Chapter 1

Reading rock exposures: how rock exposures contain evidence of how they were formed and subsequently deformed

1.1 Rock exposures are formed of minerals, rocks and fossils

Rocks and rock faces can look complicated, but this chapter will help you to 'read' rocks and rock faces, the places where rocks are exposed at the Earth's surface. It will introduce you to the clues rocks contain about how they were formed and what they can tell us about the Earth in the past.

You will find that many of the descriptions in this chapter, and elsewhere in the book, use the words 'usually', 'mostly' and 'generally'. This is because the the world is a complex place and the world of Earth science is complex too. So, while we can identify important principles that can be applied most of the time, there are often specific instances when they don't apply. This means that the Earth scientist has to be wary of applying rigid rules, but needs to look at all the evidence before interpreting a rock, a process or an environment (Figure 1.1). The more experience you have of this approach, the easier it becomes. This book will give you the basic outline of 'how the Earth works' and help you to use the evidence you can find for yourself to see 'how the Earth works'

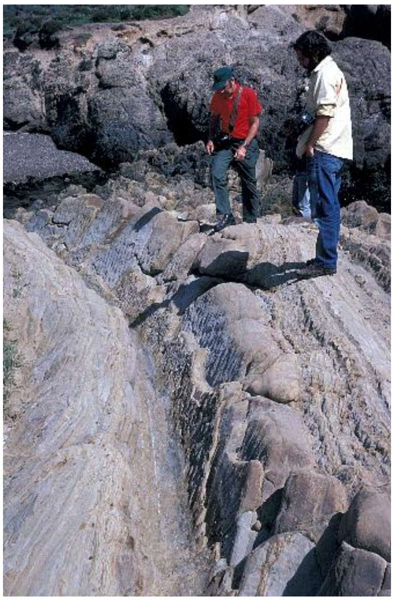


Figure 1.1: Studying a rock exposure.

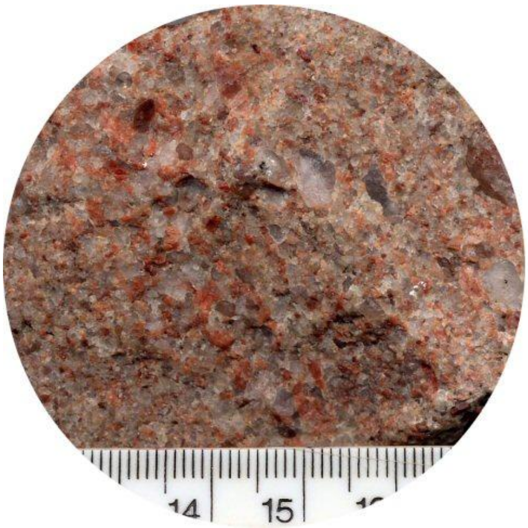


Figure 1.2: A sandstone rock - made of minerals.

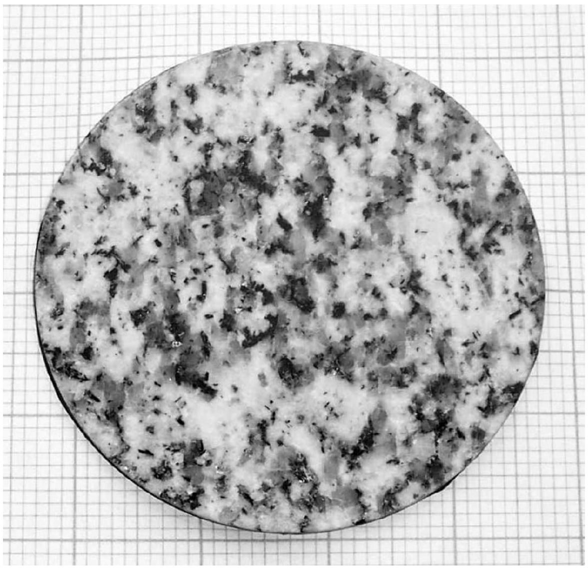


Figure 1.3: A granite rock - made of minerals.

for the different processes, rocks and landscapes you want to study.

If you look closely at a rock, either as a specimen in your hand, or forming a rock face out-of-doors, you will see that it is made of many 'bits' called grains. The grains can range from the very smallest, that are much too fine to see unless you have a microscope, to the largest grains that can be bigger than a house. See figures 1.2 and 1.3 for examples. All these grains are of just three types: they are either minerals, or other pieces of rock, or pieces of organisms, called fossils. So a definition of the term **'rock'** is 'A naturally occurring material composed of minerals, fragments of rock, or fossils.'

To make sense of this definition, we need to understand what minerals and fossils are, as well as how one rock can be made of pieces of other rocks. **Minerals** can be simply defined as 'Naturally occurring inorganic compounds or elements' whilst **fossils** are 'Traces of organisms preserved in rocks (usually more than 10,000 years old)'. But these definitions just tell you what they are; we need much more insight into how they form and what they can tell us, if we are to be able to 'read the rocks' successfully.

The chemical composition of a rock depends on the minerals it contains. The ways in which the mineral grains are fitted together is the **texture** of the rock. Rocks also have larger scale features, called structures, that will be discussed later.

1.2 Minerals are formed in a number of geological environments

Minerals are naturally formed elements and compounds and, as the first Chapter of 'Basic Books in Science' Book 5, 'Atoms, Molecules and Matter: The Stuff of Chemistry', tells

Figure 1.4: A crystal of the mineral diamond.

Figure 1.5: Crystals of the mineral calcite.

us, everything is built of atoms. A collection of atoms of just one kind is called a chemical element. When atoms combine together they form molecules and molecules of different chemical elements combined together are called compounds.

An example of a mineral formed of one chemical element is diamond, which is made entirely of the element carbon (designated: C). **Diamond** (Figure 1.4) is of course a rare mineral, which is why it is so expensive. A much more common form of carbon in the Earth's crust is the molecule calcium carbonate (formed of one atom of carbon, C, with one atom of calcium, Ca, and three atoms of oxygen, O, designated as $CaCO_3$). The most common form of the compound calcium carbonate in the crust is the mineral **calcite** (Figure 1.5).

The pictures show that both these minerals have clear shapes, called crystal shapes. This is because atoms of chemical elements in all minerals are bonded together like building blocks to form an atomic structure, and the atomic structure of each mineral is different. 'Bonds' are the forces that join the atoms of chemical elements together. In diamond, the carbon atoms form a symmetrical three-dimensional crystal structure with strong bonding in all directions, which is why diamond is so hard. However, in calcite, the atomic structure is less symmetrical, and so the shapes of the crystals are different. As some of the bonds are weaker, calcite is not as hard as diamond and can break much more easily.

In the natural world, chemical compounds are not as pure as they might be in a chemistry laboratory, so that minerals often contain traces of other elements which change their structure and properties. Thus a more complete definition of a mineral is: 'A naturally occurring inorganic compound with a definite chemical composition, a definite atomic structure, and physical properties which vary between known limits'.

One of the physical properties which quite often varies is colour, particularly in the paler-coloured minerals so that, whilst calcite is usually colourless or white, it can have grey, yellow, blue, red, brown or black tints, depending on which trace elements it contains.

Even diamond can have different colours, although colourless diamonds are usually the most valuable.

We use the properties of minerals to identify them, and the most useful properties are their colour, which depends on their chemical composition, and their shape, **hardness**, and the way they break (**cleavage**), which depend on their atomic structures. Certain minerals also have other properties, which help us to identify them. We can use how heavy they feel (their density), their surface appearance (their lustre) or the colour of their powder (their **streak** left as a scratch on a white tile), whilst some react with acid, are soluble or are magnetic.

Common minerals that you can identify using these properties are listed at the end of this section.

Minerals form in only five common ways and you can usually use the clues they contain to find out how they crystallised. They form by:

- crystallising from molten rock as it cools
- recrystallising due to increases in heat and/or pressure
- crystallising from evaporating water
- crystallising from liquids flowing through the pores in rocks
- crystallising from hot fluids that cool as they flow through rocks

1.2.1 Igneous rocks

When rock becomes very hot, more than 600°C, the minerals begin to melt and nearly all minerals have melted by a temperature of 1800°C. As the minerals melt, the atoms and molecules they contain are released to form a 'bath' of liquid called **magma**. This liquid mixture of atoms and molecules is usually less dense than the surrounding rock and so tends to rise. As magma rises, it cools down, so that minerals begin crystallising again. As cooling continues, the first crystals to form become larger, as more molecular building blocks come together and the atomic structures grow. The longer the liquid has to cool, the larger the crystals become. Eventually all the liquid crystallises and the rock has become a solid mass with a texture of randomly-orientated interlocking crystals. Rocks formed by crystallising from magma are called **igneous rocks**. Most magmas crystallise underground, but if magma flows to the surface it is called **lava**. So solidified lavas are also igneous rocks.

The most common minerals found in igneous rocks are quartz, a compound of silicon and oxygen (SiO_2) with a simple but well-bonded atomic structure, and feldspar, another silicon/oxygen (silicate) compound but with extra aluminium, sodium, potassium and calcium, and with an atomic structure that is nearly as well bonded as quartz. A third common mineral is mica, another silicon/oxygen compound with extra elements added; this is poorly bonded, particularly in one direction, and so has a strong cleavage, is soft and layers are easily broken off. In igneous rocks, quartz is grey, feldspar is white or

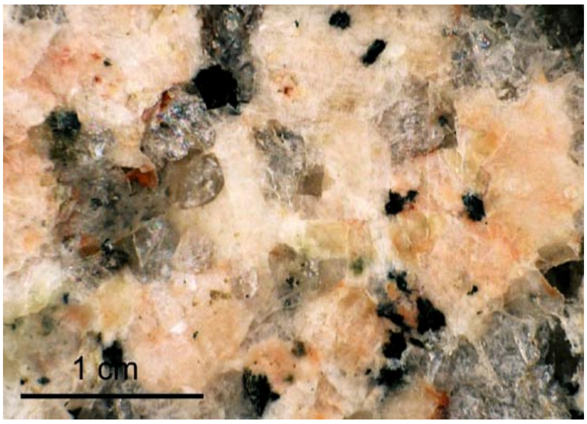


Figure 1.6: A close up view of a piece of the igneous rock, granite.

Figure 1.7: A close up view of a piece of the metamorphic rock, gneiss.

sometimes pink, and mica is black or colourless. You can see all these minerals in the photograph of granite (Figure 1.6), which also shows how the minerals have crystallised to fill all the space, in a tough interlocking texture.

1.2.2 Metamorphic rocks

When rocks are heated or come under great pressure, the minerals in them can be recrystallised without melting and this recrystallisation process is called **metamorphism**. During metamorphism the sizes and shapes of the original minerals can be changed or original minerals can be changed into new minerals. The result is a tough rock with an interlocking texture. If the metamorphism is caused by high temperature alone, then the new metamorphic minerals have random orientations. However, if the metamorphism is caused by high temperatures and pressures in the roots of mountains during mountain-building episodes, the metamorphic minerals recrystallise at 90° to the direction of the pressure. In high pressure/temperature metamorphic rocks, many of the minerals are therefore aligned into a rock texture of sheets or bands, as seen in the photograph in Figure 1.7.

The minerals quartz, feldspar and mica found in igneous rocks are also common in metamorphic rocks, but a mineral that is not frequently found in other rocks, but is common in higher grades of metamorphic rock, is garnet. **Garnet** (Figure 1.8) is another silicate mineral, where the silicon/oxygen has combined with aluminium and calcium, iron or magnesium. It has a strongly-bonded structure making it a hard mineral. Due to this hardness and its pleasing red or pink colours, it is sometimes used as a semi-precious gem mineral.

If the metamorphosed rock was originally made of calcium carbonate (i.e. limestone), then the rock produced by metamorphism is also composed of the calcium carbonate mineral, calcite. The new rock, formed of interlocking calcite crystals, is called **marble** (Figure 1.9).

Figure 1.8: A red garnet crystal in a metamorphic rock.

Figure 1.9: Marble, a metamorphic rock made of interlocking calcite crystals.

Figure 1.10: Salt deposited by evaporation of a drying lake.

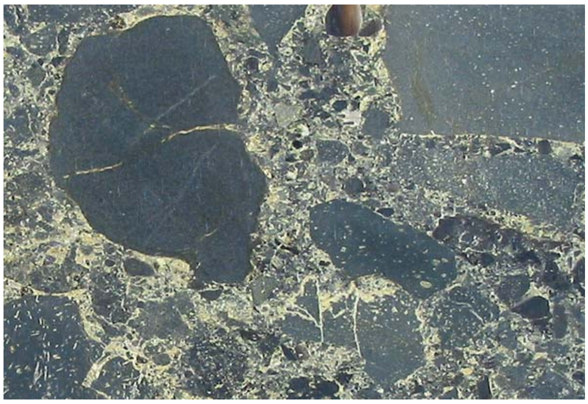


Figure 1.11: A sedimentary rock, showing the fragments 'glued' together by a natural cement.

1.2.3 Evaporites

When salt water evaporates as lakes or arms of the sea dry out, salts crystallise out as deposits that cover the old lake or sea bed (Figure 1.10). These are called **evaporite deposits** and the most common of these is the compound of sodium and chlorine, **halite**. The sodium chloride (NaCl) molecules are weakly bonded, so halite is a mineral with a very low hardness. The sodium chloride bonds are broken in water so that halite is soluble and easily dissolves again. Because of this solubility and its salty taste, a refined form of halite is used as table salt in many of the foods you eat. If you want to crystallise your own salt crystals, see the 'Salt of the Earth: Who can make the biggest salt crystal?' activity on the `http://www.earthlearningidea.com` website.

1.2.4 Sedimentary rocks

Sedimentary rocks are made of sediments which are fragments of other rocks or fossils, although sometimes sediments can also be laid down by the evaporation of water. Often the grains of sediment are single crystals which have been broken and worn down. These have often become **rounded** during transportation and **sorted** into grains of similar sizes. After they have been deposited, water flows in the **pore spaces** between the grains. This slow-flowing water dissolves minerals from some parts of the deposit and they recrystallise between the grains in other parts of the deposit, 'gluing' the grains together. This natural 'glue' is called **cement** and changes soft sediments into harder sedimentary rocks (Figure 1.11).

The most common cementing minerals in sedimentary rocks are quartz and calcite. As the pore spaces become filled with cement, the tougher the rocks become. Try the 'Make your own rock' activity on the `http://www.earthlearningidea.com` website to find out how this cementing process works.

1.2.5 Veins and ores

Sedimentary rocks are cemented by cool fluids. However, magmas can produce hot watery fluids, whilst in other areas where rocks are heated, the water in the rocks can dissolve minerals. This watery hot liquid containing dissolved minerals is a **hydrothermal fluid** that, being less dense than the surroundings, rises through pore spaces and cracks in the rock. As it rises it cools and minerals crystallise out of solution in the pore spaces or as coatings on the sides of the cracks. Mineral-coated cracks are called mineral veins. These can be thin sheets that fill a fracture such as a joint in a rock, or more complex sheets filling faults or other fractures and spaces.

Quartz and calcite commonly crystallise in mineral veins, but sometimes more unusual minerals can form (Figure 1.12). If mineral deposits contain metal minerals and the deposits are rich enough to be worth mining, the deposits are called **metal ores**. **Ore minerals** include the red iron oxide (Fe_2O_3) mineral hematite, and the shiny grey, lead sulfide (PbS) mineral galena. In deposits of ore minerals, the uneconomic minerals that have to be thrown away, like quartz and calcite, are called **gangue minerals**.

Figure 1.12: Crystals of minerals in a mineral vein.

Figure 1.13: The red ore mineral hematite with the colourless gangue mineral, calcite

The most valuable minerals are the precious gemstones like diamond, which are hard and rare. Diamond is found in certain volcanic deposits, or has been eroded from them and deposited in rivers and beaches. Semi-precious stones which are fairly hard and rare include garnet and the coloured forms of quartz found in some mineral veins, like purple amethyst quartz.

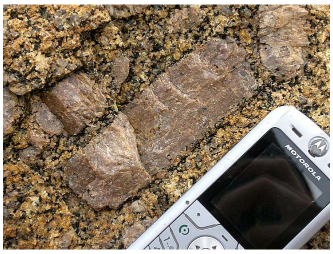
Metallic minerals like hematite (Figure 1.13) and galena, can form valuable ore deposits, but one of the most valuable metal minerals is gold. Gold is an un-reactive element that rarely combines with other elements and so is usually found on its own as native gold. **Native gold** can be found in mineral veins or, like diamond, eroded from veins and deposited in sediments.

Minerals can be identified by their properties; some important minerals are shown in figures 1.14 and 1.15.

Calcite, $CaCO_3$ - usually white or colourless, can form good 'dog tooth' crystals, good cleavage, fairly low hardness, reacts with dilute acid, is the main mineral in marble and in some sedimentary rocks (limestone) and is found as a gangue mineral in veins

Diamond, C - usually colourless, forms good crystals, extremely hard, rare

Feldspar, silicate - white or pink, can form good rectangular crystals, good cleavage, hard, common in igneous and some metamorphic rocks



Garnet, silicate - usually red or pink, often forms ball-shaped crystals, no cleavage, hard, found mainly in medium and high grade metamorphic rocks

Gold, Au - gold colour, usually irregular shape, feels very dense, low hardness (soft), rare. The picture shows a very unusual example of a large gold nugget, rounded because it was found in stream sediment.

Figure 1.14: Some important minerals and their identification properties

Hematite, Fe_2O_3 - earthy red, metallic lustre, feels dense, red streak, usually hard, often found in irregular masses. The picture shows its reddish colour and metallic lustre

Halite, NaCl - white or pink, cube-shaped crystals, good cleavage, low hardness (very soft), soluble in water, forms rock salt deposits. The picture shows pink halite crystals in a rock salt deposit.

Galena, PbS - silvery grey, metallic lustre, cube-shaped crystals, good cleavage, feels very dense, grey streak, low hardness (soft), found in mineral veins; the picture shows silvery-grey cubic galena crystals with their metallic lustre, with a pale gangue mineral.

Mica, silicate - usually colourless or black, forms platy crystals, good cleavage in one direction, low hardness (soft), common in igneous and some metamorphic rocks

Quartz, SiO_2 - usually grey, white or colourless, can form good hexagonal crystals, no cleavage, hard, common in sedimentary, igneous and metamorphic rocks and as a gangue mineral in veins

Figure 1.15: Important minerals - 2

1.3 Sedimentary rocks - formed by a range of surface processes in a variety of environments

You can 'read' sedimentary rock exposures by understanding that these rocks were once loose sediments that have been cemented or compressed into sedimentary rocks. They therefore contain the same clues that sediments do, to the ways in which they were originally moved and laid down. These clues include the size, shape and chemical makeup of the grains, the layers these grains form, like the cross bedding in Figure 1.16, and the fossils and overall sequences that the rocks contain. Since most sedimentary rocks have layers, called **beds** (see Figure 1.17), you can usually use this to distinguish them from igneous and metamorphic rocks. The good 'rock detective' can read sedimentary rock clues of composition (chemical makeup), texture (grain size, shape and arrangement) and **structure** (including structures like bedding and **cross bedding**) to discover the environment in which the sediments were first deposited.

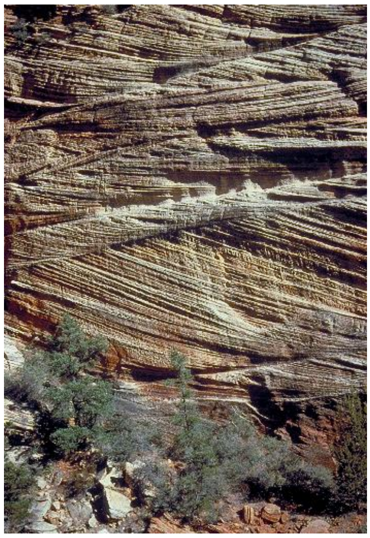


Figure 1.16: A sedimentary rock face showing some of the clues into how the sediments were originally laid down.

Sediments are generally eroded from higher land and deposited in lower areas. Such lower areas include puddles, ponds, lakes, gutters, streams, rivers, valleys, general lowland areas, coastlines and shallow and deep seas. These lower areas span a range of latitudes, from polar to temperate to arid and tropical. Deposits of boulders, gravel, sand or mud can be laid down at all latitudes but calcium carbonate deposits (limestones) and evaporites are formed mainly in the tropics.

Most sediments are formed from the weathering and erosion of rocks, which could be of sedimentary, igneous or metamorphic type. Deposits near these source rocks have not been moved very far and so the make up of the sediment is similar to the mineral composition of the source rocks. If the source rocks contained quartz, feldspar and mica, so will the sediment. Similarly, there has been little chance for the sediment to become sorted into different grain sizes or for the grains to be worn down. Thus, sediments near source rocks usually have mixed compositions and mixed sediment sizes, whilst the individual sediment fragments have sharp corners and are angular, as in Figure 1.18, which shows a **breccia** (a sedimentary rock of mixed sizes with angular grains).

Gravity moves these sediments downhill and they become picked up by water and moved along streams and into rivers. Minerals like mica, and later, feldspar, become broken down. The corners of individual grains are worn away and they become rounded and reduced in size. The abrasion of the grains to smaller sizes and rounded shapes is called **attrition**. As rivers and later, coastal currents, carry the grains along, they are sorted into different sizes, since fast flows can carry large grains whilst small grains only settle

Figure 1.17: Sedimentary rocks showing their layers or beds. The surface of each bed is called a bedding plane.

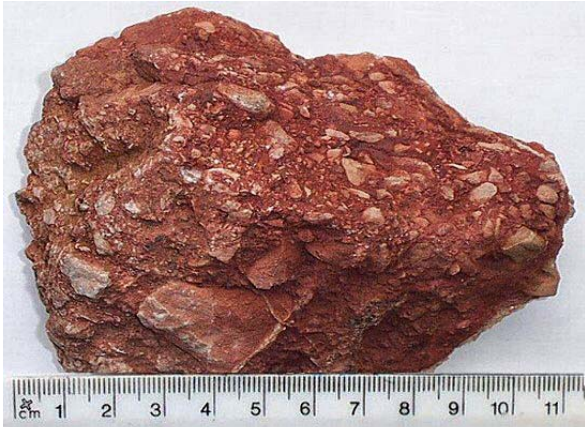


Figure 1.18: A sedimentary rock made of angular fragments, a breccia.

in quiet conditions. So, we find boulders, pebble deposits, sand deposits and deposits of mud in different parts of rivers and coastal areas, depending upon the speeds of current flow. The proportion of quartz in sediments increases as they are carried along, since mica and feldspar are broken down physically and chemically. The chemical breakdown of mica and feldspar produces **clay minerals** so there is a high proportion of these very fine-grained minerals in quiet areas of deposition.

So, we can use the sizes, shapes and compositions of the grains to tell us about how the grains were transported and deposited. Poorly-sorted angular sediments of mixed compositions are found near source areas, whilst well-rounded gravel deposits, well-sorted, quartz-rich sands, and muds rich in clay minerals, are found far from their sources. The gravels can become cemented into **conglomerates** (Figure 1.20), the sands into **sandstones** (Figure 1.21) and the muds can later be compressed into **mudstones** (Figure 1.19) and **shales** (shales are weak mudstones that tend to fall apart in your hand).

Deserts are mostly very dry but sometimes have torrential storms, so we can find a wide range of deposits. There are accumulations of angular fragments that could become breccias, whilst dried rivers can have pebble and sand deposits that could become conglomerates and sandstones (Figure 1.22). Dried up lake beds can have muds and salt deposits (as in Figure 1.23) that could be preserved as mudstones with evaporite layers. These deposits often contain more clues as well.

Most of the sands have layers that will become sandstone bedding. When sands are deposited in fast flowing rivers, they are usually laid down in small underwater **dunes** as sloping layers that are seen as the cross beds in many sandstones. Since the cross beds always slope downstream, they tell you the direction of the current flow that laid down the sand (as in Figure 1.24). As water currents slow down, the surface of the sand often forms into current ripple marks that have a shallow up-stream slope and slope more steeply downstream. These **asymmetrical ripple marks** or **current ripple marks** tell you the direction of the current that deposited them and these can be preserved in

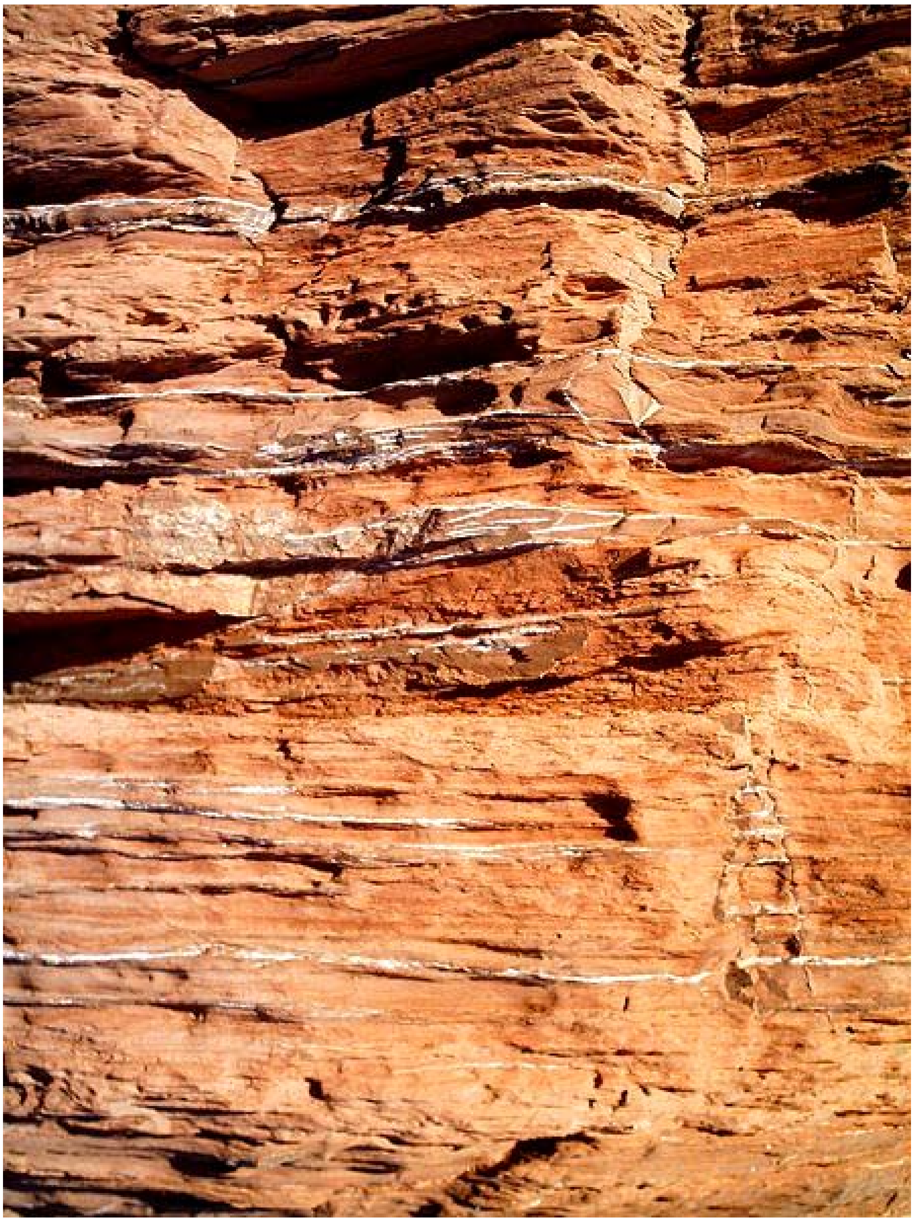


Figure 1.19: The fine grained sedimentary rock mudstone, in a cliff face.

Figure 1.20: A close up view of a piece of conglomerate, a sedimentary rock made of rounded pebbles cemented together.

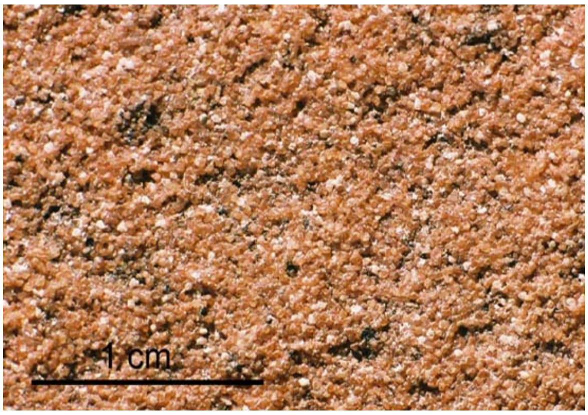


Figure 1.21: A close up view of a piece of sandstone, made of cemented sand grains.

Figure 1.22: The angular fragments of sediment in a desert that could become a breccia.

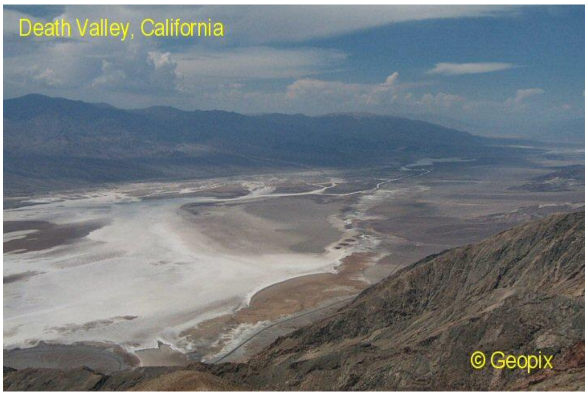


Figure 1.23: Dried up lake bed with white salt deposits in a desert environment - Death Valley, California.

rocks. Many of these desert deposits are red, because the weathering processes there tend to concentrate red hematite on the surfaces of sediment grains. You can investigate how water currents move and deposit sand using the 'Mighty river in a small gutter: sediments on the move' activity from the http://www.earthlearningidea.com website.

Each time mud settles out from quiet water, in a desert lake, for example, it forms a layer. These layers build up into series of **laminations**. If the mud dries out the surface shrinks and it breaks into **mudcracks** (desiccation cracks) with polygonal shapes (Figure 1.25). Since mudcracks only develop in dried mud, mud cracked mudstones show that the mud could not have been deposited in the sea - which would not dry out. The mud sometimes carries other clues that it dried out, such as small pits made by raindrops or footprints of the creatures that lived there at the time. If salt was deposited on the drying lake floor, the mud can preserve the shapes of the cube-shaped crystals. The next mud layer can make casts of these shapes, preserving the shapes of the salt crystals in mud. All these clues of an environment that was once wet but dried out, can be preserved as the mud becomes mudstone.

Figure 1.24: Cross bedding in sandstones. The water flow direction here was from left to right

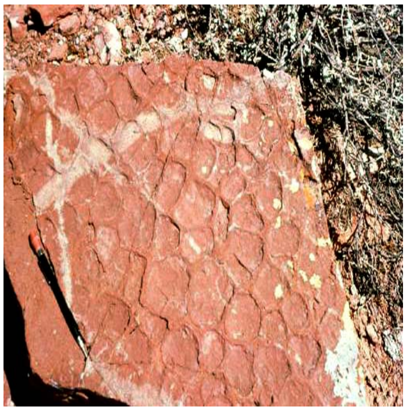


Figure 1.25: Ancient mud cracks preserved in mudstone.

If desert lakes have sandy floors then waves can form **symmetrical ripple marks** in the sand in the same way as they do on beaches and in shallow seas (see Figure 1.26). Unlike current ripple marks, these **wave ripple marks** are symmetrical, with an equal slope on either side. They are usually straight as well, and are parallel to the waves that formed them so, since waves are parallel to shorelines, they can show you the direction of the shoreline of the lake or coast when the ripples formed. Try making your own ripple marks using a washbowl (current ripples) or a tank (wave ripples) as shown on the http://www.earthlearningidea.com website.

Windy deserts are famous for wind-formed sand dunes. The winds were not strong enough to pick up larger particles, so they formed the sand into dunes, and blew any mud-sized particles out of the area. The sand is therefore well-sorted and has become rounded and quartz-rich as it was blown along (Figure 1.27). The grains often have red hematite-stained surfaces. Sand is laid down by the wind on the sloping fronts of the dunes in large cross beds, much larger than the cross beds found in rivers. Preserved dune cross bedding can be a metre or more high. So, desert dune deposits have well sorted, red-coloured quartz sands deposited in large sets of cross bedded sandstone (see Figure 1.28).

Figure 1.26: Ancient wave ripple marks preserved in sandstone.

Figure 1.27: Modern sand dunes, made of well-sorted sand deposited in sloping layers on the front of the dune - Death Valley, California.

Figure 1.28: Ancient dune cross bedding preserved in a sandstone cliff face.

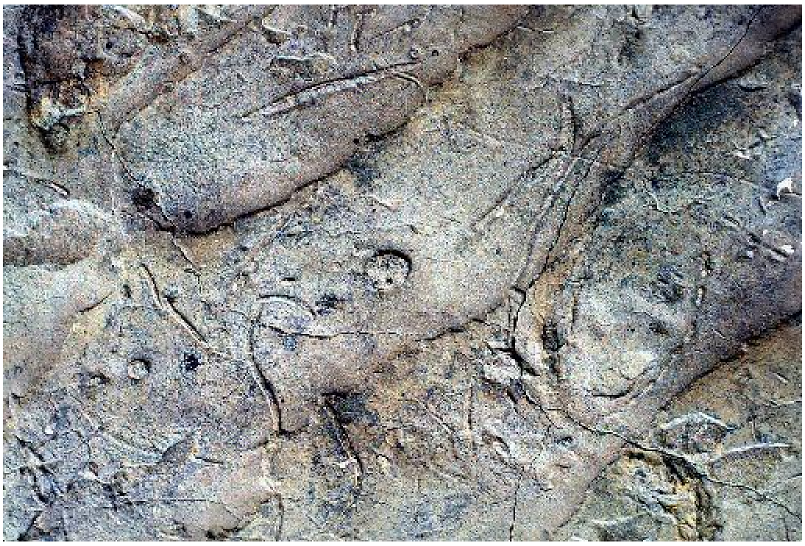


Figure 1.29: Tracks, trails and burrows preserved on a ripple-marked sandstone.

In coastal regions, gravel is deposited on shingle beaches, sand is laid down in sandy beaches and on much of the shallow sea floor, and mud is deposited on tidal flats. Beach gravel is usually easily distinguished from other sorts of gravel deposit because the pebbles have become well-rounded by the continuous pounding of waves. Beach and shallow sea sands can also be distinguished from other sorts of sand because, although they can also have bedding, cross bedding and wave ripple marks, they contain much fossil debris too, because coasts and shallow seas have great variety of life and many of the organisms have hard parts that can become fossilised. Similarly, although coastal muds can dry out and have mudcracks, footprints and rain pits, they also contain abundant life. The evidence of life can be preserved in tidal muds, not only as fossil fragments, but also as whole fossils and burrows, feeding tracks and trails (as shown in Figure 1.29).

When rivers that are loaded with sediment reach the sea, the sediment often builds out into the sea as a delta. The delta is mostly formed of sand but mud is deposited on the

Figure 1.30: Coal seams in an opencast coal quarry.

edges of the delta in the deeper regions. If the delta is in a tropical region, the delta top becomes covered in vegetation in the swampy tidal conditions. When the swamp trees die through natural life cycles, they fall and build up into thick sequences of organic material. These are the conditions in which coal forms and there are many examples in the geological record of delta sediments with coal deposits on the top. The muds of the foot of the delta are preserved as shale, above this are the thick sands of the delta itself, often deposited in large cross beds. On top of the delta sands are coal deposits, found as coal seams (see figure 1.30). Coal mining industries across the world are based on deposits like these.

Shallow tropical seas have some of the greatest variety of life on Earth and many of the organisms that live there, from microscopic algae to giant clams, have hard parts made of calcium carbonate (often calcite, but also a different form of calcium carbonate, called **aragonite**). When the organisms die, this material builds up on the sea floor as deposits that will become **limestones**. These calcium carbonate deposits are often rich in fossil debris including broken shells, pieces of coral and other carbonate debris. Since most types of coral are only found in tropical seas today, if they are preserved, we know that the environment must originally have been a tropical shallow sea. The photograph (Figure 1.31) shows a limestone made of fragments of crinoid, another animal that was common in some ancient shallow seas. Since calcium carbonate reacts with dilute acid, it is easy to distinguish limestone from other sedimentary rocks, by the one-drop **acid test**.

In shallow tropical sea areas where there is strong evaporation, another type of limestone can form. It is made of spherical grains of carbonate sand that are rolled around by the waves and tidal currents. The carbonate grains grow as the water evaporates and tiny crystals of aragonite crystallise onto the surfaces of the tiny balls. The balls are called

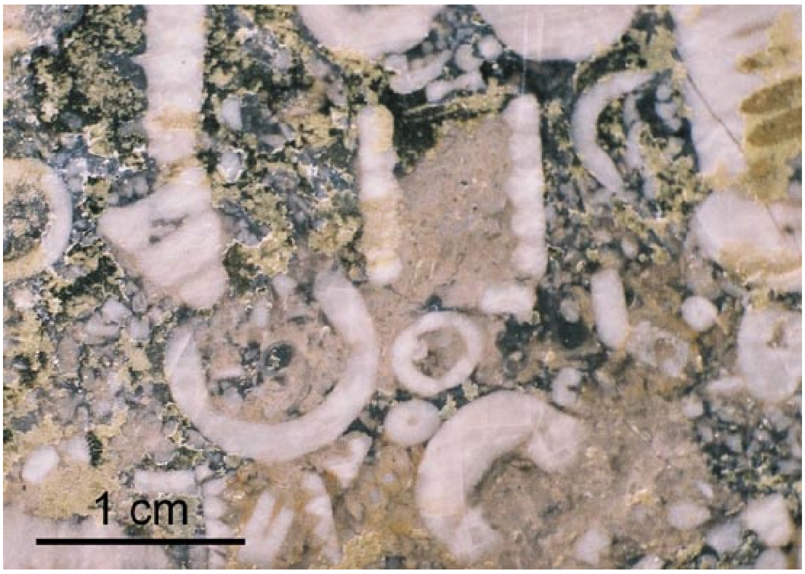


Figure 1.31: Close up view of a piece of limestone, formed of fossil fragments cemented together.

oolites and, when cemented into rock, become **oolitic limestone** (see Figure 1.32). You can easily distinguish limestones from other sedimentary rocks since, being calcium carbonate, they react with dilute acid. So the one-drop acid test causes violent fizzing.

The only sediment that is normally deposited in deep sea areas is fine-grained mud. Most of the mud was originally fine-grained sediment blown from deserts or was brought into the sea by rivers. This mud contains the remains of tiny organisms that live as plankton near the surface of the ocean and sink to the bottom when they die. The mud builds up on the deep sea floor very slowly and, when it is compressed by the overlying layers, it becomes mudstone or shale. However, where rivers bring large amounts of sediment into the sea, in deltas for example, this sandy and muddy sediment builds up on the edge of the ocean basin. An earthquake can trigger a slide of this material, which flows down into the deep sea as a billowing cloud of muddy sediment called a **turbidity current**. Turbidity currents can flow at more than 100 km per hour across thousands of square kilometers of deep ocean floor depositing flat sheets of sediment called **turbidites**; there is an example of one in Figure 1.33. Each turbidite has the coarser material, usually sand, at the bottom and becomes finer grained upwards to the mud at the top. This is a **graded bed**, coarser at the bottom and becoming finer upwards. Usually many turbidite layers build up into thick turbidite sequences on the deep ocean floor as stacks of graded beds of different thicknesses.

In cold areas, such as polar regions and high mountains, ice can erode, transport and deposit sediments. Since the ice carries material of all sizes and the particles do not become sorted or rounded as they are carried along by the ice, deposits of melting ice are easily recognised by their mixture of boulders, sand and clay. This ice-deposited mixture is called **glacial till** (Figure 1.34)and can be found across many areas of northern Europe

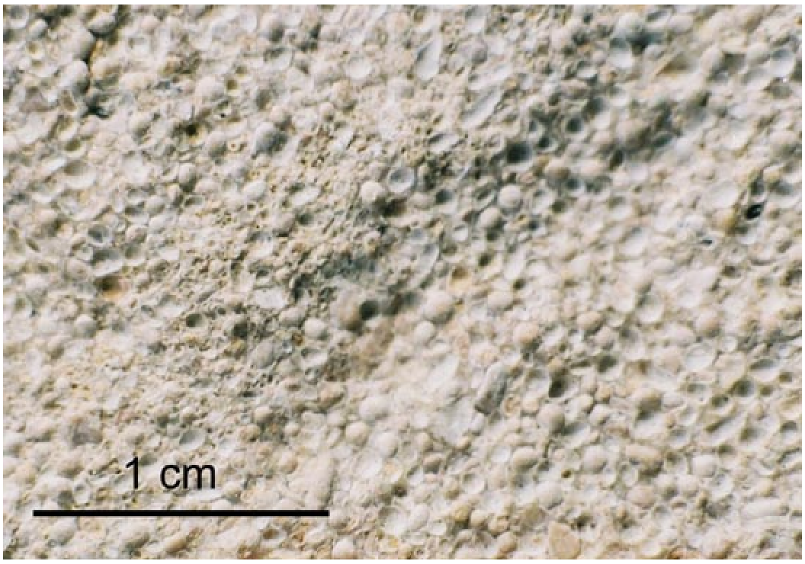


Figure 1.32: Close up view of a piece of oolitic limestone, made of tiny balls of oolite cemented together.

Figure 1.33: This sequence of turbidites has become tilted by folding of the rocks.

Figure 1.34: A melting glacier depositing glacial till.

and America, showing that these areas were covered by ice sheets in the past. Other clues to past glaciation include glacial U-shaped valleys and rocks scratched by the ice dragging rock fragments across their surfaces.

The fossils that sedimentary rocks contain can be valuable in helping us to interpret the environment in which the sediments were laid down. Where possible we use the principle that the 'present is the key to the past' (the Principle of **Uniformitarianism**) to help us to understand the ancient environment. Coral is a good example of an environmental indicator, since most coral only grows in shallow tropical seas today so, if we find fossil coral, we can assume that the sediments were also deposited in shallow tropical seas (there is a picture of a fossil coral in Figure 1.35). Similarly, land plants only grow on land or in the swampy conditions of some coastal areas, so fossil land plants, such as those in Figure 1.36, show not only that the area was land, but also, if the growth was very luxuriant and produced lots of organic material (as in coal), that the area was tropical as well. Tracks, trails and footprints are important, since they tell us that there were living things in the area and so it was not too hot, or cold, or dry, or lacking in oxygen or food for organisms to live there. Since most tracks, trails and footprints are made on dried-out land or in shallow water, they provide further clues to the environment.

When animals have become extinct, we cannot use the 'present is the key to the past' principle so easily. Nevertheless, we know that the modern relatives of fossils like trilobites (Figure 1.37) and ammonites (Figure 1.38) only live in the sea and so the sediments in which they are found are most likely to have been deposited in the sea. Likewise, the modern equivalents of trilobites live in shallow waters, indicating that most trilobites also lived in shallow sea conditions.

Sediments become sedimentary rocks after they become buried, through a process called **diagenesis** that normally takes millions of years ('millions of years' is usually abbreviated to **Ma**). As the sediments become buried by more and more sediments, they become

Figure 1.35: A fossil colonial coral preserved in shallow tropical sea sediments.

Figure 1.36: Fossilised trees that must have grown on the land.

Figure 1.37: A fossil trilobite, found in shallow sea sediment.

Figure 1.38: Fossil ammonites, indicating sediments that were deposited in the sea.

compressed as the grains are compacted together and water is squeezed out. Meanwhile waters flow through the pore spaces between the grains and minerals crystallise from the water as cement. The effect of the two processes of compaction and cementation is that the rocks become harder and the porosity is reduced. **Porosity** is a measure of the amount of pore space in the rock and, even after compaction and cementation, can still be as high as 20% in some sandstones and limestones. Since the pore spaces are large enough for water and gas to flow through, these rocks are also permeable (**permeability** is a measure of how quickly a fluid can flow through a rock). Most mudstones and shales are surprisingly often very porous, but because the pores are too small to allow fluids to flow through, they are often **impermeable**, i.e. the fluids cannot flow through the rock.

The porosity and permeability that remains in the coarser grained sedimentary rocks after diagenesis is vital, because these rocks can contain underground water supplies and oil and gas, stored in the gaps between the grains. You can investigate for yourself how rock porosity and permeability work using the 'The space within: the porosity of rocks' and the 'Modelling for rocks: What's hidden inside - and why?' practical activities from the http://www.earthlearningidea.com website. These show that, if you drop water on to permeable porous rocks, it will soak in, but if a drop of water is put onto an impermeable rock, it will stay on the surface. You can also distinguish a permeable from an impermeable rock by putting them both in water. You might see a few small air bubbles on the surface of the impermeable rock but many more bubbles rise from the permeable one as air bubbles rise out of the pore spaces at the top, and water flows in to fill the spaces at the bottom.

Try to get a feel for how scientists use the clues from sedimentary rocks and fossils to interpret the environments in which the sediments were deposited by using the 'What was it like to be there - in the rocky world? and 'What was it like to be there - bringing a fossil to life' activities from the http://www.earthlearningidea.com website.

1.4 Igneous rocks - formed from molten rock by a range of processes

Rock faces made of igneous rock can usually be distinguished from sedimentary rock faces because they have no layers, and so are described as '**massive**' (Figure 1.39). Since igneous rocks are made of interlocking crystals, they are normally tougher than most sedimentary rocks and are also impermeable, so that water runs off them, unless they are cracked or fractured.

Igneous rocks are formed from once-liquid magma. Although there may seem to be a wide variety of them, they are affected by just two main variables, their chemical composition and where in the Earth's crust they cool down and solidify.

The composition of a magma depends primarily on the composition of the rock that was melted and on how much melting took place. Rocks are formed of a mixture of minerals and these have different melting points. In general, minerals rich in iron and magnesium have high melting points, while minerals rich in silicon have low melting points. As a rock is heated, it is the minerals with the lowest melting points, those rich in silicon, that melt first. This is **partial melting**, but can also be called fractional melting, since the fraction of the rock with the lowest melting points melts first. This is similar to fractional distillation, when a mixture of liquids is heated and the different fractions change from liquid to gas at different temperatures and boil off in turn. The only difference between the processes of fractional melting and fractional boiling, is that fractional melting is a solid to liquid change that depends on the melting points of the materials involved, while in fractional distillation liquid becomes gas and depends on boiling points.

Beneath the Earth's crust is the mantle and all the material that now forms the crust originally came from the mantle. Mantle rock is relatively rich in iron and poor in silicon; it is called an **ultramafic** rock ('mafic' meaning magnesium, 'ma' and iron, 'fic'). When this becomes heated, it partially melts and, since the silicon-rich minerals melt first, the melt is richer in silicon than the original mantle rock. The new magma rises and penetrates the rock above. This is a **mafic** melt, found mainly at mid-oceanic ridges, above the mantle source rock. Mafic melts cool to form mafic rocks. Where these mafic rocks become heated again, as can happen if they are carried to oceanic trenches, they partially melt, producing a magma which is even richer in silicon, called an **intermediate** magma. If, in turn, the rocks formed by intermediate magmas are partially melted, as can happen if they are carried beneath continents, a melt even richer in silicon is formed, called a **silicic** melt. This shows that the four main different compositions of magma are produced by the partial melting of the others in sequence: ultramafic rocks partially melting to mafic melts; mafic rocks partially melting to intermediate melts; and intermediate rocks partially melting to form silicic melts and rocks.

Being hot and therefore less dense than the surrounding rocks, magma rises and intrudes into the rocks above. If it doesn't reach the surface but cools down beneath the surface, it cools slowly, since the rocks above are very efficient insulators. When magmas cool slowly there is time for the crystals to grow and so slow-cooling, deep magmas have large crystals. These are **coarse-grained igneous rocks** with easily-visible crystals. Magmas nearer

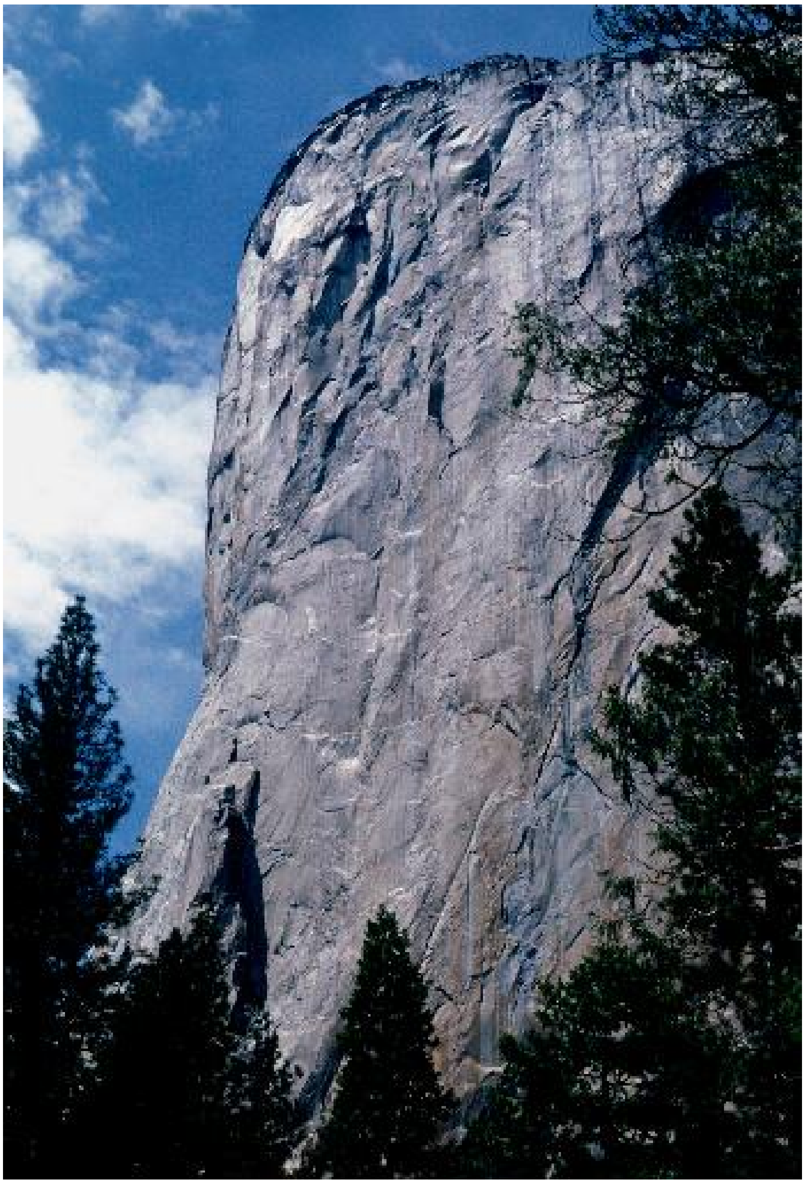


Figure 1.39: 'Massive' igneous rocks; a granite rock face without layers.

Figure 1.40: A close up view of a piece of gabbro.

the surface cool more quickly, forming medium-grained rocks with individual crystals that are harder to see. If magmas reach the surface, they are extruded and can flow out as lavas. Such **extrusive** igneous rocks cool down very quickly, either in the air or under water, forming fine-grained igneous rocks. The crystals in **fine-grained igneous rocks** are almost impossible to see without using a hand lens.

These processes show how igneous rocks of different compositions and grain sizes are formed. Their colour can help us to distinguish them, since the richer in silicon a rock is, the paler it is. Some igneous rocks are more common than others. The coarse-grained ultramafic rock that forms much of the mantle is called **peridotite**. A coarse-grained mafic rock, dark in colour, is called **gabbro**, shown in Figure 1.40, and its fine-grained equivalent, that forms many lava flows, is dark-coloured **basalt** (Figure 1.41). The fine-grained rock forming intermediate lavas is **andesite**, which is paler in colour than basalt. Fine-grained silicic lavas are uncommon, but the coarse-grained equivalent is very common. This coarse-grained, pale-coloured, silicic igneous rock is **granite** (Figure 1.42).

This fairly simple picture is complicated by how lavas erupt, since the composition affects the viscosity of the magma and this in turn affects the type of eruption. Mafic magmas that produce basalts are relatively rich in iron but poor in silica. Silica-poor magmas have low viscosity and these runny lavas can flow quickly out of cracks in the ground, called volcanic fissures (Figure 1.43), and spread over wide areas. Small bubbles of volcanic

Figure 1.41: A close up view of a piece of basalt with gas holes, called vesicles.

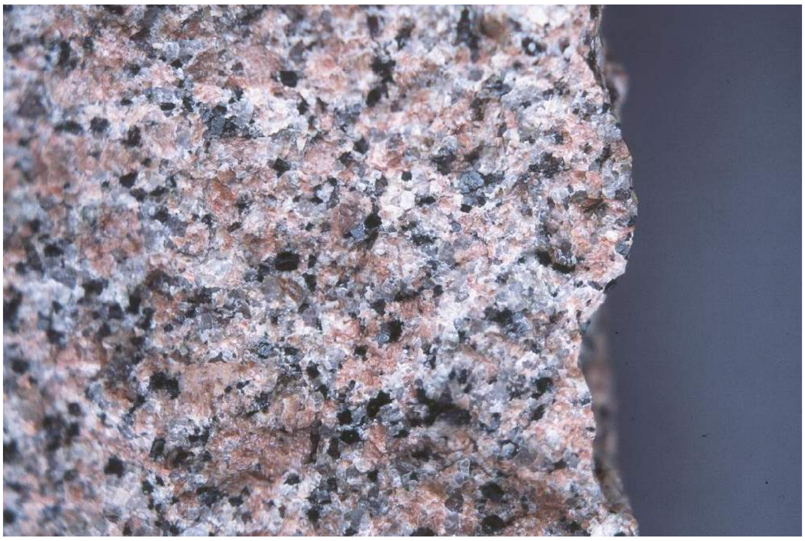


Figure 1.42: A close up view of a piece of granite.

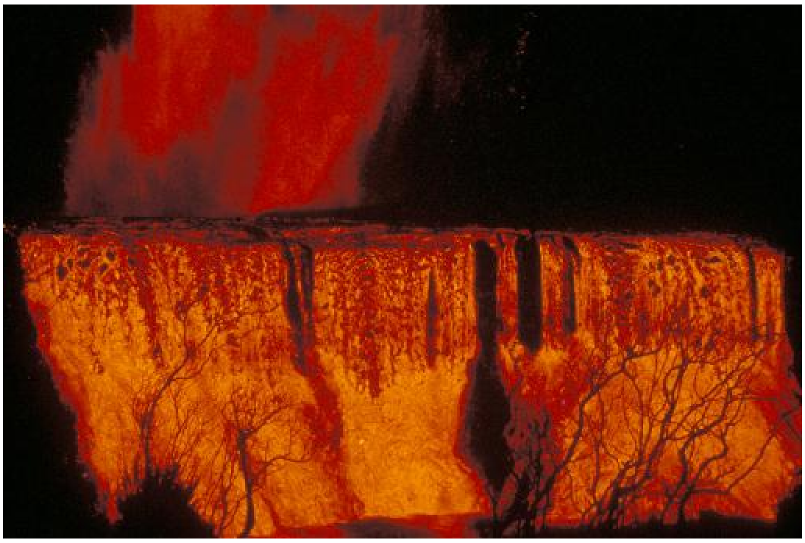


Figure 1.43: Basalt erupting from a fissure, with its low viscosity, running like water into a hollow.

gas are often trapped in basaltic lavas as they cool, and become preserved as spherical **vesicles**; there is an illustration of this in Figure 1.41. In very active volcanic areas, many sheets of basalt can be laid down in this way, as tabular basalts over wide areas. Some of these shrink as they cool, breaking into the vertical polygonal columns of **columnar basalt** (shown in Figure 1.44). If basalts are erupted under water, they are often extruded as thin tongues of lava that cool quickly into pillow-like shapes. Each pillow has a very fine-grained (chilled) edge, is slightly coarser inside, and also often contains vesicles. As the eruptions continue, the pillows build up, so a sequence of **pillow basalts** (shown in Figure 1.45) shows that the eruption must have happened underwater.

Intermediate magmas are more viscous than mafic ones. While they can flow out of central volcanic vents as thick andesite lavas, they often solidify in the vent. Pressure builds up on this volcanic plug and eventually the volcano erupts violently, often ejecting **volcanic blocks** and huge quantities of **volcanic ash** high into the air. These solid eruption products are called pyroclastics (*pyro* = fire-formed; *clastic* = broken material). The blocks and ash rain down on the surrounding area, often producing the typical cone-shape of many central vent volcanoes, as shown in the photograph in Figure 1.46. Andesitic volcanoes are much more dangerous than basaltic ones and some eruptions have killed thousands of people. To simulate magma viscosity, try the 'See how they run: investigate why some lavas flow further and more quickly than others' activity on the `http://www.earthlearningidea.com` website.

Silicic magmas are even more viscous than intermediate ones, which is why silicic lavas are rarely found. Instead, silicic volcanoes tend to have very violent eruptions, erupting enormous volumes of ash from central vents catastrophically (Figure 1.47).

Igneous magmas that don't reach the surface often **intrude** through the rocks above as large upside-down raindrop shapes. These **intrusive** large drop-shapes, that can be

Figure 1.44: A basalt flow that cooled and fractured into vertical columnar basalt.

Figure 1.45: A basalt pillow lava exposed in a rock face.

Figure 1.46: A central vent volcano erupting intermediate magma violently as a billowing ash cloud.

Figure 1.47: Deposit of volcanic ash with small bombs in the upper layer.

tens of kilometres across, are called **plutons**. Since plutons cool slowly, their rocks are always coarse-grained. As they cool, the plutons heat up the surrounding rock, producing metamorphic zones that can be hundreds of metres wide, called **metamorphic aureoles**.

If the magma remains liquid, it can intrude higher into the crust, forcing its way through cracks in the rock and along bedding planes as sheets of magma. The cooling of this magma is much quicker, so the igneous rocks produced are normally medium-grained. The edges of these intrusions cool even faster, forming fine-grained **chilled margins** and heating the edges of the surrounding rock to form thin metamorphic zones called **baked margins**. The sheet-like igneous intrusions that cut across rocks are called **dykes** (Figure 1.48) whilst sheets along bedding planes are **sills**.

This range of igneous processes allows us to identify the igneous rocks and structures you might find in a rock face according to first principles. Plutons are deep slow-cooling intrusions which are often silicic (made of pale-coloured coarse-grained granite) or mafic (darker coloured coarse-grained gabbro). Dykes and sills are shallower tabular intrusions that are usually formed of medium-grained rocks. Mafic rocks that reach the surface in volcanic fissures produce tabular or pillow basalts. Intermediate magmas erupt from central vents forming andesitic lavas but also deposits of volcanic blocks and ash, whilst silicic magmas produce huge volumes of volcanic blocks and ash in highly explosive eruptions.

1.5 Metamorphic rocks - formed by heat and pressure in metamorphic processes

Metamorphic rocks have been changed from their original rocks by great heat and/or pressure deep in the Earth's crust, but have not been heated up enough to melt them (if they

Figure 1.48: Igneous dykes cutting through the surrounding rock and also cutting through each other.

had become molten, they would have formed igneous rocks). So metamorphism is defined as the recrystallisation of rocks by heat and/or pressure, without complete melting. Metamorphic recrystallisation forms new minerals and new rocks with different properties. The recrystallisation produces interlocking crystals, like those found in igneous rocks. Whilst igneous rocks have randomly orientated crystals, most metamorphic rocks have crystals that are aligned. Being formed of interlocking crystals, both igneous and metamorphic rocks are generally tough and impermeable, so that water drains off them unless there are cracks for the water to sink into. Because metamorphic rocks are 'waterproof' some have been used for roofing buildings in the past, like the slates excavated from the slate quarry shown in Figure 1.49.

Just two major factors affect the sorts of rocks produced by metamorphism: the composition of the original rock that they came from, and the amount of heat and pressure to which they have been subjected. However, there are two main sorts of metamorphism, the metamorphism caused by baking by hot igneous intrusions, and the metamorphism caused by great heat and pressure deep in the roots of mountains during **mountain-building episodes** (also called **tectonic** episodes, since the rocks are affected by tectonic activity). Metamorphism by baking is caused by igneous intrusions and is called **thermal metamorphism**, while metamorphism caused by both increased heat and pressure is much more widespread and is called **regional metamorphism**.

The common minerals that are most changed by metamorphism are the clay minerals found in mudstones and shales. As metamorphism increases, the clay minerals recrystallise into a range of new minerals, so the greatest variety of metamorphic rocks is

Figure 1.49: An old slate quarry, showing metamorphic slates without the bedding planes usually seen in sedimentary rocks. The slates were cut from the quarry benches seen in the photo.

caused by the metamorphism of fine-grained sedimentary rocks. Whilst the quartz and calcite common in sandstones and limestones are also affected by metamorphism, their composition doesn't change, although the crystals often recrystallise into different shapes.

When magma intrudes cracks and bedding planes in mudstones, it will cool down to form dykes and sills. As the magma cools, the mudstones at the edges of the intrusions are heated and thermally metamorphosed, forming a thin fringe of metamorphic rock, called a baked margin. With much bigger intrusions, like the plutons that can be several kilometres across, the 'baked margins' are much wider because the plutons originally contained much more heat and took much longer to cool. These 'baked margins', which can be tens or hundreds of metres across, are called metamorphic aureoles. As the mudstone is metamorphosed, the clay minerals recrystallise into new metamorphic minerals scattered through the rock with random orientation (Figure 1.50). If the dykes, sills or plutons have intruded limestone, the calcite crystals in the limestone will have recrystallised into a random texture of interlocking calcite crystals, forming the metamorphic rock, marble. Thin, baked margins of marble fringe dykes and sills, whilst much larger zones of marble are found in metamorphic aureoles. Similarly, when sandstones are intruded, the quartz crystals recrystallise into a tougher rock of interlocking grains, called **metaquartzite**. Thin metaquartzite bands fringe sheet intrusions and wider metaquartzite zones are found in metamorphic aureoles. We can easily distinguish marble from other types of metamorphic rock because, like the limestone it originally came from, it is composed of calcium carbonate that reacts to the one-drop acid (dilute HCl) test.

Figure 1.50: Mudstone metamorphosed in a metamorphic aureole, showing the new randomly orientated metamorphic minerals.

Rocks containing randomly orientated metamorphic minerals must have been produced by thermal metamorphism, but metamorphic rocks with aligned minerals are produced during regional metamorphism. Under the high temperatures and great pressures caused by tectonic episodes, when new minerals crystallise and original minerals recrystallise, they do so at right angles to the pressures affecting the rocks. So the new and recrystallised old minerals become lined up and parallel to each other. If the new minerals are flat platy minerals, as many new regional metamorphic minerals are, then the new rocks which form will develop a new metamorphic layering, called **foliation**.

When mudstone or shale are regionally metamorphosed, the clay minerals recrystallise into very fine grained new metamorphic minerals. The new rock develops a slaty foliation (cleavage) and is called **slate**, whilst any fossils in the original rock are either deformed or destroyed. Because of the alignment of the new minerals, slates can easily be split along the foliation, which is called **slaty cleavage** (Figure 1.51). This is why slates can be split into the thin waterproof sheets used to make roofs for buildings. Since the direction of the new slaty cleavage is often different from the direction of the original bedding in the sedimentary rocks, we can sometimes see both the bedding and the cleavage in slates, running in different directions (Figure 1.52). See how 'fossils' can be deformed by pressure using the `http://http://www.earthlearningidea.com` activity, 'Squeezed out of shape: detecting the distortion after rocks have been affected by Earth movements'.

Slate is a low-grade metamorphic rock, since it is formed at relatively low temperatures and pressures. As metamorphic pressures and temperatures increase, the new metamorphic minerals in the slates grow in size and some new minerals, like garnets, can form, producing a new coarser-foliated, medium-grade metamorphic rock, called **schist** (Figure 1.53). Continued metamorphism causes the minerals to separate into foliated bands pro-

Figure 1.51: Slate, a low-grade metamorphic rock, formed by the metamorphism of mudstone or shale. The new slaty cleavage can be seen in this photo.

ducing a high-grade metamorphic rock, called **gneiss** (Figure 1.54). Gneiss can also be formed by the high-grade metamorphism of granite.

When mudstones are regionally metamorphosed, the metamorphic sequence is from slate, to schist to gneiss. When limestones are regionally metamorphosed, marble is formed, as it is when limestones are thermally metamorphosed (Figure 1.55). In regionally metamorphosed marbles, the calcite crystals can be aligned, allowing them to be distinguished from marbles produced by thermal metamorphism. Similarly, regional metamorphism of pure sandstone forms metaquartzite that can have aligned grains, helping us to distinguish it from thermally-metamorphosed metaquartzite (Figure 1.56). You can simulate how metamorphic rocks are formed using the `http://http://www.earthlearningidea.com` activity, 'Metamorphism - that's Greek for 'change of shape', isn't it?: What changes can we expect when rocks are put under great pressure in the Earth?'.

Regionally metamorphosed areas are the roots of mountain chains that have become exposed by the erosion of the rocks above. You can trace the progression of metamorphism, if the margins of the region are composed of mudstones, limestones and sandstones, containing fossils. Moving inward, broad regions of slate are found, sometimes with deformed fossils. These are followed by schist zones, where any fossils have been destroyed, and finally zones of gneiss. Meanwhile, limestones become marbles, progressively destroying fossils, while sandstones become metaquartzites. So as you move in from the margins, original structures and fossils are progressively lost, the crystal size of the minerals becomes larger, and the rocks tend to become more compact and tougher.

Figure 1.52: Bedding and cleavage seen in slate, formed from mudstone by regional metamorphism. The bedding runs steeply from the lower right to the upper left and is shown by the different bands of colour. The cleavage is the series of cracks that slope up gently from right to left. The tectonic forces that caused the metamorphism were at right angles to the cleavage, from top to bottom of the photo.

1.6 Deformation in rocks - geological structures

Rocks in the Earth's crust can be put under pressure by three different sets of forces: they can be compressed by compressional stresses; they can be pulled apart by tensional stresses; or they can, under pressure, slide past one another due to shear stresses. There is a fourth type of stress, caused by turning forces, but these are not geologically very common.

Not only do three types of stress commonly affect crustal rocks, but rocks respond in three different ways. The rock can absorb the stress, like a rubber ball, but when the stress is removed, it bounces back. This is called **elastic behaviour** and, although as we shall see, this is important in earthquakes, the effects cannot be seen in rocks, since they have sprung back elastically to their original shape and size. The rocks may bend and flow, like clay does when you mould it with your hands. This is **ductile behaviour**, the rocks remain distorted into folds that can be seen clearly. Finally the rocks may break; this is **brittle behaviour** and results in different forms of rock fracture.

Whether a rock bends, breaks or behaves elastically under stress depends on a range of different factors, including what the rocks are made of, how great the stress is, the temperature of the rocks, how deeply buried they are, and for how long the stresses are applied. The result is that the same rock can behave differently at different times and under different conditions and can show the effects of both ductile and brittle behaviour; so folded rocks can be fractured as well.

Figure 1.53: Schist, a medium-grade metamorphic rock, formed by the continued metamorphism of slate, by higher temperatures and pressures. The foliation gives this flat surface of aligned platy minerals.

Figure 1.54: Gneiss, a high-grade metamorphic rock, formed by the continued metamorphism of schist by even higher temperatures and pressures, or by the high-grade metamorphism of granite. The high pressures deformed the bands of gneiss into tight folds before this rock was intruded by a narrow dyke.

Figure 1.55: Marble, produced by the metamorphism of limestone either by high temperatures in thermal metamorphism, or by high temperatures and pressures in regional metamorphism. This specimen clearly shows the interlocking crystal texture.

Figure 1.56: Metaquartzite, produced by the metamorphism of pure quartz sandstone, by either thermal metamorphism or regional metamorphism. This photo shows the interlocking quartz grain texture that makes it such a tough rock.

So we have three types of stress and three types of behaviour, and you might anticipate seeing the results in rocks in nine different ways. However, as we don't see the effects of elastic behaviour, this cuts this number down to six. When rocks are pulled apart by tensional stresses and flow, the main effect is that the layers become thinner and it is usually impossible to tell that this has happened. We rarely see the ductile effect of shear stresses in rocks either, so we are left with only four common effects to examine below.

When rock sequences are strongly compressed and flow, we see the effects as **folding** of the layers. The rocks may be deformed into gentle open folds or into a series of tight folds. The place where the rock has bent most is called the **fold axis**. Since there is never just one layer that is bent, but many, if you join the axes of all the folded layers together, a plane is formed and this is called the **axial plane surface**. Compressional stresses always act at right angles to the fold axial plane surfaces, so we can use the shape of the folds to work out the directions of the stresses that deformed them. The directions of axial planes are measured as compass directions; folds with north/south axial planes must have been formed by east/west stresses.

Where rocks are folded, you often can't see the folds themselves, you just see the sides or **limbs** of the fold, as layers that were once horizontal but are now sloping. The direction of downward slope is called the **dip** and can be measured as an angle to the horizontal. At right angles to the dip direction is the **strike**, which is measured with a compass. So rocks that dip towards the east have a north/south strike, which is parallel to the directions of the fold axial plane surface. This may sound complicated but you can make a simple model to explain it by using your hands. Put the sides of your hands, with the little fingers together, into a downfold, with both hands sloping at the same angle. The fold

Figure 1.57: Folded rocks, with a downfolded syncline to the left and an upfolded anticline to the right. The axial plane surfaces of these folds go into and out of the cliff face and the compressional stresses that deformed them came from the left and right. These are tight folds and you can see their scale from the street light in front of them.

axial plane surface is vertical and runs through the join between your hands. Each hand forms the limb of the fold and dips towards the centre. The strike direction is parallel to your fingers and to the fold axial plane surface.

If rocks in a cliff face are dipping towards the left and you followed them to the left far enough, you would find them bending up again. The rocks are therefore forming a downfold, this is called a **syncline**. Similarly, if you followed the dipping rocks to the right far enough, you would eventually find them bending down again. Upfolds like this are called **anticlines**. So dipping rocks are just the visible parts of much larger folds that become synclines in one direction and anticlines in the other.

You can use these principles on any dipping rocks to work out where the synclinal and anticlinal axial planes are. For example, in Figure 1.57, the synclinal axial plane surface must be to the left and the anticline to the right. You can also work out the directions of the stresses that caused this tilting; here, they must also have come from the left and right. So, you should now be able to use these principles on any tilted or folded rocks you find, to work out the directions of the stresses that caused them. Since folds are three dimensional, you can not only see them in vertical cliff faces but also on horizontal surfaces too, as in the aerial view in Figure 1.58. The same principles are used to work out the stress directions wherever you see folded or tilted rocks. Folding results in **crustal shortening**, where the crustal rocks have been compressed and take up less space than they used to.

When rocks are compressed and fracture, compressional **faulting** occurs as one slab of rock slides up over the other, again resulting in crustal shortening. The sliding surface is called the **fault plane** and fault planes in compressional faults usually dip at around

Figure 1.58: A region of folded rocks seen from the air. An anticline makes the ridge in the distance and there is a syncline in the left foreground. The fold axial plane surfaces run up and down the photo and the compressional stresses came from left and right.

45°(between 30 and 60°). These faults, shown in Figure 1.59, are called **reverse faults** (because they have moved in an opposite direction to the more common 'normal' faults, described below). When compressional forces are very great, slabs of rock can be forced great distances over the rocks beneath, when the sliding surface usually has a much lower slope of 10°or less. These types of reverse faults are called **thrust faults** (Figure 1.60), and sometimes rock can be moved tens of kilometres along them. You can make your own folds and compressive faults using sand and flour in a small plastic box using the 'Himalayas in 30 seconds!' activity on the `http://www.earthlearningidea.com` website.

When rocks are pulled apart by tensional stresses, they can fracture to form **normal faults**. They are called normal faults because they are the most commonly seen types of faulting (Figure 1.61). The rocks fracture, usually along a steep fault plane of 60°or more, and the rocks on one side slide down relative to the rocks on the other. The tension was at right angles to the fault plane and the result is **crustal extension**, with the rocks taking up more space than they did originally. Try making tensional faults in a box using the `http://http://www.earthlearningidea.com`, 'A valley in 30 seconds' activity.

Shear stresses cause slabs of rock to move sideways across the Earth in relation to the rocks on the other side. These are called **strike-slip faults** because rocks striking across the land surface have slipped sideways relative to the rocks on the other side (Figure 1.62). The fault planes of strike-slip faults are usually vertical. These result in neither crustal shortening nor crustal extension and are most easily seen by viewing rock sequences from above.

Where you can see layers on each side of a fault that used to match up, but are now broken, it is easy to tell there is a fault there and to work out which type of fault it is and how far the rocks have been moved (the **fault displacement**). However, if you can't see

Figure 1.59: Reverse faults. The compressional stresses came from left and right in this photo, and the rocks on the left were moved up over the rocks to the right.

Figure 1.60: A thrust fault. You can tell that the rock sequence at the top has been thrust over the rocks at the bottom from the left to the right because the lower rocks have been bent into a drag fold as the upper rocks were forced over them.

Figure 1.61: A normal fault. This steep fault was caused as the rocks were pulled apart by tension to left and right, allowing the rocks on the right to slide down relative to the rocks on the left.

Figure 1.62: A strike-slip fault. The strike of the rocks is from lower left to upper right. Along the fault, the rocks in the background have been moved to the right relative to those in the foreground.

Figure 1.63: The rocks of this quarry wall have horizontal bedding planes and at least two sets of vertical joints.

layers that have been dislocated, it is much harder to tell the type of the fault and work out the amount of fault displacement. In rocks that have no layers, you may not be able to tell there is a fault there at all, you might just see a fracture and not know that there had been fault movement along it. If this is the case, you have to look for other clues of fault movement, such as bits of fractured rock along the fault plane (**fault breccia**).

Faults are defined as fractures with movement of the rocks on either side. Rocks can have other sorts of fractures too and the most common of these is **joints**. Joints are straight fractures where the rock has not moved up, down or sideways on either side, and are usually caused when an area of the crust has been put under pressure and later the pressure is released. Most joints are vertical or near-vertical and form parallel sets. If, at different times in the past, the crust has been put under pressure from different directions, then two or more joint sets can be formed, running in different directions, which break the rocks up into large blocks. Joints are important: not only are they the most common fractures seen in rocks, but they are also pathways for underground fluids. They control the ways in which rocks break in quarries and mines as well, which is why they were first called joints, looking as they do like the joints in a cut stone wall (as shown in Figure 1.63). Minerals may have crystallised from fluids flowing along joints in the past, to form mineral veins. Joints are also formed as igneous rocks cool and contract; many of the joints in plutons probably formed in this way, as did the columnar joints in basalts.

A structural feature that tells you a lot about the geological history of an area is an **unconformity**. Unconformities form when a rock sequence has been buried and hardened into rock. If it is then uplifted and the rocks above are eroded away, a surface is eroded across the rocks, called an unconformity surface, as shown in Figure 1.64. Later, another

Figure 1.64: An unconformity. Older rocks beneath are overlain by much younger rocks above.

sequence of sediments is deposited on this surface and becomes hardened into rocks. Unconformities are most easy to see when the older rocks beneath the unconformity have been folded or tilted, and so have a different angle of dip from those above; these are called **angular unconformities**. Unconformities tell us a tale: the first formed rocks were buried, became hard and then were uplifted by a mountain-building episode. Over millions of years the rocks above were eroded and an erosion surface was cut across them, like the near-horizontal surfaces found on rocky coastal foreshores today. Later, a sequence of sediments was deposited on top of this surface and it too became buried, hardened and possibly tilted. Finally, the whole sequence was uplifted again and the rocks above eroded away to allow you to see the unconformity as it is today.

1.7 Rock exposures contain evidence of the sequence of geological events that formed and deformed them

Rocks can tell us the tales of how they formed and were changed if we apply a series of principles to them, as for the unconformity story above. These are called 'Principles' if they generally apply, but there can be certain unusual circumstances in which they don't. If they always apply, we call them 'Laws' instead. Many of them have very complex-sounding names, but the ideas are very simple.

We understand how many surface rocks and fossils formed by applying the 'Principle of

Uniformitarianism' that the 'present is the key to the past'. Most of the different types of sediment and volcanic rock we find in the rock record are being formed on Earth today, so we can investigate modern Earth processes to get good insights into how they were originally formed. Similarly, many fossils have modern living relatives today, so we can examine these to find out how the fossil organisms probably lived and died. It is more difficult to determine how extinct organisms lived and how rock-forming processes worked that don't seem to be happening on the planet today, but even for these, the 'Principle of Uniformitarianism' can give us clues if we examine similar organisms and processes.

The five principles below are called '**Stratigraphic Principles**' since they apply to **strata**, or sedimentary and volcanic layers. Two of these are the '**Principle of Original Horizontality**' and the '**Principle of Lateral Continuity**'. 'Original Horizontality' states that most sedimentary and volcanic rocks were originally laid down in near-horizontal layers. This means that if we find them in layers that are not horizontal, they must have been tilted by tectonic activity. There are unusual instances when sediments and volcanic layers are not deposited flat, as in current bedding and dune bedding, or the layers in scree slopes or on the sides of volcanoes, which is why 'Original Horizontality' is a principle and not a law. 'Lateral Continuity' states that sedimentary and volcanic rocks once formed laterally continuous layers over wide areas. We know that they cannot have been laterally continuous over the whole Earth, since they must either have hit the edge of the area where they were deposited or died out laterally, so this too is a principle and not a law.

The next principles are vital in sequencing rock events. The '**Principle of Superposition of Strata**' states that the rocks on top are the youngest. This is because rocks are deposited in layers that build up over time, so the oldest is at the bottom and the youngest at the top. Since it is possible for such sequences to be turned over by intense folding or for older rocks to be thrust over younger ones by intense compression causing thrust faulting, 'Superposition of Strata' remains a principle and not a law. The '**Law of Cross-Cutting Relationships**' states that anything that cuts across anything else must be younger. The 'anything' can include faults, joints, dykes, plutons and unconformity surfaces. This law applies in all circumstances and to all rocks, since something cannot be cut until it is first formed. However, sometimes you have to examine the evidence very carefully to be sure that something is indeed cutting something else, and it doesn't just look that way. The '**Law of Included Fragments**' states that anything included in a rock must be older than the rock that includes it. Included fragments are such things as the pebbles found in conglomerates and the pieces of surrounding rock called xenoliths, sometimes included in plutonic igneous rocks. This is a law since, for anything to be included, it must be older, but sometimes you have to examine the evidence very carefully to be sure. See for yourself how these stratigraphic principles work, by trying the http://www.earthlearningidea.com activity, 'Laying down the principles'.

The '**Law of Faunal Succession**' applies to fossils, and so is not a stratigraphic principle, in spite of the fact that it is used to sequence strata. It states that groups of fossil animals follow one another in time in a predictable sequence and we now know that plant fossils can be used in the same way. It was by using the 'Law of Faunal Succession' that geologists were first able to sequence rocks, divide up the geological time scale and make the first proper geological maps. Fossils have now been sequenced all over the world to provide very detailed evidence for the relative ages of rocks. When fossils appear at the same time in rocks across the world, we can use these fossils to say that the rocks must have been formed at the same time too. This method of identifying rocks of the same age in different areas is called **correlation** and rock correlation using fossils has been essential in linking together the geological sequences of different areas, regions, continents and across the world. We now know that the reason for the 'Law of Faunal Succession', that fossils are always found in the same sequence across the world, is evolution.

A huge range of fossils can be found in rocks, but only a few of these can be used for correlation. The best fossils for correlation have these key features:

- they were common, so many of them could be fossilised;
- they were easily preserved, usually because they had hard parts, and so are frequently found;
- the fossil group evolved quickly over time, meaning that the fossils in different beds are slightly different;
- they were widespread, so are found in many rocks across the world;
- they are found in many different rock types, such as sandstones, limestones and shales;
- and they can be easily identified, at least by experts.

Two types of fossil that fit these requirements are graptolites and cephalopods.

Graptolites are now extinct, so it is difficult for us to know how they lived. However, this doesn't matter if we are just using them for correlation. They were small colonies of animals that were strung together in saw-blade-like shapes. Each animal lived in a small living chamber called a **theca** and the string of animals is called a **stipe** (see Figure 1.65). Their important properties are:

- they were very common in ancient seas;
- they were made of a hard organic material that readily fossilised;
- they evolved quickly over time, so have many different forms, in particular the shapes of the stipes changed, as you can see in Figure 1.66, and the shapes of the thecae were also very varied;
- they lived right across the world's oceans;
- as they floated in the sea, they could be found in sandy, muddy and limestone environments, and so are preserved in different rock types;
- experts can easily identify them from their stipe orientations and the shapes of their thecae.

Figure 1.65: Graptolites. The largest fossil here had four stipes and the small thecae or living chambers can be seen along each stipe.

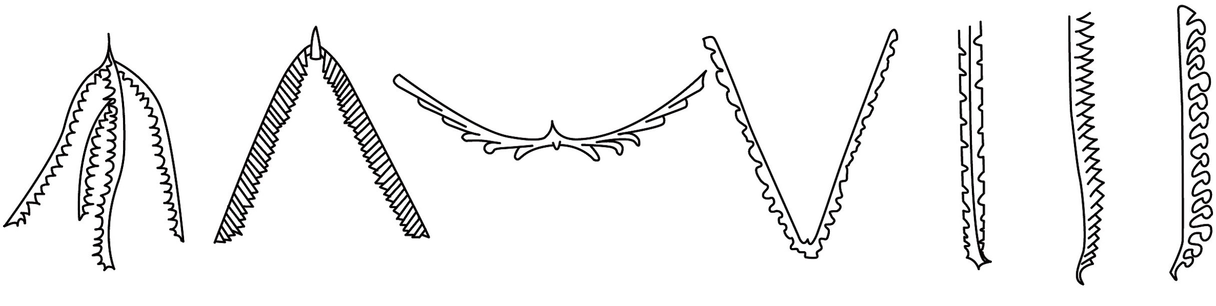


Figure 1.66: How graptolites evolved, from the four-stiped form common around 480 million years ago on the left to the single-stiped form common around 420 million years ago on the right.

Cephalopods are not extinct, but the sorts with coiled shells that are useful in correlation, nearly became extinct and only one type can be found alive today (it is shown in Figure 1.67). Nevertheless, we know from these living animals that they lived in the sea, floating in the water, and could move around by squirting jets of water backwards. The animal living in the coiled shell had tentacles, like an octopus, that could grab passing small animals for food. When they were attacked they could withdraw into their shells for protection. Inside the coiled shells there were chambers of gas that helped the animal to float in the water. Where the chamber walls joined the outside of the shell they formed a line, called a **suture line**. You can see these if you peel away the outer shell, exposing the chambers and chamber walls beneath (as has been done in Figure 1.68). In the earliest shelled cephalopods, the chamber walls were straight lines, but as the animals evolved, the lines became more and more complex, and it is these complex suture lines that help us to identify the different types. The cephalopods with the most complex suture lines, that lived between about 200 and 65 million years ago, are called ammonites (Figure 1.69).

Shelled-cephalopods have all the properties that make them excellent fossils for correlation:

- like graptolites, they also were very common in ancient seas;
- their hard shells were easily fossilised;
- their suture lines evolved quickly, and three of the main shapes are shown in Figure 1.70;
- they lived in all seas and oceans;
- since they floated in the sea, and continued floating even after they died, their shells can be found in all sorts in environments and sediments;
- experts can easily identify them from their suture lines.

With all these properties, it is not surprising that cephalopods have been key fossils in helping us to unravel the geological history of the Earth.

A final rule that helps us to work out geological histories of rock sequences is the fact that rocks cannot be deformed or metamorphosed until they have first been formed. So we now have all the principles and laws needed for working out how rocks were first formed and the sequence of changes that happened to them after formation. When all the changes have been put into order, the geological history of the rock sequence has been explained.

These methods were used to work out the geological time sequence of all rocks on Earth, as shown in Figure 1.71. This is the international geological time scale. The principles above only help us to put things into order, to work out the **relative ages** of the events, but they do not tell us how old the rocks and events are. To do this, we need another technique that will give us the **absolute ages** of the rocks.

Geologists have been able to work out the relative ages of rocks and events for more than 200 years, but an absolute dating technique has only been widely available for about 50 years. This is **radiometric dating** and is based on the fact that radioactive atoms decay

Figure 1.67: A shelled cephalopod.

Figure 1.68: A shelled cephalopod fossil with the outer shell removed so that the chambers and the suture lines can be seen. The suture lines here have fairly simple shapes, so this must be a goniatite.

Figure 1.69: A coiled cephalopod fossil with the outer shell removed to show very complex suture lines. Complex sutures like these show it must be an ammonite.

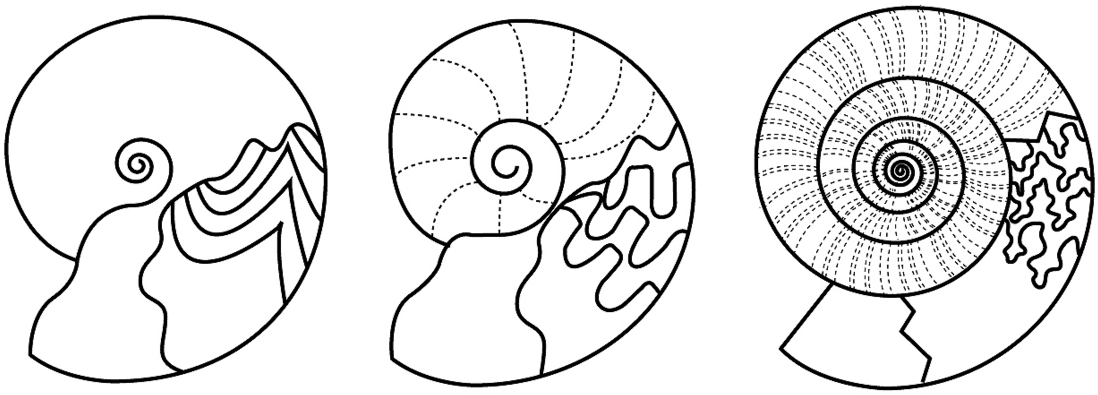


Figure 1.70: The change of ammonoid suture lines over time, from the simple sutures of goniatites, common around 350 million years ago in the left hand diagram, to the more complex sutures of ceratites, common around 240 million years ago in the centre to the very complex sutures of ammonites, found in 190 million year old rocks in the right hand diagram. Ribs on the outsides of the fossils are shown as dashed lines.

at fixed rates over time. So, if we can find a rock or mineral that contains radioactive atoms, and we can find out what proportion of them has decayed, and we also know the rate of decay, we can work out the age of the rock. For example, potassium (with the chemical symbol, K) has a radioactive component and is commonly found in igneous rocks like granite. When the granite first forms, the potassium-containing minerals contain only potassium. But the radioactive part of potassium decays to argon (Ar) over time. So, if we measure the amount of potassium that should have been in the rock, and the amount of argon it now contains, we can work out the age of the rock. Half the radioactive potassium would decay to argon in 1260 million years, so an igneous rock with half radioactive potassium/half argon must have crystallised 1260 million years ago.

Unfortunately we can only apply this method to igneous rocks and some metamorphic rocks that contain potassium (and the unusual sedimentary rock, greensand). This is true of other radiometric methods as well - we can only apply them to certain sorts of rocks. This means, for example, that we usually can't find the absolute ages of sedimentary rocks and fossils directly. If a bed of sedimentary rock containing a useful correlation fossil happened to have a lava flow below and above it, and we could work out the radiometric ages of the lava flows, we would have a good idea of the absolute age of the rock and its fossil. However, this is very unusual and, even when it does happen, the ages of the lava flows may be several million years apart, so that we can only find the approximate age of the rock and fossil.

Nevertheless, geologists have been working on this problem for many years, and now have a good idea of the ages of key correlation fossils and the rocks in which they are found. These ages have been added to the international geological time scale, so as well as knowing the sequence of global geological events, we also know when in the geological past they actually happened (Figure 1.71).

You can try the rock sequencing principles yourself through the http://www.earthlearningidea.com activity, ‘What is the geological history?’. Even better, try them on a nearby rock exposure in a cliff or a quarry instead.

Eon	**Era**	**System**	**Age, millions of years ago (Ma)**
			0
Phanerozoic	Cenozoic	Neogene	
			23
		Paleogene	
			65
	Mesozoic	Cretaceous	
			146
		Jurassic	
			200
		Triassic	
			251
	Palaeozoic	Permian	
			299
		Carboniferous	
			359
		Devonian	
			416
		Silurian	
			444
		Ordovician	
			488
		Cambrian	
			542
Precambrian	Proterozoic		
			2,500
	Archaean		
			3,800
	Hadean		
			4,600

Figure 1.71: The geological time scale, used internationally