

INTRODUCING METAMORPHISM

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Preface

Students of geology do not normally start with metamorphic rocks. They would normally learn about them later in their coursework, having first become familiar with topics like the internal structure of the planet, the theory of plate tectonics, the chemistry of minerals, the nature of igneous and sedimentary rocks, and the polarizing microscope. 'Introducing metamorphism' to a general readership, for whom this background knowledge may be limited, presents challenges.

In an attempt to meet those challenges, some relevant foundation material is included as three appendices. The first deals with the Earth's interior, the second with minerals, and the third with the polarizing microscope. The content of each appendix, while widely available from other sources, has been prepared with metamorphism in mind, and is cross-referenced in the text. Readers who are new to geology might like to look through these three appendices before starting chapter 1.

This book is not a comprehensive account of metamorphism, but is biased towards the author's interests and experience. The aim is to give an overall sense of what metamorphism entails through describing a broad selection of metamorphic rocks, through explaining the methods (particularly polarized light microscopy) used to investigate them, and particularly through addressing all kinds of questions about how the rocks came to be the way they are found today.

Following a general introduction (chapter 1), the text is organized into four chapters. The first of these, chapter 2, is easily the longest, and is largely descriptive. The other three focus on interpretation. Chapter 3 asks about the processes that lie behind features seen on the scale of a hand specimen or smaller, chapter 4 extends the questioning to processes that relate to the geological setting and timing of metamorphism, and chapter 5 aims to quantify the pressure and temperature conditions during metamorphism in two specific case studies which the author knows well, and continues to find fascinating.

All terms in **bold font** in the appendices and in the main text are defined in a Glossary at the end of the book.

1 Introduction

1.1 What is metamorphism?

Metamorphism (from the Greek words *meta* = change and *morphos* = form) is a geological process that changes pre-existing **igneous** and **sedimentary rocks** into new rocks – metamorphic rocks – that look quite different from the rocks they started out as. Metamorphism can even change pre-existing metamorphic rocks into new, and different looking, metamorphic rocks. When rocks are changed like this they are said to become metamorphosed, but one should note that the related word *metamorphosis* has no place in geology. It applies to other kinds of change, such as from a tadpole to a frog, and is not a synonym for metamorphism.

Metamorphism takes place underground, usually deep underground at a high temperature. Since it cannot be watched as it happens, it is often portrayed as being more difficult to grasp than the formation of igneous or sedimentary rocks. Igneous rocks can be seen being made when, for example, volcanic **lava** cools and hardens. The formation of sedimentary rocks can be watched as grains of sand, for example, are moved by rivers and deposited on the seabed. Yet the very fact that metamorphic rocks originate ‘out of sight’ makes them all the more intriguing. The detective work involved in figuring out how they were made can be both challenging and rewarding.

1.1.1 Protoliths

An igneous or sedimentary rock that becomes metamorphosed is called a **protolith** (from the Greek words *proto* = first and *lithos* = stone). While there are many kinds of protolith, only six are important for the purpose of introducing metamorphism. These are the three common kinds of sedimentary rock, called **sandstone**, **shale** and **limestone**, the two common kinds of igneous rock, **basalt** and **granite**, and the important but less well-known rock called **peridotite** (Fig. 1.1).

The first five protoliths are common in the Earth’s **crust**, whereas the sixth one, peridotite, is the main kind of rock in the Earth’s **mantle**. Readers who would like to remind themselves

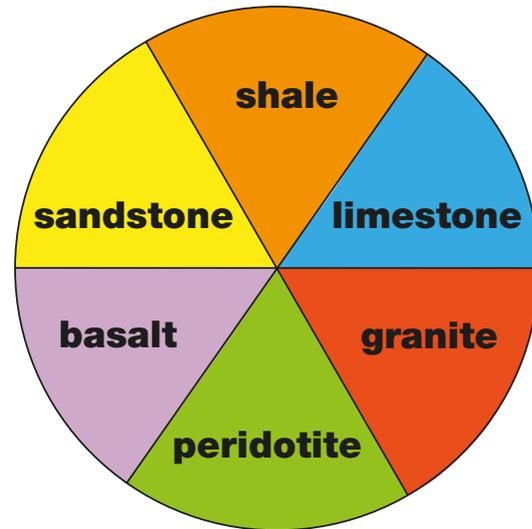


Figure 1.1 The six main protoliths.

of the nature of the Earth’s crust and mantle should turn now to section A1.1 of Appendix 1. Peridotite is important because it is the primary rock from which the other five protoliths have been derived, over time, by geological processes that operate at the surface and beneath it. These processes are touched on in this chapter, and are explained in more detail in Appendix 1 and chapter 2. At the end of chapter 2 they are summarized in a flow chart called the **rock cycle**.

When a protolith becomes metamorphosed, its appearance generally changes in two distinct ways. Firstly, it usually develops new **minerals**, and secondly, it always develops a new **texture**.

1.1.2 Changes to the minerals

Minerals are a rock’s ingredients. They are the solid chemical compounds from which the grains (individual particles) of a rock are made. Examples of minerals include quartz, garnet

and mica. Each has its own distinctive features and its own chemical composition. Most metamorphic rocks are aggregates of two, three, four or more different minerals. The list of minerals in a rock is known as the rock's **mineral assemblage**.

The minerals in a rock change because, during metamorphism, the pressure and temperature beneath the surface change. The existing minerals become unstable together and react chemically with each other to produce new minerals that are stable together under the new pressure and temperature conditions. The meaning of mineral stability is explained in chapter 3.

The number of different minerals that can occur in metamorphic rocks is enormous, but only about two dozen of them account for most of the rocks described in this book, with a further ten, called **accessory minerals**, being widespread but never abundant. These minerals are introduced one at a time in chapter 2, as their names arise. An emphasis is placed on how they can be recognized and on which chemical elements each contains.

Just ten chemical elements are needed to make the main two-dozen minerals, namely hydrogen (symbol H), carbon (C), oxygen (O), sodium (Na), magnesium (Mg), aluminium (Al), silicon (Si), potassium (K), calcium (Ca), and iron (Fe). The way in which these elements team up to make minerals is explained from first principles in Appendix 2.

It is helpful to know the elements in each mineral because then, if the minerals in a rock can be recognized, the overall chemistry of the rock can be estimated. This allows the protolith to be worked out, because each kind of protolith has its own distinctive chemical signature, which generally does not change during metamorphism.

However, there is an important exception to this constancy of chemical composition. This is a change in the amounts of water (H_2O) and/or carbon dioxide (CO_2) that are chemically bound in the rock's minerals. Minerals that contain chemically bound water are described as hydrous, and their chemical formulae contain the so-called hydroxyl group, (OH). Minerals that contain chemically bound carbon dioxide are known as carbonate minerals, and their chemical formulae have the carbonate group, (CO_3). Both water and carbon dioxide are examples of **fluids** (the term *fluid* covers liquids *and* gases), and they can be added to rocks, and can be lost from rocks, during metamorphism. Many examples of such gains and losses will be encountered in the book.

Perhaps the best evidence that fluid really was present during metamorphism is that it can be seen today, sealed up inside tiny cavities within mineral grains in some metamorphic rocks. Such fluid-filled cavities are known as **fluid inclusions** (Fig. 1.2). They are thought to have formed when small quantities of fluid became trapped inside mineral grains as they grew larger during metamorphism.

Since fluids can carry dissolved salts, then the introduction of fluids to rocks, and their loss from rocks, during metamorphism will cause any dissolved elements to move into, and out of, the rocks. If the resulting chemical changes to the rocks are substantial, the process is called **metasomatism** (pronounced meta-**soh**-ma-tism; from the Greek words *meta* = change and *soma* = body).



Figure 1.2 A fluid inclusion within a transparent wafer of rock seen through a microscope. The image is about 10 microns (10 thousandths of a millimetre) wide. The fluid is water. It is trapped in a cavity with a roughly triangular outline inside a grain of metamorphic quartz from an emerald deposit in Columbia. The water was very hot when it became trapped, and has since cooled and contracted, causing it to separate into liquid water and a bubble of water vapour. Also present in the water is a cube-shaped crystal of sodium chloride (square outline) that was precipitated during cooling. It shows that the liquid is not pure water but salty water. *Photo courtesy of Bruce Yardley.*

1.1.3 Changes to the texture

The second kind of change a rock undergoes during metamorphism is a change in its texture. The term texture needs a bit of explanation. To most people texture is the way cloth or other fabric feels when handled – whether, for example, it is silky, supple or coarse to touch. When applied to rocks, texture has a different and very specific meaning. It refers to the sizes and shapes of the mineral grains, and to the way the grains are orientated and distributed within the rock. A great variety of textures may be seen among metamorphic rocks, and these will be described and interpreted in chapter 2 and elsewhere in the book. For instance, grains may be too tiny to be seen with the naked eye or large enough to be spotted easily; grains with an elongate shape may be aligned parallel to each other or they may be randomly orientated; grains of a particular mineral may be evenly distributed throughout a rock, like dispersed raisins in a well-mixed fruit cake, or they may be concentrated in discrete clusters or layers.

During metamorphism the texture automatically changes when a new combination of minerals replaces an old one, but it can also change when existing mineral grains grow in size, and change their shapes, through a process called **grain growth** or **recrystallization**. With grain growth, small grains merge to become larger ones, for reasons that will be explained in chapter 3.

On the basis of their texture, rocks fall into two broad groups, known as **foliated** rocks and non-foliated rocks. In foliated rocks, grains of minerals with elongated shapes are aligned, and the rocks may also be layered in the same direction as the mineral alignment. Non-foliated rocks have no preferred orientation of elongated grains, and are not layered. Whether or not a rock is foliated depends on a factor called directed stress, which will be introduced below, in section 1.2.

1.1.4 Naming metamorphic rocks

Metamorphic rocks have a somewhat confusing choice of names. In fact, an individual rock may be given two or more different names that are equally correct. The choice arises because a rock's name can be based on its protolith, or its minerals, or its texture, or on some combination of these three.

Names based on the protolith are, perhaps, the simplest. Any metamorphic rock can be named by adding the prefix **meta** to the name of the protolith giving, for example, metalimestone, metashale, and metabasalt. Most metamorphosed

sedimentary rocks, however, already have their own widely used names relating to the protolith: **marble** is nearly always used instead of metalimestone, **metapelite** (pronounced meta-**pee**-lite) is often used as an alternative to metashale, and **metapsammite** (the 'p' is silent) is a popular alternative to metasandstone.

Names based on the rock's texture include several that relate to a foliated texture. These are **slate** or **phyllite** (pronounced **fill**-ite) if the foliated rock is very fine-grained, **schist** if it is medium or coarse-grained, and **gneiss** (pronounced nice, the 'g' is silent) if it is both coarse-grained and layered. In many cases these four kinds of foliated rock are metapelites.

Names based on a rock's minerals can be created by hyphenating one or more of the prominent minerals in the rock to a name like schist, or gneiss, or simply 'rock', giving, for example, garnet-mica-schist, quartz-feldspar-gneiss, or quartz-garnet-rock. A few rocks have special names that relate to their main minerals, like quartzite (made largely from quartz) or amphibolite (made mostly from a mineral called amphibole).

Examples of rocks with all the above names, and with additional names, will be described in chapter 2.

1.2 Metamorphic rocks – made under mountains

Since metamorphic rocks are formed beneath the Earth's surface, they are seen today only because, after they were made, the land became elevated and then eroded away to expose them. The great majority of metamorphic rocks fall under the heading of **regional metamorphic rocks** because they occur over extensive regions. These rocks were made, and are still being made, deep within the continental crust in the roots of major mountain ranges (Fig. 1.3).

A minority of metamorphic rocks are not from mountain roots, but were made in local settings beneath the ground, such as close to bodies of hot, shallow magma, or within deep fault zones, or below the impact sites of giant meteorites. Metamorphism in these local settings is introduced separately, after this section on metamorphism under mountains.

1.2.1 Mountain building

Mountain building is a protracted process; its formal name is **orogenesis**, or simply **orogeny** (from the Greek word *oros* = mountain). Orogeny is commonly preceded by the accumulation of many layers of sediment (mud, sand, and calcium



Figure 1.3 Mountains in the Mont Blanc area of the Western Alps. Metamorphism is happening today in the hot, deep roots of this and all other major mountain ranges.

carbonate shells) in a place where the Earth's surface continues to subside over a long period of geological time. The weight of the upper layers of sediment exerts a huge downward pressure on the layers below. Also, the temperature gets higher with increasing depth because heat is continually produced by traces of **radioactive** uranium, thorium and potassium that are present in all rocks. However, the heat and pressure at this stage of burial will generally not be sufficient to cause metamorphism, but only enough to convert the loose accumulated sediments into hard sedimentary rocks. Mud becomes shale or mudstone, sand becomes sandstone, and shells become limestone, in a process known as **lithification** (from the Greek word *lithos* = stone). An additional geological factor is normally needed to induce metamorphism. This factor is the onset of a strong compressive force that arises when the local tectonic plates change their relative motions and begin to converge. The nature of **tectonic plates** and the theory behind their movement is outlined in section A1.2 of Appendix 1.

The compressive force is called **directed stress**, and it acts like a giant vice; it crumples and buckles the layers of

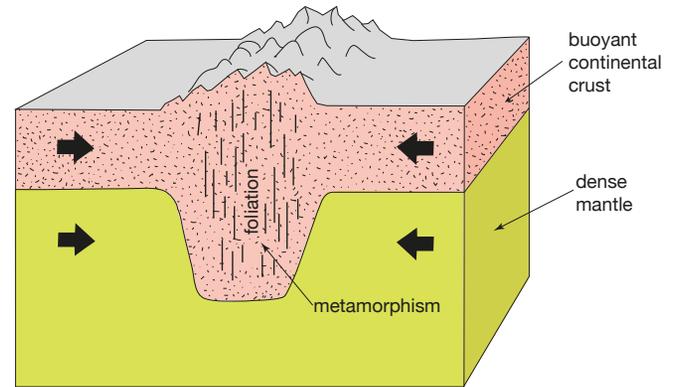


Figure 1.4 Simplified block diagram showing continental crust and underlying mantle in a place where orogenesis (mountain-building) is in progress. The continental crust thickened by directed stress (compressive forces shown by the black arrows) is buoyant. It 'floats' in the mantle, holding the mountains up.

sedimentary rock (and also any igneous rocks that might have been injected into them as magma), squeezing them simultaneously upwards to form a mountain range and downwards, deeper into the mantle, to form the mountain's root (Fig. 1.4). It also squeezes the pre-existing rocks of the continental crust beneath the sediments. The entire continental crust might be doubled in thickness during this process.

As noted in Appendix 1, continental crust is less dense than the mantle, so it is buoyant. The mountain's root behaves like the underwater portion of an iceberg. Just as an iceberg floats in the sea with its top towering above the waterline, so the thickened continental crust floats, as it were, in the mantle with its top sticking up as mountains. This analogy must not be taken too far, though. While the mountains and their roots are imagined as floating in the mantle, the mantle is not liquid; it is not composed of molten rock, as some school textbooks suggest when inadvertently confusing *mantle* with *magma*. The mantle is known to be solid because, as is explained in section A1.1 of Appendix 1, seismic S-waves can travel through it.

1.2.2 Directed stress, pressure and temperature in a mountain's roots

As the rocks in the mountain's deep root become hot and squeezed, so they become metamorphosed. As the new

minerals grow, the directed stress that causes the crust to become much thicker also causes the newly formed metamorphic rocks to become foliated. Slate, phyllite, schist and gneiss commonly develop. Images of these rocks are shown in chapter 2. The plane of the foliation is represented by vertical dashes in Figure 1.4, and is usually roughly at right angles to the direction in which the mountain's root is being squeezed.

Directed stress should not be confused with a related term, **lithostatic stress**. Lithostatic stress is the same thing as **pressure** (abbreviated 'P'). Pressure increases with depth and has the effect of making a rock denser by squashing it equally in all directions into a smaller volume. Thus, high pressure favours mineral combinations that are denser than those formed at low pressure. Pressure in metamorphism is usually measured in **kilobars** (kbar); a depth of 35km corresponds roughly to a pressure of 10kbar. Directed stress, in contrast, does not affect a rock's volume. Instead it causes a rock to change in shape by being flattened or sheared, as captured in the cartoon of stressed and pressurized divers in Fig. 1.5. The technical word for a change in shape is **strain**; so directed stress causes strain.

The temperature (abbreviated 'T') of metamorphism is denoted in a loose way by the term **grade**; metamorphism can be low-, medium-, or high-grade. Most geologists use the term

low-grade for temperatures between about 300°C and 500°C, medium-grade for temperatures between about 500°C and 700°C, and high-grade for temperatures between about 700°C and 900°C, as shown by the coloured areas in Figure 1.6. Slate and phyllite are low-grade rocks, schist is typically a medium-grade rock, and gneiss is usually a high-grade rock.

Describing rocks as low-, medium- or high-grade is not the only way of referring to the temperature conditions of their formation. Two other ways are by stating the so-called metamorphic **zone**, and by stating the so-called metamorphic **facies** (pronounced either fash-eez or fay-sheez). Metamorphic zones are based on key indicator minerals, like garnet, that develop in metapelite (metashale) as the grade increases. Metamorphic facies are based on distinctive combinations of minerals that develop in rocks of basaltic composition as the grade increases, and also as the pressure increases. Zones and facies will both be introduced fully in chapter 2. For the present, the names of three important facies – greenschist, amphibolite (pronounced am-**fib**-a-lite) and granulite – are shown in Figure 1.6. They correspond respectively to low-, medium- and high-grade conditions in rocks formed at intermediate pressures. Rocks formed at high pressures are different, and have their own facies names, as will be explained in chapters 2 and 4. Ways in which the P and T conditions of metamorphism can be quantified are considered in chapters 3 and 4, and especially in chapter 5.

Metamorphism has lower and upper temperature limits. At temperatures below about 300°C metamorphism barely gets started, and any changes tend to be imperceptible to the unaided eye. In sedimentary rocks the process of change at these temperatures is called **diagenesis** (pronounced dye-a-genesis) rather than metamorphism. Diagenesis accompanies and follows lithification. Between roughly 200°C and 300°C it is sometimes called very-low-grade metamorphism, and is also called **anchimetamorphism** (pronounced anki-metamorphism). This temperature range is referred to as the **anchizone** (Fig. 1.6).

At the upper end of the temperature scale, high-grade metamorphism causes some rocks, particularly metapelite, to partially melt, and a kind of gneiss called **migmatite** is produced. The nature of migmatite is described in chapter 2, and its origin is explained in chapter 3.

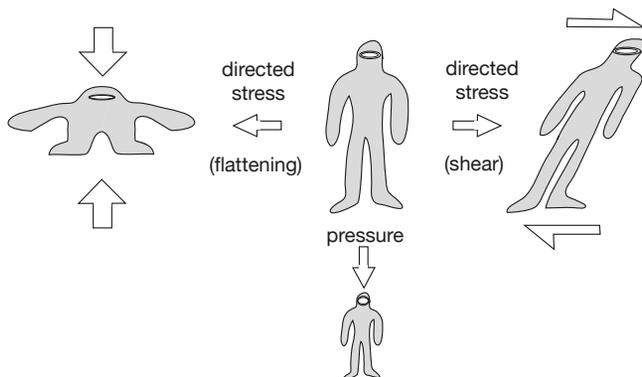


Figure 1.5 Cartoon highlighting the distinction between pressure (sometimes called lithostatic stress) and directed stress. The former causes a reduction in volume, the latter causes strain (change in shape), by flattening or shearing, with no change in volume.

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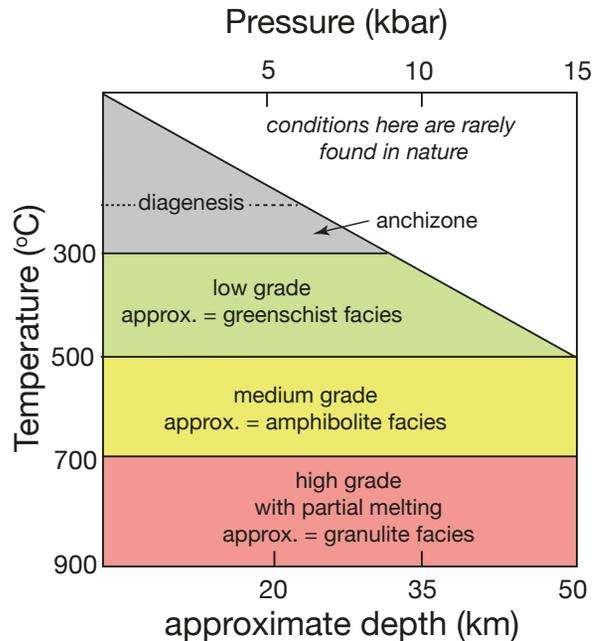


Figure 1.6 Diagram showing the approximate range of pressure (P) and temperature (T) conditions at which metamorphism occurs in the continental crust. Temperature bands are shown for low-grade, medium-grade, and high-grade metamorphism. P–T conditions in the white triangle are rare to absent in nature because temperature almost always increases with depth at more than about 10°C per km (equivalent to 500°C at 50km). For an explanation of other terms, see the text.

1.2.3 Exhumation of a mountain's roots

Mountain building is over when, after perhaps a few million years, the convergent **plate** motions cease. Erosion will then steadily reduce the height of the mountains, and the debris will be carried downhill to accumulate as sediment in neighbouring low-lying places. As the mountains are stripped away, their buoyant roots will continually adjust to the reduced load by floating upwards and maintaining the mountains, though at a lower height than before. Erosion will come to an end only when the root has entirely gone, and the continental crust has been restored to its original stable thickness, typically of about 35km. A huge volume of rock, as indicated in Figure 1.7, will by then have been removed, and the topography will have been reduced almost to sea level. An enormous swathe

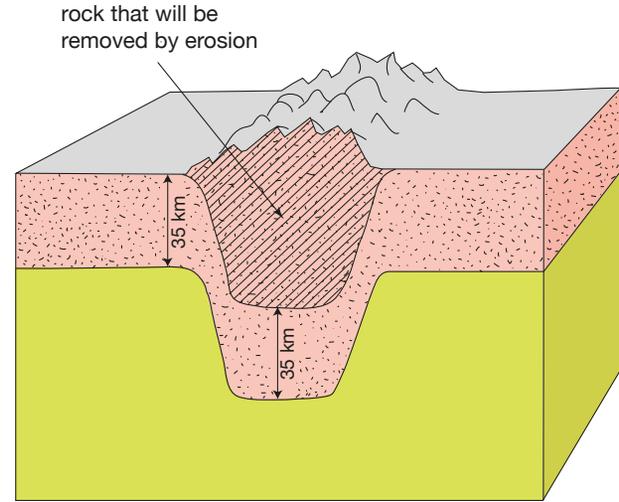


Figure 1.7 The same block diagram as in Figure 1.4, but here showing the volume of rock (diagonal lines) that will be removed by erosion once the lateral compression is over. After erosion is complete, metamorphic rocks from a considerable depth will be at the surface.

of metamorphic rocks that were formerly deep within the root will have become exposed at the surface. This swathe of rocks is known as a regional **metamorphic belt** or an **orogenic belt**. Rocks that were formed at higher grades, like schist and gneiss, will tend to be in the centre of the belt, with lower-grade rocks, like slate and phyllite, located towards the margins.

Metamorphic belts are typically tens to one or two hundred kilometres wide and several hundred, up to one or two thousand, kilometres in length. Metamorphic belts of various ages can be examined in different parts of the world today. One of them, called the Grampian belt, provides many of the examples of metamorphic rock described here. The Grampian belt extends diagonally in a NE to SW direction across Scotland and north-western Ireland. It is the vestige of a mountain range that existed about 470 million years ago.

1.3 Metamorphism in local settings

Contact, hydrothermal, dynamic and shock metamorphism are four kinds of metamorphism that, in contrast to regional metamorphism, are quite restricted in their geographical extent. They are briefly introduced here, and aspects of them are developed in later parts of the book.

1.3.1 Contact metamorphism

Contact metamorphism happens next to bodies of hot igneous rock beneath the ground. The heat from a body of molten granite 10km in diameter will, for example, affect the surrounding cool rock (called the **country rock**) and bake it up to a distance of about 1km from the contact. A granite body like this is called a **pluton** (pronounced **ploo-ton**) from Pluto, Greek god of the underworld. New metamorphic minerals and textures appear in the surrounding baked zone, which is known as a **metamorphic aureole**, a **thermal aureole** or a **contact aureole** (Fig. 1.8). Temperatures are obviously highest right next to the granite and decrease outwards into the cool country rock. A common product of contact metamorphism is a fine-grained, tough and somewhat flinty rock without foliation called **hornfels**. Examples of hornfels derived from shale and basalt are among the rocks described in chapter 2, and

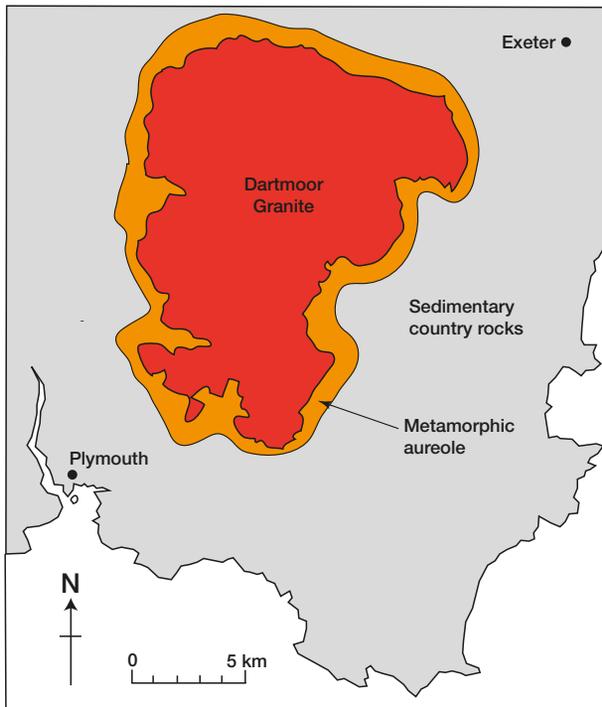


Figure 1.8 Map showing the thermal aureole surrounding the 280 million-year-old pluton of granite exposed at Dartmoor in Devonshire, England. The country rock is composed of sedimentary rocks that were deposited about 400 million years ago. *After R. Mason.*

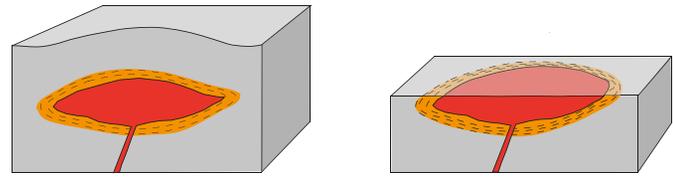


Figure 1.9 Simplified block diagrams of a granite pluton (red) and its metamorphic aureole (orange) shortly after it was intruded (left) and today (right), after the overlying rock and the top of the pluton have been stripped off by erosion.

further examples of contact metamorphic rock are discussed in the first section of chapter 4. Contact metamorphism generally happens at depths of a few kilometres so, as with orogenic belts, the land surface must rise and the overlying rocks must become stripped off by erosion for the pluton and its metamorphic aureole to be exposed today (Figs 1.8, 1.9).

1.3.2 Hydrothermal metamorphism

Hydrothermal metamorphism is caused by hot water. It happens in volcanic regions where shallow bodies of underground magma heat the nearby cold **groundwater**. Groundwater is rainwater or seawater that has drained down and filled any voids and crevices beneath the surface. The heated water is less dense than nearby cold water, so it rises (red arrows in Fig. 1.10), flowing through interconnected pores and cracks.

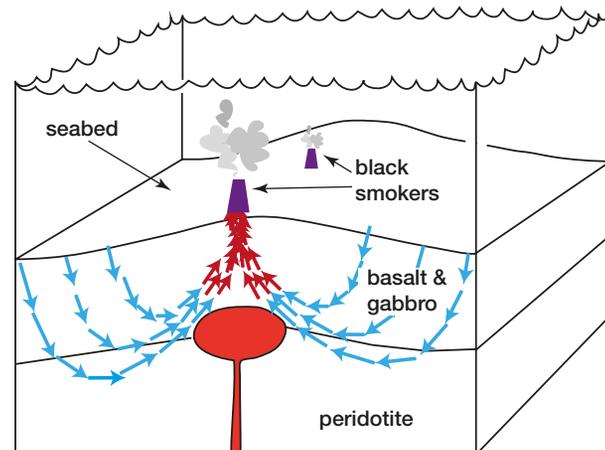


Figure 1.10 Diagram showing the flow of seawater through interconnected fractures in rock beside a sub-oceanic magma chamber at a spreading ridge. For an explanation see the text.

8 As it moves upwards it is replaced by colder groundwater (blue arrows in Fig. 1.10) in a process known as thermal **convection**. The flowing hot water reacts with the minerals in the rocks through which it passes, creating new hydrous minerals. Hydrothermal metamorphism is particularly important in the vicinity of newly formed basaltic rocks at so-called **spreading ridges** where tectonic plates are moving apart (see Appendix

1, section A1.2). Cameras in submersibles at such places have filmed rising plumes of superheated dark, cloudy water called **black smokers** streaming from vents on the seabed (Fig. 1.11). The uptake of water by basaltic rocks during hydrothermal metamorphism is discussed in the first part of chapter 3, and its implications for volcanism above subduction zones are mentioned in chapter 4.

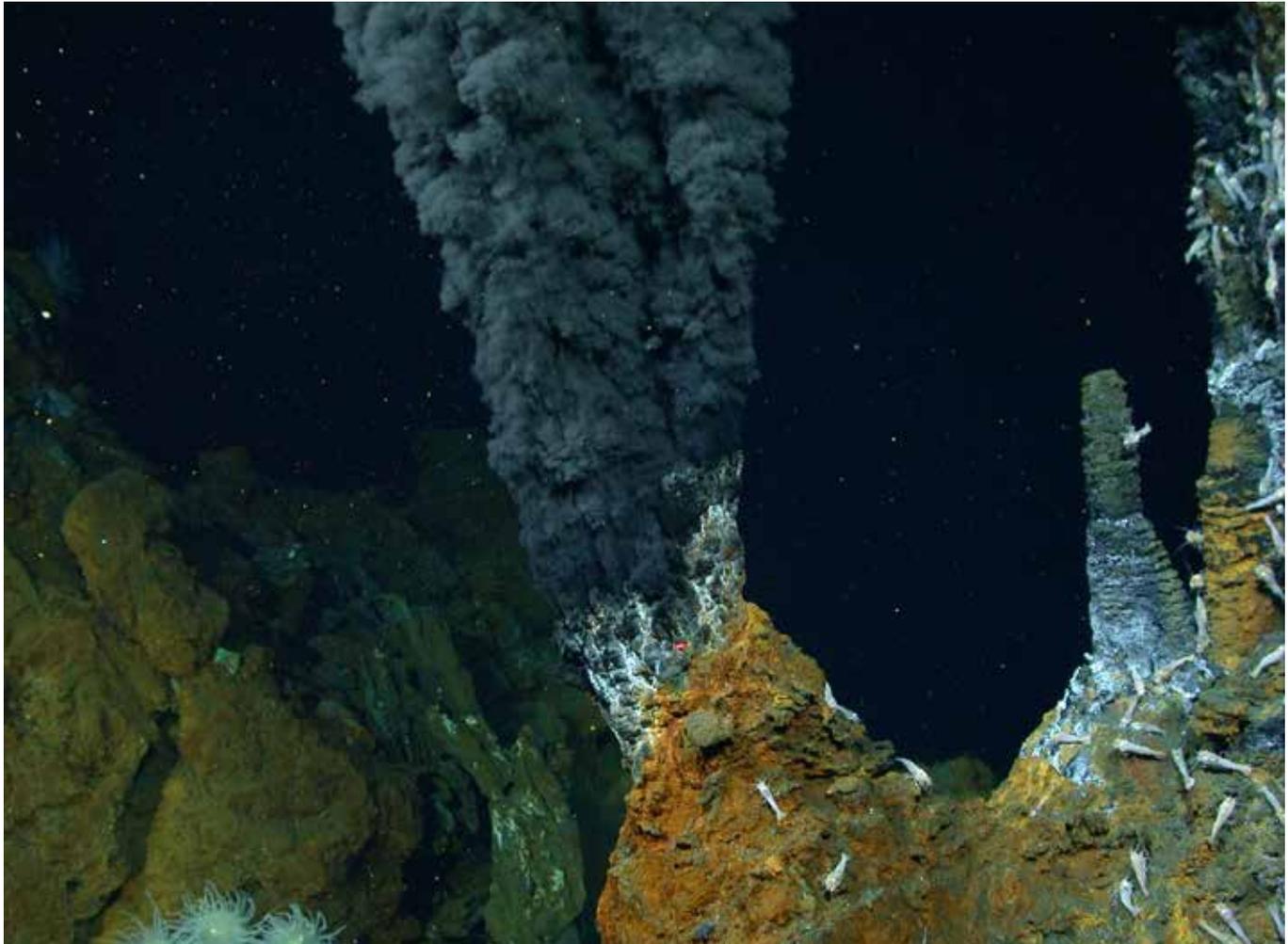


Figure 1.11 One of the Earth's hottest and deepest black smoker hydrothermal vents which is located on the Mid-Cayman Rise in the Caribbean. The temperature of the expelled H_2O -rich fluid is about 400°C . The 'smoke' (centre of field against black background) consists of dark, fine-grained particles suspended in the fluid. The particles come out of solution as the fluid cools and mixes with cold seawater. Deep-sea shrimps and other life-forms appear to thrive in this bizarre pitch-black environment. *Photo courtesy of Chris German, WHOI/NSF, NASA/ROV Jason 2012©Woods Hole Oceanographic Institution.*

1.3.3 Dynamic metamorphism

Dynamic metamorphism is a localized variety of metamorphism that occurs within major fault zones, and is caused by fault movement. In the top few kilometres of the continental crust it involves the fracturing and comminution of pre-existing rocks, in a process called **cataclasis** (pronounced cata-**clay**-sis). The products are described as cataclastic rocks, or **cataclasites**. Where the faulted rocks are quite cool and shallow they usually end up as a chaotic, loosely-bound mixture of angular broken rock fragments and rock dust called **fault breccia** (pronounced **bretch**-ya). If the product is nearly all pulverized, with few fragments, it is called **fault gouge**. Many geologists do not consider fault breccia and fault gouge to be metamorphic rocks because they are merely loose, unconsolidated mixtures of crushed rock fragments. At greater depth, however, where it is warmer and the grains are

pressed more tightly together by the weight of overlying rocks, the faulted rocks often end up as a strong, extremely fine-grained, streaky-looking rock called **mylonite** (pronounced **my**-la-nite; Fig. 1.12). A fault zone at such depths is called a **shear zone**. Shear zones are evolving at depth in the continental crust today wherever major faults are active, such as at plate boundaries (see Appendix 1) and beneath mountain ranges. Various products of dynamic metamorphism, particularly as it affects granite, are described in chapter 2, and the mechanism is discussed in the second part of chapter 3.

1.3.4 Shock metamorphism

Shock metamorphism is an extremely violent process that affects the rocks below the impact site of a giant meteorite. It is a localized and, thankfully, rare phenomenon;



Figure 1.12 Banded and slightly crumpled mylonite from a kilometre-wide, inclined shear zone that can be traced for a great distance immediately to the east of a fault line in NW Scotland known as the Moine Thrust. The Moine Thrust marks the western margin of the Grampian orogenic belt, shown mainly in yellow on the map in the photo.



Figure 1.13 Barringer Meteorite Crater in Northern Arizona. The crater is 1.2km in diameter and was excavated some 50,000 years ago by a huge explosion following the impact of a chunk of iron from space thought to have been travelling at about 20km per second and measuring 50 metres across. This is the best-preserved crater of almost 200 ancient impact craters that have so far been recognized on the Earth's surface. The severe damage caused to the rocks beneath the crater by shock waves from the explosion is known as shock metamorphism.

such impacts have not been observed in recorded history. The meteorite explodes on striking the ground because its kinetic energy (the energy due to its speed) is instantly turned into intense heat that vaporizes the meteorite and the adjacent rock below the point of impact. The explosion excavates a crater (Fig. 1.13). The rocks below the crater become crushed, shaken, and melted as intense shock waves pass through them. Shock metamorphism is like an extreme kind of dynamic metamorphism. Evidence for it is

preserved only rarely on Earth because the surface is continually being refreshed by geological processes, but evidence is widespread on the surfaces of the Moon, Mars and Mercury, and of small bodies in the Solar System. More about shock metamorphism is presented in the last section of chapter 4.

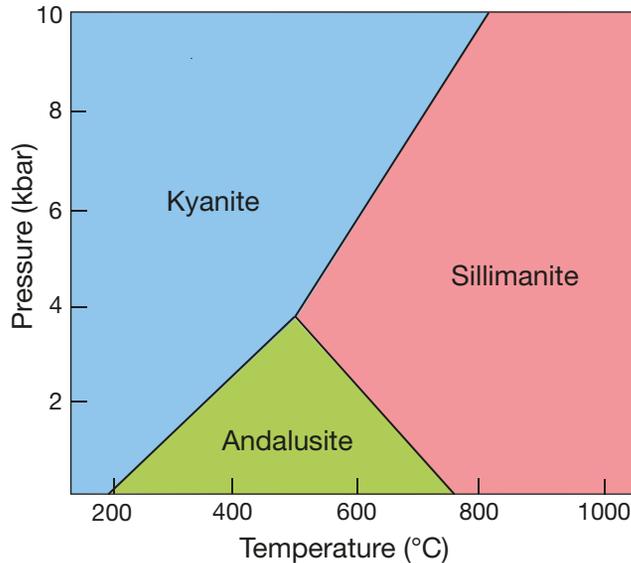


Figure 2.32 Pressure–temperature diagram for Al_2SiO_5 showing the stability fields of kyanite, andalusite and sillimanite.

for example, then it is presumed to have been metamorphosed under P–T conditions within the blue area in Figure 2.32, called the kyanite stability field. The meaning of *stability*, and the way in which the three stability fields were determined, will be explained in chapter 3. For the present, it can be noted that the Al_2SiO_5 diagram has an obvious bearing on interpreting metapelites. For example, it shows that the pressure during *regional* metamorphism (no andalusite) must have been greater than about 4kbar because the kyanite zone is next to the sillimanite zone. It also shows that andalusite hornfels must have been formed at pressures less than about 4kbar.

2.3 Marble

Metamorphosed limestone is known as **marble**. The limestone protolith is almost always made from the calcium carbonate (CaCO_3) skeletal remains of planktonic marine organisms, together with shells from larger creatures, which accumulate on the seabed before becoming buried and lithified to make hard limestone. The organisms extract calcium and CO_2 from seawater where they are in solution (calcium is a soluble product of chemical weathering, and is carried by rivers to the sea).

Incidentally, this process, operating over geological time, has continuously removed CO_2 from seawater, increased the amount of limestone in the continental crust, and kept CO_2 in the atmosphere at a steady low level. Today the rate of removal of CO_2 seems unable to keep pace with the rapid anthropogenic production of CO_2 , and the level of atmospheric CO_2 is rising alarmingly year on year.

2.3.1 Pure calcite marble

The simplest kind of marble is made entirely of the mineral **calcite** (CaCO_3). Pure calcite marble is usually white or pale grey in colour, and when it is crushed it breaks into blocky pieces with no foliation, so at first sight it can resemble quartzite. However, it is much softer and more easily abraded than quartzite; crushed marble would be of little use as railway ballast. Its softness means that marble can be scratched, scraped and chiselled with ease, making it an excellent choice of stone for carving. It has been used throughout history in sculpture and as a decorative stone in buildings. Notable in this regard is the white marble from the quarries at Carrara, perched on the mountain slopes high above the Ligurian Sea in Tuscany, northern Italy (Fig. 2.33). Michelangelo created his masterpiece *David* (Fig. 2.34) from a carefully chosen block of Carrara marble.



Figure 2.33 Marble quarry at Carrara, Tuscany, Italy, looking west. The Ligurian Sea is in the far distance. Shutterstock/Alessandro Colle.



Figure 2.34 Michelangelo's famous masterpiece *David*. It was carved from a block of marble from Carrara. *Shutterstock/javi_indy*.



Figure 2.35 A block of clear, transparent calcite obtained by splitting a larger crystal of calcite along cleavage planes in three directions. Clear calcite like this is called Iceland spar. *Shutterstock/Eduardo Estellez*.

Marble is amenable to being carved because of the crystalline properties of calcite. Examination of a large crystal of calcite shows that it can be split easily along internal cleavage planes, but instead of having a single direction of cleavage, like muscovite, calcite has three cleavage directions (Fig. 2.35).

Every grain of calcite in marble has these three directions of cleavage within it so, when the rock is being carved, it will fracture easily in any direction chosen by the sculptor. Also, a freshly broken surface of a marble specimen will glitter as it is turned and the light catches the tiny, perfectly flat, mirror-like surfaces where individual grains have parted along a cleavage plane. The grains in marble, and Carrara marble is no exception, tend to be evenly sized. This texture is sometimes described as saccharoidal (sugar-like). Another term used for even-grained rocks like this is **granoblastic**.

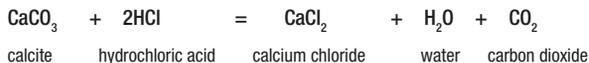
Marble is not only easy to carve, but it readily takes a shine when buffed with a polishing paste. Polished marble tiles and worktops have become something of a fashion statement in people's homes, but unwitting house-proud home owners may find themselves dismayed when they discover, too late, that polished marble, being so soft, is very easily scratched! Worse still, the polished surface of marble is easily etched and stained by weak acids like vinegar and lemon juice. Such acids react spontaneously with the calcite grains in the marble, dissolving them.

Geologists take advantage of this chemical reaction with acid when trying to distinguish marble (or its limestone protolith) from rocks that contain no calcite. The surface of the rock is wetted with a drop of dilute hydrochloric acid from a dropper bottle. If the rock contains calcite it will effervesce vigorously as CO_2 gas is released (Fig. 2.36).



Figure 2.36 Effervescence of CO_2 gas as a drop of dilute hydrochloric acid wets a piece of marble.

The chemical reaction that takes place is written:



It may come as a surprise to learn that some pure marbles are not made of calcite. Calcium carbonate also exists in nature in another crystalline form or polymorph, named **aragonite**. Aragonite has a higher density than calcite, and in exceptional metamorphic settings where the pressure is abnormally high (as will be reviewed in chapter 4), marble can be made from aragonite. In fact, many shells of marine organisms are made from aragonite rather than calcite. However, aragonite in shells, and in aragonite marble, is not as stable as calcite at the Earth's surface, and it is nearly always in the process of changing slowly into calcite.

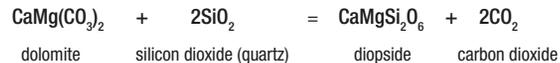
2.3.2 Impure marble

Some varieties of limestone contain the calcium–magnesium carbonate mineral named **dolomite**. Such limestone becomes dolomitic marble when it is metamorphosed. Dolomite contains equal numbers of calcium and magnesium atoms; its formula is $\text{CaMg}(\text{CO}_3)_2$. Compared with calcite, it is equally soft, but is usually cream coloured rather than white, and it does not react vigorously with dilute hydrochloric acid unless it has first been crushed to powder. Dolomite is formed by the natural replacement of exactly half the calcium atoms in calcite by magnesium atoms from seawater during a process called dolomitization. The process takes place on, or just beneath, the seabed in tropical latitudes where calcite is accumulating (e.g. as shells) and seawater is evaporating. It is an example of **diagenesis**, the name for changes that happen to rocks below the temperatures normally associated with metamorphism, as was outlined in chapter 1.

Dolomitic limestone may contain silica (SiO_2), for example as quartz sand grains or as the variety of silica known as opal, which forms the skeletons of certain planktonic organisms like radiolaria and diatoms. After such limestone has been metamorphosed, the resulting dolomitic marble will normally contain magnesium-bearing silicates like diopside and forsterite. Both minerals are introduced in Appendix 2. **Diopside** is a white mineral (see Fig. 2.37) with good cleavage, whose formula is $\text{CaMgSi}_2\text{O}_6$. It is an example of a kind of silicate known as clinopyroxene. **Forsterite** is green; its formula is Mg_2SiO_4 . It is a variety of

olivine, a mineral that is described in Appendix 1 as the main constituent of the Earth's mantle. Forsterite is the so-called magnesian end-member of olivine, as explained in Appendix 2.

Diopside is formed by a simple chemical reaction between dolomite and quartz:



Forsterite is formed by a related reaction:



An example of forsterite-diopside marble in the field is illustrated in Figure 2.37, and a thin section of the same rock is shown in Figure 2.38.

The thin section photographs (Fig. 2.38) are screen shots of VM GeoLab M05, forsterite marble. The rock is well worth exploring online. Calcite and dolomite, the two carbonate minerals, look alike; they are both colourless with high relief. However, they can be distinguished here because the surface of the thin section was treated with a pink dye, called alizarin red S, before the glass cover



Figure 2.37 A rock outcrop of forsterite-diopside-marble at Glenelg in NW Scotland. The forsterite forms small, rusty-looking granules protruding from the surface. The diopside forms white protruding grains, and it also forms the large, rounded white lump on the left. The silicates protrude because they do not dissolve as easily as the calcite and dolomite during weathering.

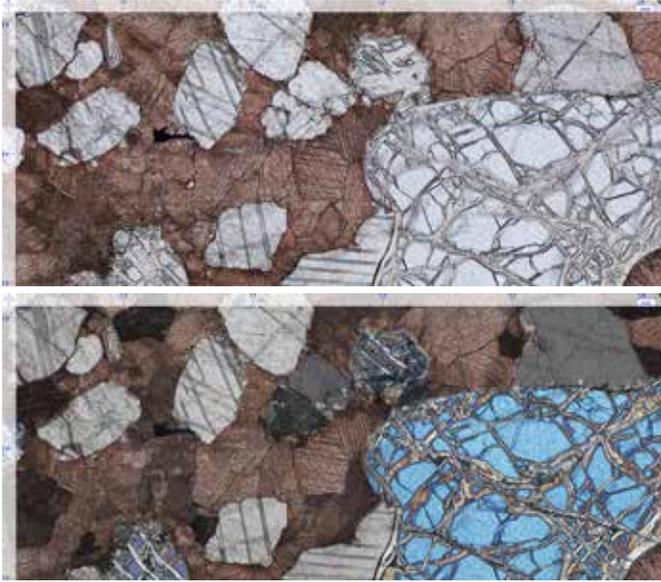


Figure 2.38 Screen shots of VM GeoLab forsterite marble, M05, in PPL (upper image) and XP. The pink colour shows where calcite has been stained. Sets of straight parallel bands are deformation twin lamellae. Forsterite (Mg-olivine) lower right is traversed by narrow veins of serpentine. The width of the image is 5mm.

slip was attached (for an account of how thin sections are made, see Appendix 3). The dye stained the calcite but did not affect the dolomite. Online examination will show how half the thin section was treated in this way. Both minerals commonly display one, two or even three sets of narrow, parallel, coloured bands that show up in XP, and may even be visible in PPL. These are known as **deformation twin lamellae**. They were created when the marble was subjected to stress and became strained. The interference colour of calcite and dolomite in XP is creamy white (often with a hint of twinkly pinks and greens), known as high-order white (see Appendix 3). The colour remains unchanged when a special filter, called a sensitive-tint plate, is inserted into the light path in the microscope. This test allows carbonates to be identified; nearly all other minerals change colour in XP when the sensitive-tint plate is inserted. Unfortunately, this useful test does not currently feature in the Virtual Microscope.

Forsterite occurs as colourless, high-relief grains with rounded outlines. It has bright interference colours in XP and any grains with roughly rectangular shapes tend to have parallel extinction. Each grain is traversed by a network of narrow channels, called veins, filled by a low-relief mineral with grey interference colours. This is the mineral **serpentine**, a water-rich sheet silicate mineral whose formula is $Mg_6Si_4O_{10}(OH)_8$ (see Appendix 2). It formed here at some late stage in the marble's history, when water seeped in and made its way along cracks in the forsterite, converting it to serpentine. The reaction is an example of what is known as **retrograde** alteration. In a few cases the alteration has progressed to the point where no forsterite (or very little) remains (Fig. 2.39, upper pair of images) and the resulting lump of serpentine, where once there had been forsterite, is known as a **pseudomorph** of serpentine after forsterite.

Diopside has the same high relief and second order interference colours as forsterite. It can be distinguished from forsterite because it does not have veins of serpentine, and it may show parallel cleavage cracks (Fig. 2.39, lower pair of images). Grains with cleavage have a very high extinction angle (see Appendix 2), i.e. they go into extinction when the cleavage lines are typically between 30° and 45° from the nearest cross hair.

The above marble was formed under high-grade conditions. At medium grade a different magnesium-silicate mineral, **tremolite**, generally occurs instead of forsterite and diopside, giving tremolite marble. It forms long white prisms (Fig. 2.40). Tremolite has the formula $Ca_2Mg_5Si_8O_{22}(OH)_2$, and is the simplest example of a clinoamphibole. Although its formula looks complicated, it is easy to remember when it is related to the crystal structure of tremolite (see section A2.4.3 of Appendix 2).

Some impure marbles occur as serpentine marble. The serpentine may have formed by the hydration of former olivine (as in Fig. 2.39) but, if water were present during the original metamorphism of the siliceous dolomitic limestone, then the serpentine may have formed directly, as a primary metamorphic mineral. Green serpentine marble has for many years been quarried and carved for use as an ornamental stone (Fig. 2.41).

Limestone may be rendered impure not only by having an admixture of quartz, but also as a result of containing a high proportion of clay minerals (mud), giving it a

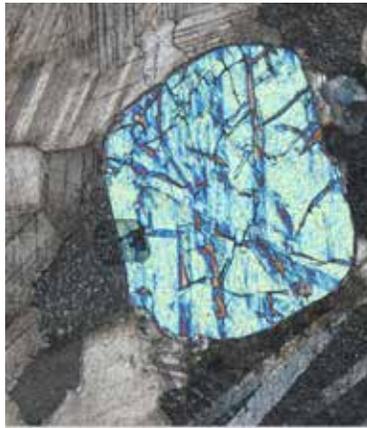


Figure 2.39 Pairs of images in PPL (left side) and XP, each about 1mm wide from forsterite marble, M05. *Top* Forsterite almost entirely replaced by serpentine, making a so-called pseudomorph of serpentine after forsterite (right of centre). *Bottom* Diopside with a hint of cleavage and no alteration to serpentine. The grain is at rotation 2 where its highly inclined extinction can be checked.



Figure 2.40 Specimen of tremolite.
Wikimedia Commons photograph by Didier Descouens.



Figure 2.41 Carved and polished green serpentine marble from Connemara, western Ireland, decorating the interior of the Museum Building at Trinity College Dublin.

transitional composition between limestone and shale. Such a sedimentary rock is known as **marl**. When marl is metamorphosed the resulting marble will contain silicates with calcium (from calcite) and aluminium (from shale), such as the minerals epidote, grossular (calcium garnet), or perhaps anorthite (Ca-plagioclase). **Epidote** is a dense yellowish green mineral with the chemical composition $\text{Ca}_2(\text{Al}, \text{Fe}^{3+})\text{Al}_2(\text{SiO}_4)_3(\text{OH})$, as outlined in section A2.4.1 in Appendix 2. The iron it contains is Fe^{3+} , and it substitutes for one of the three Al^{3+} ions in the formula, but little Fe^{3+} occurs in marbles. The appearance of epidote in thin section will be described more fully in the forthcoming section on metamorphosed basic igneous rocks. **Grossular** garnet has a somewhat related formula to epidote, $\text{Ca}_3\text{Al}_2(\text{SiO}_4)_3$, and

anorthite is the Ca end-member of plagioclase with the formula $\text{CaAl}_2\text{Si}_2\text{O}_8$ (see Appendix 2). These two minerals occur particularly in impure marbles in contact aureoles, and more will be said about them in chapter 4.

2.3.3 Metasediments with mixed compositions

Sedimentary rocks with transitional compositions exist not only between limestone and quartz sandstone, and between limestone and shale, as just described, but they also exist between quartz sandstone and shale where intermediate rocks include ordinary sandstone, muddy sandstone (called greywacke), and siltstone. Compositions in the middle of this range give rise to semipelitic metamorphic rocks.

Where the amount of ‘contaminating’ mud or quartz in limestone is considerable, metamorphic reactions may consume all the calcite and dolomite, leaving a rock made entirely from silicates. Such a rock is known simply as a **calc-silicate** rock, and, depending on its protolith, it may be dominated by Ca-Al silicates like epidote, grossular and anorthite, or by Ca-Mg silicates like forsterite, diopside and tremolite.

The full variation in the composition of the three common kinds of sedimentary rock can be represented on a **triangular diagram** with sand, carbonate and mud at its three corners (Fig. 2.42). Using this triangular diagram, the names of sedimentary rocks are shown in Figure 2.43 and the names of the

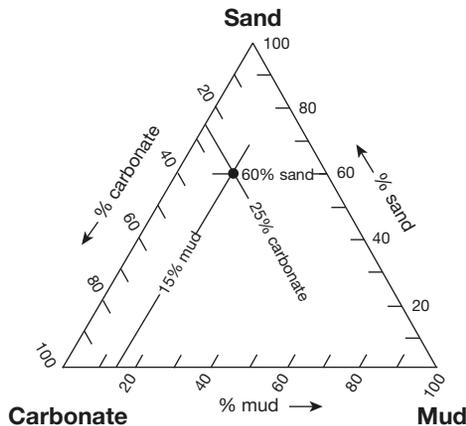
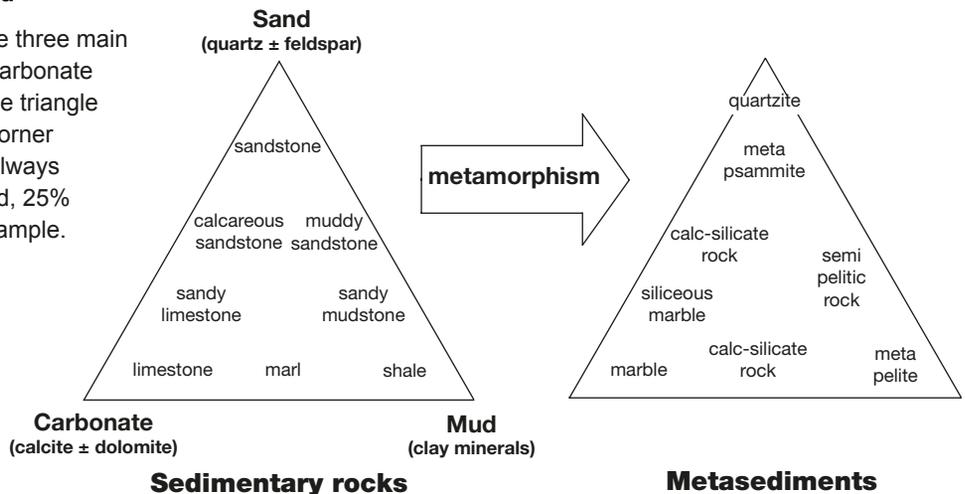


Figure 2.42 (above) Triangular diagram with the three main components of sedimentary rocks – sand, carbonate and mud – at its corners. Any point within the triangle represents some combination of the three corner components, and the sum of the three will always be 100%. A sedimentary rock with 60% sand, 25% carbonate and 15% mud is shown as an example.

Figure 2.43 (right) The same triangle as in Figure 2.42 showing, on the left, the names of sedimentary rocks with different combinations of sand, carbonate and mud, and, on the right, the names of the equivalent metasediments.



compositions, as already shown, fall near the middle of the triangle.

The diagram does not handle metapsammites or metagranites well, because the minerals in these rocks are mainly quartz, Na-plagioclase and K-feldspar. These three minerals fall outside the triangle, and so, therefore, do the compositions of the metagranites and metapsammites made from them. These last two rock types, nevertheless, can be imagined as being located at separate compositional 'corners' above the plane of the triangle, with metagranites joined to the metabasites via intermediate igneous rocks, and with quartzite and metapsammites joined to the metapelites via semipelitic compositions (Fig. 2.79).

Figures 2.78 and 2.79 together provide a useful *aide memoire* for dealing with unknown metamorphic rocks. Once a rock's minerals have been identified, then, with the help of the ACF triangle (Fig. 2.78), the approximate overall composition of the rock can be estimated, and the protolith can be inferred.

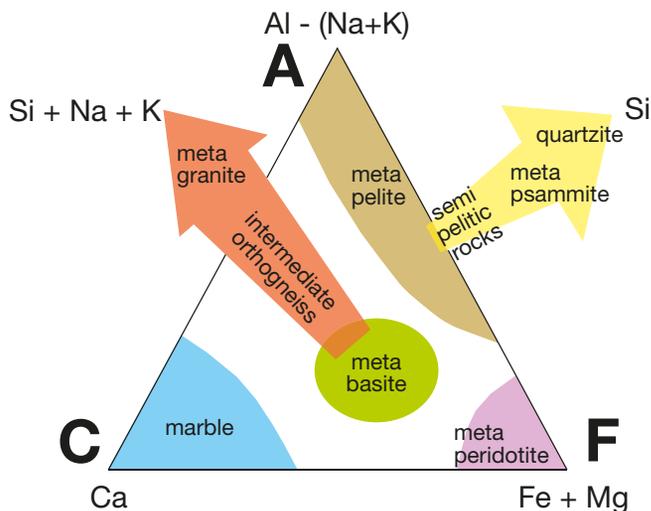


Figure 2.79 ACF triangle showing the compositional fields of four of the six protoliths, with the other two 'floating' above the triangle at the ends of the arrows shown. Metagranite is on an extension from the green metabasite field via intermediate igneous rocks, and quartzite and metapsammite are on an extension from the metapelite field via rocks with semipelitic compositions.

2.7.2 Where do the six protoliths come from?

Peridotite was said in chapter 1 to be like a 'mother rock' from which, ultimately, the other five protoliths are derived. To demonstrate its parental role and the line of descent to its offspring, a flow chart with numbered steps is presented in Figure 2.80. This diagram is the so-called **rock cycle** for the continental crust, modified here by adding partial melting of the mantle. It summarizes all the geological processes described in the book so far.

The standard rock cycle outlines how rocks in the continental crust are changed from one kind to another, or 'recycled'. It looks complicated (Fig. 2.80) but, taken step by step, it is logical and simple to follow. In the continental crust it has a main circuit (thick black line in Fig. 2.80), and several short-circuits. A convenient starting point is the **weathering** and erosion of rocks exposed at the surface (number (1) in Fig. 2.80). Weathering divides rocks into soluble and insoluble fractions, which are transported (2), mostly by flowing water, to lower levels. Soluble elements like K, Na, Ca and Mg are mainly carried in solution, while elements like Fe, Al and Si are mainly moved along as flakes of mud and grains of sand. Either way the material ends up accumulating (3), usually on the seabed, as sand and mud, and also as calcium carbonate shelly material from organisms that extract soluble Ca and CO₂ from seawater. The debris becomes buried and compressed beneath further accumulations of sediment. As it gets deeper it also gets warmer and turns into hard sedimentary rocks through the process of **lithification** (4). Sand becomes sandstone, mud becomes mudstone or shale, and calcium carbonate shells become limestone.

With deeper burial and more intense heating, assisted by orogenesis, metamorphism begins (5) and the sedimentary rocks change into rocks such as schist, marble and quartzite, which are the subject of this book. With yet further heating, schist partially melts (6), becoming migmatite, and granitic **magma** is a by-product. This magma gathers at depth, then rises through the crust because of its low density (7). The main cycle then divides: the rising magma may either travel directly to the surface and erupt as volcanoes (8), or it may form plutonic intrusions (9) that will, later, be brought to the surface by uplift and erosion (10). Either way, the granite and its volcanic equivalent will be subjected to weathering and erosion, and the cycle will begin again.

Four 'short-circuits' to the main loop are shown by thin

black lines, two of them dashed, in Figure 2.80. Firstly, sedimentary rocks can return directly to the surface by uplift and erosion (11) without becoming metamorphosed. Secondly, metamorphic rocks can be exhumed (12) without ever reaching the stage of partial melting. Thirdly, **plutonic igneous rocks** like granite can remain beneath the surface and become metamorphosed directly (13). Fourthly, volcanic rocks can similarly become buried and metamorphosed directly, without being weathered and eroded (14).

The rock cycle in the continental crust accounts for the formation of four of the protoliths, namely sandstone, shale, limestone and granite. To show how these four have descended from peridotite, and to account for the protolith *basalt*, mantle processes have been added. They are shown as green lines. Decompression melting of mantle peridotite produces basaltic magma (15), which rises into the continental crust (16) where it follows a similar path to granitic magma, via plutons or volcanoes, to the surface. At the surface, it is weathered, eroded and transported in

the normal way. Thus, basaltic magma can be thought of as a primary feedstock for the crust, and its parent rock, mantle peridotite, can be regarded as the 'mother rock' for all five other protoliths.

Two further pathways are shown in Figure 2.80. Firstly, some basaltic magma from the mantle ponds at depth in the continental crust (17) where it contributes heat for metamorphism and partial melting, making more granitic magma. Secondly, the small amount of peridotite within the continental crust is evidently added mechanically, as solid chunks (18), e.g. as ophiolite slabs, during plate convergence.

To conclude, metamorphic rocks lie in a box at the confluence of four inward streams. One brings in the three sedimentary protoliths, two streams (from plutons and from volcanoes) deliver the igneous protoliths, and one introduces peridotite from the mantle.

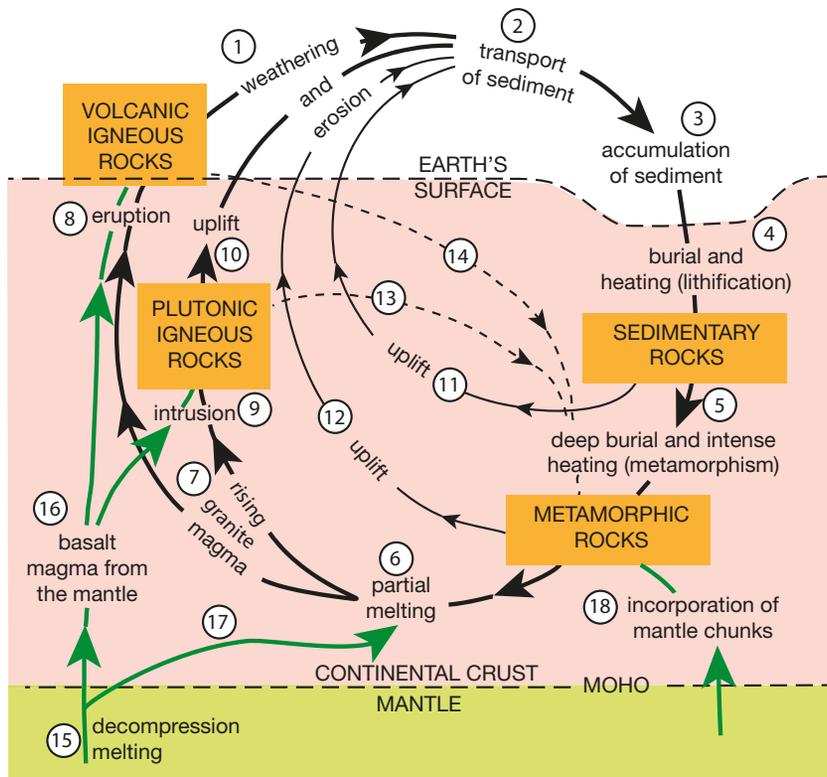


Figure 2.80 The rock cycle in the continental crust (black lines) and its links with the mantle (green lines). For an explanation see the text.

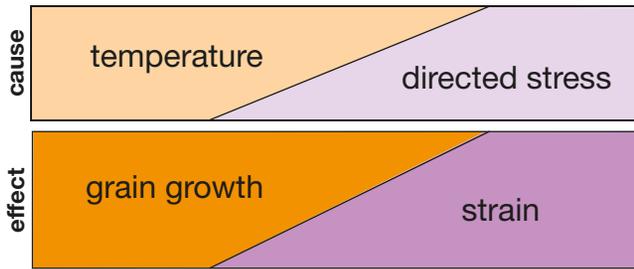


Figure 3.16 Diagram showing that temperature promotes grain growth, while directed stress promotes strain. Most rocks fall into one of three textural categories: those that have textures resulting from grain growth alone (e.g. hornfels), or strain alone (e.g. mylonite), or grain growth and strain together (e.g. schist).

example, in the discussion above on retrograde hydration, both the presence of serpentine along cracks in forsterite and the fine-grained hornblende and plagioclase along grain boundaries in eclogite point to the late ingress of water. In a third example, inclusion trails in garnet porphyroblasts (Fig. 2.26) provide a record of the early stages in the rock's history of deformation and grain growth.

The remainder of this section on textures addresses questions relating only to grain growth and strain. Further examples of how grain distribution provides evidence for a multistage history of metamorphism will be presented in chapters 4 and 5.

3.2.2 What makes grains grow?

Grain growth, which is also called **recrystallization**, is a fundamental metamorphic process. Grains can only grow in size at the expense of other grains that are getting smaller, eventually to disappear altogether. For grains to grow, atomic bonds in the smaller grains must be broken, so that the atoms are released and can then migrate and add on to larger grains. This process is favoured by high temperatures because temperature is a measure of the vigour with which atoms are vibrating and of the frequency with which bonds are being broken. The importance of high temperature is clear from the fact that grain size increases with metamorphic grade.

But while high temperature facilitates recrystallization, it does not explain *what* makes grains grow. To understand that, one needs to return to the idea of stability. It turns out that metamorphism is driven by a strong drive for a rock to

minimize its Gibbs energy, not just in its mineral assemblage, but also in its texture.

The Gibbs energy being minimized with grain growth is energy associated with the surfaces of grains, called surface energy. Surfaces are untidy places as far as atoms are concerned. Each atom on the boundary between two adjacent grains can only be bonded neatly to one of those grains. It will be out of alignment with the regular pattern of atoms in the adjoining grain (Fig. 3.17). Moreover, misfit atoms, too large or too small to be accepted by either grain, often end up stranded at the interface between two grains. This untidiness of grain boundaries gives rise to excess surface energy. The excess will be reduced if the total surface area is reduced, and this obviously happens with grain growth, when many small grains are replaced by a few large ones (Fig. 3.18).

Today it is possible to watch grain growth in real time, not in rocks but in metals, where the principles are the same but the atoms can move around much more easily. A special heated specimen stage inside a scanning electron microscope is used, and a succession of time-lapsed images is taken. Such grain growth can be viewed online at: https://www.youtube.com/watch?v=Cy_rYNc0UAY

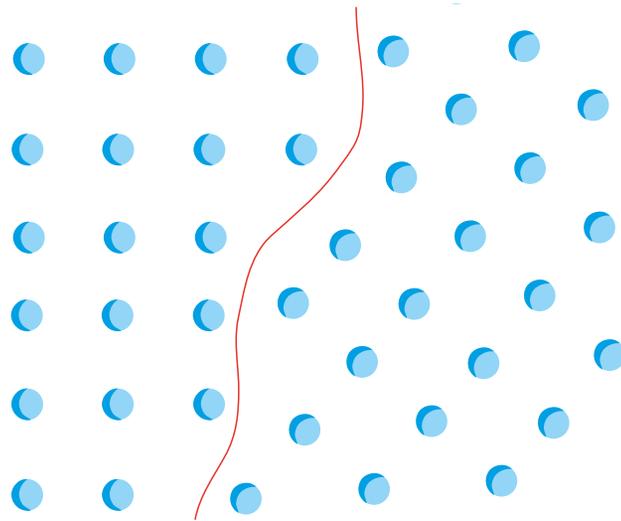


Figure 3.17 Cartoon depicting atoms at a boundary between two grains with different orientations. Atoms on the boundary cannot be bonded neatly to both grains, which results in a small excess of Gibbs energy, called surface energy.

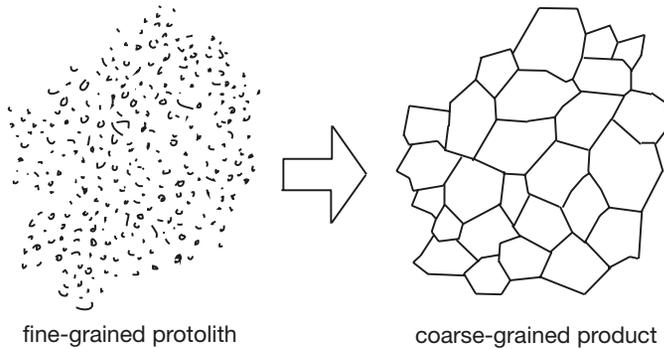


Figure 3.18 Cartoon depicting grain growth, before and after, for a rock made from just one kind of mineral.

In this youtube video a fine-grained piece of copper metal is heated, and new large grains can be seen growing at the expense of the small grains.

Two other videos provide insights into the way grains grow. In one of them Sir Lawrence Bragg, the Nobel Laureate in Physics who worked out how to interpret X-ray diffraction by crystals (see Appendix 4), is delivering the 1952 Royal Institution lecture, using rafts of equal-sized soap bubbles as a proxy for crystals, in which the bubbles are atoms: <https://www.youtube.com/watch?v=UEB39-1lmdw>.

The second video is a gem, a model in clarity of presentation, again relating to textures in metals but highly relevant to metamorphic textures: https://www.youtube.com/watch?v=uG35D_euM-0.

3.2.3 Is time, like temperature, a factor in grain growth?

While the correlation between metamorphic grade and grain size shows that temperature influences grain size, it is not the only influence. An additional factor is the *duration* of high-temperature conditions, or time. The importance of time is self-evident. It takes a long time for grains to grow to a large size. As noted above, strong chemical bonds first have to be broken so that atoms in the small, shrinking grains can move to the growing grains. But atoms are not easily able to tell which grains are good bets to join and which are not, so the process is one of trial and error, with many failed attempts by grains to grow before a few grains eventually prevail at the expense of the others.

The importance of time can perhaps be demonstrated by comparing pyroxene granulite in a regional metamorphic setting with pyroxene granulite (as hornfels) in a contact setting. Both have the same mineral assemblage, so presumably formed at roughly the same temperature. In the first case the grains are large, whereas in the second case they are small (Fig. 3.19). A plausible reason for the difference is that grains continue to grow for a very long time in a regional setting, whereas in a contact aureole the brief duration of high temperatures gives them little time to grow.

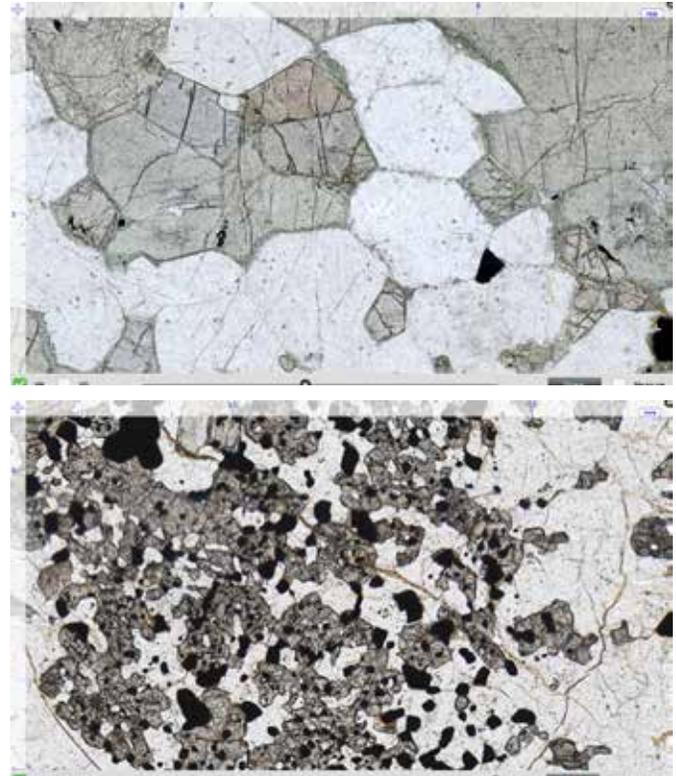


Figure 3.19 Thin sections in PPL of two contrasting pyroxene-plagioclase rocks. Upper image is a pyroxene granulite (M24) from a regional metamorphic area where grain growth lasted for millions of years. Lower image shows a pyroxene hornfels from a contact aureole (VM UK collection) where grain growth was short-lived. Both images are 2.5mm wide.

3.2.4 Does fluid have a role in grain growth?

Even at a high temperature and with ample time, it seems that grains will not grow without the help of a third factor – the presence of fluid along the grain boundaries. The importance of fluid was alluded to in section 3.1.6 above, in the context of the prograde metamorphism of metapelites. Fluid facilitates grain growth because it assists the breaking of the strong chemical bonds in the smaller grains, allowing them to dissolve. Once in solution, the atoms can move freely to a place where they can come out of solution and add to the surfaces of growing grains. The fluid in this case, by permitting easy movement of atoms without being involved in chemical reactions, is said to act as a catalyst.

An excellent field example that highlights the importance of fluid is shown in Figure 3.20, where an outcrop of pyroxene granulite gneiss that was formed about 1000Myr ago is



Figure 3.20 A narrow band of eclogite (dark) with a long, straight fracture along its centre, in pyroxene granulite gneiss near Bergen in western Norway. It appears that fluid entered the fracture and soaked into the rock, where it promoted grain growth, with eclogite replacing pyroxene granulite. *Photo courtesy of Andrew Putnis.*

traversed by a long, straight fracture on either side of which the pyroxene granulite has become darkened. The dark zone consists of eclogite that was formed much later, during an orogeny about 400Myr ago. Clearly the pyroxene granulite gneiss must have been deeply buried and subjected to eclogite-facies conditions 400Myr ago, but outside the narrow strip of eclogite it remained unchanged and metastable. To account for the limited extent of the alteration to eclogite, it seems that water-rich fluid travelled along the fracture and soaked a short distance into the hot, metastable gneiss on either side, where it gave that metaphorical ‘push’ to the perched boulder and triggered grain growth producing, in this case, the new minerals that make eclogite. The term **neocrystallization** is sometimes used to distinguish growth of new kinds of mineral grain, like this, from the simple case of recrystallization, where existing mineral grains grow. In both cases, the presence of fluid is widely thought to be essential for the process to happen.

3.2.5 What determines the shape of a grain?

Why are the grains of some minerals commonly polyhedral (i.e. having random polygonal outlines in thin section), while grains of other minerals are commonly euhedral? Polyhedral grains are usually of quartz, feldspar, calcite, dolomite, pyroxenes, epidote, and olivine. Euhedral grains include those with **platy** shapes like muscovite, biotite and chlorite, and those with **prismatic** shapes like amphiboles, staurolite, and the Al_2SiO_5 polymorphs.

Taking polyhedral grains first, the reason for the shape is simply that it provides the lowest surface area for a given volume of mineral. A polyhedral grain with flat boundaries has less surface energy than a grain of the same mineral with the same volume, but with an irregular shape and wiggly or irregular boundaries. A polyhedron is simply a stable (minimum energy) shape. It is, incidentally, the shape adopted by bubbles in foam.

For minerals with euhedral grains, the explanation is a little different. Euhedral grains, such as plates of muscovite or prisms of andalusite, have flat surfaces that are parallel to specific internal layers of atoms. Such surfaces here provide a very tidy, and therefore energy-efficient, arrangement of the layer of atoms on the outer surface. The euhedral grain in these cases has a lower overall surface energy than a polyhedron with the same volume, despite its larger surface area.

3.2.6 Why do some minerals occur as porphyroblasts?

The small number and large size of porphyroblasts can be traced back to a process called **nucleation**. A grain of garnet can only grow where atoms can add to an *existing* grain of garnet. A metapelite in the biotite zone contains absolutely no garnet. As the temperature rises and the metapelite finds itself in the garnet zone, garnet would like to grow, but with no existing garnet grains it cannot grow. New tiny grains of garnet, called nuclei, first have to start from scratch as tiny seed crystals. The formation of a crystal nucleus is known as nucleation. Nucleation of garnet can only happen when the concentration of garnet, dissolved in the intergranular fluid, is enormous – when the fluid is said to be highly **supersaturated** or to have a very large excess of Gibbs energy, G . This state is only reached when the temperature has gone some way beyond the temperature where garnet first becomes stable (Fig. 3.21).

Once a garnet nucleus has formed, the situation changes. The nucleus grows rapidly, fed by the atoms dissolved in the

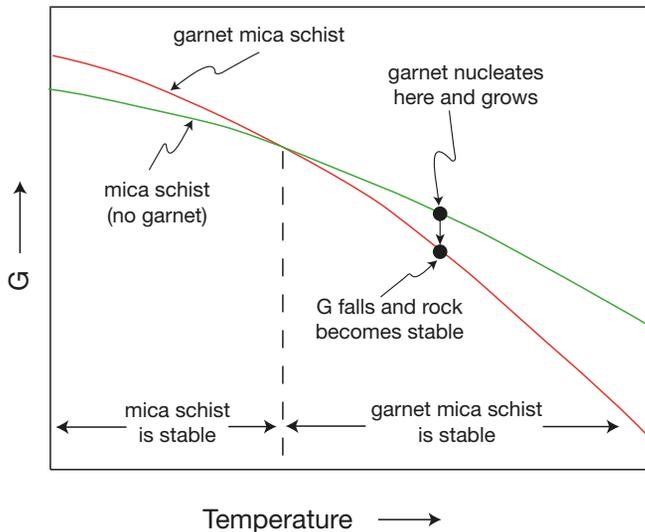


Figure 3.21 Conceptual diagram showing Gibbs energy for mica schist with no garnet (green line), which is lowest and most stable in the biotite zone, and for garnet-mica schist (red line), which is most stable in the garnet zone. Garnet can only nucleate when, with increasing temperature, the gap between the two lines becomes sufficiently large.

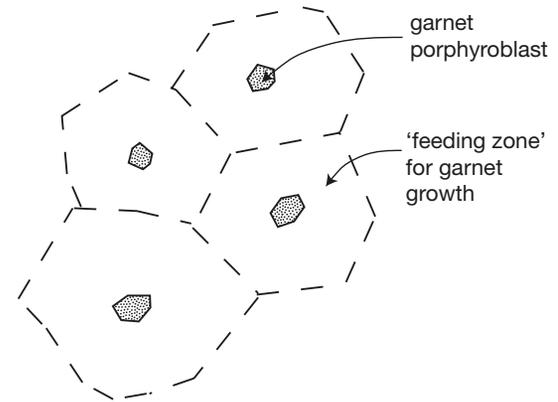


Figure 3.22 Garnet porphyroblasts. They grow large because further crystals of garnet will not nucleate in a wide 'feeding' zone around each porphyroblast where the level of excess Gibbs energy (the level of supersaturation in the intergranular fluid) is kept below the critical level needed for nucleation.

highly supersaturated fluid around it. As it grows, the level of supersaturation in its catchment volume of fluid, or feeding zone, drops below the high level needed for garnet nucleation. This prevents any further nuclei of garnet forming there (Fig. 3.22). So the few nuclei that do form grow, in the absence of competition, into very large crystals. It is a case of the 'rich getting richer'.

How does a garnet porphyroblast make space for itself as it grows? There is little evidence to suggest that it forcefully heaves aside the surrounding grains. Instead, it seems that the surrounding grains obligingly dissolve in the grain-boundary fluid just as fast as new garnet is precipitated from that fluid onto the advancing garnet crystal surface.

Why do some garnet porphyroblasts contain inclusion trails of other minerals? In some cases, the grains surrounding a growing porphyroblast, particularly grains of quartz, fail to dissolve completely, and what remains of them becomes engulfed by the porphyroblast, where they are now seen as inclusions. Very rapid growth of porphyroblasts, which can happen in the period immediately following nucleation when the level of supersaturation is still close to its peak, can leave the porphyroblasts riddled with inclusions, i.e. leave them as poikiloblasts. A case in point is the formation of poikiloblasts of cordierite in andalusite-cordierite hornfels (Fig. 2.31).

Most examples of UHP metamorphism are found in orogenic belts of Phanerozoic age (less than 542 million years old). The continental crust today is 90km thick at most. Since some UHP rocks descended to more than 200km, then the depth of subduction of continental crust in the past was, in some cases, much greater than it is today.

4.3 Shock metamorphism

Shock metamorphism, which is caused by giant meteorite impacts, was introduced in a short paragraph at the end of chapter 1. This section elaborates on the process and describes its products.

Shock happens when rock is suddenly made to move faster than sound waves (i.e. seismic P-waves) can travel through it. P-waves travel at up to 6.5km/sec in most rocks, whereas meteorites arrive at speeds typically of 20km/sec. With shock, the atoms in mineral grains are instantly pressed extremely tightly together under intense, but short-lived, pressures that are ten times to one hundred times higher than those of normal regional metamorphism, i.e. 50 to 500kbar instead of 5kbar. (Shock pressures are often quoted in gigapascals (GPa); 1GPa is 10kbar). Closest to the impact, where the shock pressures are highest, the rock will instantly vaporize, causing the explosion that excavates the crater. Further out the rock will melt. Further out again it will suffer cataclasis (mechanical disaggregation). The crater is emptied out in minutes, but almost immediately the steep walls collapse inwards, sometimes rising as a central peak, and the wall-rocks become mixed in a chaotic way with incandescent ejected material as it falls back, partly filling the initial crater with a kind of breccia called **suevite** (pronounced **soo-av-ite**).

4.3.1 The discovery of shock metamorphism

Shock metamorphism is a relatively new branch of metamorphism, dating from the 1960s. Its emergence as a discipline is mirrored in the history of scientific investigations at the Barringer Meteorite Crater in Arizona (Fig. 1.13). This is one of the best-preserved craters on Earth, having formed only about 50,000 years ago in a desert region. During the first half of the twentieth century geologists argued over whether the crater had an explosive volcanic origin or a meteorite impact origin. One supporter of impact was Daniel Barringer, a mining engineer who linked the crater's origin with chunks of iron-nickel meteorite found in the surrounding desert. He

believed that a very large and valuable mass of metal, with a diameter approaching that of the crater, lay buried beneath the crater floor, so he obtained mineral rights and spent a fortune drilling down in search of it, but to no avail. Only later did it become clear that the crater was large because of an impact explosion, and that the meteorite had largely vaporized.

In 1960 the impact theory was strengthened when a piece of sandstone ejected from the crater and picked up in the desert was found to contain the minerals coesite and stishovite. These are two different dense polymorphs (crystalline varieties) of SiO₂ that require enormously high pressures to form (as seen for coesite in UHP metamorphism), consistent with the pressures produced, albeit briefly, by impact-induced shock waves. The evidence that finally convinced any remaining sceptics still favouring a volcanic origin came in 1963, when Eugene Shoemaker of the United States Geological Survey demonstrated that features produced in shock-damaged rocks from nuclear bomb testing sites could be matched in detail with features seen at Barringer Crater. Today the crater is marketed as a major tourist attraction. It is owned by Daniel Barringer's descendants, who generously donate a significant fraction of the profits to support research on impact cratering and meteorites.

By now almost 200 impact structures have been identified on Earth, including craters and deeply eroded remnants of craters. The two largest, Vredefort in South Africa and Sudbury in Canada, are each more than 250km in diameter. They are also the two oldest, having been formed, respectively, at just over 2000Myr ago and close to 1850Myr ago.

4.3.2 Products of giant impacts

The shock-damaged rocks at an impact site, and in the excavated ejecta, display one or more of the distinctive features known as shatter cones, pseudotachylite, dense polymorphs of silica and carbon, planar deformation features in quartz and other minerals, mineral grains that have been turned into glass (called diaplectic glass), and large-scale melt sheets.

Shatter cones are an unusual fracture pattern in the form of aligned sets of long, nested, striated cones (Fig. 4.21). The tips of the cones were once thought to be directed towards the impact, but that interpretation is now refuted because examples of shatter cones have been discovered that point in many different directions within a single outcrop.



Figure 4.21 Shatter cones in sandstone at Sudbury, Ontario. The notebook is about 10cm wide. *Photo courtesy of Gavin Kenny.*

Pseudotachylite was introduced in chapter 2 (Fig. 2.68) as a product of dynamic metamorphism. At impact sites it is a dark, flinty-looking rock that appears to have been injected as a liquid into cracks that opened up in the host rock. The injections vary from broad dyke-like sheets down to millimetre-wide veins. In some large sheets, angular and rounded fragments of the host rock are embedded in the pseudotachylite, giving a breccia (Fig. 4.22). Here the pseudotachylite was produced *in situ* apparently by frictional melting as adjacent blocks of rock were jostled and shaken violently against each other.

The dense high-pressure polymorphs of SiO_2 , coesite and stishovite, originally found at Barringer Crater have since been found at many impact sites, but identifying them is not straightforward and requires careful analysis using XRD. Also, shock-generated diamond has been found in carbon-bearing crater rocks. Coesite and diamond are not, of course, unique to shock metamorphism. As noted in the previous section, they are known in ultra-high pressure metamorphic rocks.

Planar deformation features in quartz are believed to be diagnostic evidence of shock metamorphism. They are parallel sets of dark lines visible under a microscope in individual grains or in a thin section (Fig. 4.23). Known simply as PDFs, they are found not just at impact sites, but also in tiny fragments of ejecta from distant craters. They are found, for



Figure 4.22 Zone of breccia in impact-shocked granitic gneiss near Vredefort, South Africa. The dark material injected between the blocks is pseudotachylite. Angular blocks of gneiss appear to be in place. Rounded blocks may have been transported. Width of field is about 5 metres. *Photo courtesy of H. Jay Melosh.*

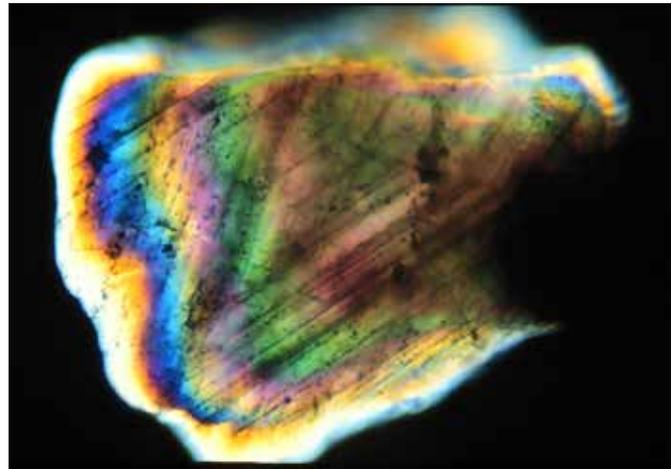


Figure 4.23 A shocked quartz grain, 0.32mm across, seen in XP through a polarizing microscope. It shows two strong sets of planar deformation features that partly overlap. This grain is from deposits within the Chicxulub crater, recovered from the drill hole Yucatán-6, located 50km from the centre of the crater. *Photo courtesy of Alan Hildebrand.*

example, in the worldwide ‘boundary clay’ that comprises fall-out from the 66Myr old Chicxulub crater in Mexico. This crater, at 180km in diameter, is the third largest on the planet,

but remains hidden beneath younger sedimentary layers. Its discovery in 1990 can be attributed to serendipitous action by a geologist named Alan Hildebrand. It is implicated by many as the principal cause of the mass extinction event at the end of the Cretaceous Period that saw the demise of the dinosaurs.

Shock-melted minerals are another impact-related feature. In this case, the shock pressure and associated heating are sufficiently intense to cause total melting of individual mineral grains leaving, for example, pure silica glass or pure feldspar glass after the shock wave has passed. The glass is known as diaplectic (pronounced dye-a-plek-tic) glass. **Diaplectic glass** formed from plagioclase is known as maskelynite (pronounced mass-ke-lin-ite).

Finally, in very large impact craters such as those at Sudbury and Vredefort, the heating was so intense that a lake of melted rock, known as a melt sheet, was produced. At Sudbury it cooled slowly, and is now an enormous layer of coarse-grained igneous rock. Since the melting happens almost instantly, much of the liquid gets ejected as spray, along with pulverized and shock-damaged rock and mineral fragments, in the impact plume. The liquid coalesces into droplets that freeze to glass and fall back to the Earth's surface, possibly hundreds or even thousands of kilometres from the crater, as glass beads known as **tektites**. Small spherical tektites are called spherules. Thin layers of spherules, interleaved with normal sedimentary layers, have been reported from many places. Each of these so-called **spherule beds** is a testimony to some ancient far-off impact event. An example is shown in Figure 4.24, which is a thin section of a spherule bed of late Triassic age in south-west England, near Bristol. The spheres, which are up to 1mm across, are no longer glassy, having been altered to secondary minerals, including green chlorite. They are cemented by a form of K-feldspar, which has been dated by the ^{40}Ar – ^{39}Ar method. The age is consistent with the spherules having been derived from the Manicouagan impact crater in eastern Canada, which is now deeply eroded. During the Triassic, the north Atlantic Ocean had yet to open up, so the impact site would have been much closer than it is today.

4.3.3 Extra-terrestrial shock metamorphism

While shock metamorphic effects are preserved in only a tiny fraction of the Earth's rocks, they are widespread on other bodies in the Solar System, where, in the absence of water and an atmosphere, and without plate tectonic recycling, shock

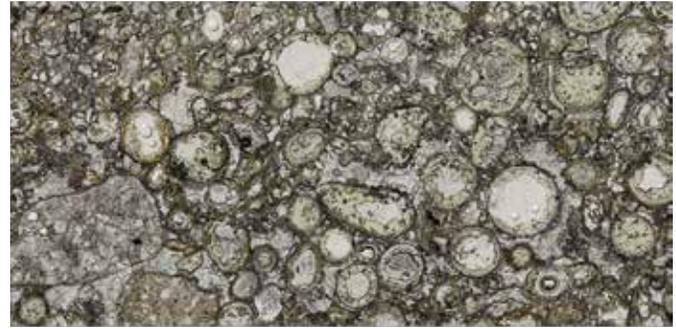


Figure 4.24 Thin section in PPL of the Triassic Meteorite Impact Deposit in the UK Virtual Microscope collection. The spherules, now altered to chlorite and other secondary minerals, are believed to have originated as glass droplets from the Manicouagan Impact Crater in Canada.

damage does not get healed. Giant impacts have affected the Moon, Mercury, Mars and most asteroids.

Meteorite craters are well known on the Moon. The lunar surface seen from Earth appears blotchy, with large rounded or irregular dark areas surrounded by light areas. These are known, respectively, as the lunar maria ('seas') and the lunar highlands. The maria are covered in smooth basalt lava flows with few craters, but the lunar highlands are very heavily cratered. Consistent with the cratered topography, samples collected from the highlands during the Apollo programme were found to consist almost entirely of breccia, produced by repeated large impacts into the original rock of the lunar surface. This rock is the igneous rock known as **anorthosite** (pronounced an-or-tha-zite) which is made largely of Ca-rich plagioclase, making it light in colour. A sample of anorthosite numbered 60025 from the Apollo 16 landing site is one of the oldest lunar rocks known. A thin section of it in XP in Figure 4.25 shows obvious cataclastic damage in a single large grain of plagioclase; the twin lamellae (stripes) have been off-set along miniature faults, and the grain as a whole has been fragmented, with each small piece shifted a little out of alignment with its neighbour. The overall effect is a little like a mosaic, so the plagioclase is said to have been mosaicized. In another sample, 78235 from the Apollo 17 site, shock has caused *localized* melting of plagioclase (Fig. 4.26). The melt (now maskelynite glass) has a swirling blue and brown pattern in PPL, and is isotropic in XP. Some unmelted but mosaicized (highly strained) plagioclase just beside the glass implies



Figure 4.25 Thin section in XP of lunar anorthosite (plagioclase rock) 60025, showing part of a large mosaicked grain of plagioclase traversed by a band of finely crushed plagioclase.

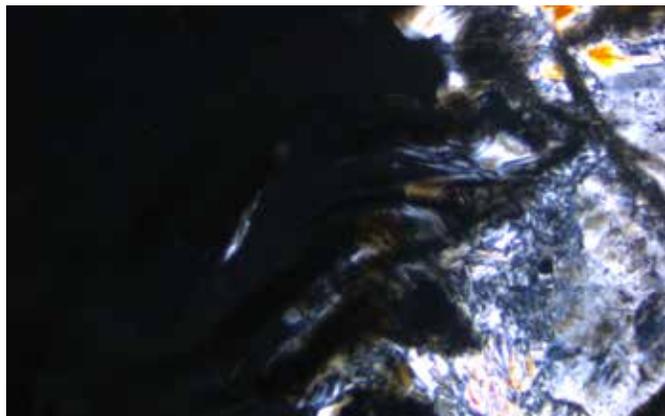
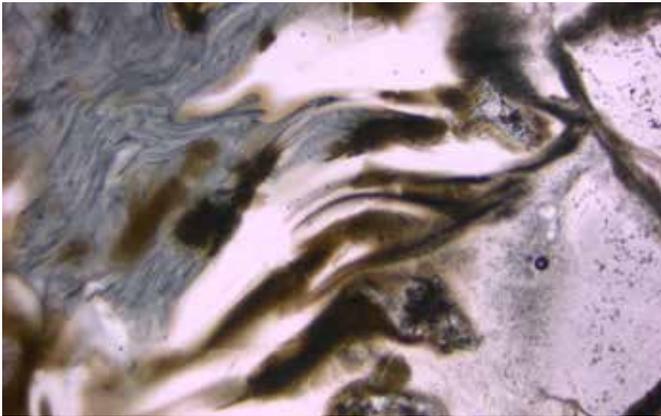


Figure 4.26 Shocked anorthosite in Apollo sample 78235 in PPL (top) and XP. For a description and comment see the text. Width of field is 2mm.

that the melting process, and hence the effect of the passing shock waves, was local and arbitrary. Almost all the cratering of the Moon's surface happened more than 4000Myr ago. The circular maria are actually giant impact basins, subsequently flooded with basalt lava.

Meteorites also show evidence of shock. They fall into a few dozen distinctive groups, and each group is thought to come from a single asteroid source. One group, known as the L chondrite group, includes many meteorites that have suffered intense shock. Figure 4.27 shows a thin section image of one of these meteorites in which the shock caused local melting. Fragments enclosed within the resulting dark glass contain clear olivine that has been partly transformed to a blue mineral called **ringwoodite**. Ringwoodite has the same composition as olivine; it is a dense high-pressure polymorph of $(Mg,Fe)_2SiO_4$. The transformation of olivine to ringwoodite is attributed to enormously high shock pressures caused when the L-chondrite asteroid collided catastrophically with another asteroid and broke up. The collision happened around 470Myr ago based on potassium–argon dating of shocked L-chondrites. Heat from the shock event drove off the argon gas that had been accumulating until that time in the meteorite's minerals,

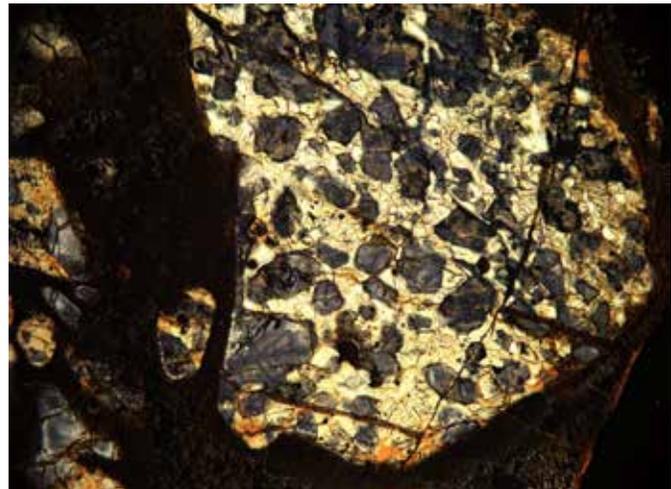


Figure 4.27 Thin section in PPL of a pale-coloured olivine-rich fragment containing blue crystals of ringwoodite enclosed in dark, shock-melted glass in a meteorite called Taiban. Ringwoodite has the same composition as olivine. It is a high-density polymorph of $(Mg,Fe)_2SiO_4$. Width of field is 0.5mm. *Photo courtesy of Ed Scott.*



Figure 5.16 Part of a thick layer of eclogite in the Glenelg Lewisian (tinged red by garnet), which has been altered to amphibolite along shear zones (greyish green), presumably as a result of ingress of water during Grampian reworking.

is conspicuous with its protruding brown pebble-like grains of weathered forsterite (Fig. 2.37). The eclogite occurs as layers and lenses enclosed in grey quartz-feldspar-hornblende gneiss. It is rare to find eclogite that has not been partly altered to amphibolite (Fig. 5.16). This alteration is presumed to have occurred during the Grampian orogeny.

Another widespread feature is the development of zones of mylonite, which run the length of the Glenelg Lewisian strip parallel to the Moine thrust. The mylonite zones may have formed during the Grampian orogeny when dry Lewisian rocks, unable to recrystallize, succumbed to intense stress.

5.2.2 The calcite–dolomite solvus geothermometer

The calcite–dolomite geothermometer is applicable to marble containing both calcite and dolomite, such as the marble from Glenelg. It is based on the limited atomic substitution of Mg^{2+} for Ca^{2+} in calcite. The amount of substitution is limited because Mg^{2+} ions (radius 0.066nm) are considerably smaller than Ca^{2+} ions (0.1nm) so a Mg^{2+} ion would tend to ‘rattle uncomfortably’ in a site that would normally hold a large Ca^{2+} ion (see Appendix 2 for an account of ionic radii). The higher the temperature the more that Mg will enter calcite and substitute for Ca. The reciprocal substitution of Ca for Mg in dolomite also occurs, but to a far lesser extent. As with Na and K substitution in feldspars (Fig. 5.5),

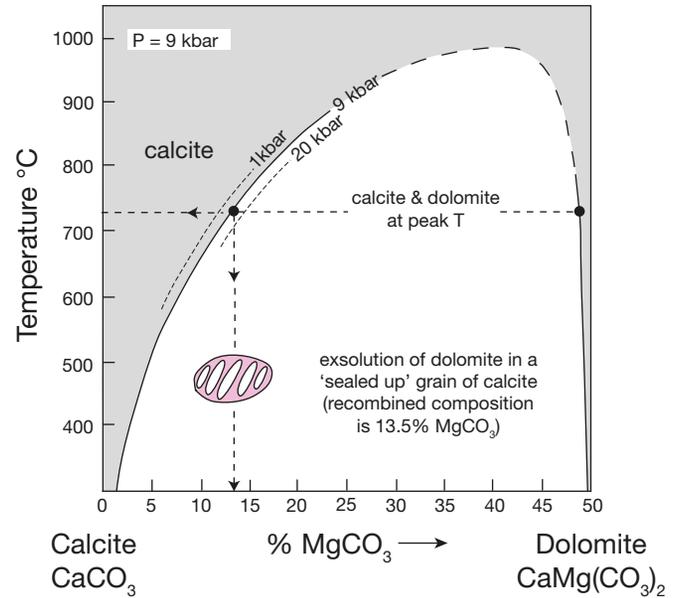


Figure 5.17 The calcite–dolomite solvus at 9kbar. For an explanation of its use in geothermometry see the text.

the increase in Mg substitution in calcite with temperature can be illustrated as a solvus curve (Fig. 5.17). This curve has been the target of several careful experimental studies, and its position on the calcite side of the solvus is known quite precisely at three different pressures. The curve for a pressure of 9kbar is highlighted in Figure 5.17.

Where marble contains calcite and dolomite in equilibrium, then the compositions of both minerals will fall exactly on the solvus curve. Assuming that the calcite retains its Mg during cooling, its composition will record the peak temperature, which can simply be read off the temperature axis of the solvus diagram.

A problem, however, is that Mg tends to escape from Mg-bearing calcite during cooling. This has happened in the Glenelg marble, where the composition of calcite is commonly about 6% MgCO_3 , equivalent to a temperature of only about 550°C. This is presumed to be the temperature where movement of Mg ions becomes too sluggish for them to leave their calcite host.

Fortunately, in a few places the Mg did not get away. In these places tiny lamellae (plates) of dolomite grew inside

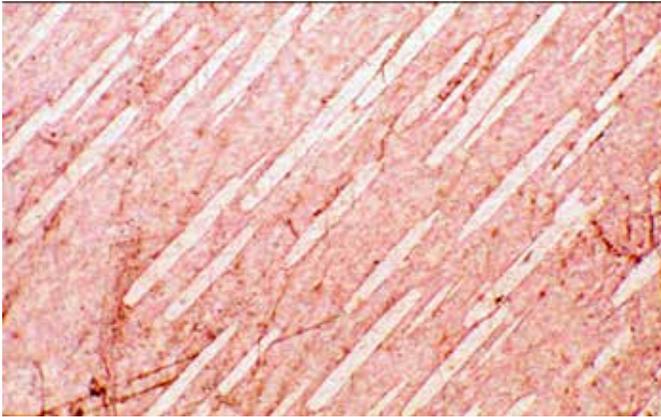


Figure 5.18 Thin section in PPL showing exsolved plates of dolomite, edge-on in a single grain of calcite (stained pink) trapped within a huge crystal of forsterite (not included in the image) from a sample of Glenelg marble. Width of field is about 1mm.

the calcite by the process of exsolution (in an equivalent way to exsolution in feldspar to make perthite). As the dolomite lamellae grew, they mopped up the Mg ions that were escaping from the adjacent cooling calcite. Thus the overall carbonate grain (residual calcite combined with the new dolomite lamellae) retained its full quota of Mg from when the calcite was at the peak temperature. An example is shown in Figure 5.18 – a calcite grain completely enclosed by a huge porphyroblast of forsterite (similar to the grain in Fig. 2.38). The composition of the original calcite can be obtained by measuring the compositions of the dolomite plates and of their calcite host, and measuring the volume percentage (which is the same as the percentage area in a thin section) of each. Such studies in marble from Glenelg consistently point to 13.5% of MgCO_3 (on a molecular basis) in the original calcite. This implies a metamorphic temperature of about 720°C at 9kbar. The temperature would be a little *lower* if the pressure were higher because the solvus decreases slightly with increased pressure (Fig. 5.17).

5.2.3 The clinopyroxene–albite–quartz geobarometer

The composition of clinopyroxene coexisting with albite and quartz serves as a geobarometer. In this case the composition changes with changing pressure (and also, to a small extent,

with temperature). The clinopyroxene in eclogite is a variety called omphacite, which has **coupled atomic substitution** between augite, $\text{Ca}(\text{Mg,Fe})\text{Si}_2\text{O}_6$ and jadeite, $\text{NaAlSi}_2\text{O}_6$ (see Appendix 2). As Ca^{2+} is replaced by Na^+ , so simultaneously ($\text{Mg}^{2+}, \text{Fe}^{2+}$) is replaced by Al^{3+} . The extent of this substitution depends on pressure because Al^{3+} ions are much smaller than Mg^{2+} and Fe^{2+} . Pressure compresses the large oxygen ions, making the octahedral spaces smaller, and more suitable for the smaller Al^{3+} ions. Na^+ has no problem swapping for Ca^{2+} ; they are the same size.

If quartz is present with clinopyroxene, then there is a maximum amount of Al^{3+} and Na^+ (i.e. of jadeite) that can be accommodated, and it is this maximum value that changes with pressure. It is shown as a kind of ‘pressure solvus’ in Figure 5.19.

Eclogite is a metabasite which, by definition, never contains albite, so how can this geobarometer be applied? Fortunately, present among the rocks at Glenelg are some rare metadiorites that contain all three minerals. The omphacite in these rocks consistently has ~ 45% of the $\text{NaAlSi}_2\text{O}_6$ (jadeite)

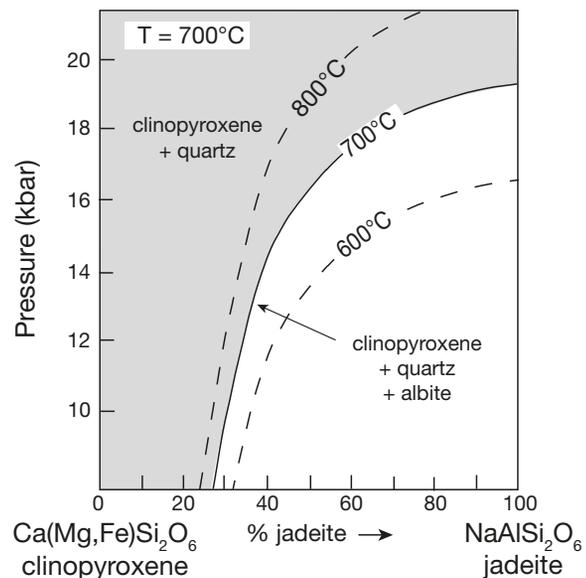


Figure 5.19 The ‘pressure solvus’ at 700°C for limited substitution (grey area) of jadeite in clinopyroxene, in the presence of quartz and albite. The approximate positions of the ‘pressure solvus’ at 600°C and at 800°C are also shown.

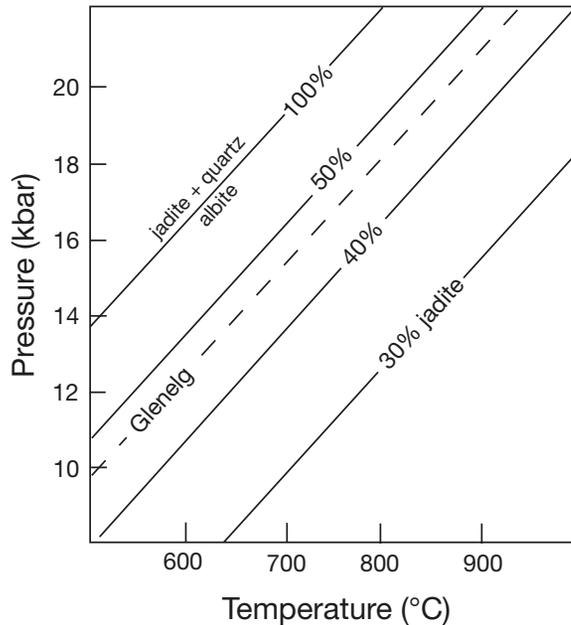


Figure 5.20 Experimentally determined pressure–temperature diagram showing the percentage of jadeite in clinopyroxene that coexists with quartz and albite. The Glenelg rocks formed under conditions, along the dashed line, for 45% jadeite.

end-member. This constrains the P–T conditions to a line on a pressure–temperature diagram (Fig. 5.20). The line has a relatively low slope, which crosses the steep line for the Mg-bearing calcite at a high angle and gives a good ‘fix’ on pressure and temperature (Fig. 5.21).

5.2.4 The garnet–clinopyroxene Fe–Mg exchange thermometer

The garnet–clinopyroxene Fe–Mg exchange thermometer that was used to estimate the temperature of metamorphism in the Slishwood rocks has been applied also in the Glenelg region. Ten mineral pairs were analysed. Their K_D values were found to be remarkably constant, with an average value of 6.4 ± 0.4 . This is pleasing because there is an enormous range in Fe/Mg in the minerals analysed, and there is considerable variation in the size of the correction needed to deal with Fe^{3+} ions in the garnet and clinopyroxene. The K_D value of 6.4 ± 0.4 defines a line on a pressure–temperature diagram (Fig.

5.21, and also shown on Fig. 5.9) which passes right through the intersection of the lines corresponding to calcite with 13.5% $MgCO_3$ and omphacite with 45% jadeite, indicating that the Glenelg rocks were metamorphosed at about $700^\circ C$ and 15kbar. The intersection of all three lines so precisely at one point must be, to some extent, fortuitous because each line has significant errors. A more realistic estimate of the P–T conditions is $700 \pm 25^\circ C$ and 15 ± 1 kbar, and is shown by the orange ellipse in Figure 5.21.

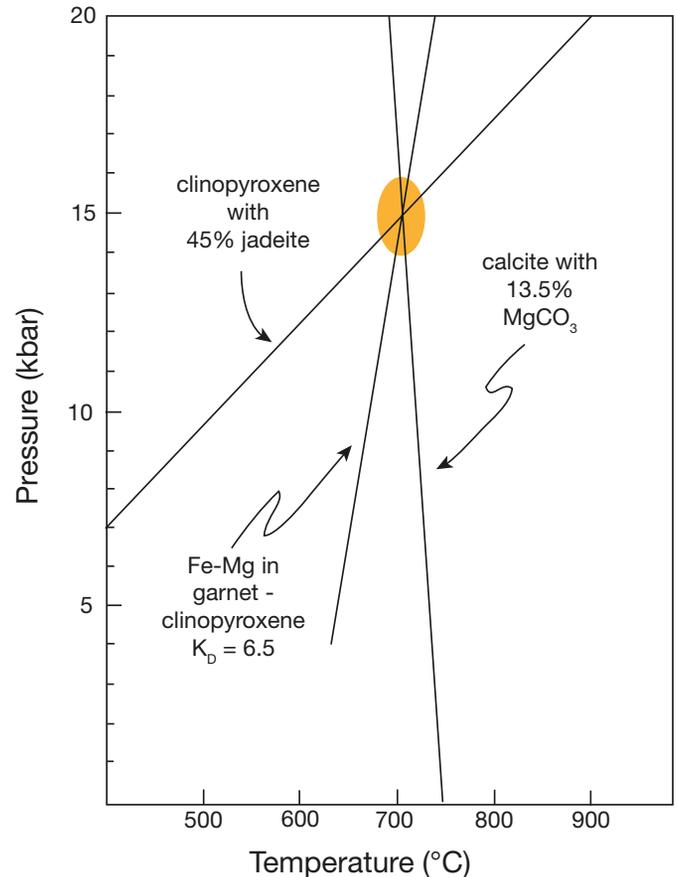


Figure 5.21 Pressure–temperature diagram showing the estimated conditions for the Glenelg high-grade rocks (orange ellipse). For an explanation see the text.

Appendix 1 The Earth's interior

Metamorphism is a process that takes place beneath the surface, often at considerable depth, so as a background to understanding how the process works, it is helpful to have a clear idea about the nature of the planet's interior. Important in this regard, and discussed in this short appendix, are the Earth's continental crust, its oceanic crust, the mantle beneath both, and the so-called tectonic plates that combine all three.

A1.1 The continental crust, the oceanic crust, and the mantle

The **continental crust** was discovered a little over 100 years ago by the Croatian seismologist Andrija Mohorovičić. He carefully analysed many records of the time taken for sound waves (called seismic P-waves) to travel from an earthquake source to a recording station at various distances away, and found that the waves travel at speeds of up to 6.5km per second through rocks between the surface and a depth of roughly 35km, and travel at 8km per second through rocks at a depth greater than about 35km. An example of the kind of observation made by Mohorovičić is shown in Figure A1.1. The jump in speed from about 6.5 to 8km per second is now recognized beneath continents and shallow seas throughout the world. Rocks in the upper 35km or so, with P-wave velocities up to 6.5km per second, comprise the continental crust. The rocks beneath the continental crust comprise the **mantle**.

The boundary between the continental crust and the mantle is called the **Mohorovičić discontinuity** in honour of its discoverer. Most people just call it the **Moho** (pronounced **mo-hoe**). The depth to the Moho (i.e. the thickness of the continental crust) varies. The thickness of 35km, stated above, is only an average value. It is greater under mountain ranges, where it is locally as much as 90km, and can be 20km or less near the margins of continents.

The Moho has also been recognized globally beneath the deep ocean floor. The crust there is only about 7km thick on average, and is called the **oceanic crust** (Fig. A1.2).

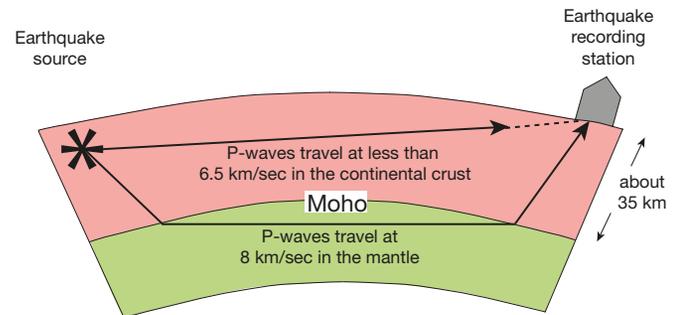


Figure A1.1 Cartoon, not to scale, of a vertical section through the outer part of the Earth in a continental region showing two paths taken by seismic P-waves travelling from an earthquake source to a recording station over a hundred kilometres away. At this distance, waves travelling via the mantle (green) arrive at the recording station before waves travelling directly, but more slowly, through the continental crust. The positions of the two arrow heads show the distances travelled by the P-waves in the same time.

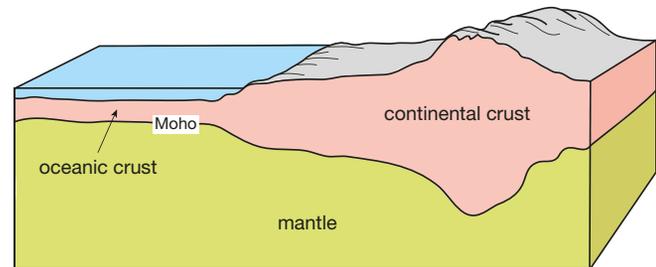


Figure A1.2 Cartoon to illustrate the thickness variation of the crust. The crust comprises the outer layer of the Earth, above the Moho, where seismic P-waves travel at 6.5km per second or less. It is typically about 35km under continents, much thicker under mountain ranges, but only about 7km under the deep oceans.

What is the crust made of? The continental crust is believed to consist of normal sedimentary, igneous and metamorphic rocks, based on the kinds of rock seen in very deep boreholes

and in places where the crust has been deeply eroded. The oceanic crust, in contrast, is believed to consist mainly of the dark-coloured igneous rocks called **basalt** (fine-grained), **dolerite** (medium-grained) and **gabbro** (coarse-grained) since these rocks are regularly recovered in deep-sea dredges and drill cores.

What is the mantle made of? It has not been possible to drill down through the continental crust to sample it. The deepest borehole (in the Kola Peninsula in north-western Russia) reached a depth of just over 12 kilometres, only about one-third of the way to the Moho. However, four lines of circumstantial evidence strongly suggest that the mantle is made of **peridotite**, a dense, green rock made largely from the mineral **olivine**. Olivine has the chemical formula $(\text{Mg,Fe})_2\text{SiO}_4$ (this mineral formula is explained in Appendix 2).

Firstly, nodules (lumps) of peridotite are found in basalt (Fig. A1.3). Since peridotite is very rare in the crust, the nodules are believed to be detached fragments of the mantle through which the basaltic magma flowed rapidly on its ascent to the surface. Secondly, the speed of seismic P-waves in peridotite, measured experimentally, is about 8km per second, the same as the speed observed in the mantle. Thirdly, in a few parts of the world massive slabs of what appears to be oceanic crust and mantle, called **ophiolite** (pronounced **oh-fia-lite**) are exposed at the surface. The mantle part of an ophiolite



Figure A1.3 A broken lump of peridotite believed to have been brought up from the mantle entrained in flowing basaltic magma. A skin of black basalt is still attached to part of its surface. It consists mostly of large grains of the dense, green, glassy-looking mineral called olivine.

slab is made of peridotite. Fourthly, common meteorites, called **chondrites**, are made mostly of olivine. Chondrites are fragments of tiny planets (asteroids) from the asteroid belt, beyond the orbit of Mars. They have much the same chemical composition as the Sun (known from so-called absorption lines in the spectrum of sunlight), so they are thought to be made of the primordial (olivine-rich) starting material from which the Sun and the planets were made when the solar system was formed.

Why are the upper surfaces of oceanic crust and continental crust at different heights relative to sea level? The answer relates to density. Peridotite is about 3.3 times heavier than water (its density is 3.3 grams per cubic centimetre: g/cm^3), whereas most rocks in the crust are between 2.6 and 3 times heavier than water (average density 2.8 g/cm^3). Thus, the crust, whether continental or oceanic, is buoyant relative to the mantle, and ‘floats’ on top of it. The surface of most of the continental crust is close to sea level. Where the continental crust is thicker, it ‘floats’ with its surface above sea level, locally forming mountains whose base extends deep into the mantle. The oceanic crust is so thin that it ‘floats’ on the mantle with its surface well below sea level (Fig. A1.2).

A1.2 Plate tectonics

The continents are believed to have changed their positions over the course of geological time. Their movement is understood today in terms of the theory of **plate tectonics**. The theory emerged from the old hypothesis of continental drift, an idea proposed about a century ago by the German geophysicist and meteorologist Alfred Wegener. Wegener was fascinated by the way the opposing shorelines of the Atlantic Ocean appear to fit together almost perfectly, like pieces of a giant jig-saw puzzle. He imagined that the continents on either side were not only floating in the mantle, but were able to move through the mantle, like ice floes (sheets of floating ice). However, his ideas failed to gain acceptance, largely because geophysicists like Sir Harold Jeffreys in Cambridge pointed out that the mantle was simply far too strong to allow slabs of floating continental crust to drift through it in this way.

In the 1950s and 1960s it gradually became clear from studies of rock magnetism that Wegener and Jeffreys were both, in fact, right. The continents had, indeed, moved, but they had not moved *through* the mantle but had moved *with*

the mantle. It was the mantle, or rather an outer layer of the mantle, that had moved.

This outer layer of mantle is now recognized as forming a planet-wide carapace of cool, strong rock about 100km thick, which is known as the **lithosphere** (meaning sphere of rock). Beneath the lithosphere, the mantle is hot and weak. Although it is solid it can easily be deformed and it behaves, over a long period of time, as a material that is mechanically weak and can flow very slowly, more like plasticine than rigid rock. This deeper part of the mantle has been named the **asthenosphere** (meaning 'sphere without strength').

The lithosphere is divided into separate pieces, like crazy paving, by deep cracks that extend through to the asthenosphere; it comprises a dozen or so 'paving slabs'. These slabs are not flat, of course, but curved to fit the Earth's spherical shape. They are called **tectonic plates**, or simply plates (*tectonic* comes from the Greek word *tektonikos*, meaning *to do with building*). There are eight large plates, four smaller ones, and several that are smaller still.

The plates are easily picked out on a world map of earthquake locations, because their edges are associated with earthquakes (Fig. A1.4). Earthquakes are caused because the plates are slowly sliding over the asthenosphere independently of each other, and their edges are repeatedly snagging, and being released in jolts.

The Earth's buoyant continental crust 'floats' in the mantle, just as Wegener imagined and as described above. It resembles several separate and extensive sheet-like rafts. However, in contrast to Wegener's ideas, the mantle in which the rafts 'float' is the strong lithosphere; the rafts are firmly embedded in it. As the plates move, so the rafts of continental crust simply ride along with them passively, like passengers on a travelator. Two of the eight large plates, named the Pacific Plate and the Nazca Plate, are covered entirely by oceanic crust. The other six plates carry both continental crust and oceanic crust, and each is named by the continent (or continents) within it (Fig. A1.4). The relative movement of plates means that they are converging in some places, and moving apart in others, as is

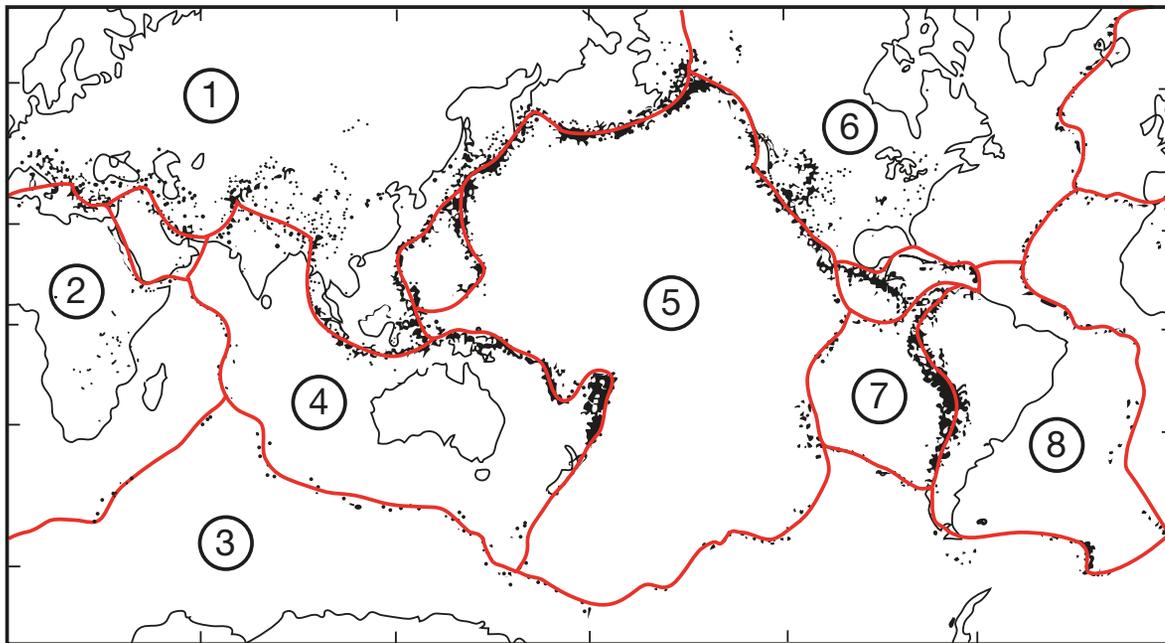


Figure A1.4 Map of the world with the locations of recorded earthquakes shown as tiny black dots. The dots, which have been 'joined up' with a red line, mark the edges of plates. Eight major plates are numbered, and named (1) the Eurasian, (2) the African, (3) the Antarctic, (4) the Australian/Indian, (5) the Pacific, (6) the North American, (7) the Nazca, and (8) the South American plates. Four minor plates (not numbered) are the Arabian, Philippines, Cocos, and Caribbean plates.

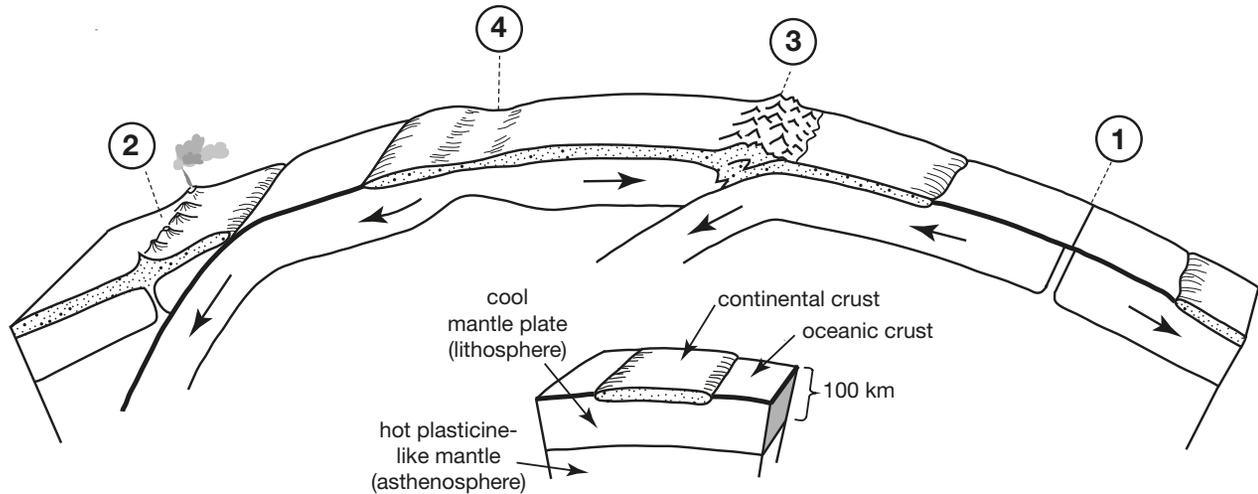


Figure A1.5 A hypothetical cross-section (not to scale) through the outermost 200 kilometres of the Earth. It includes parts of four lithospheric plates, about 100km thick. Oceanic crust is shown by a thick black line on the plate's upper edge; continental crust by a stippled outer capping. The plates are rigid and slide slowly over hot, weak mantle called the asthenosphere (see labelled inset drawing). Three kinds of boundary are numbered: (1) A mid-ocean ridge where plates are moving apart and new oceanic lithosphere is created. (2) A subduction zone where plates are converging and one of the plates, capped with oceanic crust, is subducted ('dives' at an angle into the mantle) beneath the opposing plate. (3) A subduction zone where both converging plates carry a cap of continental crust. (4) In a separate setting, not at a boundary, the lithosphere becomes stretched and thinned. More details of the processes at locations (1) to (4) are given in the text.

shown in Figure A1.5, which is a hypothetical vertical section through several plates and their mutual boundaries.

A1.2.1 *What happens where plates move apart?*

Where plates are moving apart, the hot asthenosphere creeps slowly upwards from a depth of over 100km to fill the widening gap between them (Fig. A1.5 position 1 and Fig. A1.6). As it rises, it partially melts to make a slush of basaltic liquid mixed with residual (unmelted) grains, mostly of olivine. Eventually the basaltic magma separates from the residual solids, rises towards the surface, then cools and hardens in the form of basalt pillows, dolerite dykes, and gabbro plutons, to make new oceanic crust. The mantle immediately below the oceanic crust consists of peridotite from which basaltic magma has been extracted. A line of volcanic activity known as a **spreading ridge** or **mid-ocean ridge** persists on the ocean floor between the diverging plates.

What causes the rising asthenosphere to melt? To answer this question one needs to know two things. Firstly, the

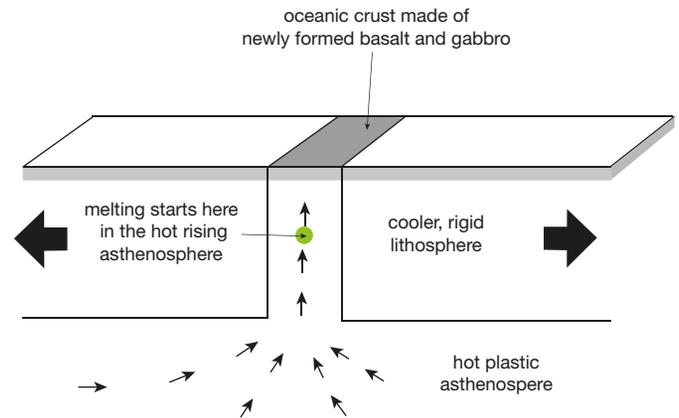


Figure A1.6 Cartoon showing the upwelling of hot asthenosphere to fill the widening gap between two plates that are moving apart. The asthenosphere starts to melt as it loses pressure, and the resulting basaltic magma separates out, rises further and cools to make new oceanic crust (shaded grey).

Appendix 2 The chemical formulae of minerals

Minerals are made from chemical elements. Over 90 different elements exist in nature, from the lightest element, hydrogen, to the heaviest, uranium. As is stated in chapter 1, only ten of these elements are needed to make the two-dozen or so minerals that comprise the vast majority of metamorphic rocks. These ten are hydrogen (H), carbon (C), oxygen (O), sodium (Na), magnesium (Mg), aluminium (Al), silicon (Si), potassium (K), calcium (Ca), and iron (Fe). Eight of these elements account for around 99% by weight of the **continental crust** (Fig. A2.1). The other two, hydrogen and carbon, are included in the tiny sector labelled 'others' in Figure A2.1. Two elements, oxygen and silicon, are especially abundant.

A2.1 How are chemical formulae of minerals written?

The atoms of elements combine together in simple proportions to make minerals. The chemical formula of a mineral is a kind of shorthand notation for writing down these proportions. For example, the formula for **quartz** is SiO_2 . The subscript '2' means that there are two atoms of oxygen for every one atom of silicon. The mineral **kyanite** has the formula Al_2SiO_5 , which shows that it contains aluminium, silicon and oxygen atoms in the proportions 2:1:5. The mineral **dolomite** has the formula $\text{CaMg}(\text{CO}_3)_2$. The brackets here, followed by the subscript '2', mean that there are two lots of '(CO₃)' for one atom each of calcium and magnesium. Thus (CO₃) must be multiplied by 2 to get the overall proportions of the elements in dolomite: calcium, magnesium, carbon and oxygen are in the proportions 1:1:2:6.

What fixes the numbers? Why is kyanite's formula written Al_2SiO_5 , and not Al_3SiO_5 ? The reason is that elements occur as electrically charged atoms called **ions**, and not as neutral atoms. Ions have to combine in strictly defined ratios that are dictated by the electrical charge. Neutral atoms of any particular element, such as silicon, have a central nucleus with a fixed number of particles called protons (14 for silicon), each proton having one positive charge. The nucleus is surrounded by an equal number of minuscule particles called electrons

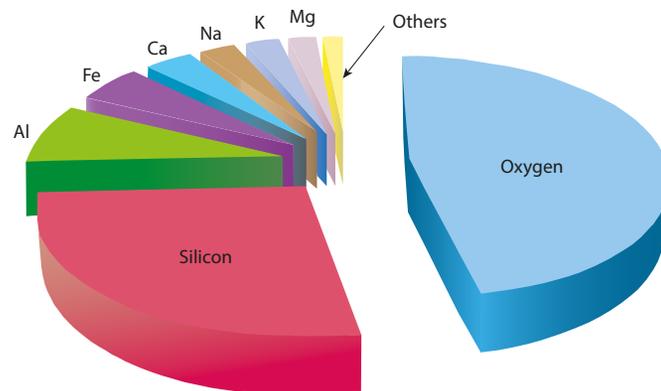


Figure A2.1 Proportions by weight of the eight most abundant elements in the continental crust. Two others, hydrogen and carbon, are included in 'others', and all ten are needed to make the common minerals in metamorphic rocks.

'orbiting' the nucleus, each with one negative charge. The neutral atoms become ions by spontaneously losing or gaining one or more electrons. Loss of electrons will leave an ion with more protons than electrons, so it will be positively charged. Conversely, gain of electrons will convert a neutral atom into a negatively charged ion. The charge is simply the number of electrons lost or gained; it is written as a superscript numeral with a plus or minus sign following the element symbol. The ten ions that are most important in minerals, in order of increasing mass, are:

H^+ , C^4+ , O^{2-} , Na^+ , Mg^{2+} , Al^{3+} , Si^{4+} , K^+ , Ca^{2+} and Fe^{2+} .

Only one of these elements, oxygen, occurs in the form of negatively charged ions. The other nine form positively charged ions. Since opposite charges attract, the positive ions are attracted to, and become tightly bound to, the negative oxygen ions by so-called **ionic chemical bonds** to make minerals.

In writing the chemical formula of a mineral, the sum of charges on the positive ions must be exactly equal to the sum of negative charges on the oxygen ions. Continuing

with the example of kyanite, its formula, Al_2SiO_5 , has two Al^{3+} ions and one Si^{4+} ion giving a total of 10 positive charges, and it has five O^{2-} ions that give 10 negative charges. The positive and negative charges are therefore balanced and overall the formula is neutral. To give another example, in the mineral calcite, whose formula is CaCO_3 , Ca^{2+} and C^{4+} together provide 6 positive charges, while three O^{2-} ions give 6 negative charges. This simple process of balancing the negative and positive charges works for every mineral formula. If the charges do not balance, then the formula is incorrectly written.

The hydrogen ion, H^+ , is not normally treated as a separate ion in minerals because it almost always teams up with O^{2-} to make the **hydroxyl ion** (OH^-). So (OH^-) is usually taken as a single ion with just one negative charge rather than as separate O^{2-} and H^+ ions. Another point to note is that iron atoms in nature can occur as Fe^{3+} ions as well as Fe^{2+} ions, by losing three electrons instead of two.

The number of charges on an ion can be worked out from its position the **Periodic Table**, which lists all the elements in the order of the number of protons that each element has in the atomic nucleus. The first element, hydrogen, has one proton per atom, the second, helium, has two protons per atom, lithium has three, beryllium has four, and so on, increasing by one proton per element. The list is arranged

in a table with rows and columns. The number of protons increase going along the rows. Elements with similar properties lie in the same column. Elements in the first column have ions with one positive charge, those in the second column have ions with two positive charges, and so on. The Periodic Table for the first 20 elements, shown as ions, is set out in Figure A2.2.

To help memorize the Periodic Table, the element symbols can be strung together and read aloud as long 'words'. The first two 'words' sound like 'her-helly-beb-cernoff-knee' and 'nam-gall-see-puss-clar'.

A2.2 Minerals whose composition can vary

In many minerals the composition is not fixed, as it is in quartz, kyanite and calcite, but it can vary within limits. For example, the mineral **olivine** can have any intermediate composition between pure magnesium olivine, Mg_2SiO_4 , and pure iron olivine, Fe_2SiO_4 (see Fig. A2.3-A). The formula of olivine is written $(\text{Mg,Fe})_2\text{SiO}_4$ to indicate this possible variation in Fe and Mg. Iron and magnesium are said to substitute for each other, so the chemical variation is called **atomic substitution**. Olivine is described as a **solid solution series** between two end-members. The **end-members** have their own names. Mg_2SiO_4 is called **forsterite** and Fe_2SiO_4 is called **fayalite**.

ionic charge							
+1	+2	+3	+4	+5/-3	-2	-1	0
H^+ hydrogen							He helium
Li^+ lithium	Be^{2+} beryllium	B^{3+} boron	C^{4+} carbon	N^{3-} nitrogen	O^{2-} oxygen	F^- flourine	Ne neon
Na^+ sodium	Mg^{2+} magnesium	Al^{3+} aluminium	Si^{4+} silicon	P^{3-} phosphorus	S^{2-} sulphur	Cl^- chlorine	Ar argon
K^+ potassium	Ca^{2+} calcium						

Figure A2.2 Periodic table showing the first 20 elements. The nine elements highlighted in yellow, along with element number 26, iron (Fe), are the ten important elements in minerals. The charge on each ion is determined by the column in which the element sits. Of the highlighted elements, only oxygen occurs as negative ions.

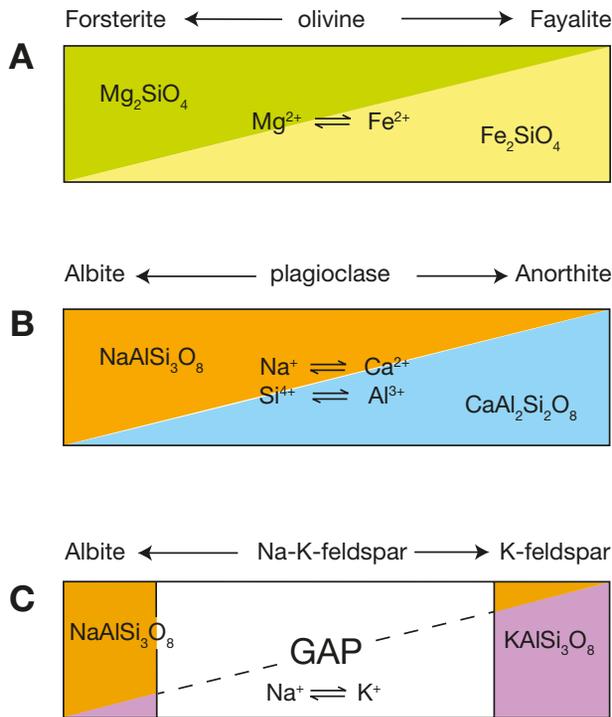


Figure A2.3 Three styles of atomic substitution. (A) Simple substitution of Fe^{2+} for Mg^{2+} in olivine. (B) Coupled substitution of Ca^{2+} for Na^+ and Al^{3+} for Si^{4+} in plagioclase. (C) Limited substitution of K^+ for Na^+ and *vice versa* in Na- and K-feldspars.

Atomic substitution between Fe^{2+} and Mg^{2+} happens because the two ions are the same size. Ions behave like small rigid spheres with a definite radius. A list showing the radii of common ions is presented in Figure A2.4, where Fe^{2+} and Mg^{2+} are both seen to have a radius of about 0.07nm. One nm (nanometre) is one-millionth of a millimetre, so atoms are pretty small!

Another example of atomic substitution occurs in the mineral **plagioclase feldspar** (usually just called plagioclase), which is the most abundant mineral in the continental crust. Plagioclase is a solid solution series between the end-member called **albite**, $NaAlSi_3O_8$, and the end-member called **anorthite**, $CaAl_2Si_2O_8$, (Fig. A2.3-B). The ions Na^+ and Ca^{2+} each have a radius of about 0.10nm (see Fig. A2.4), but they have different charges. Nevertheless, they do substitute freely for each other because a second atomic substitution, of Al^{3+} for

Ion	Radius in nanometres	Coordination number	Coordination polyhedron	Packing geometry
C^{4+}	0.015	[3]		
Si^{4+}	0.04	[4]		
Al^{3+}	0.05	[4] [6]		
Mg^{2+}	0.066	[6]		
Fe^{2+}	0.074	[6]		
Ca^{2+}	0.10	[8]		
Na^+	0.10	[8]		
K^+	0.14	[12]		

Figure A2.4 Radii of the main positively charged ions in common minerals and the way each is surrounded by a polyhedron of large, negatively charged O^{2-} ions (and also $(OH)^-$ ions). The number of oxygen ions in a polyhedron is called the coordination number of the positive ion at its centre. The smaller the radius of a positive ion, the smaller is its coordination number, and the larger is its charge.

Si^{4+} , takes place simultaneously, keeping the overall charge neutral. Al^{3+} and Si^{4+} ions have slightly different radii (see Fig. A2.4) but they are close enough for them to substitute for each other. This style of atomic substitution, where two separate substitutions take place in tandem to keep the charge balanced, is called **coupled atomic substitution**.

A third kind of substitution is called **limited atomic substitution**. It occurs, for example, in **potassium feldspar** (K-feldspar), $KAlSi_3O_8$. Here, Na^+ can substitute for K^+ , but only to a limited extent because the two ions are of different sizes. The radius of Na^+ (about 0.10nm) is much less than the 0.14nm for K^+ (see Fig. A2.4). In the same way, there is limited substitution of K^+ for Na^+ in albite, $NaAlSi_3O_8$. Both these examples are shown together in Figure A2.4-C, where the range of compositions extends a short way out from each end, with a gap in the middle.

The width of the gap shown in Figure A2.4-C is not fixed, but depends on the temperature. The gap gets smaller (the limit of atomic substitution increases) as the temperature rises, and eventually the gap closes altogether (at about

700°C). The reason for this is because the whole crystal structure vibrates more vigorously at higher temperatures, and each kind of feldspar becomes less 'fussy' about accepting ions of the wrong size.

Limited atomic substitution is widespread among minerals, and since it depends on temperature, it has been used in estimating the temperature at which a rock is metamorphosed, as shown in chapter 5.

A2.3 How are atoms (ions) stacked together?

In all minerals the atoms are packed closely together in a pattern called a **crystal structure**, which is a pattern that is repeated regularly in three dimensions with negative and positive ions next to each other. The negative oxygen ions (and also the $(\text{OH})^-$ ions) are larger than most others, having a radius of 0.14nm. They cluster together, as shown in Figure A2.4, to make 'cages' or **polyhedra** out of three, four, six, eight, or twelve atoms. A positively charged ion fits snugly into the centre of a polyhedron (Fig. A2.4). The number of oxygen atoms in a polyhedron is called the **coordination number** of the central positive ion, and it depends on the central ion's size.

Carbon, the smallest, has a coordination number of 3; it sits between three oxygens giving the **carbonate ion**, $(\text{CO}_3)^{2-}$. Silicon, Si^{4+} (radius 0.04nm) has a coordination number of 4; it fits nicely inside a **tetrahedron** (four touching oxygen atoms) so all silicate minerals contain $(\text{SiO}_4)^{4-}$ tetrahedra (the traditional plural of tetrahedron). Mg^{2+} and Fe^{2+} ions (radius about 0.07nm) need more space. They have a coordination number of 6, so are surrounded by six oxygens, making an octahedron (a polyhedron with six corners and eight triangular faces; Fig A2.4). Na^+ and Ca^{2+} , the two ions with a radius of about 0.1nm that substitute for each other in plagioclase, have a coordination number of 8. They fit between eight touching oxygens. K^+ is the same size as an oxygen ion, so it is surrounded by 12 oxygens in a large polyhedron.

Aluminium, Al^{3+} (radius of 0.05nm) is something of a misfit. It is bigger than Si^{4+} but smaller than Mg^{2+} . In many minerals, including feldspar, Al^{3+} ions go into a tetrahedron, while in some minerals, like kyanite, Al^{3+} is in an octahedron. Aluminium's dual behaviour gives rise to a style of coupled atomic substitution where two Al^{3+} ions simultaneously replace one Si^{4+} ion in a **tetrahedral site**, and one Mg^{2+} ion in an **octahedral site**. This kind of substitution is always limited. Octahedral

aluminium is favoured by high pressure, because high pressure compresses the oxygen atoms and reduces the space at the centre of an octahedron, making it better suited for the small Al^{3+} ions.

A2.4 Classification and properties of silicates

Most of the abundant minerals in rocks are composed largely of the elements silicon and oxygen. They are known as **silicate minerals**, or simply as silicates, and they are divided into five main groups. All silicates have silicon–oxygen tetrahedra, and each of the five groups has its own distinctive arrangement of these tetrahedra. In the first group the SiO_4 tetrahedra are independent units, not touching each other. In the other four groups neighbouring tetrahedra are joined at their corners by sharing oxygen atoms. They are joined to make long single chains, or pairs of parallel cross-linked chains (double chains), or sheets, or three-dimensional frameworks. Single chains, double chains and sheets are illustrated in Figure A2.5.

A2.4.1 Silicates with independent tetrahedra

Seven of the common silicates have crystal structures with independent tetrahedra. They are olivine, garnet, kyanite,

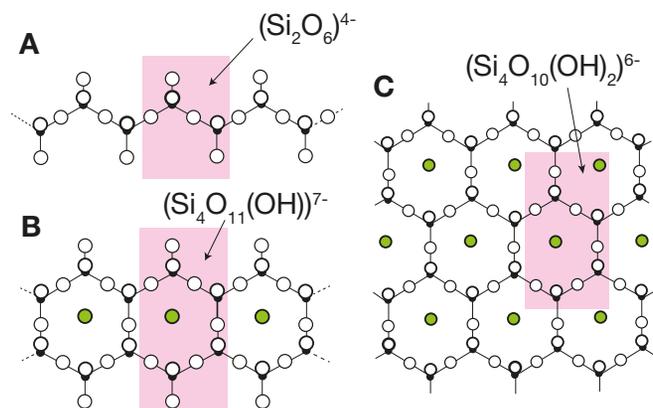


Figure A2.5 Arrangements of SiO_4 tetrahedra in (A) single chain silicates, (B) double chain silicates and (C) sheet silicates. Small black spheres are silicon atoms. Each is linked to four large white spheres, which are oxygen atoms. Green circles in (B) and (C) are hydroxyl $(\text{OH})^-$ ions. In each group, the repeat unit is highlighted by a pink rectangle in which the numbers of atoms can be added up to give its formula and negative charge.

Appendix 3 Minerals under the microscope

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The standard way of examining rocks in close detail is to prepare them as **thin sections** and to look at them through a **polarizing microscope**. Anyone with access to the internet can study thin sections of rock at their leisure using the **Virtual Microscope**. This is an Open University website (virtualmicroscope.org) that provides free access to images of several hundred different rocks. Users can zoom in and out, move around the thin section, and adjust the viewing conditions almost as if they were looking through a real microscope. Many of the images of thin sections in this book are taken from the Virtual Microscope, and most of the metamorphic rocks in the Irish University Rocks collection (the GeoLab collection) were commissioned with the book in mind.

This appendix explains how thin sections are made; it describes the operation of a polarizing microscope and the features of minerals that can be seen when using it, and it suggests a practical approach to identifying common minerals in a thin section.

A3.1 Thin sections

A thin section is a slice of rock, just 30 microns (0.03mm) thick, cemented to a glass microscope slide with epoxy resin. Most minerals at this thickness are translucent (they let light through), so they can be examined using a microscope with the illumination from below.

The preparation of thin sections requires high-precision machinery and a dedicated workshop (Fig. A3.1). First, using a water-cooled diamond-tipped circular saw, a slice of rock a few millimetres in thickness is cut from a rock specimen. A rectangular tablet, typically about 2cm by 4cm, is trimmed from the slice of rock, and one face of it is ground perfectly flat on a lapping wheel. The flattened face is then cemented firmly to a glass microscope slide using clear epoxy resin. The bulk of the tablet is skimmed off with a thin-bladed circular saw leaving a veneer of rock, perhaps 200 microns (0.2mm) thick, attached to the glass slide. The veneer of rock is then ground away gently on a lapping wheel until a thickness of just 30



Figure A3.1 Equipment used, and one of the intermediate steps, in preparation of thin sections. Top left: water-cooled rock saw. Top right: rectangular rock tablet glued to a glass microscope slide. Bottom left: precision circular saw for skimming off most of the tablet. Bottom right: grinding wheel for the high-precision lapping of the skimmed rock, down to 30 microns.

microns remains. The thin section is now almost finished. Final preparation is in one of two ways. The traditional way is to make a covered thin section. Here, a hot, sticky coating of a natural resin called Canada balsam is smeared over the surface and a glass cover slip is gently pressed onto the resin, avoiding entrapment of air bubbles (Fig. A3.2). The covered section is then baked to cure the resin. The second way is to make a polished thin section. In this case the surface is not covered, but simply polished to a mirror finish using gamma alumina or diamond paste on a polishing lap.

Covered thin sections give the best images under the microscope because they show a feature called **relief**. Relief is the apparent roughness of the surface of a mineral grain. Minerals that look flat and featureless, like quartz, are said to have low relief, whereas those that stand out with bold outlines and a rugged surface, like garnet, are said to have high relief (Fig. A3.3).

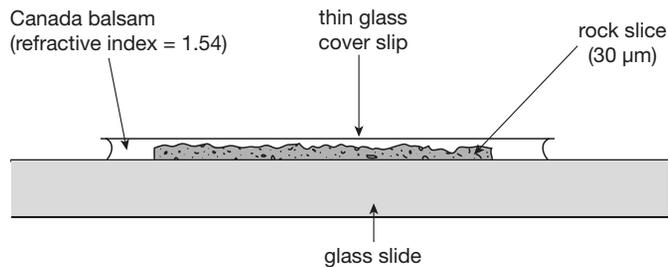


Figure A3.2 Cross-section through a standard covered thin section.

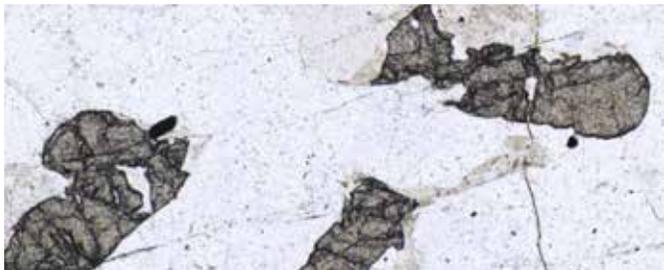


Figure A3.3 Contrasting examples of relief. This microscope image (a screen shot of the Virtual Microscope rock, GeoLab M06) is 2mm wide and shows three grains of garnet surrounded by quartz in a covered thin section. Garnet has high relief: it looks greyish and roughened, with bold margins and dark cracks. Quartz has low relief: its surface is so flat and featureless it can hardly be seen.

Relief depends on an important property of a mineral called **refractive index**. This is the ratio of the speed of light in a vacuum to its (lower) speed through the mineral, so it is a number greater than unity. Minerals have low relief if their refractive index is close to that of Canada balsam, which is 1.54. The refractive index of quartz is about 1.55 and so quartz has low relief. In contrast, garnet's refractive index is about 1.8, so garnet has high relief.

Polished thin sections, on the other hand, can be examined in several different ways. As well as being visible under a regular polarizing microscope, they can be looked at using a reflected light microscope, and they are also amenable to study using an instrument called a **scanning electron microscope (SEM)**. An SEM produces wonderful images, and also permits the *in situ* chemical analysis of tiny parts of mineral grains. The technique is described in Appendix 4.

The method for preparing thin sections was developed, incidentally, in the mid-nineteenth century by Henry Clifton Sorby (Fig. A3.4). Sorby was a brilliant and independently wealthy geologist, biologist, metallurgist, and microscopist who was elected as a Fellow of the Royal Society at the age of 31. Sorby's method of making thin sections revolutionized the

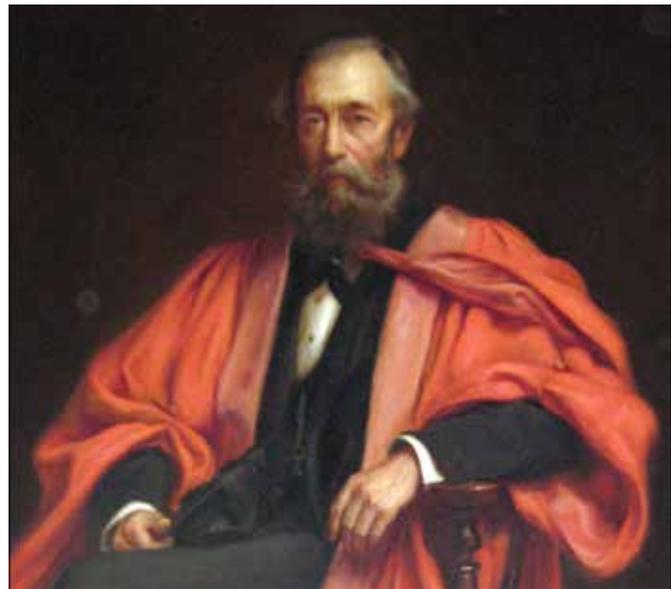


Figure A3.4 Henry Clifton Sorby (1826–1908) who, amongst his many scientific achievements, invented the process for making thin sections of rock. *Source: Wikipedia.*

which has been displaced and now rests on top of, or lies within, continental crust.

order (of interference colour) [121]: see interference colour.

orogenesis (orogeny) [3]: the process of mountain building (a mountain-building event).

orogenic belt [6]: tract of country where orogenesis is now in progress or where it took place in the past. In the second case it is the same thing as a metamorphic belt.

orthoamphibole [114]: amphibole with only octahedral sites between the chains, and having the formula $(\text{Mg,Fe})_7\text{Si}_8\text{O}_{22}(\text{OH})_2$.

orthogneiss [41]: gneiss derived from an igneous protolith.

orthopyroxene [33, 113]: pyroxene with only octahedral sites between the chains, and having the formula $(\text{Mg,Fe})_2\text{Si}_2\text{O}_6$, also written $(\text{Mg,Fe})\text{SiO}_3$.

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paragneiss [41]: gneiss derived from a sedimentary protolith.

parallel extinction [123]: the case where the elongate outline, or the trace of a single set of cleavage planes, of a mineral grain in a thin section is orientated perfectly up-and-down or perfectly sideways when the grain is in extinction. Where the orientation at extinction is diagonal (inclined), the mineral is said to have inclined extinction.

partial melting (= anataxis) []: a process whereby metamorphism at a high temperature (high grade) turns a rock into a very hot 'slush' of liquid and residual solid grains. If the rock is peridotite, the liquid is usually basaltic, and if the rock is pelitic the liquid is usually granitic.

pelite [14]: obsolete term for shale or mudstone, but sometimes used instead of metapelite.

peridotite [1, 44, 104]: a rock composed largely of the mineral olivine. It is the main rock of the mantle.

Periodic Table [110]: a table in which the chemical elements are listed in the order of their atomic number (the number of protons each element contains), starting with hydrogen. The elements are laid out in rows (periods) and columns (groups) and the position of each element reflects its chemical behaviour.

perthite [91]: a variety of feldspar comprising finely interdigitated layers of albite and K-feldspar, formed by the process of exsolution.

petrography [11]: the description of rocks. It is the first stage in the overall study of rocks, known as petrology. The second stage, interpretation, is called petrogenesis.

phyllite [3, 16]: a foliated metamorphic rock that is transitional between slate and schist. It is usually, though not always, derived from shale during low-grade metamorphism.

pillow basalt [30]: basalt that resembles a pile of pillows, each draped on those below. It can only form where basaltic lava is erupted under water. The shapes of the pillows may be preserved during metamorphism.

plagioclase [30]: a framework silicate and the most abundant variety of feldspar. It forms a solid solution series between a sodium end-member called albite ($\text{NaAlSi}_3\text{O}_8$) and a calcium end-member called anorthite ($\text{CaAl}_2\text{Si}_2\text{O}_8$).

plane polarized light [120, 122]: see PPL.

planar deformation features (PDFs) [85]: a product of damage caused by shock metamorphism, seen as sets of dark parallel bands in quartz and other minerals.

plate (= tectonic plate) [6]: one of the dozen or so huge curved 100-km-thick slabs of cool, strong uppermost mantle, known as the lithosphere. Together the plates cover the entire surface of the planet. They move slowly relative to each other. Most plates are partly capped by continental crust, and partly by oceanic crust.

plate tectonics [104]: a theory of the Earth based on plates.

platy [65]: a term describing crystals with a broad flat shape, like mica.

pleochroism (adjective pleochroic) [19, 123]: the case where a mineral grain in thin section, viewed in PPL, changes colour as the microscope stage is turned.

pluton [7]: a term commonly used for a body of coarse-grained intrusive igneous rock, such as granite, that is not a sill or dyke, and is usually several kilometres across.

plutonic igneous rock [51]: a coarse-grained rock that solidified slowly as a pluton or other large intrusive body.

poikiloblast [19]: a porphyroblast with many inclusions.

polarizing microscope [118]: a microscope with two polarizing filters and a rotating stage, used to examine thin sections of rock.

polyhedron (adjective polyhedral, plural polyhedra) [62, 112]: a three-dimensional shape bounded by a number of flat surfaces. Examples are a tetrahedron (4 surfaces – the minimum), a cube (6), an octahedron (8), a dodecahedron (12) or shapes with any other number of flat surfaces. Polyhedra are the shapes adopted by bubbles in foam, by mineral grains in many metamorphic rocks, and, on the scale of atoms, by clusters of oxygen atoms surrounding positive ions.

polymorph [23, 113]: one of two or more different minerals each having the same chemical formula.

porphyroblast [18]: a mineral grain in a metamorphic rock, often with a distinct crystal shape, that has grown to a conspicuously larger size than the other grains in the rock.

porphyroclast [42]: a mineral grain in mylonite or other cataclastic rock that, having survived crushing, is conspicuously larger than the other grains in the rock.

potassium feldspar (K-feldspar) [111]: feldspar with the formula KAlSi_3O_8 . It is an essential mineral in granite.

PPL [11]: plane polarized light, produced when only one polarizing filter is in the light path in a polarizing microscope.

prehnite-pumpellyite [39]: a pair of hydrous, calcium-rich silicate minerals that form together, for example in basic igneous rocks, at low grade and give their names to a metamorphic facies.

pressure [5]: a force that increases with depth and tends to reduce the volume (increase the density) of rock by squashing it equally in all directions. It is also known as lithostatic stress, and is conventionally expressed in units called kilobars (kbar).

prismatic []: a term describing crystals (e.g. hornblende, andalusite) with a long, thin shape.

prograde [21, 56]: metamorphic changes that develop during increasing grade (temperature) are described as prograde.

protolith [1]: the original igneous or sedimentary rock from which a metamorphic rock is derived.

psammite [13]: obsolete term for sandstone, sometimes used instead of metapsammite.

pseudomorph [27, 70]: a grain of one mineral that has been replaced by another mineral, or minerals, yet has retained its original shape.

pseudotachylite [43, 85]: a dark, glassy product of intense dynamic