Brahmaputra delta of Bangladesh is tide-dominated.

Marine Environments

Marine environments include the continental shelf, slope, and rise, and the deep seafloor (Table 6.2). Much of the detritus eroded from continents is eventually deposited in marine environments, but other types of sediments are found here, too. Much of the limestone in the geologic record, as well as many evaporites were deposited in shallow marine environments.

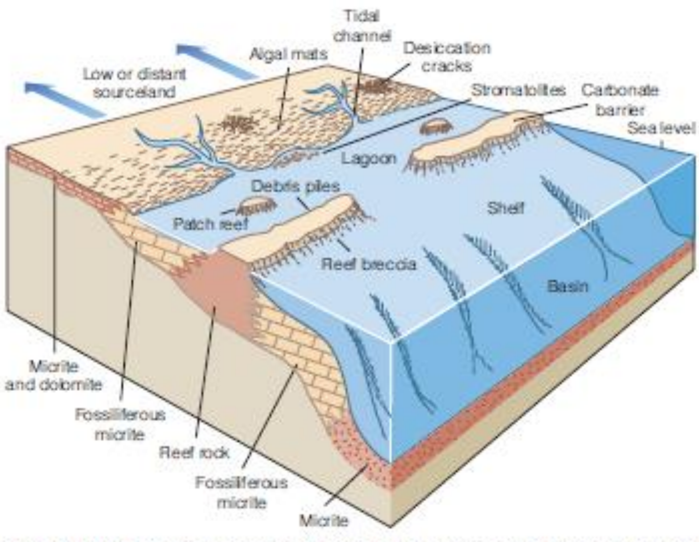


• **Figure 6.18 Depositional Environments on the Continental Shelf, Slope, and Rise** Submarine canyons are the main avenues of sediment transport across the shelf. Turbidity currents carry sediments beyond the shelf, where they accumulate as submarine fans.

Detrital Marine Environments The **continental shelf**, the gently sloping area adjacent to a continent, consists of two parts. A high-energy inner part is periodically stirred up by waves and tidal currents. Its sediment is mostly sand shaped into large cross-bedded dunes, and bedding planes are marked by wave-formed ripple marks. Of course, marine fossils are typical, as is bioturbation. The lowenergy outer part of the shelf has mostly mud with marine fossils, but at the transition

between the inner and outer shelf, layers of sand and mud intertongue. Much of the sediment derived from continents crosses the continental shelf and is funneled into deeper water through submarine canyons. It eventually comes to rest on the **continental slope** and **continental rise** as a series of overlapping submarine fans (Figure 6.16). Once sediment passes the outer margin of the shelf, turbidity currents transport it. So sands with graded bedding are common but so is mud that settled from seawater.

Carbonate Depositional Environments Limestone and dolostone are the only widespread carbonate rocks, and we already know that most dolostone is altered limestone. Thus, our discussion focuses on the deposition of sediment that, when lithified, becomes limestone. In some respects, limestone is similar to detrital sedimentary rocks—many limestones are made up of graveland sand-sized grains and microcrystalline carbonate mud called *micrite*. Some limestone forms in lakes, but by far, most of it was deposited in warm, shallow seas on carbonate shelves (Figure 6.18). In any case, deposition occurs where little detrital sediment is present, especially mud. Carbonate barriers form in high-energy areas and may be reefs, banks of skeletal particles, or accumulations of spherical carbonate grains known as *ooids*—limestone composed mostly of ooids is called *oolitic limestone*.



▸**Figure 6.18** **Carbonate Depositional Environments** Although some limestone forms in lakes, most of it is deposited on carbonate shelves similar to the one in this illustration. However, the carbonate barrier may be reefs of corals, clams, and algae, or it may be made up of oolitic sand or skeletal-fragment sand. Deposition of limestone in environments much like these is now taking place in southern Florida and the Persian Gulf.

Evaporite Environments The most of the extensive deposits were formed in the seas. Although locally abundant, evaporites are not nearly as common as sandstone, mudrocks, and limestone.

Paleogeography

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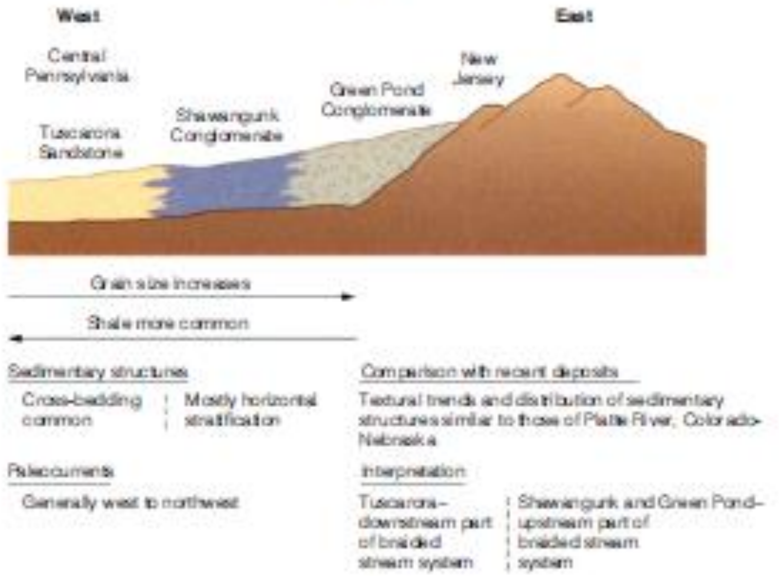

References: Wicander, R., and Monroe, J. S. 2016. Historical geology-Books-Cole.

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Lecture FIVE

Sedimentary Rocks – The Archives of Earth History

Part (2)

by

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Geometry of Sedimentary Rocks

The three-dimensional shape, or geometry, of a sedimentary rock body may be helpful in environmental analyses, but it must be used with caution because the same geometry can be found in more than one environment. Moreover, geometry can be modified by sediment compaction during lithification, and by erosion and deformation. Nevertheless, it is useful when considered in conjunction with other features.

Some of the most extensive sedimentary rocks in the geologic record resulted from marine transgressions and regressions (see Figure 5.8). These rocks cover hundreds or thousands of square kilometres but are perhaps only a few tens to hundreds of meters thick. That is, they are not very thick compared to their dimensions of length and width and thus, have a *blanket* or *sheet geometry*.

Some sand deposits have an *elongate* or *shoestring geometry*, especially those deposited in stream channels or barrier islands. Delta deposits tend to be lens shaped when viewed in cross profile or long profile, but lobate when observed from above. Buried reefs are irregular, but many are long and narrow, although rather circular one also exist.

Fossils—the Biologic Content of Sedimentary Rocks

We defined fossils as the remains or traces of prehistoric organisms, and we discussed how geologists use fossils in some aspects of stratigraphy to establish biostratigraphic units.

Fossils are also important constituents of some rocks, especially limestones that may be composed largely of shells of marine-dwelling animals such as brachiopods, clams, and corals (see Figure 2.15a and c below), or even the droppings (pellets) of these organisms. Fossils are not present in all sedimentary rocks, but if they are, they are important for determining depositional environments.

We must consider two factors when using fossils in environmental analyses. First, did the organisms in question live where they were buried, or were their remains transported there? Fossil dinosaurs, for example, usually indicate deposition in some land environment such as a river floodplain, but if their bones are found in rocks with clams, corals, and

sea lilies, we must assume a carcass was washed out to sea. Second, what kind of habitat did the organisms originally occupy? Studies of a fossil's structure and its living relatives, if any, are helpful.

▸Figure 2.15 Varieties of Limestone

(a) Limestone with numerous fossil shells is called fossiliferous limestone.

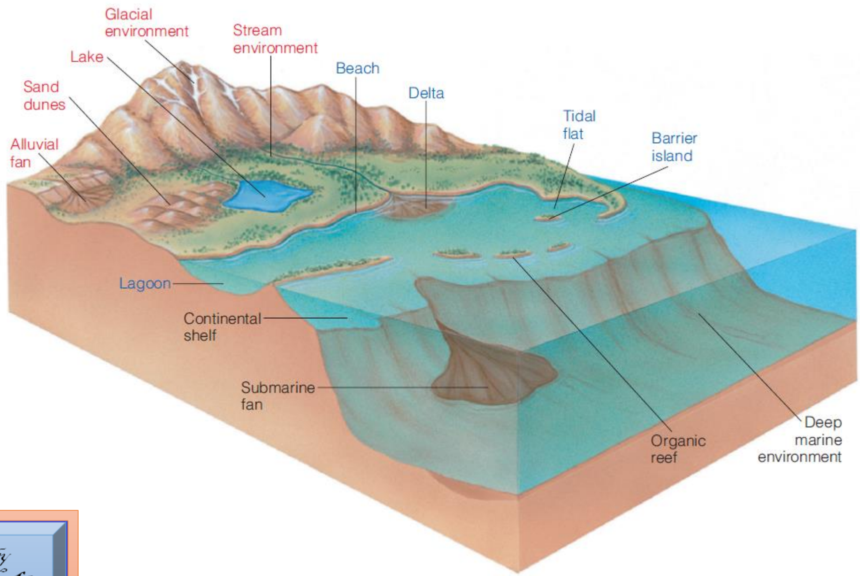
(c) Coquina is limestone made up entirely of broken shells. Compare with the fossiliferous limestone in (a).

Clams with heavy, thick shells typically live in shallow, turbulent water, whereas those with thin shells are found in low-energy environments. Most corals live in warm, clear, shallow marine environments where symbiotic bacteria can carry out photosynthesis.

Depositional Environments

We defined a ***depositional environment*** *as any area where sediment accumulates, but more specifically it entails a particular area where physical, chemical, and biological processes operate to yield a distinctive kind of deposit.* Geologists recognize three broad areas of deposition *continental*, *transitional*, and *marine*—each of which has several specific environments (Figure 6.9).

▸Figure 6.9 Depositional Environments Continental environments are shown in red type. The environments along the seashore, shown in blue type, are transitional from continental to marine. The others, shown in black type, are marine environments.



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Continental Environments

Deposition on the continents—that is, on land—takes place in fluvial systems (rivers and streams), lakes, deserts, and areas covered by or adjacent to glaciers (Table 6.2). The sedimentary deposits in each of these environments show combinations of features that allow us to determine that they were in fact deposited on land and to distinguish one from the other.

Fluvial refers to river and stream activity and to their deposits. These fluvial deposits accumulate in either braided stream systems or meandering stream systems.

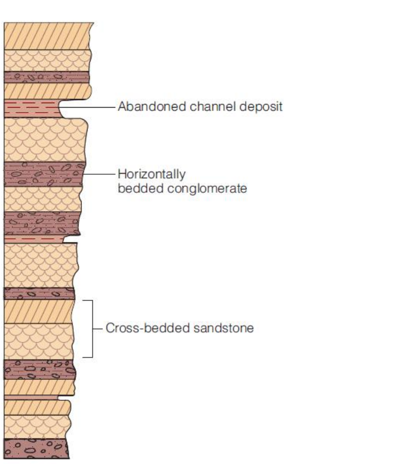
Desert environments are commonly inferred from an association of features found in **sand dune, alluvial fan,** and **playa lake** deposits (Fig. below).

Varves represent yearly accumulations of sediment—the light layers form in spring and summer and consist of silt and clay, whereas the dark layers consist of clay and organic matter that settled when the lake froze over. Dropstones liberated from icebergs might also be present in glacial lake deposits

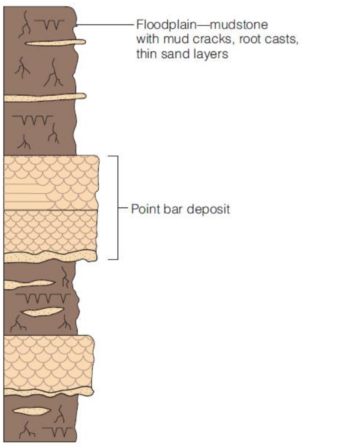
►Figure 6.10 **Fluvial Deposits of Braided and Meandering Streams**

(a) Sand and gravel bars in this braided stream in South Dakota divide the channel into several parts.

(c) The Sacramento River in California is a meandering stream with a single, sinuous channel. The large sand deposit is a point bar.



(b) Braided stream deposits are mostly gravel and cross-bedded sand with subordinate mud.



(d) The deposits of meandering streams are mostly fine-grained floodplain deposits with subordinate sand bodies.

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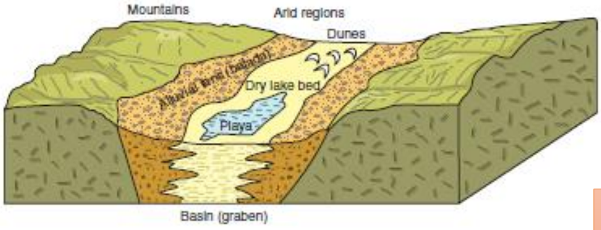
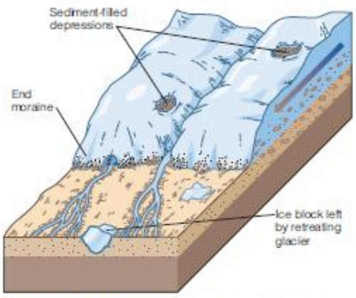
▸ Figure 6.11 Deposits of Sediment in a Desert Environment



(a) Block diagram showing the association of deposits found in a desert environment.

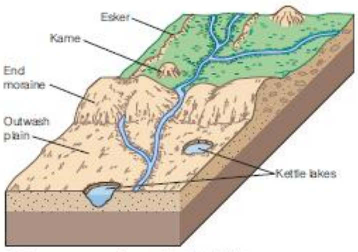
(b) A playa lake near Fallon, Nevada. Notice the deposits of rock salt that were forming when this image was taken in June 2007.

▸ Figure 6.12 Glaciers and Their Deposits



(a) Glaciers transport huge quantities of sediment that is deposited as moraines and outwash.

(c) This ridge of gravel is a small end moraine that was deposited in 1995 by the Exit Glacier near Seward, Alaska. The gravel layers in the foreground are outwash.



(b) The same area as shown in (a), but with the ice gone.

(d) Dropstone in glacial lake deposits showing varves in Canada.

Transitional Environments

Transitional environments include those in which both marine processes and processes typical of continental environments operate (Table 6.2). For instance, deposition where a river or stream (fluvial system) enters the sea yields a body of sediment called a **delta,** but the deposit is modified by marine processes, especially waves and tides. Small deltas in lakes are also common, but the ones along seashores are much larger, far more complex, and many deltas are important economically because they form the reservoir rocks for petroleum and natural gas.

On broad continental margins with abundant sand, long **barrier islands** lie offshore separated from the mainland by a lagoon. Barrier islands are

common along the Gulf Coast and much of the Atlantic Coast of the United States, and many ancient deposits in those areas formed in this environment. Notice in Figure 6.15a that a barrier island has several subenvironments: beach sands grade offshore into finer deposits of the shoreface, dune sands contain shell fragments and thus differ from dune sands of deserts, and fine-grained lagoon deposits (mostly mud) have marine fossils and bioturbation.

Tidal flats are present along many coastlines where part of the shoreline environment is periodically covered by seawater at high tide and then exposed at low tide. Many tidal flats build or prograde seaward and yield a sequence of rocks grading upward from sand to mud. One of their most distinctive features is *herringbone crossbedding* sets of cross beds that dip in opposite directions.

TABLE 6.2 **Summary Chart of Rocks and Features in Depositional Environments**

Environment	Dominant Type of Rocks
Continental Environments	
Fluvial	
Braided stream	Mostly horizontally bedded conglomerate and cross-bedded sandstone; mudrocks not common
Meandering stream	Mostly mudrocks deposited on floodplains; subordinate but distinctive lenticular sandstones deposited in point bars
Desert	
Alluvial fan	Poorly sorted conglomerate from debris flows and sandstone- conglomerate-filled channels
Sand dune	Well sorted, rounded sandstone with large-scale cross-beds
Playa lake	Laminated mudstone/siltstone; evaporites, rock salt, rock gypsum, and others
Glacial	
Outwash	Much like braided stream deposits
Moraines	Unsorted, nonstratified deposits of sand and gravel
Transitional Environments	
Delta (marine)	Mudrocks and sandstone in coarsening upward sequences; associated rocks of marine origin; fossils of marine and land-dwelling organisms
Beach	Rounded sandstone with variable soring, commonly with shells or shell fragments, wave-formed ripple marks, and small-scale cross-bedding
Barrier Island	
Beach	As above
Sand dunes	Much like desert dunes but with sand-sized shell fragments
Tidal flat	Mudstone and sandstone in fining-upward sequences; distinctive herringbone cross-bedding in sandstone
Marine Environments	
Continental shelf	
Inner Shelf	Mostly cross-bedded sandstone with wave-formed ripples, marine fossils, and bioturbation
Outer Shelf	Mostly mudrocks with subordinate sandstone; marine fossils and bioturbation
Continental slope and rise	Turbidite sequences in submarine fans with graded bedding in sandstone and mudrocks
Carbonate shelf	Limestone (dolostone). Limestone varies from coquina (made of shell fragments) to oolitic limestone to micrite (carbonate mud). Cross-beds, mud cracks, ripple marks common; marine fossils.
Deep-ocean basin	Pelagic clay and calcareous and siliceous oozes.
Evaporite environments	Rock salt and rock gypsum the most common, but others including potassium and magnesium salts may be present.

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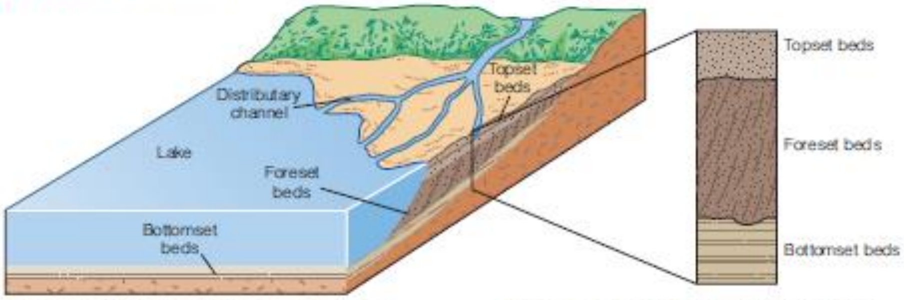
▸**Figure 6.13** Origin of a Delta



(a) Surface view and internal structure of the simplest type of prograding delta.

(b) Coarsening upward vertical sequence of deposits that result from a prograding delta.

▸**Figure 6.14** Stream-, Wave-, and Tide-Dominated Deltas

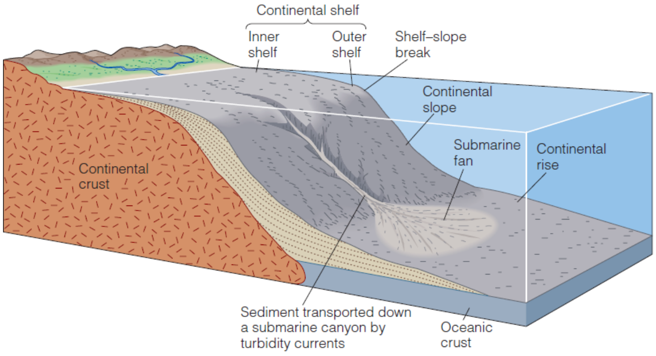
(a) The Mississippi River delta on the U.S. Gulf Coast is stream-dominated.

(b) The Nile delta of Egypt is wave-dominated.

(c) The Ganges-Brahmaputra delta of Bangladesh is tide-dominated.

Marine Environments

Marine environments include the continental shelf, slope, and rise, and the deep seafloor (Table 6.2). Much of the detritus eroded from continents is eventually deposited in marine environments, but other types of sediments are found here, too. Much of the limestone in the geologic record, as well as many evaporites were deposited in shallow marine environments.



• **Figure 6.18 Depositional Environments on the Continental Shelf, Slope, and Rise** Submarine canyons are the main avenues of sediment transport across the shelf. Turbidity currents carry sediments beyond the shelf, where they accumulate as submarine fans.

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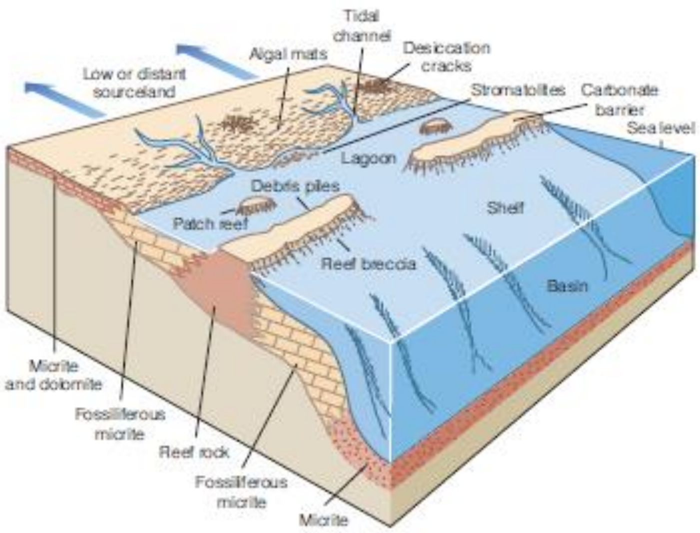
<u>Detrital Marine Environments</u> The **continental shelf**, the gently sloping area adjacent to a continent, consists of two parts. A high-energy inner part is periodically stirred up by waves and tidal currents. Its sediment is mostly sand shaped into large cross-bedded dunes, and bedding planes are marked by wave-formed ripple marks. Of course, marine fossils are typical, as is bioturbation. The lowenergy outer part of the shelf has mostly mud with marine fossils, but at the transition

between the inner and outer shelf, layers of sand and mud intertongue. Much of the sediment derived from continents crosses the continental shelf and is funneled into deeper water through submarine canyons. It eventually comes to rest on the **continental slope** and **continental rise** as a series of overlapping submarine fans (Figure 6.16). Once sediment passes the outer margin of the shelf, turbidity currents transport it. So sands with graded bedding are common but so is mud that settled from seawater.

<u>Carbonate Depositional Environments</u> Limestone and dolostone are the only widespread carbonate rocks, and we already know that most dolostone is altered limestone. Thus, our discussion focuses on the deposition of sediment that, when lithified, becomes limestone. In some respects, limestone is similar to detrital sedimentary rocks—many limestones are made up of graveland sand-sized grains and microcrystalline carbonate mud called *micrite*. Some limestone forms in lakes, but by far, most of it was deposited in warm, shallow seas on carbonate shelves (Figure 6.18). In any case, deposition occurs where little detrital sediment is present, especially mud. Carbonate barriers form in high-energy areas and may be reefs, banks of skeletal particles, or accumulations of spherical carbonate grains known as *ooids*—limestone composed mostly of ooids is called *oolitic limestone*.



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<u>Evaporite Environments</u> The most of the extensive deposits were formed in the seas. Although locally abundant, evaporites are not nearly as common as sandstone, mudrocks, and limestone.

<u>Paleogeography</u>

Paleogeography deals with Earth's geography of the past. In the chapters on geologic history, paleogeographic maps show the distribution of continents at various times; they have not remained fixed because of plate

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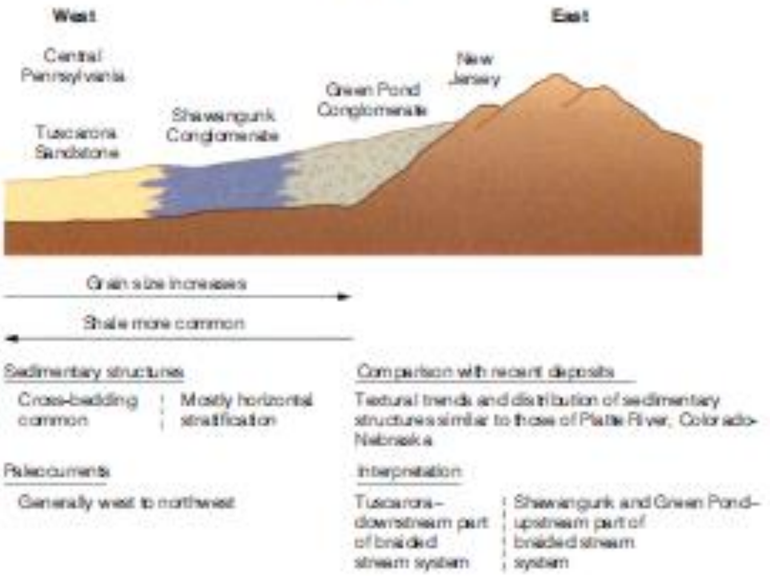
▸ Figure 6.20 **Lower Silurian Strata in the Eastern United States** Simplified cross section showing the probable lateral relationships for the Green Pond Conglomerate, Shawangunk Conglomerate, and Tuscarora Sandstone and the criteria used to interpret these rocks as braided stream deposits.



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Lecture SIX

Evolution

Theory and the Fossils Records

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Evolution: what does it mean?

Biologic evolution is the process whereby organisms have changed since life originated - that is, they have descended with modification from their remote ancestors.

Plant and animal breeders practice **artificial selection** by selecting those traits they deem desirable and then breeding plants and animals with those traits, thereby bringing about a great amount of change (Figure 7.3).

What is the Significance of Natural Selection?

We can summarize the salient points on **natural selection**, a mechanism that accounts for evolution, as follows:

1. Organisms in all populations possess heritable variations for example, size, speed, agility, visual acuity, digestive enzymes, color, and so on.
2. Some variations are more favorable than others; that is, some variant types have a competitive edge in acquiring resources and/or avoiding predators.
3. Those with favorable variations are *more likely* to survive and have more offspring, thereby passing on their favourable variations.

The central concept of modern biology is that living species have come into being as a result of the evolutionary transformation of quite different forms of life that lived long ago. Indeed, it is often maintained that very little of what is now known about life would make sense in any context other than that of organic evolution. It is important to remember, however, that while the broad definition of evolution is “change,” organic evolution does not encompass every kind of biological change. The term refers only to changes in populations, which consist of groups of individuals that live together and belong to the same species.

Adaptations

Adaptations is the specialized features of animals and plants that perform one or more useful functions. Each individual organism possesses many adaptations that function together to equip it for its particular way of life. In other words, a species may develop a useful new feature with which to perform a function, but the evolution of the feature will sometimes be constrained by the structure of the ancestral organism.

Evolution can operate only by changing what is already present. The action of evolution, i.e. is remodelling rather than new construction.

Darwin Evidences of Evolution:

1. **Geographic evidences:** Darwin read Charles Lyell's ***Principles of Geology*** during the voyage of the *Beagle* and became convinced that the **uniformitarian** approach to geology was valid. While his adherence to the uniformitarian view of Earth's history provided a framework for his acceptance of evolution, it was his observation of the geographic distributions of living things that ultimately led him to conclude that many different forms of life possess a common biological heritage. See the example of (Figure 7-7).
2. **Anatomical evidences:** He weighed other evidence indicating that one type of organism has evolved from another, he found that certain anatomical relationships seemed to build an especially compelling case. One such piece of evidence was the remarkable similarity of the embryos of all vertebrate animals. Darwin also recognized a different type of evidence that pointed to the validity of biological transformation in nature. Darwin saw no reason why animals should be anatomically straitjacketed in nature. The question was, what could bring about such biological changes under natural conditions? This question led to Darwin's second great contribution. The **first**, of course, had been his amassing of an enormous amount of evidence indicating that species had evolved in nature. The **second** was his conception of a mechanism through which evolution could have taken place. The mechanism Darwin proposed was **natural selection**—a process that operates in nature but parallels the *artificial selection* by which breeders develop new varieties of domestic animals and plants for human use.

Rates of Origination

One unique contribution of fossils to biological science is the ability they afford us to assess rates of evolution and extinction. It is only through data derived from the fossil record, for example, that we have been able to measure the rates at which new species, genera, and families have appeared and disappeared within large groups of animals and plants.

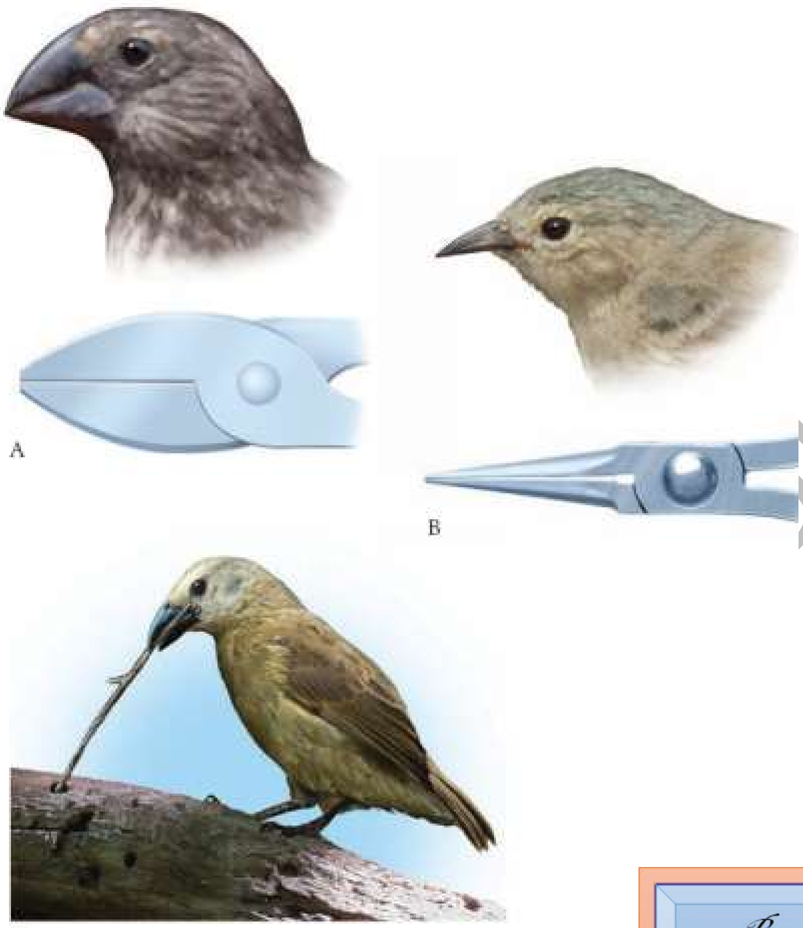

FIGURE 7-7 **Three of the finch species that Darwin observed in the Galápagos Islands.** A. The large tree finch's parrotlike beak operates like heavy pliers to crush fruits and buds. B. The warbler finch's beak operates like needle-nose pliers to catch insects. C. The woodpecker finch, which excavates tree bark with its chisel-like beak, uses a cactus needle as a tool to probe for insects.

Evolutionary radiations are the undergoing of remarkably rapid evolutionary expansion which have produced many new genera or species during brief intervals of time. The word *radiation* refers to the pattern of expansion from some group of ancestral adaptive traits to the many new adaptive traits represented by the descendant taxa.

Evolutionary Convergence as a group of organisms undergoes an evolutionary radiation, some of the taxa that arise may come to resemble taxa that evolved separately, in other radiations. **Evolutionary convergence**—the evolution of similar forms in two or more different taxonomic groups—offers convincing evidence that biological form is adaptive. This principle is strikingly illustrated by the similarity between

many of the marsupial mammals of Australia and the other kinds of mammals that live in similar ways on other continents (Figure 7-15).

FIGURE 7-13 **Several examples of over 500 species of cichlid fishes that evolved in Lake Victoria, Uganda, within the last 13,000 years.** First row (left to right): *Yssichromis pyrrhocephalus*, a pelagic zooplanktivore; *Haplochromis sp.* "all red weed scraper"; *Pundamilia nyererei*, a zooplanktivore of rocky reefs; *Ptyochromis xenognathus*, a oral shelling snail eater; Second row: *Paralabidochromis sp.* "rockkribensis", an insectivore of rocky reefs; *Tridontochromis sp.*, an unknown prawn eater; *Platytaeniodus degeni*, a snail and detritus feeder from sandy and muddy bottoms; *Paralabidochromis sp.* "chilotes complex", a lobed-lip insectivore from rocky reefs; Third row: *Haplochromis cyaneus*, a picker from rocky reefs that feeds on chironomid larvae living in filamentous algae; *Lithochromis sp.* "yellow chin", a zooplanktivore from rocky reefs; *Neochromis omnicaeruleus*, a rocky reef algal scraper; *Harpagochromis thereuterion*, a semipelagic zooplankton and small fish eater living over rocky reefs; Fourth row: *Lithochromis sp.*, a recently discovered zooplankton and insect feeding species from rocky reefs; *Gaurochromis hiatus*, a demersal insectivore from muddy bottoms; Enterochromis paropius, a demersal detritivore from muddy bottoms; *Neochromis omnicaeruleus*, an orange-blotched female morph of this rocky reef algal scraper. (Courtesy of Ole Seehausen, EAWAG [www.eawag.ch], University of Bern.)

FIGURE 7-15 **Evolutionary convergence between marsupial mammals of Australia and placental mammals of other continents.** Although each of the marsupials is more closely related to a kangaroo than to its placental counterpart in the other column, these pairs of mammals have converged in body form and way of life. (After G. G. Simpson and W. S. Beck, Life, Harcourt, Brace & World, New York, 1965.)

Extinction

Fossils provide the only direct evidence that life has changed substantially over long spans of geologic time. They also offer the only rigid evidence that millions of species have disappeared from Earth, or suffered **extinction**. Extinction results from extreme impacts of the limiting factors that normally hold populations in check. These limiting

factors, as we saw before, include predation, disease, competitive interactions with one or more other species, and restrictive conditions of the physical environment. They also include chance fluctuations in the number of individuals in a population. Population declines resulting from one or more of these factors have led to the extinction of most of the species of animals and plants that have inhabited Earth; in fact, of all the species that have existed in the course of Earth's history, only a tiny fraction remain alive today.

Species have also disappeared by evolving to the point at which they have been formally recognized as different species. In this process, known as **pseudoextinction**, a species' evolutionary line of descent continues, but its members are given a new name.

Rates of extinction: Extinction rates have varied greatly within most large groups of animals and plants over the course of geologic time, and they have varied just as greatly from taxon to taxon.

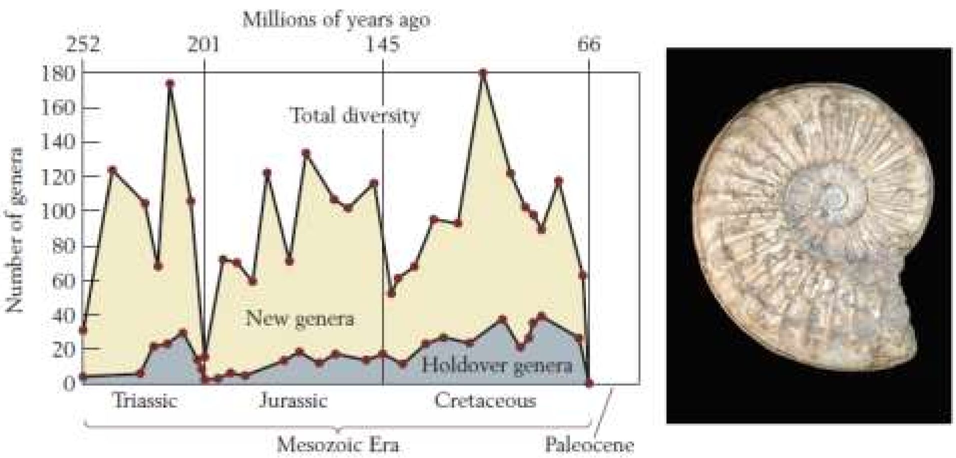

FIGURE 7-16 **The appearance and disappearance of ammonoid genera through time.** The turnover rate of genera was high throughout the ammonoids' history. Data plotted for the many ages of the Mesozoic Era show that few genera present during one age were still present in the next. (After W. J. Kennedy, in A. Hallam, ed., *Patterns of Evolution*, Elsevier, Amsterdam, 1977; photo, Sabena Jane Blackbird/Alamy.)

FIGURE 7-18 **The Siberian tiger, a large-bodied subspecies on the verge of extinction.**

Evolutionary Trends

By examining the evolutionary history of any higher taxon that has left an extensive fossil record, we can observe long-term evolutionary trends—general changes that developed over the course of millions of years. Some of these changes affected form, but others simply affected body size. Some overall trends for taxonomic groups resulted from evolution, but extinction contributed to some by preferentially eliminating certain kinds of species to change the groups' composition.

Animals tend to evolve toward larger body size

A general tendency for body size to increase during the evolution of a group of animals is known as **Cope's rule**, after Edward Drinker Cope, a nineteenth-century American paleontologist who observed this phenomenon in his studies of ancient vertebrate animals. Numerous factors may cause a group of animals to become larger in body size as the group evolves, but all have to do with the tendency of large individuals to produce more offspring than smaller ones. Within species in which males fight for females, for example. Within other species, larger animals may produce more offspring because they are better equipped to obtain food or to avoid predators (Fig. 7-20).

This evolutionary trend cannot continue indefinitely in any animal group, however, because at some point a further increase in body size will inevitably cease to be advantageous. A four-legged animal the size of a large building, for example, could not run, or even stand, because its weight would greatly exceed the strength of its limbs. Indeed, many animals could not gather sufficient food or move efficiently if they were appreciably larger than they are.

Evolution is irreversible

A complex evolutionary transition that has resulted from several genetic changes is highly unlikely to be reversed by subsequent evolution. This principle is called **Dollo's law**, for Louis Dollo, the Belgian paleontologist who proposed it early in the twentieth century. Dollo's law reflects the fact that it is extremely unlikely that a long sequence of genetic changes in a population will be repeated in reverse order. Thus evolution occasionally produces a species that crudely resembles an

ancestor, but it never perfectly duplicates a species that has disappeared. In other words, once a species has evolved into another or has been eliminated by extinction, it is gone forever.

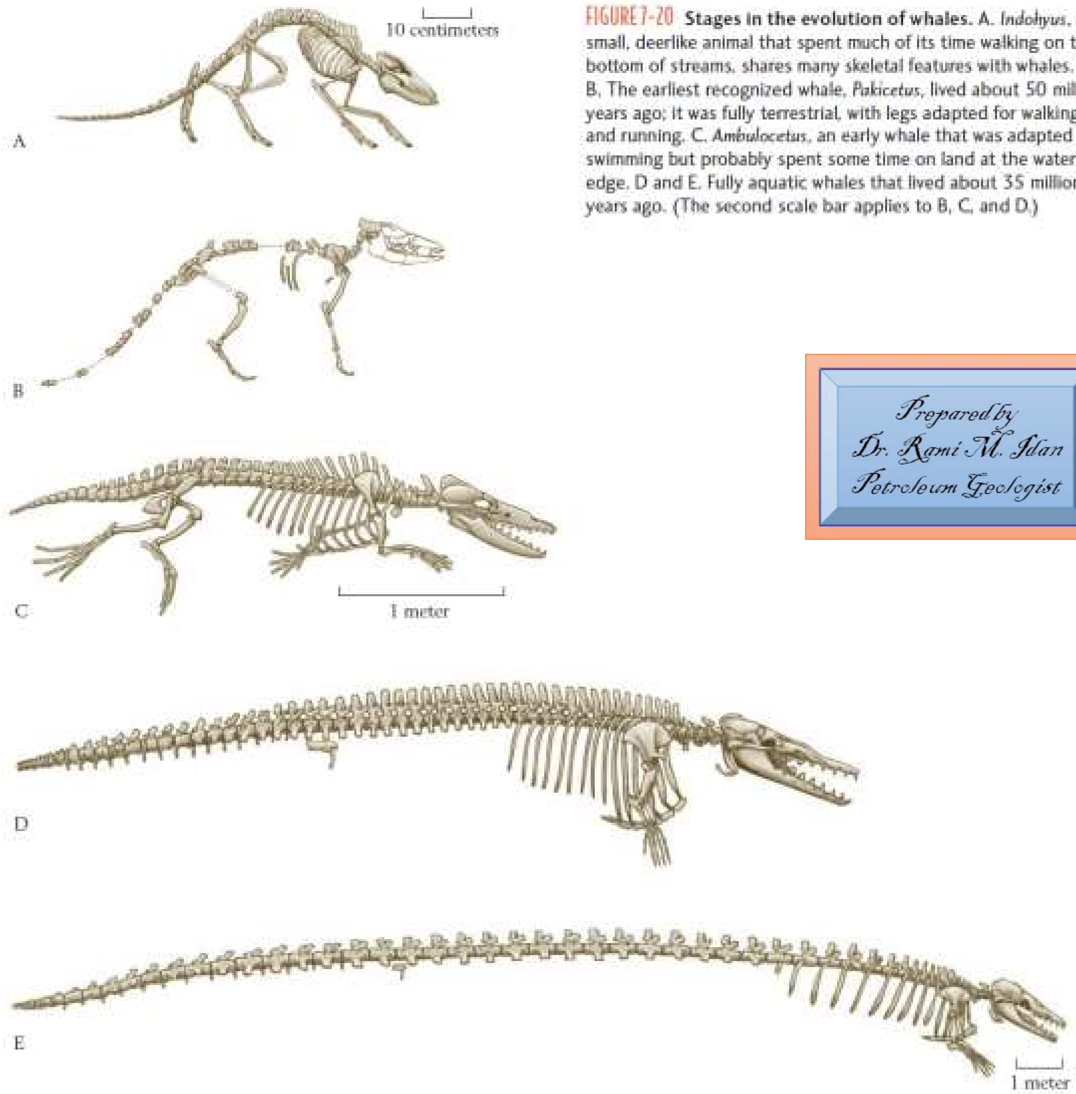

FIGURE 7-20 **Stages in the evolution of whales.** A. *Indohyus*, a small, deerlike animal that spent much of its time walking on the bottom of streams, shares many skeletal features with whales. B. The earliest recognized whale, *Pakicetus*, lived about 50 million years ago; it was fully terrestrial, with legs adapted for walking and running. C. *Ambulocetus*, an early whale that was adapted for swimming but probably spent some time on land at the water's edge. D and E. Fully aquatic whales that lived about 35 million years ago. (The second scale bar applies to B, C, and D.)

References:

- Wicander, R., and Monroe, J. S. 2016. Historical geology, 8th Edition-Books-Cole, Cengage Learning products are represented in Canada by Nelson Education, Ltd.450 pp.
- Stanley, Steven M., Luczaj, John A., 2015, Earth System History, 4th Edition, W. H. Freeman and Company, New York. A Macmillan Higher Education Company, 617 pp.

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Lecture SEVEN

Orogeny through time: an overview

(Synopsis)

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Introduction

Orogenesis is the most complex of tectonic processes and interpreting ancient mountain belts. As originally defined, an ***orogeny is simply a period of mountain building.*** To field geologists the term ***orogeny represents a penetrative deformation of the Earth's crust associated with phases of metamorphism and igneous activity along restricted, commonly linear zones and within a limited time interval***. It is eventually a mountain building event, which results from compression of continental *crust* and addition of *igneous* material to it.

Orogen: A body of *rocks* affected by mountain building.

Orogenesis: The process of mountain building.

Changing approach to the study of orogens

The major disagreements that have arisen from the study of looking at orogens in the past are documented in several works. Based on a mixture of spiritual contemplation and observation, many nineteenth century theories on the origin of mountain belts could not conceive of any large movements in the Earth to produce orogenic belts. The important concept of lateral compression gained credence as late as the middle nineteenth century. It was only in the twentieth century that Earth scientists first suggested and then established with the plate tectonic paradigm that large horizontal movements were responsible for Cenozoic orogens. As elaborated elsewhere plate tectonics unified several long-lived theories such as those of ***geosynclines*** and ***continental drift*** and reduced old-fashioned ideas such as worldwide ***orogenic cycles*** and a contracting Earth. Shortly after the acceptance of the **Plate Tectonic Theory**, modern and ancient mountain belts were analysed in terms of global tectonics. However, partisans of primary vertical tectonics resisted the plate tectonics paradigm and maintained that ancient orogens were better explained by contraction of intracontinental mobile zones (evolved from rift zones i.e. geosynclines) between stable regions emphasized differences between the Alpine and the Hercynian orogens, leading to the widely used classification

of orogens as either Alpino-type or Hercyno-type depending principally on the amount of *ophiolites*, *high-pressure metamorphic rocks* and *granites*. Application of the plate tectonic concept was however more fruitful in that geologists could show that major characteristics of an orogen (namely deformation, metamorphism and igneous activity) record stages of the plate tectonic history of the orogen, i.e. successively subduction, obduction *(the sideways and upwards movement of the edge of a crustal plate over the margin of an adjacent plate)*, collision and eventually post-collisional intra-continental deformation. Many articles and books, dedicated to the link between plate tectonics and mountain have convincingly revealed that modern orogenic belts occur principally at convergent plate boundaries and result from collision between continental, arc-derived or oceanic crustal blocks.

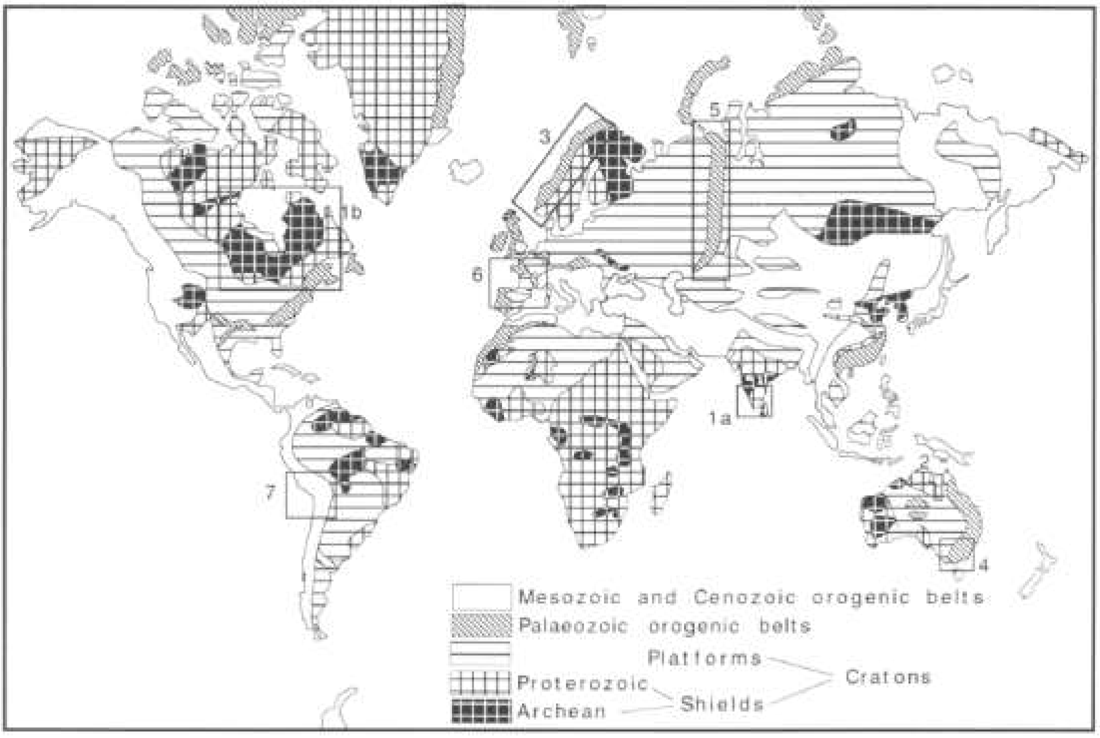

Fig. 1. World map of orogens distinguished by their age. Orogens coved in this volume are boxed. 1, Archaean orogens of Choukroune *et al.* a, Dharwar craton and b, Superior province; 2, Mount Isa Terrain, O'Dea *et al*; 3, the Scandinavian Caledonides, Milnes *et al.* and Rey *et al.*; 4, the Lachlan fold belt, Gray; 5, the Urals, Puchkov; 6, the Variscides, Rey *et al.*; 7, the Central Andes, Lamb *et al.* Adapted from Miyashiro *et al.* (1979) and Condie (1982).

Recent techniques

Since the 1960s the impact of geodynamics and geophysical data on the study of orogens has been enormous. Rapid advances in technology and increasingly powerful computers have ❶generated completely new data sets (e.g. geochronology and deep seismic profiling) which must be

reconciled with more traditional field observations and ❷allowed numerical modelling of complex Earth systems and processes whose results can be compared with and constrained by factual data.

We summarize below the more important of these modern techniques and models:

1. ***Geochronology:***

❶*Isotopic techniques* permit the dating of crystalline rocks and therefore have become a prerequisite to understanding the crystalline axes of all orogens. ❷*Radiometric dating* has also resulted in the division of large cratons into provinces surrounded by younger orogenic belts with an age progression away from the oldest central craton.

2. ***Seismicity and deep reflection seismic profiling:***

The layered configuration of stable lithospheric plates verified by seismic studies stands as a fundamental concept that rules the mode in which an orogen may evolve.

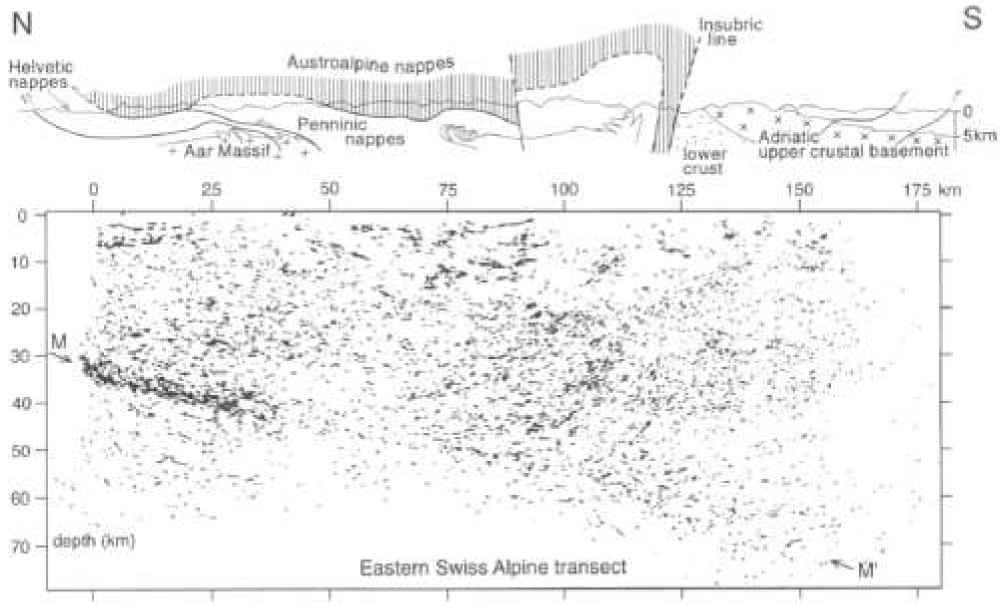

Fig. 2. Depth converted line drawing of the NFP20 deep seismic profile through the Central Alps. Adapted from Valasek (1992). The corresponding surface geology is shown above the profile. M–M' marks the trace of the European Moho which clearly plunges down below the Adriatic plate.

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3. :Rheology

The science of the *deformation* and *flow* of matter. The mechanical properties of lithospheres have been explored from inferred mineralogical stratification, temperature gradient and pressure conditions in the plate.

4. Lithospheric modelling of orogenesis:

Two modelling approaches have been used to investigate mountain building processes: numerical and analogue modelling. Both point to the fundamental control exerted on the deformation style by the strength of the crust and its coupling with the rigid or ductile mantle at its base.

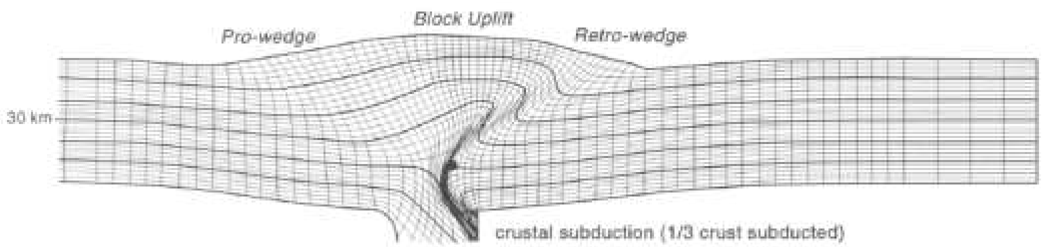

Fig. 3. Lagrangian grid showing deformation of the crust generated by a two dimensional plane strain finite element model of a small compressional orogen. Redrawn from Beaumont & Quinlan (1994, fig. 6) by kind permission of Blackwell Science, Oxford. This model represents a cold, single layer crust where one third of the crust is subducted. The crust is modelled with the rheology of wet feldspar. Decoupling of the lower one third of the left-hand crust occurs at the singularity (black spot). The model is shown after 2 Ma and 40 km of convergence. These geometries are comparable to those seen on deep seismic lines through the Pyrenees (ECORS) and the Gulf of Bothnia, Svenofennide transects (Beaumont & Quinlan 1994) and possibly with the Central Alps (compare with Fig. 2).

Orogenic Case Histories

1. **The Hercynian Orogeny**
2. **The Laramide orogeny:**
3. **Alpine Orogeny**

   The position of the continents by 50 million years ago looked quite similar to that of today. This orogeny occurred mainly between 65 and 2.5 million years ago, although it is still active today. It saw the collision of the African and Eurasian plates, and the closure of the Tethys Ocean as oceanic lithosphere was subducted northwards beneath the Eurasian Plate, leaving today what we now know as the Mediterranean Sea.

- **The reference original paper at:**
- http://sp.lyellcollection.org/content/specpubgsl/121/1/1.full.pdf
- **youtube at:**
- https://www.youtube.com/watch?v=kT6RxtWDeh0