S276 Geology
Science: Level 2

Book 2
Magmas and Mountains

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Chapter 4 Intrusive igneous rocks

This chapter considers igneous bodies produced by intrusion of magmas that did not reach the surface. It begins by looking at the pathways taken by magma in the shallow sub-volcanic level, and then investigates deeper, larger bodies of intrusive rock that represent solidified magma chambers.

4.1 Minor intrusions

Magma that erupts onto the Earth's surface and produces extrusive igneous rocks clearly comes from below the ground, but what form does the magma transport system take, and can this system be recognised in the geological record, on maps and in the field? By their very nature, it is impossible to observe directly the subsurface magma conduits supplying an ongoing volcanic eruption. So, as with many aspects of geology, various bits of indirect evidence have to be pieced together in order to solve the problem.

■ What approaches could be used to gather evidence about the shallow roots of volcanoes?

□ There are three ways of doing this: interpreting observations of active vents made during or after an eruption; interpreting geophysical signals such as earthquakes and surface deformation that occur during an eruption; and interpreting rocks exposed by erosion of old volcanoes.

In some eruptions, such as those along the rift zones of Hawaii's volcanoes and in Iceland, activity occurs along a row of craters that can be many kilometres long (Figure 4.1a). The simplest explanation for this is that directly beneath the eruptive fissure must lie a long narrow vertical intrusion feeding magma upwards to the active vents.

■ What name is given to an intrusion with this shape?

□ A dyke.

Recall that a dyke is a discordant curtain-like intrusion (Figure 4.2), intruded in a near-vertical plane, whereas a sill is a sheet-like body, originally intruded horizontally, that is mostly concordant with the bedding of the strata it intrudes.

Many of the eruptions in Iceland involve fire fountaining from vents located along fissures. Others erupt from vents at centres of activity associated with hydrothermal activity and calderas located at the centre of fissure swarms (Figure 4.1b). The central volcanoes are thought to overlie persistently hot magma chambers. Each central volcano and its associated fissure swarm define a separate volcanic system.

One of northern Iceland's volcanic systems, Krafla, was active between 1975 and 1984, with much seismic activity, ground deformation and fire fountaining. The activity was cyclic, with slow inflation of the ground above the central volcano followed by fast deflation and migration of earthquakes away from the centre of ground deformation. This pattern was repeated 21 times, with fissure eruptions happening 9 times during periods of deflation and earthquake migration.

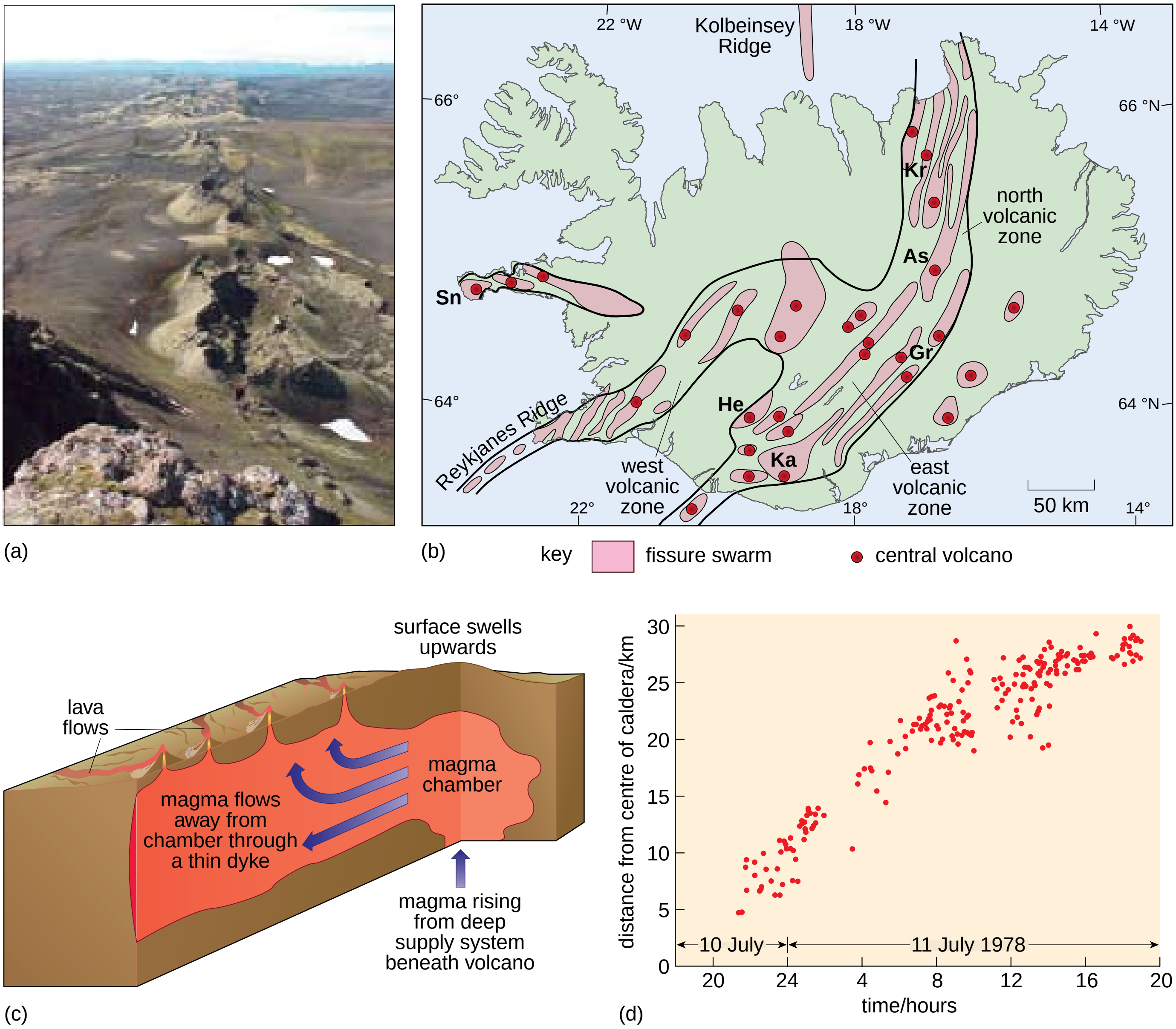


Figure 4.1 (a) Aerial view of the 27 km-long Laki crater row, southern Iceland, which erupted basaltic scoria and lava in 1783–1784 in one of recorded history's largest volcanic eruptions (involving about 15 km^3 of magma). (b) Map of Iceland showing locations of central volcanoes and fissure swarms arranged in volcanic zones across the country. Selected volcanic systems identified by abbreviations: As, Askja; Gr, Grimsvotn (source of the Laki eruption); He, Hekla; Ka, Katla; Kr, Krafla; Sn, Snaefellsjokull. (c) Interpretive model of vertical magma rise beneath a volcano and lateral emplacement of a dyke, followed by eruption along a crater row. (d) Plot of earthquake location, measured northwards from the centre of the Krafla caldera during 10–11 July 1978.

Figure 4.2 Erosion has exposed the three-dimensional view of a dyke, which forms the jagged crest of the ridge emanating from a larger igneous intrusion known as Ship Rock (New Mexico, USA). The dyke varies from 0.5 to 5 m in width. A second dyke is visible in the background, trending to the left of the photograph.

The activity at Krafla can be explained if the volcanic system comprises a shallow magma chamber into which more magma is injected from below (Figure 4.1c). This causes the chamber to expand and the ground to swell upwards until the strength of the enclosing rocks is exceeded. At this point, fractures open up in the surrounding rock, allowing magma to escape from the chamber into a newly forming dyke. The ground surface deflates and the dyke extends in length by propagating away from the chamber, causing small earthquakes wherever the crust is broken apart at the tip of the advancing dyke (Figure 4.1d). In some cases magma breaks through to the surface at points above the dyke, giving rise to fire fountains.

Question 4.1

For the case of the dyke emplacement event at Krafla (Figure 4.1d), estimate the approximate speed at which the dyke advanced. Was the speed constant?

Dykes rarely occur in isolation, and many areas of the crust have been repeatedly intruded by dykes, by magma intruding either from below or laterally from a shallow chamber. For example, deeply eroded areas of eastern Iceland expose parallel dykes up to a few metres wide cutting vertically through basaltic lava flows and making up to 8% of the crust. These appear to be the subsurface equivalent of fissure swarms that characterise present-day volcanic systems such as Krafla. The parallel orientation of the dykes indicates that the path of least resistance taken by the intruding magma was always in the same direction. A reasonable explanation for this is that the crust was consistently being pulled apart in the direction perpendicular to the strike of the dykes. The large array of dykes is known as a **dyke swarm**.

In the absence of external influences, dykes will be radially distributed about a central source, as found for example in the eroded cores of stratovolcanoes in regions where the crust has not been subject to tensional forces in a preferred direction (Figure 4.3a). Radial dykes may reach the surface of the flanks of an active volcano (in which case they may act as sources for eruptions), but are more clearly seen after a long period of deep erosion.

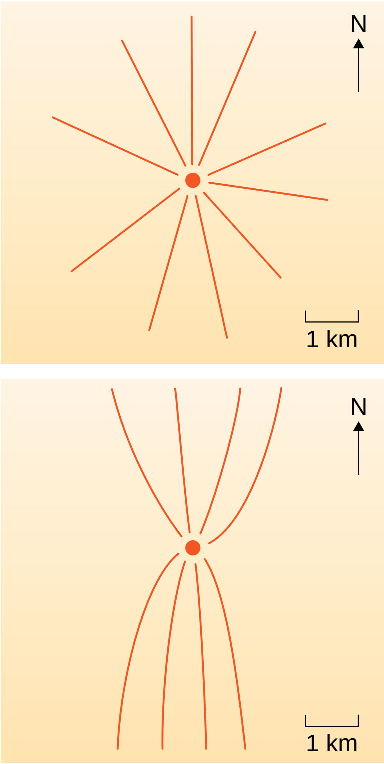


Figure 4.3 Schematic maps of the outcrop patterns of: (a) radial dykes around a volcano intruded in the absence of regional stress; (b) a dyke swarm associated with a central volcano in a region of east–west extension.

Dykes are important means by which magma moves through the crust and to the surface. But other types of intrusion can be found, especially horizontal intrusions, which are known as sills. Sills may be metres to hundreds of metres

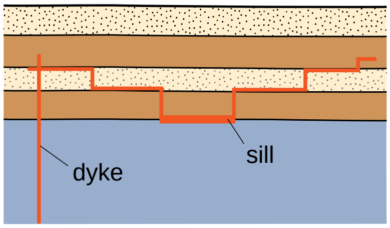


Figure 4.4 Cross-section showing the typical relationship between a sill (here fed by a dyke on the left) and horizontal strata intruded by it. The sill is generally concordant, but is locally discordant where it cuts up or down between bedding planes.

thick, and usually occur only at very shallow levels within the crust where mafic magma has been injected into horizontal or gently dipping sedimentary strata within a kilometre or so of the surface (Figure 4.4).

If dykes intrude as vertical ‘walls’ of magma because the surrounding rock can be most easily pushed (or pulled) aside horizontally, then sills intrude because it is energetically easier to lift the overburden of rocks than to shoulder aside the country rock.

When dykes and sills intrude, the enclosing country rock becomes heated by the magma, but the total amount of heat escaping from a minor intrusion is generally insufficient to cause extensive contact metamorphism of the rocks it invades. Thus, sedimentary rock may be baked and become hard and brittle in a zone a metre or so wide against the contact with a dyke or sill, but growth of new, metamorphic minerals is uncommon.

In contrast, the consequences of cooling within the intrusion can be more noticeable. Intrusions lose heat to the country rock and therefore solidify from the outside in. As the magma cools and solidifies into igneous rock it shrinks, with the result that contraction cracks develop, penetrating into the intrusion to generate hexagonal columns arranged parallel with the cooling direction (and perpendicular to the cooling surface). This is why dykes and sills often have regular six-sided columnar joints arranged perpendicular to their margins (Figure 4.5). For the same reasons, columnar joints can be found in some lava flows.

Figure 4.5 Vertical columnar joints in a 30 m-thick dolerite sill at Kilt Rock, Skye, formed by thermal contraction as the sill cooled through its roof and floor. The contact between the base of the sill and underlying horizontal sandstone strata is roughly halfway down the cliff. A second sill is also exposed near the base of the cliff, but columnar jointing is much less well developed. Reproduced with the permission of the British Geological Survey © NERC. All rights Reserved.

The loss of heat from a dyke or sill that has been emplaced into cold country rock is sufficiently rapid to affect the texture of the igneous rock within the minor intrusion. The magma at the margin is cooled so rapidly that only a few small crystals have time to form, and the texture of the igneous rock at the contact is often glassy. Minor intrusions are therefore characterised by **chilled margins** and interiors with medium grain size. A thin section through the chilled margin at the base of the Whin Sill in NE England is shown in Figure 4.6. The field of view is too small to show any of the medium-grained, completely crystalline interior of the sill.

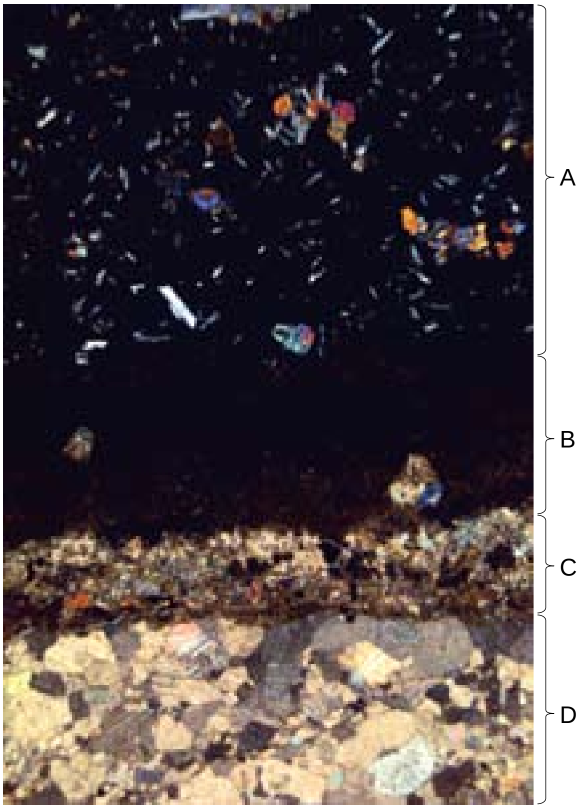


Figure 4.6 Thin section through the base of the Whin Sill (viewed between crossed polars), showing an area about 1.4 mm long. In region A, crystals of pyroxene and plagioclase feldspar occur in a fine-grained and partly glassy matrix. Region B is the actual chilled margin, which is glassy throughout except for a couple of olivine crystals that presumably grew within the magma before it was emplaced. Regions C and D are within the country rock, which in this case is limestone. You should recognise the pastel interference colours of the calcite crystals in D. In region C (only about 0.2 mm wide), some new metamorphic minerals have begun to grow because of the intense heat from this >50 m-thick sill.

Question 4.2

Figure 4.7 shows black and white images of thin sections of two samples from the Cleveland Dyke, County Durham, England, collected from where the dyke is about 20 m wide. Look at these, and then attempt the following.

(i) Decide whether the average grain sizes of (a) and (b) in Figure 4.7 are coarse, medium, fine or glassy.

(ii) Which of the two thin sections comes from closer to the margin of the dyke?

(iii) How would you explain the presence of the relatively long (0.3 mm) plagioclase feldspar crystal visible in each specimen?

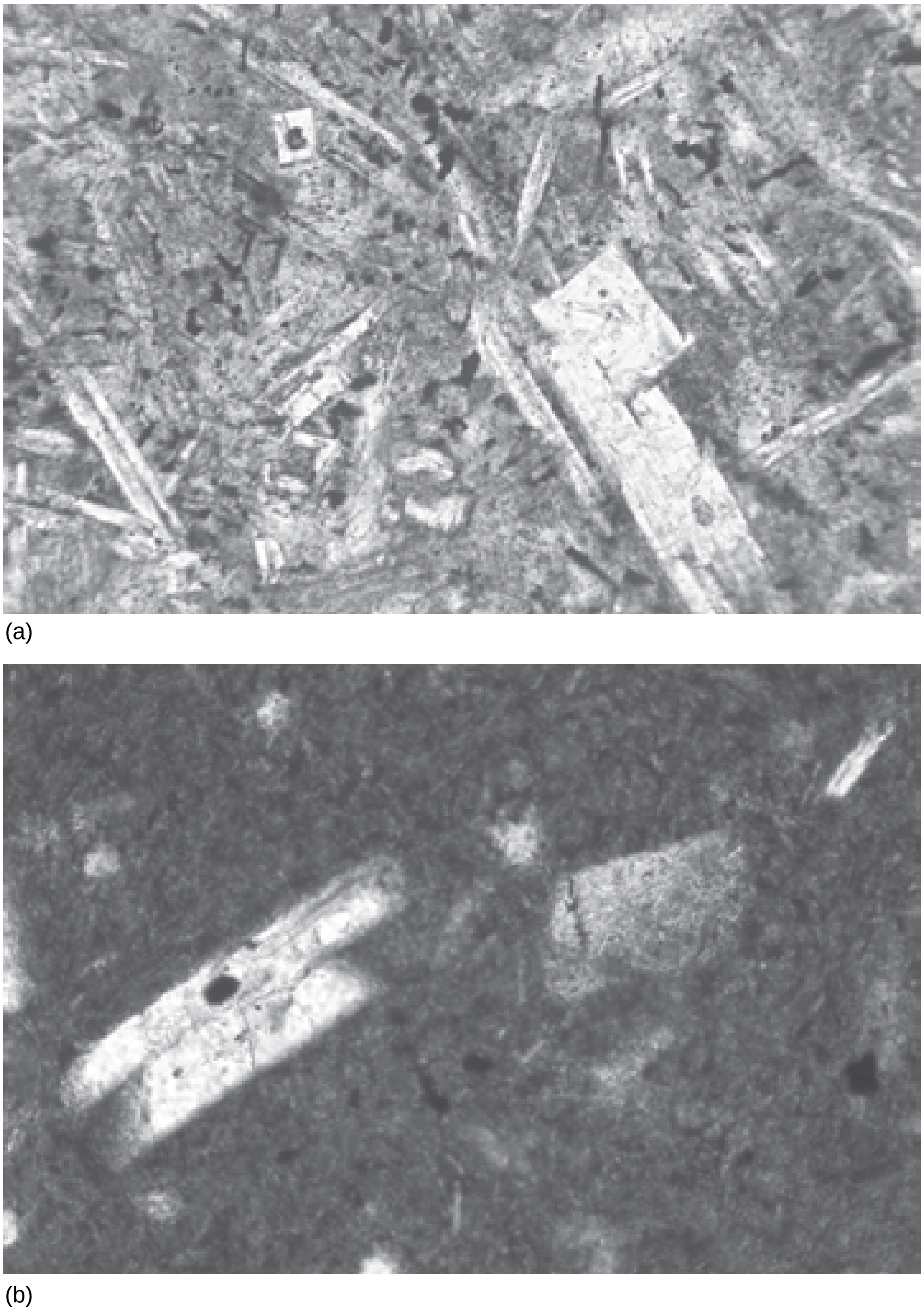


Figure 4.7 (a) and (b) Thin sections from two parts of the Cleveland Dyke (plane-polarised light, photographed in black and white) (field of view 0.7 mm across). For use with Question 4.2.

Sills and dykes are minor intrusions that are characteristically planar in shape. In outcrop, dykes typically define straight lines that cut across topography as a result of being vertical, whereas sills follow contours as a result of being flat-lying. There are some minor intrusions, however, that can have other outcrop patterns. Cone sheets and ring dykes are dykes with approximately circular outcrop patterns (so they depart from the definition of a dyke as a planar body). **Cone sheets** converge downwards towards the top of the magma chamber. They are usually narrow features a few metres or less in width. Conversely, a **ring dyke** dips steeply outwards and is typically anything from a hundred metres to a couple of kilometres wide. The geometry of these two kinds of minor intrusion is illustrated in Figure 4.8; study this and then attempt Question 4.3.

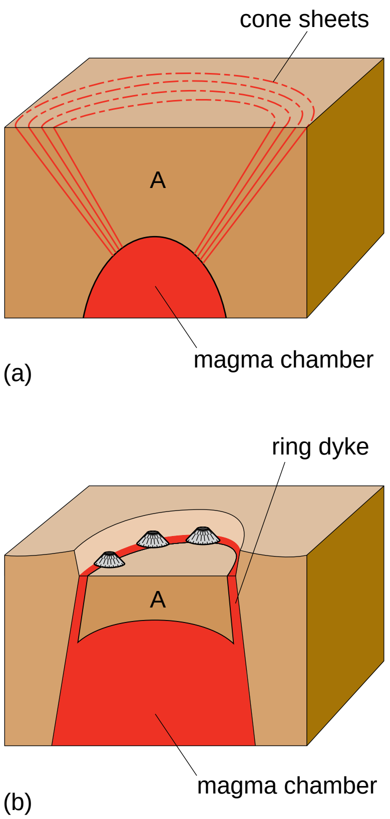


Figure 4.8 Diagrammatic cross-sections through (a) a series of cone sheets and (b) a ring dyke. The circular depression bounded by the ring dyke in (b) is an example of a caldera.

Question 4.3

Bearing in mind that emplacement of dykes of any kind requires displacement of the country rock, how do the blocks of the rock labelled A on Figure 4.8 move relative to the magma chamber during emplacement of (a) cone sheets, (b) a ring dyke? In each case, decide whether this implies that the minor intrusion was created by forceful injection or passive intrusion of magma.

Activity 4.1 Minor intrusions on geological maps

In this activity, you will use geological maps to deduce how and when some different types of minor intrusion were emplaced in Britain's geological past.

4.2 Plutons

The minor intrusions discussed in the previous section are typically 0.1 to 100s of metres thick. Larger bodies of intrusive igneous rock, of kilometre-size or more, are known as plutons and are produced by solidification of substantial bodies of magma at depths of several kilometres or even tens of kilometres.

■ Bearing in mind the depth at which a pluton crystallises, would you expect it to be fine, medium or coarse grained?

☐ A pluton crystallises deep below the Earth's surface, therefore it cools slowly and the crystals have time to grow to a large size. Plutons are typically coarse grained.

■ Bearing in mind its grain size, what rock type would you expect to find in a pluton of felsic composition?

☐ Granite, according to the classification scheme you met in Table 6.2 of Book 1.

■ Why are plutonic rocks such as granite now exposed at the surface?

☐ They are exposed as a result of erosion.

Next, we will look at examples of granite plutons and their relationships with the country rock in cases representing different depths within the crust, beginning with 'deep' granite plutons emplaced at mid-crustal levels. At this depth (about 20 km), most continental crust has a felsic or intermediate composition and the temperature and pressure are generally sufficient for the rock to be regionally metamorphosed to felsic and intermediate gneisses and schists with well-developed metamorphic foliation. When a gneiss is subjected to conditions at which melting begins, a small amount of partial melting can turn its texture into that of a migmatite (Book 1, Figure 8.5) containing a pervasive network of granite veins. The deepest granite plutons therefore tend to consist of elongate masses of granite aligned parallel to the metamorphic foliation of the country rock (Figure 4.9a), which is a sign that within these hot crustal regions granite magmas can form, collect into larger bodies and start to move around, locally intruding the migmatite. Often the final stage of deformation of the country rock outlasts the intrusion of the granite, so that the granite itself has a foliation imposed upon it.

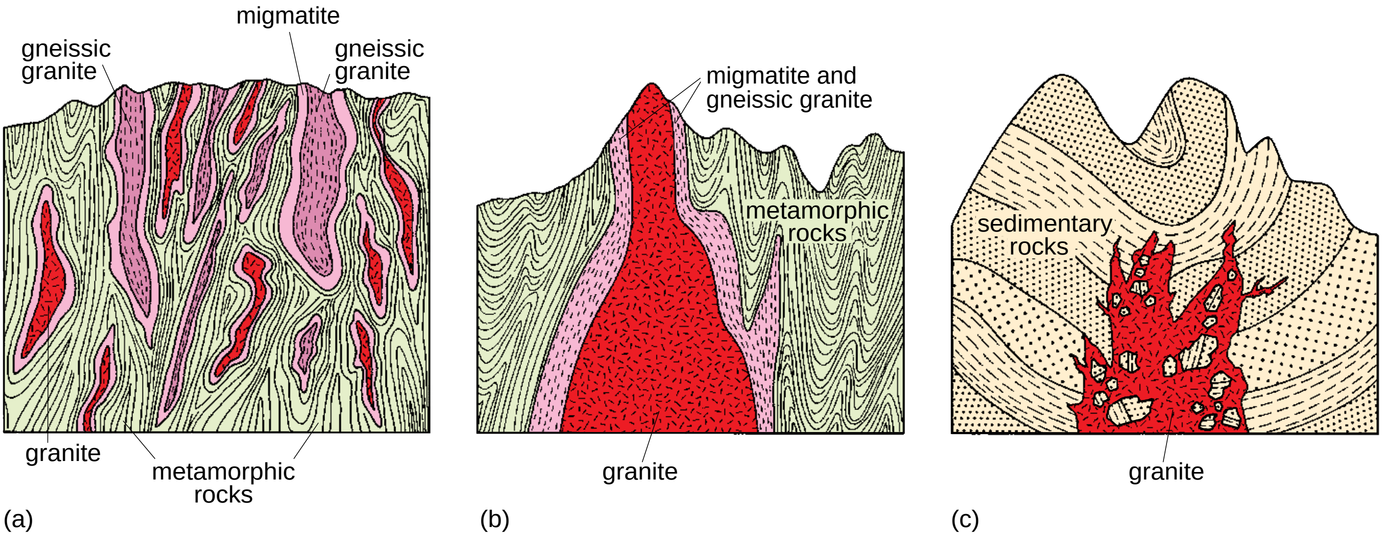


Figure 4.9 Schematic cross-sections of granite intrusions emplaced at different crustal depths and now exposed by erosion: (a) deep (more than about 20 km) mid-crustal intrusions associated with migmatites; (b) a concordant pluton intruded at about 10 to 20 km depth; (c) a discordant pluton intruded at shallow depth (less than about 10 km).

At somewhat shallower depths it is more common to find a single large pluton rather than many small ones (Figure 4.9b), but the plutons still tend to be parallel to the foliation within the country rock, rather than cross-cutting it. In other words, these plutons tend to be concordant to the foliation of metamorphosed country rock, rather than discordant with it. This is because at this depth the country rock was hot enough (and therefore deformable enough) for a pluton-sized mass of granite to force its way upwards by pushing the country rock aside (possibly aided by assimilation of country rock into the magma). A pluton that has risen in this way is described as a **diapir** and might be envisaged as a balloon of magma creeping upwards through the crust until it can move no further.

Plutons believed to have been emplaced at still shallower levels in the crust (<10 km depth) can usually be seen to be discordant. Their edges cut across the fabric of the country rock, which at this depth is usually sedimentary bedding (Figure 4.9c). When intruding into cold, brittle crust, a shallow-level pluton cannot push the country rock aside and force its way up diapirically, as may sometimes be the case deeper down. Nor can it simply melt its way upwards by assimilating all the country rock in its path, because this would require more heat than the pluton contains.

Given that assimilation cannot be a major factor, space must somehow be generated within the crust for a pluton to occupy. As you have seen, the middle crust may be deformable enough to be pushed aside and deform around a diapir. However, plutons in the upper crust can often be seen to occupy spaces made available by fracturing and fault movement where the crust was being pulled apart. It is this pulling apart that often opens up space into which magmas can intrude (Figure 4.10). A further example of fault-controlled intrusion is a **ring fracture** that allows a cylinder of country rock to subside while the magma rises up the fracture and collects in an overlying magma chamber where it solidifies as a pluton (Figure 4.11). In addition, particularly shallow plutons may be accommodated by a certain amount of updoming of the surface.

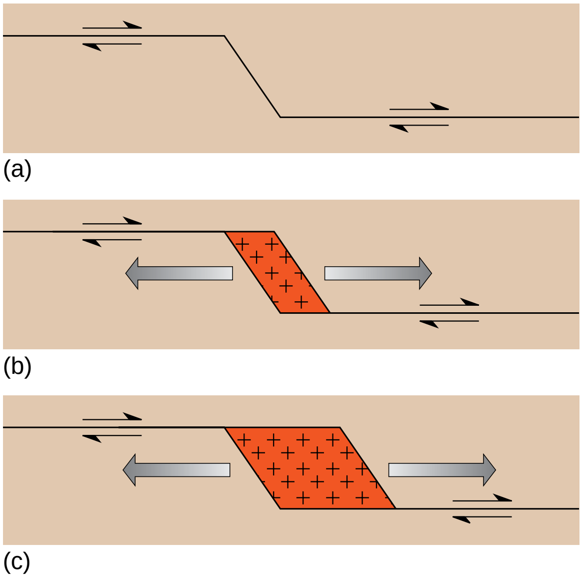


Figure 4.10 Sketch maps showing how movement along parallel faults can open up space that can be occupied by magma intruding from below.

Field evidence suggests one more process by which plutons may sometimes intrude the shallow crust. Figure 4.12a is a photograph of the roof zone of a pale granite pluton that has been intruded at shallow level into darker country rock, resulting in angular blocks of country rock being broken off into the intruding magma. A chunk of country rock enclosed within an intrusion is an example of a **xenolith** (xeno, pronounced ‘zeno’, is Greek for ‘foreign’, and the xenoliths are ‘foreign rocks’ in the sense that they are unrelated to the origin of the igneous rock in which they are found).

The process by which xenoliths are plucked away from the country rock by the magma is known as **stoping** (pronounced ‘stow-ping’). Some xenoliths may sink to the bottom of the pluton. Others can sometimes be identified near the centres of plutons, where they tend to have more rounded shapes, and sometimes a recrystallised fabric, indicating that the heat from the surrounding magma was sufficient to cause them to become soft and mushy. In extreme cases, xenoliths are no more than ghostly dark patches (Figure 4.12b), showing that they have become almost entirely assimilated into the granitic magma. It can be difficult to distinguish these types of xenolith from patches of other magmas that have been incompletely mixed with the main granite magma.

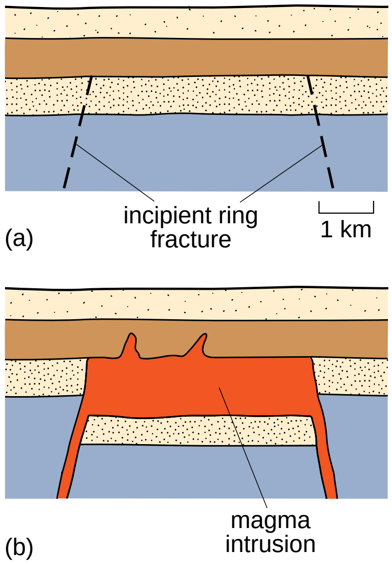


Figure 4.11 Cross-sections to show emplacement of a shallow-level pluton by subsidence of a cylinder of country rock bounded by a ring fracture. (a) A ring fracture forms, which would be approximately circular in plan view, but in this example does not reach the surface. (b) The cylinder of country rock bounded by the ring fracture subsides, and magma rises up to occupy an overlying magma chamber.

■ You have seen that the heat from a granite can have extreme effects on xenoliths that find their way into the heart of the intrusion. But what effect, if any, would you expect to see in the country rock in contact with the granite?

□ You would expect heat to have been conducted from the granite into the country rock, and this could have caused contact metamorphism.

The zone of contact metamorphosed rock around a shallow pluton is described as its metamorphic aureole. This is alternatively known as a ‘contact aureole’ or ‘thermal aureole’ and can be of the order of a kilometre wide. The heat of

(a)

(b)

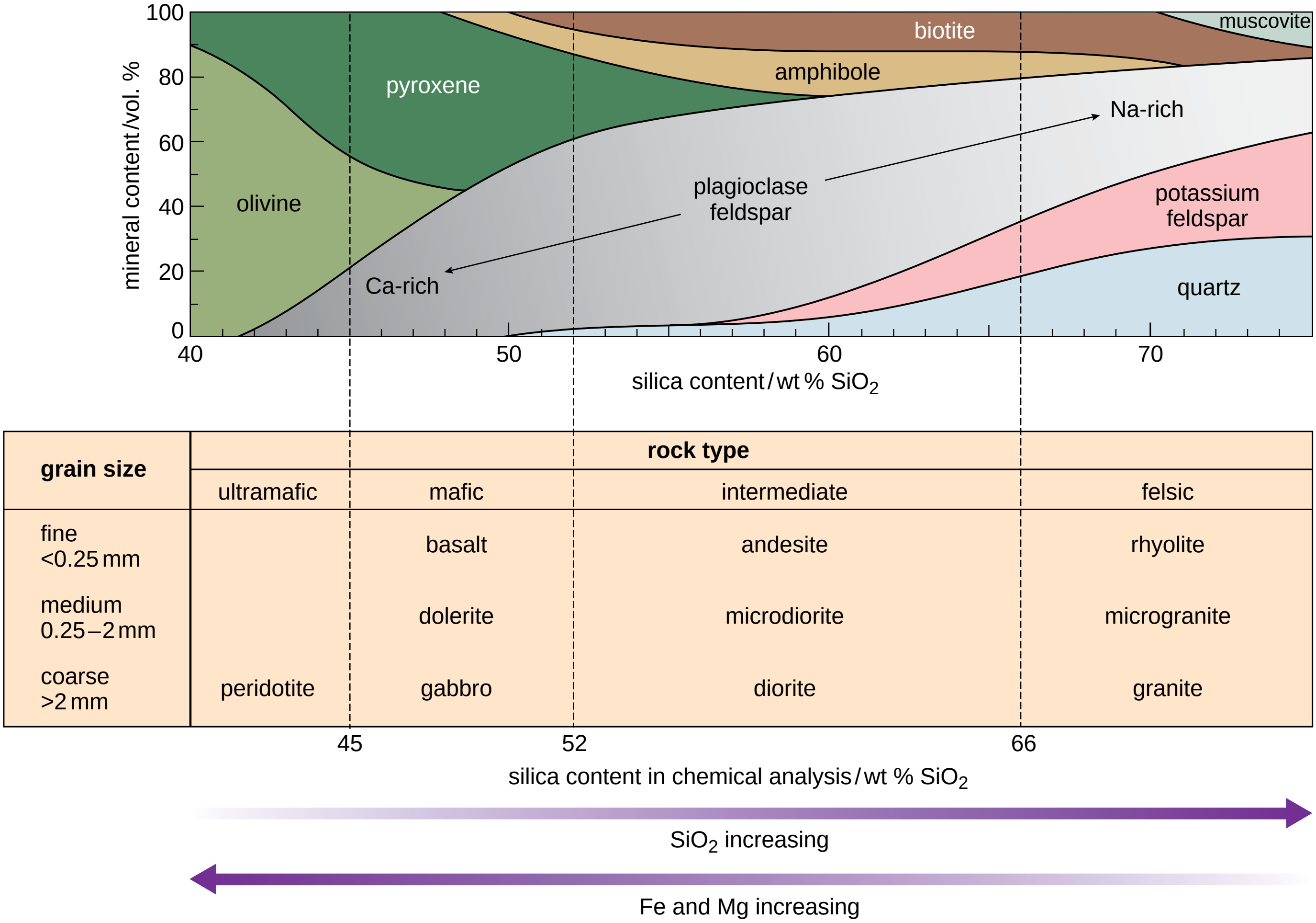
Figure 4.12 Examples of xenoliths in two granite plutons. (a) Part of a granite pluton, near its upper contact, showing an abundance of dark angular xenoliths broken from the enclosing country rock. (b) A xenolith with indistinct margins in granite. (Vertical marks are drill holes related to quarrying.)

the pluton may also cause water within the crust to convect by hydrothermal circulation, in which hot water is expelled through the roof of the intrusion and colder water is drawn in through its sides. This fluid dissolves and reprecipitates certain elements, some within the pluton and some within the country rock, and the process may continue long after the pluton has solidified. Such movement of elements by hot solutions (sometimes referred to as hydrothermal fluids) can cause further mineralogical changes in both the pluton and the country rock – an effect that is described as **metasomatism**. It may also lead to deposition of ore minerals, particularly in veins within and around the pluton, sometimes in economic quantities.

To describe and understand the geology of plutonic rocks, it becomes necessary to have a classification scheme for the different rock types that are encountered. The one in Figure 4.13 is based on a compositional classification according to silica content as reflected by the abundances of quartz, feldspars and mafic minerals. You met this diagram in Book 1 and it expresses a gradation between felsic and ultramafic compositions. Although silica content is the most important variable in igneous rock composition, it is not the only one. Perhaps the next most important factor (and one which can vary independently of silica content) is the relative abundances of the alkali metals sodium (Na) and potassium (K).

■ What common mineral group contains large amounts of Na and K?

□ The feldspars, particularly the sodium-rich plagioclase feldspar albite ($NaAlSi_3O_8$) and potassium feldspar ($KAlSi_3O_8$).



grain size	rock type			
	ultramafic	mafic	intermediate	felsic
fine <0.25 mm		basalt	andesite	rhyolite
medium 0.25–2 mm		dolerite	microdiorite	microgranite
coarse >2 mm	peridotite	gabbro	diorite	granite

Figure 4.13 Classification of igneous rocks according to mineralogy and SiO_2 content.

Plagioclase feldspar exists as a solid-solution series of any composition between the two end-members albite ($NaAlSi_3O_8$) and anorthite ($CaAl_2Si_2O_8$).

■ According to Figure 4.13, which types of coarse-grained igneous rocks have plagioclase feldspar, quartz and alkali feldspar adding up to more than 50% of the rock?

☐ All granites and many diorites have these three minerals in abundances that add up to more than 50%.

The importance of quartz, plagioclase feldspar and alkali feldspar has led geologists to devise the classification system for coarse-grained igneous rocks shown in Figure 4.14. This requires the geologist to be able to distinguish and estimate the relative proportions of the three felsic minerals alkali feldspar, plagioclase feldspar and quartz. This is best done using a petrological (polarising) microscope, as alkali feldspar and plagioclase feldspar are not always easily distinguished in hand specimen. Figure 4.14a is often known as a **QAP diagram**, after the initial letters of the three minerals on which it is based.

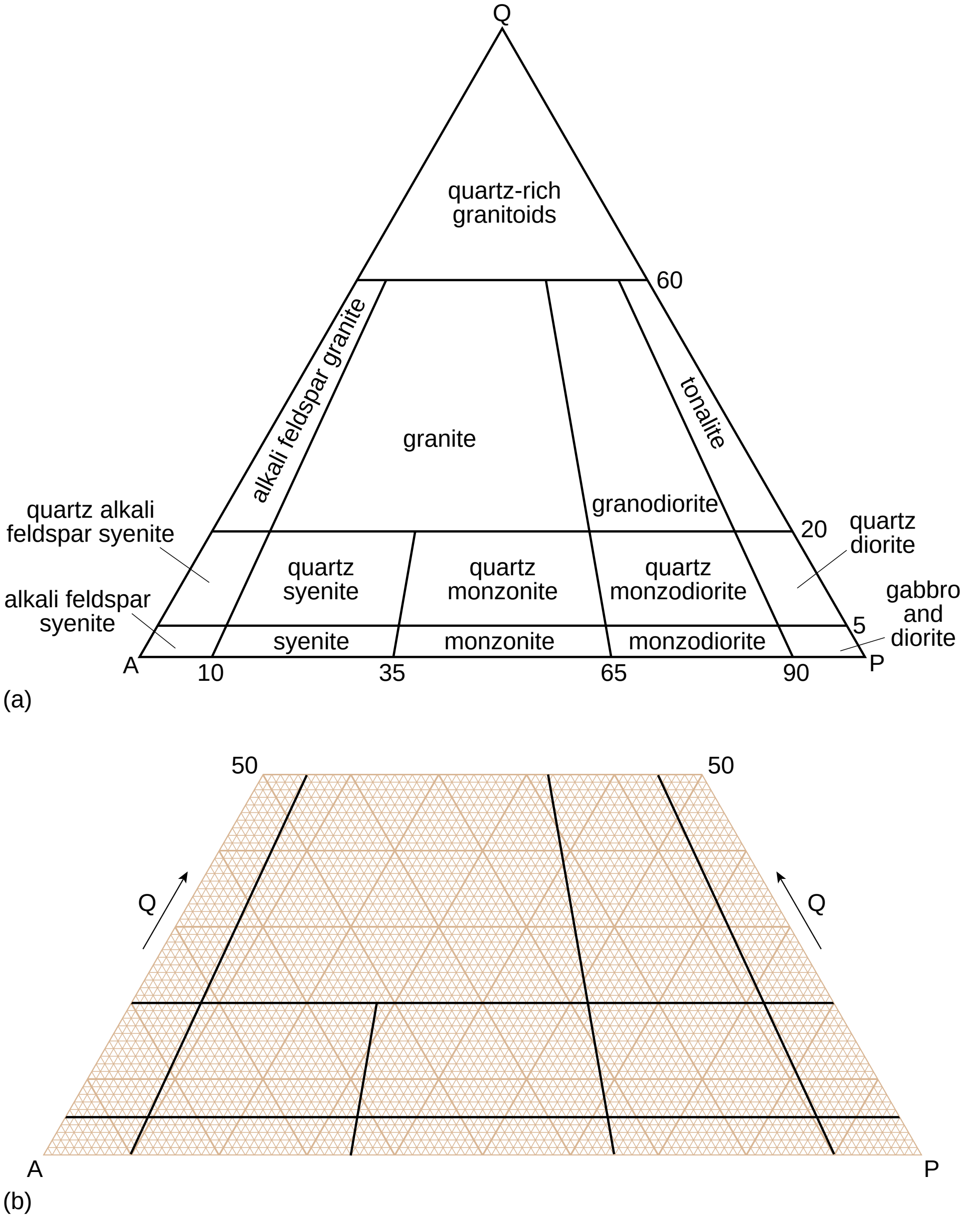


Figure 4.14 (a) The QAP diagram for plutonic igneous rocks according to the relative percentages of quartz (Q), alkali feldspar (A) and plagioclase feldspar (P). Medium-grained rocks take the name of their coarse-grained equivalents with the prefix ‘micro’ added (e.g. microsyenite), except that a ‘microgabbro’ is usually called a dolerite. Gabbro is distinguished from diorite by the composition of its plagioclase feldspar, which is >50% albite in diorite and <50% albite in gabbro. (Note: all these rock types have other minerals present, notably mafic minerals, but this classification is based only on the quartz and feldspar content.) (b) The lower half of (a) with the boundaries between the fields of the different rock types superimposed on a regular grid of triangular plotting paper.

■ A plutonic rock contains 17% quartz, 13% alkali feldspar, 42% plagioclase feldspar, 7% hornblende, 16% biotite and 5% accessory minerals. Calculate the relative percentages of quartz, alkali feldspar and plagioclase feldspar. According to Figure 4.14, what is the name of the rock?

☐ The percentage quartz value (Q) is:

$$\frac{17}{17+13+42} \times 100\% = 24\% \text{ (to 2 significant figures)}$$

The percentage alkali feldspar value (A) is:

$$\frac{13}{17+13+42} \times 100\% = 18\% \text{ (to 2 significant figures)}$$

The percentage plagioclase value (P) is:

$$\frac{42}{17+13+42} \times 100\% = 58\% \text{ (to 2 significant figures)}$$

The rock is therefore classified as a granodiorite.

To reassure yourself that you can use Figure 4.14 to classify igneous rocks, try Question 4.4.

Question 4.4

What name would you give to igneous rocks containing the relative proportions of quartz, alkali feldspar and plagioclase feldspar given below, bearing in mind the grain size? (Note: the values given here have already been recalculated so as to exclude the proportions of other minerals.)

(a) Coarse-grained: 3% quartz, 72% alkali feldspar, 25% plagioclase feldspar.

(b) Medium-grained: 30% quartz, 15% alkali feldspar, 55% plagioclase feldspar.

(c) Coarse-grained: 25% quartz, 5% alkali feldspar, 70% plagioclase feldspar.

(d) Coarse-grained: 3% quartz, 2% alkali feldspar, 95% plagioclase feldspar (of composition 30% anorthite, 70% albite).

Very few plutons are composed of a single rock type. There is usually a range of compositions that is sufficiently broad to cover two or more of the rock types in Figure 4.14 – and Figure 4.15 shows an example of this.

The Rogart intrusion (Figure 4.15) displays another feature that is characteristic of many granitic plutons, in that it consists of successive units intruded into one another. The first unit to be emplaced in the Rogart intrusion was the one mapped as tonalite. The granodiorite was then intruded into the centre of this mass, with a steeply dipping contact. This in turn was intruded by the unit mapped as granite. It is common for the compositions of successive intrusions to change in this manner, from generally dioritic or tonalitic (intermediate) at first to generally granitic (felsic) towards the end.

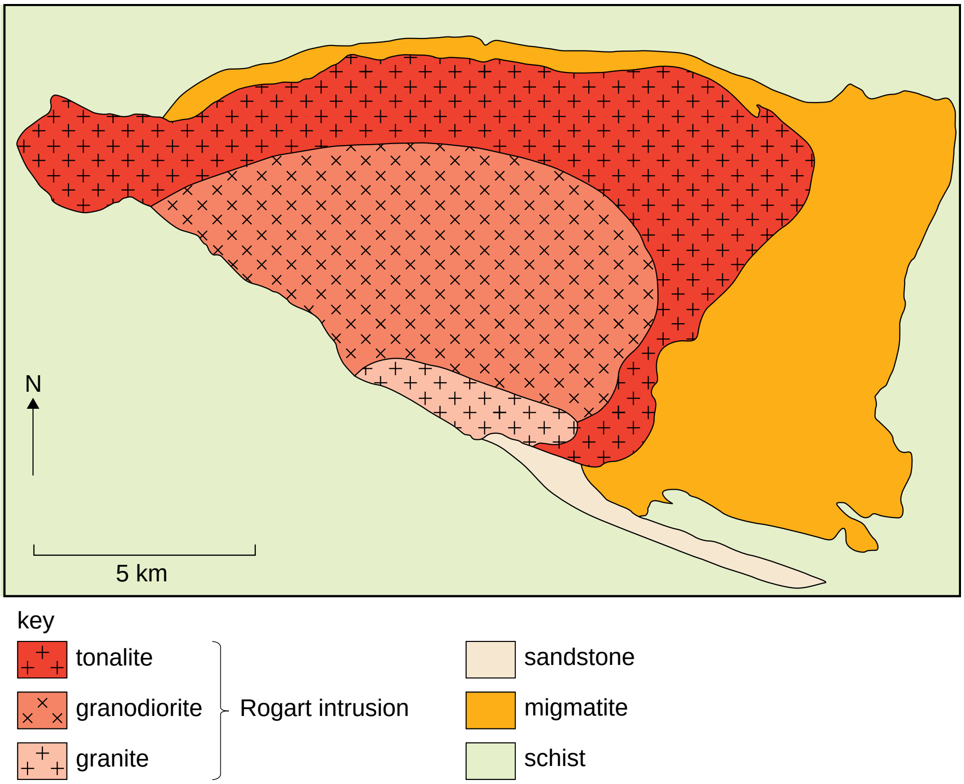


Figure 4.15 A geological map of the Rogart intrusion, Sutherland, Scotland (centred near grid reference NC (29) 6907). Contacts between the different plutonic rock types are gradational over tens to hundreds of metres.

The final stages of crystallisation of a plutonic rock are controlled by the fact that magma can hold a greater percentage of volatiles than can solid rock. As crystallisation proceeds, the remaining melt becomes progressively richer in volatiles. The volatile elements (or any other elements) that are not incorporated into the growing crystals are said to be incompatible elements. The last fraction of a per cent to crystallise may be exceptionally rich in volatiles and other incompatible elements and form veins of coarse-grained pegmatite (Book 1, Figure 6.13a). Alternatively, sudden escape of the volatiles can trigger rapid crystallisation of a fine- to medium-grained sugary-looking mixture of quartz and alkali feldspar referred to as aplite (Book 1, Figure 6.13b).

Our account of plutons has so far concentrated on granites and related intermediate rock types. Mafic magmas also form plutons. In the UK, intrusions of gabbro are found in western Scotland, notably on the islands of Rum, Mull and Skye, and were emplaced about 60 million years ago.

■ Which minerals are common in gabbro?

☐ Plagioclase feldspar, pyroxene and olivine.

In some cases the rock shows a striking horizontal layering, defined by different proportions of minerals. World-class examples of this are found on Rum, where vertical variation in the proportions of these minerals (and in the composition of minerals within their solid-solution series) takes place on a scale of metres to tens of metres (Figure 4.16); it is an example of a **layered intrusion**. The cause of the layering is not always clear, but in some cases crystals have accumulated from a body of cooling basaltic magma, with early-formed crystals sinking to the floor of the magma chamber similar to sedimentary grains settling through water. In other cases, new drafts of magma have entered the chamber and, being denser than the resident magma, have spread over the chamber floor, solidifying into a layer of rock with a different composition (and different mineral proportions) to the original magma.

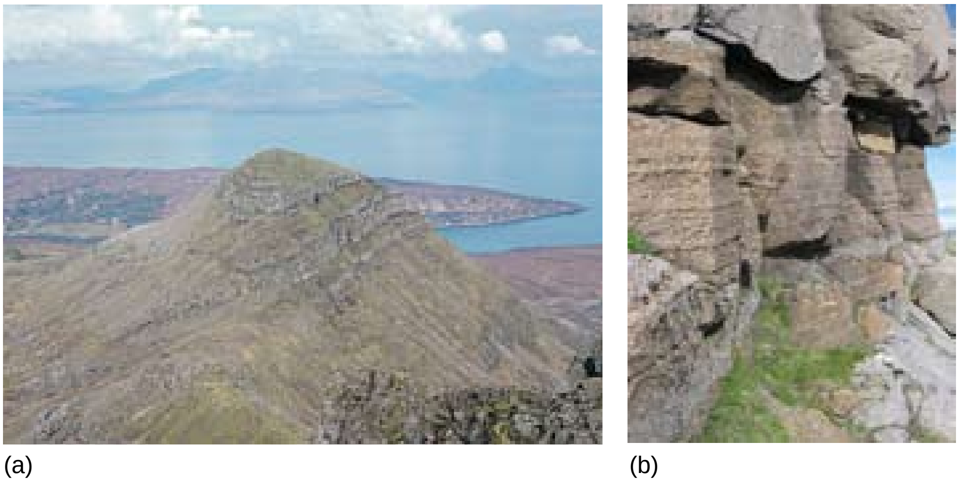


Figure 4.16 (a) View of Hallival, Rum, Scotland showing prominent horizontal layering of mafic and ultramafic rock. The resistant layers are gabbro, whereas the more easily weathered layers are richer in olivine. Note the rocks forming the low-lying area behind the mountain are also layered, and dip to the left (towards the northwest), but these are the sedimentary country rocks into which the gabbro was intruded. Reproduced with the permission of the British Geological Survey © NERC. All rights Reserved. (b) Well-developed mineral layering in peridotite, sandwiched between gabbro layers, Rum, Scotland. Peridotite layer is 2.5 m thick.

4.3 Batholiths

Most geologists would regard the Rogart intrusion as a single pluton, even though it is a concentric complex constructed by the arrival of successive pulses of magma. Likewise, layered intrusions with evidence of several separate batches of magma being intruded over time are still considered to be plutons. Sometimes two or more neighbouring plutons overlap (the younger intruding the older pluton) in a more complicated manner than the concentric pattern exemplified by the Rogart intrusion. Such a combination of plutons is referred to as a **batholith** (others may define a batholith as any pluton or association of plutons exceeding 100 km^2 in surface area). Whereas the geology of the UK contains individual plutons exposed at the surface, some parts of the world have thousands of square kilometres covered almost totally in plutonic rocks. These include the Sierra Nevada mountains of the USA (Box 4.1), and long tracts of Alaska, Peru, Chile and Southeastern Australia. The geological maps of these places are dominated by red – the conventional colour used to depict felsic intrusive rocks. In detail, individual plutons, distinguished on the basis of different mineralogy and with intrusive contacts that cut across structures in older plutons, have built the upper crust in these regions. They record pulses of magmatism on timescales that allowed one plutonic body to have solidified before the next pulse of intruding magma arrived.

Box 4.1 The Sierra Nevada batholith

An extensive part of eastern California is underlain by granite – almost 40 000 square kilometres of it (Figure 4.17a). Its outcrop underpins much of the Sierra Nevada mountain range and the highest peak in the United States outside Alaska, that of Mount Whitney (4420 m above sea level), which is named after the first Director of the California Geological Survey, Josiah Whitney. Much of the area is rugged wilderness with excellent exposures of granite, granodiorite, tonalite and other plutonic rock types making spectacular scenery in the Yosemite, Kings Canyon and Sequoia National Parks (Figure 4.17b). The granite outcrops of the Alabama Hills, outside the town of Lone Pine, have provided backdrops for generations of classic movies from early Lone Ranger westerns to Star Trek.

The igneous rocks were emplaced through Jurassic and Cretaceous times, with successive intrusions intruding the cores of earlier intrusions. The youngest (late Cretaceous) intrusions form large (1000 km^2) multiple intrusions, exemplified by the Tuolumne Suite whose rocks are exposed throughout Yosemite National Park. They show a roughly concentric arrangement, from tonalite or granodiorite margins to granite centres exposed in dramatic three-dimensional relief thanks to the forces of glacial erosion that have carved valleys more than 1000 m deep (Figure 4.17b).

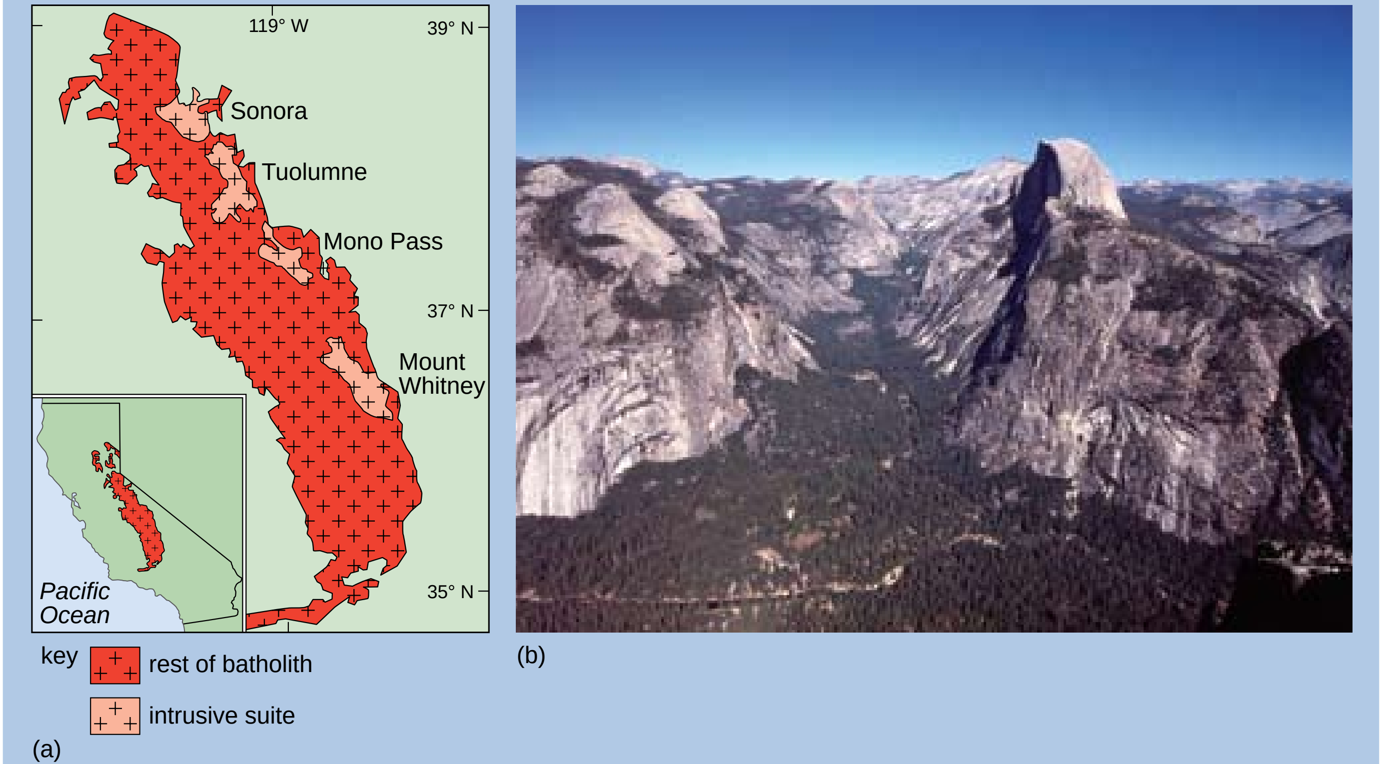


Figure 4.17 (a) Map showing the Sierra Nevada batholith and its location within California (inset). The Sonora, Tuolumne, Mono Pass and Mount Whitney intrusive suites are roughly of the same age (late Cretaceous) and were among the last intrusions in the batholith. (b) Tenaya Canyon, Yosemite National Park, exposes the Tuolumne intrusive suite.

The floors of plutons and batholiths occur deep within the crust, and tend not to be exposed. On some map cross-sections you will see a pluton or batholith displayed as a steep-sided body whose base is conveniently assigned to lie below the greatest depth shown on the cross-section. This reflects the fact that, in most cases, what happens at the base is genuinely not known.

When studying igneous intrusions, and indeed almost any aspect of geology, the rocks exposed depend on the depth to which erosion has penetrated into the crust. For instance, removing a few hundred metres from a volcanic system may expose an array of minor intrusions, whereas erosion of several kilometres may expose a pluton which, with further erosion, might reveal a batholith. When studying plutonic rocks, a geologist is seeing the culmination of long, complex and possibly multiple processes of melting, intrusion and crystallisation. It is also likely that many shallow plutons will have fed volcanic eruptions.

Activity 4.2 Plutons on geological maps

In this activity, you will use geological maps to deduce how and when different plutons were emplaced in Britain's geological past.

4.4 Summary of Chapter 4

1. Dykes fill extensional fractures, whereas the emplacement of sills occurs when magma pressure can lift the overlying rocks and allow the magma to spread along a bedding plane or other plane of weakness.
2. Plutons are intrusive igneous bodies a kilometre or more in size. They tend to be coarse grained, whereas minor intrusions are fine or medium grained and have chilled margins.
3. Plutons that were intruded at great depth in the crust tend to be concordant with the structure of their country rock, whereas shallower plutons tend to be discordant and produce metamorphic aureoles.
4. Diapiric rise of plutons is not usually effective at <10 km depth. Plutons emplaced at shallower levels than this are probably fed either from below by dykes or along fault planes, in which case fault movements create the space occupied by the pluton, with some assistance from updoming and stoping.
5. Plutonic rocks can be classified according to the proportions of the felsic minerals quartz, alkali feldspar and plagioclase feldspar.

4.5 Objectives for Chapter 4

Now you have completed this chapter, you should be able to:

4.1 Recognise minor intrusions, plutons and batholiths on geological maps and cross-sections, giving reasons for your diagnosis and suggesting, if appropriate, how they might be related.

4.2 Suggest reasoned explanations for the relationships between an intrusion and its country rock.

4.3 Describe probable emplacement mechanisms for different kinds of intrusion.

4.4 Classify igneous rocks on the basis of their felsic mineral content using the QAP diagram.

Now try the following questions to test your understanding of Chapter 4.

Question 4.5

On the Bedrock UK Maps, find the following igneous intrusions and identify each one as (i) either a pluton or a minor intrusion, (ii) either concordant or discordant, and (iii) specify the age of each as fully as possible on the evidence of the map: (a) Unit *G* between NS (26) 8718 and NY (35) 3388, (b) Unit *CP* centred near SX (20) 6578.

Question 4.6

While visiting a city centre shopping district, you become distracted by an attractive rock embellishing the facade of a shop. The rock's polished surface clearly reveals the coarse-grained interlocking crystalline texture of a plutonic igneous rock. You estimate that the rock has 10% quartz, 60% feldspar and 30% ferromagnesian minerals. There are equal amounts of alkali feldspar and plagioclase feldspar. What rock name should you give it, according to Figure 4.14?

Question 4.7

Complete Table 4.1, which is an aid to distinguishing sills from lava flows, by inserting ticks to indicate the features you might see in cross-section in the field when studying an ancient volcanic area. Insert crosses to indicate features you would definitely not expect to find.

Table 4.1 For use with Question 4.7.

	Sill	**Lava flow**
chilled margin at top		
chilled margin at bottom		
columnar joints		
concordant top with local discordance		
concordant base with local discordance		
rubbly top		
rubbly bottom		
baking of overlying rock immediately above the contact		
baking of underlying rock immediately below the contact		

Chapter 5 The origins of magmas

Volcanic and intrusive rocks are formed when magma solidifies, but why does the molten magma exist in the first place?

5.1 What causes rocks to melt?

You might think that the answer to this question is obvious, and that magma must be a result of heating. In fact, although sufficient heating of a rock will yield magma, heating is not the only, or even the main, cause of melting in the crust and upper mantle.

Let's think about what happens during melting by applying what is known about the reverse process – crystallisation of liquid magma. When a magma solidifies as a result of cooling, different minerals begin to crystallise at different temperatures. In melting, the converse is the case (irrespective of whether melting is caused by heating or some other factor). In any rock consisting of more than one mineral, melting starts by some of the minerals reacting chemically with each other to produce a liquid. As melting proceeds, with for example increasing temperature, each mineral contributes to the melt at a different rate until they are entirely used up. Minerals disappear sequentially in the reverse of the crystallisation sequence. When conditions are such that complete melting is not attained, there are crystals of at least one mineral still surviving. This behaviour is termed **partial melting**, and is typical of the way magmas are produced in the Earth.

For a rock of a particular composition, there is a temperature at which melting will begin and a higher temperature at which melting will be complete. These two temperatures vary according to pressure, and it is conventional to show this relationship on a graph of pressure (P) against temperature (T), with a line joining all points where melting begins (called the **solidus**) and another line joining all the points where the last crystals dissolve and melting is complete (called the **liquidus**). Such a P–T diagram is plotted for mafic rock in Figure 5.1. This is a phase diagram, with the solidus and the liquidus bounding areas in which the material is in different states. It is based on the results of many laboratory experiments in which powdered samples of the rock are held at different high pressures and temperatures, and then quickly returned to room temperature and pressure and examined to see whether they had melted or still contained crystals.

Figure 5.1 is plotted for anhydrous conditions (i.e. no water present), and this is a reasonable approximation to natural conditions because many basaltic magmas have very low water contents, which is why they erupt in effusive or only mildly explosive styles. Phase diagrams such as this are helpful for interpreting the processes that rocks undergo at high pressure and temperature within the Earth. Question 5.1 tests whether you can read and understand the information given by this diagram.

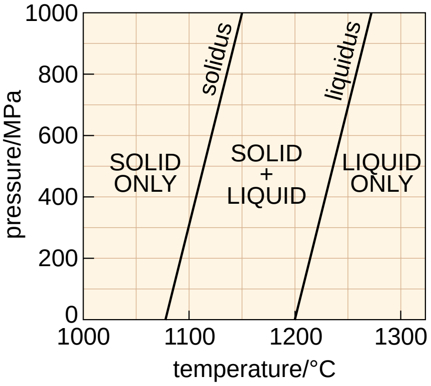


Figure 5.1 Phase diagram showing the solidus and liquidus for a typical anhydrous mafic rock. Pressure is shown in megapascals (1 MPa = 10^6 Pa). (Elsewhere you may find kilobars (kbar) used to denote pressure: 1 kbar (a thousand times atmospheric pressure at sea level) is equal to 100 MPa.)

Question 5.1

(a) Using Figure 5.1, decide whether material of mafic composition consists of solid, liquid, or both, under the following *P*–*T* conditions: (i) atmospheric pressure (0.1 MPa), 1100 °C; (ii) 1000 MPa, 1100 °C; (iii) 400 MPa, 1150 °C; (iv) 200 MPa, 1250 °C.

(b) If an anhydrous piece of gabbro is heated up keeping the pressure constant at 0.1 MPa, at what temperature does it (i) begin to melt, and (ii) become completely molten?

As the temperature increased from 1080 °C to 1200 °C (in the answer to Question 5.1b), the sample would consist of a decreasing proportion of solid and an increasing proportion of liquid. Just to the right of the solidus it would be mostly solid rock containing a tiny amount of melt, and just before it reached the liquidus it would be mostly melt with a few crystals dispersed in it.

Pressure also has an influence on partial melting.

Question 5.2

With reference to Figure 5.1, describe what would happen if we took an anhydrous piece of basalt at 1000 MPa and 1100 °C and gradually decreased the pressure to 0.1 MPa without changing the temperature.

The phase diagram of basalt shows that partial melting can be brought about by decreasing the pressure without changing the temperature. This process is called **decompression melting** and illustrates that a source of heat is not always necessary to cause melting.

It is also important to realise that when a sample is partially molten, the composition of the liquid is different from the average composition of the remaining solid (because some minerals have contributed disproportionately to the melt), and that both are different to the composition of the starting material. However, the average composition of the melt plus remaining solid must always be identical to the composition of the starting material.

■ Mafic rock consists of olivine, pyroxene and plagioclase feldspar. Of these minerals, olivine has the highest melting temperature (at a given pressure). This means that during partial melting of gabbro, some of the olivine would survive as crystals after everything else had been consumed by melting. If the melt were to be squeezed out, leaving these crystals behind, what name would you give to the rock formed by these crystals? (See Figure 4.13.)

☐ A rock consisting of olivine and little else would be ultramafic in composition and would be described as a peridotite.

Partial melting of mafic material could therefore leave a solid residue of ultramafic composition. On the other hand, the very first melt to form would be intermediate in composition because most of its constituents would be from pyroxene and the least calcium-rich plagioclase. Whether or not all the plagioclase and pyroxene had melted would depend both on their compositions and on the final *P*–*T* conditions. Broadly speaking, the closer the final conditions are to the solidus, the richer the melt is in silica, but the smaller the compositional change in the solid residue. Conversely, if the final *P*–*T* conditions lay only just to the left of the liquidus, the amount of solid residue would be smaller and it would have a more ultramafic composition, whereas the melt would be only slightly richer in silica than the starting material.

This leads us to an important generalisation, as follows:

> When conditions are such as to end up with partial rather than total melting, the melt is richer in silica than the starting material. It is common for partial melting of ultramafic starting material to yield a mafic melt, for partial melting of mafic starting material to yield an intermediate melt, and for partial melting of intermediate starting material to yield a felsic melt.

In addition to pressure and temperature, there is a third factor controlling melting to consider. Figure 5.1 is plotted for anhydrous conditions. This means that there is no water dissolved in the magma, the solid contains no hydrous minerals, and water does not permeate the rock. As soon as water is added to the system, the whole situation changes. Water has the effect of lowering the melting temperatures, so that both the solidus and liquidus are displaced to the left on a *P*–*T* phase diagram. Figure 5.2 shows a solidus and liquidus plotted for water-saturated conditions, which are defined as those in which the liquid phase contains as much dissolved water as it can hold. This depends on the pressure, with high pressures allowing more water to be dissolved in the liquid than at low pressures. The amount of water that can be dissolved varies from about 0.1 per cent (by mass) at atmospheric pressure, to many per cent at several hundred MPa. If there is more water present than the magma can hold at a given pressure, the excess water will form a separate phase (usually in the form of vapour bubbles).

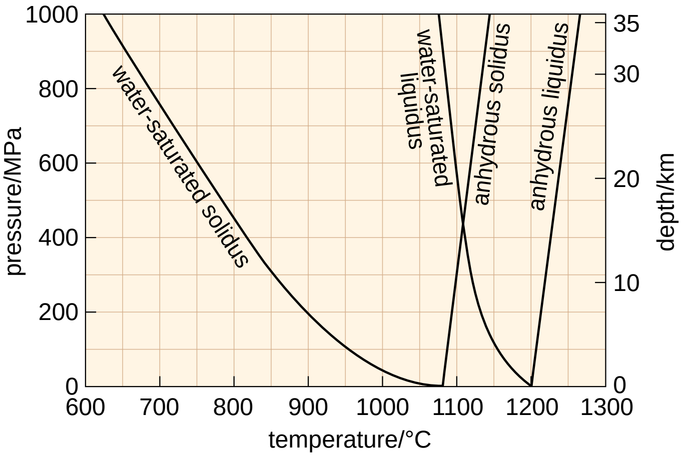


Figure 5.2 Phase diagram for mafic rock showing the solidus and liquidus for water-saturated conditions, in addition to the anhydrous lines shown in Figure 5.1. In hydrous conditions but with insufficient water for saturation, the phase boundaries would lie between the anhydrous and water-saturated extremes. The vertical scale on the right indicates the approximate depth in continental crust corresponding to the pressures on the opposite scale.

On Figure 5.2, the water-saturated phase boundaries occur at lower temperatures than the anhydrous phase boundaries, except at vanishingly small pressure. This is because if there is no confining pressure on the magma, hardly any water can dissolve in the melt, so it is virtually anhydrous.

To explore an important consequence of hydration, you should now try Question 5.3.

Question 5.3

According to Figure 5.2, what would happen if you took anhydrous mafic rock at 1000 MPa and 1100 °C and (without changing *P* or *T*) added sufficient water to make the conditions water-saturated?

Thus, melting can occur simply as a result of the addition of water (or indeed any other volatile). Water saturation has an even more extreme effect on the melting behaviour of felsic rock, as shown in Figure 5.3. This draws attention to the fact that the chemical composition of the material influences the melting temperature and pressure: the greater the silica content of a magma, the lower the melting temperature.

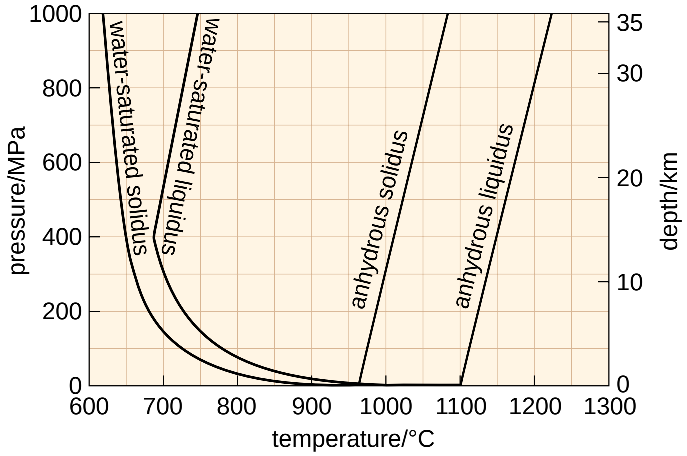


Figure 5.3 Anhydrous and water-saturated phase boundaries for felsic rock. Compare with Figure 5.2. Note that in both cases the water-saturated phase boundaries meet the equivalent anhydrous phase boundaries at zero pressure.

5.2 Crystallisation of magma

Crystallisation can be thought of as the reverse of melting, so the phase diagrams in Figures 5.1 to 5.3 can be used to show the conditions that cause magmas to crystallise.

■ The processes of heating, decompression and addition of water can cause rocks to undergo partial melting, so how likely is it that (i) cooling, (ii) increase in pressure, or (iii) removal of water will cause a magma to crystallise?

☐ The phase diagrams indicate that all three of these processes can shift a magmatic liquid from a position on (or even above) its liquidus to a position below the liquidus, and therefore into a stability field where some crystals are present.

You have already seen in Chapter 4 that cooling at the edge of an igneous intrusion causes rapid crystallisation and the formation of small crystals, whereas farther from the contact cooling is slower so fewer but larger crystals grow producing a coarser-grained rock. Although both the fine-grained and coarse-grained rocks will have started at the same temperature and ended at the same temperature, the rate at which the temperature changed will have been different. A fast cooling rate yields finer-grained igneous rock than slower cooling. The reasons for this are best understood after remembering that two separate processes occur. Starting with a liquid, the first step in producing crystals involves the gathering together of a sufficient number of the right atoms to make a tiny speck of a crystal; this is the process of nucleation and the initial speck is called a crystal nucleus (plural nuclei). Crystal nuclei are only a few nanometres in size. The second process is growth of the crystals by the addition of layers of atoms to the nucleus.

For crystals to nucleate or grow, the liquid must be at some temperature below the liquidus temperature. There must also be sufficient time spent at these lower temperatures for nuclei and crystals to achieve a reasonable size, otherwise the liquid will cool so quickly that it quenches to a glass without crystallising. In this case, simply following the temperature path across the stability fields on a phase diagram does not predict the correct outcome. This is because the phase diagram is drawn from the results of experiments that have been carried out over long enough times for the system to reach equilibrium. Rapid cooling can cause disequilibrium, with different outcomes.

With this background, we can consider the texture of an igneous rock in terms of the time spent below the liquidus and the amount of cooling beneath the liquidus. The difference between the liquidus temperature and the actual temperature is called the **undercooling**, often denoted by the symbol ΔT (pronounced ‘delta T’). The rates at which nuclei form (measured in terms of the number of nuclei forming in a cubic millimetre per second) and at which crystals grow (lengthen) depend on ΔT. You might expect that these rates increase with increasing ΔT, but this is only partly true. At large undercoolings, the atoms in the liquid move very slowly within the melt structure, for similar reasons as to why viscosity increases with falling temperature. Experiments have shown that at large undercoolings, nucleation and growth rates are both small. Graphs in which the nucleation rate and growth rate are plotted against ΔT illustrate these features (Figure 5.4). Particularly important features are that the peak growth rate occurs at lower undercooling than the peak nucleation rate, and that at large undercoolings, both the nucleation and growth rates are very small.

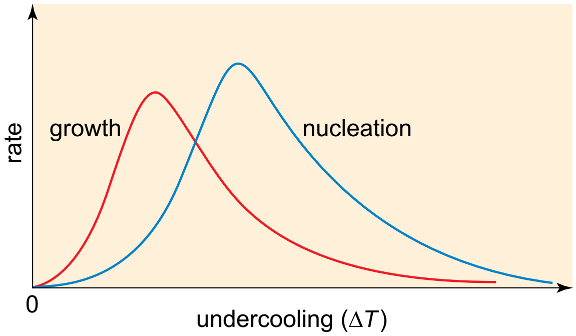


Figure 5.4 Schematic graphs of nucleation rate and crystal growth rate as functions of undercooling. There are no scales on the axes, partly because crystal nucleation and growth rates are measured in different units so the relative heights of the two curves are not significant, but note that the peak growth rate occurs at lower ΔT than the peak nucleation rate.

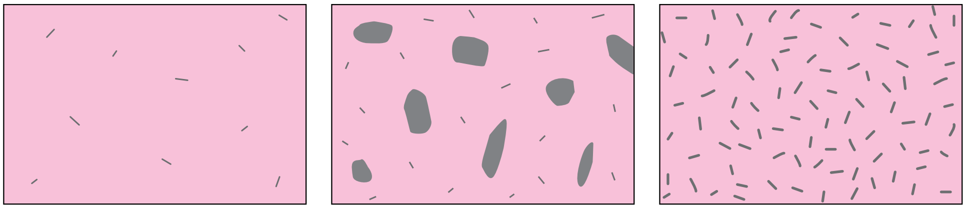
Based on Figure 5.4, it is apparent that magmas may crystallise under three regimes:

- very low undercooling, where nucleation rate is low but growth rate is high
- moderate undercooling, where nucleation rate is high but growth rate is low
- high undercooling, where both nucleation and growth rates are very low.

These three regimes lead to different textures.

Question 5.4

Look at the three sketches of igneous rock textures in Figure 5.5 and decide which one was produced under each of the nucleation and growth rate conditions listed in Table 5.1.



(a) (b) (c)

Figure 5.5 Sketches of three examples of igneous rock textures showing crystals (grey) embedded in glass (pink). For use with Question 5.4.

Table 5.1 For use with Question 5.4.

Nucleation rate and growth rate	Undercooling	Example in Figure 5.5
nucleation rate low, growth rate high	low	
nucleation rate high, growth rate low	moderate	
nucleation rate very low, growth rate very low	high	

We can now consider the circumstances that determine the size of undercooling a magma experiences at the time of crystallisation, and the cause of the undercooling. The simplest case is, of course, where hot magma is intruded or erupted into a colder environment, and cooling proceeds from the outside of the magma inwards. This is the situation that applies to an intruded sill (Figure 4.6) or a pillow lava. The greatest undercooling is imposed at the contact so here the finest-grained rock is formed. At the edge of a pillow lava, where basalt at a typical temperature of 1150 °C is brought into contact with cold water, the undercooling can be so large that no nuclei form and the liquid is quenched to a glass in the outermost 1 or 2 cm of the pillow. In the interior, the temperature changes slowly and the magma spends a longer time at temperatures not far below the liquidus (low undercooling), allowing phenocrysts to grow.

But cooling is not the only way to achieve sub-liquidus conditions and cause crystallisation within a magma. An alternative process to cooling is to remove water from a magma by decompressing it.

■ Why can decompression of a hydrous magma lead to undercooling (assuming that the temperature does not change during this process)?

☐ Once the pressure falls to a small enough value, the magma becomes saturated in water and bubbles will form with further decompression, removing water from the liquid. At a given pressure, the liquidus temperature of water-poor magma is higher than that of water-rich magma (Figures 5.2 and 5.3), so the magma will become undercooled once it starts to exsolve water.

The eruption of magma has the potential to induce crystallisation by virtue of it losing water, but how effective this will be depends on the amount of time available. A rapidly rising (rapidly decompressing) magma may not have sufficient time for nuclei or crystals to grow, in which case it will erupt as a supercooled liquid. Alternatively, more slowly erupting magma may grow large amounts of tiny crystals. Interestingly, the growth of many crystals causes the magma to increase its viscosity. This in turn causes the magma to flow more slowly driving yet more crystallisation, which is an example of a feedback process that might result in a change in the character of the eruption from explosive to effusive, or even cause it to stop if the magma became so crystal-rich as to become immobile.

5.3 Chemical composition of magma

Igneous rocks and magmas come in a wide range of compositions, from silica-poor ultramafic rocks to silica-rich felsic rocks. How is this range explained? Partial melting yields liquids which are different in chemical composition from the rock that was melted. This means that the composition of the starting material and the amount of partial melting will both determine the composition of the resulting liquid magma, known as a **primary magma**. Once a primary magma has been formed by partial melting it will normally migrate upwards, intruding into colder rocks and start to cool down, which has the effect of inducing crystallisation. Crystallisation also has the effect of changing the chemical composition of the liquid part of a magma because the early-formed crystals have a different chemical composition from the primary magma. In the case of a magma that is slowly cooling in a subterranean magma chamber, phenocrysts will grow in the magma. If these phenocrysts do not move while further cooling and crystallisation of the magma occurs, then by the time the magma has completely solidified it will still have the same composition as the original magma.

■ How will the composition be affected if the first-formed crystals become separated from the melt?

□ The remaining melt will be more felsic than the original melt, whereas the separated crystals will form a rock less felsic (more mafic) than the original melt.

Such physical separation of the crystal fraction from the melt fraction in a partly crystallised magma is a very important way in which the composition of a magma can change. The process is termed **fractional crystallisation**. It can occur by the residual liquid being squeezed out of the space between the crystals, or by crystal settling if magma is stored for a long period in a magma chamber. The result of fractional crystallisation is a liquid whose chemical composition is more felsic than the original melt, and a body of separated crystals called a **cumulate**. So some igneous rocks are cumulates, and include certain types of layers in layered intrusions (e.g. Figure 4.16b).

■ What types of minerals would you expect to form cumulate rocks?

□ Cumulates form from early crystallised minerals, which are those that crystallise at high temperature. These are mafic ferromagnesian minerals, such as olivine and pyroxene, and calcium-rich plagioclase feldspar.

Another way in which magma composition can change is if the magma has enough heat to melt some of the rock it passes through. If this new melt is able to mix into the main magma body, then this will develop a new composition that is a weighted average of its initial composition and that of the new material that has been mixed in. This process is described as **assimilation**.

■ Do you think it likely that a rising body of felsic magma would assimilate mafic rocks through which it was rising? (Refer to Figures 5.2 and 5.3.)

☐ Mafic rocks begin to melt at a much higher temperature than felsic rocks, so a felsic magma would not normally be hot enough to melt a mafic rock.

The converse is more common, in that a rising body of mafic magma can assimilate rocks of intermediate and felsic composition, and a rising body of intermediate magma can assimilate rocks of felsic composition. In the process, the magma becomes more felsic than its initial composition. Both fractional crystallisation and assimilation therefore cause the silica content of a magma to increase, and the magma is said to evolve in composition. In relative terms, felsic magmas are said to be more evolved than intermediate magmas, which are more evolved than mafic magmas.

5.4 Magmatism and plate tectonics

We conclude our discussion of igneous processes with a look at the relationships between the origin of magmas, magma evolution and the broader workings of the Earth’s interior, particularly in relation to plate tectonics.

5.4.1 Magmatism at divergent plate boundaries

Mafic magma, which abounds at divergent (constructive) plate boundaries, is what you would expect to find as a consequence of partial melting of the ultramafic mantle, and the mantle is the only potential melt source in such a setting. It is also natural that magma should escape upwards, because mafic magma is less dense than ultramafic rock. However, why the mantle should partially melt is not quite so obvious.

At mid-ocean ridges, two oceanic plates diverge and asthenospheric mantle is drawn upwards from beneath the plate boundary at a rate sufficient to plug the gap (Figure 2.8).

■ What will happen to the pressure experienced by this asthenospheric mantle as it is drawn upwards beneath the divergent plate boundary?

☐ The pressure must drop as the depth decreases.

To see the implications of this, look at the *P*–*T* phase diagram for material of ultramafic composition (Figure 5.6). The mantle beneath mid-ocean ridges is typical of normal upper mantle and has a very low water content, so it is the anhydrous solidus that is relevant in controlling partial melting in this setting. The diagram shows a likely path followed as asthenosphere is drawn upwards beneath a divergent plate boundary.

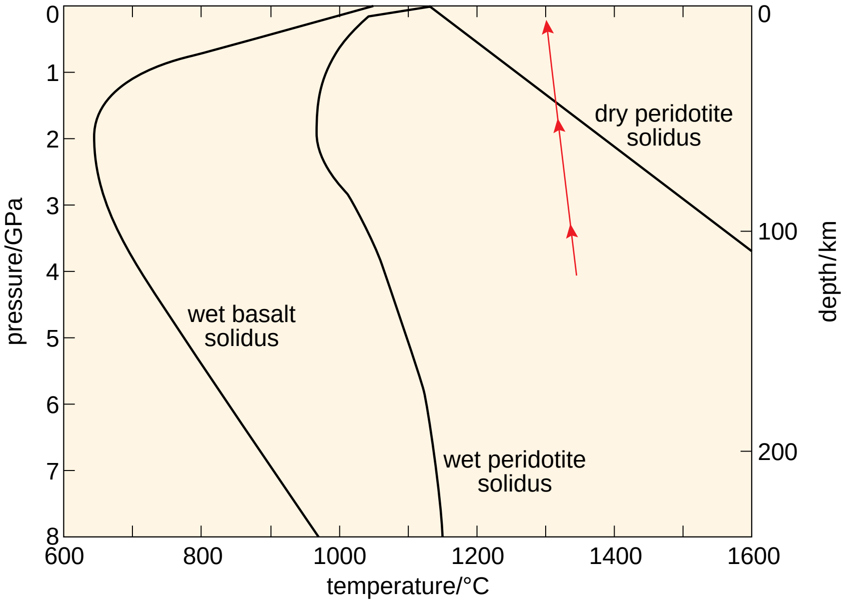


Figure 5.6 The path of changing pressure and temperature (arrowed red line) followed by mantle upwelling beneath a divergent plate boundary, and the locations of the solidus of dry (anhydrous) peridotite, wet peridotite and wet basalt. Note that this diagram has been drawn with pressure increasing down the vertical axis, and with pressure shown in gigapascals (1 GPa = 1000 MPa).

■ According to Figure 5.6, what will happen to upwelling anhydrous mantle when it reaches a depth of about 45 km?

☐ This is the depth at which it crosses the anhydrous (dry) solidus, so it will begin to melt.

The exact depth at which partial melting begins below a particular divergent plate boundary depends on factors such as the starting temperature, and the speed and cooling rate of the upwelling asthenosphere. However, the important point is that although upwelling mantle cools slightly as it rises, the drop in pressure is such that it can rarely avoid crossing the solidus into the partially molten field. Mafic magma at divergent plate boundaries is therefore generated by partial melting of the ultramafic mantle entirely as a result of decompression melting, a process you met in Question 5.2.

Sea-floor spreading induces partial melting of the underlying mantle, releasing basaltic magma that rises to the surface. ‘Fieldwork’ at mid-ocean ridges involves submersible dives and has found basaltic lava flows that are too young to have become covered in the slowly accumulating layer of deep-sea sediment that blankets the ocean floor at greater distances from the ridge. Pillow lavas are particularly common (Figure 5.7). To find out about deeper portions of the oceanic crust geologists must rely on seismic studies, a limited amount of drilling, and sampling (by submersibles and dredging) at places where deeper levels are exposed (e.g. at transform faults and fracture zones).

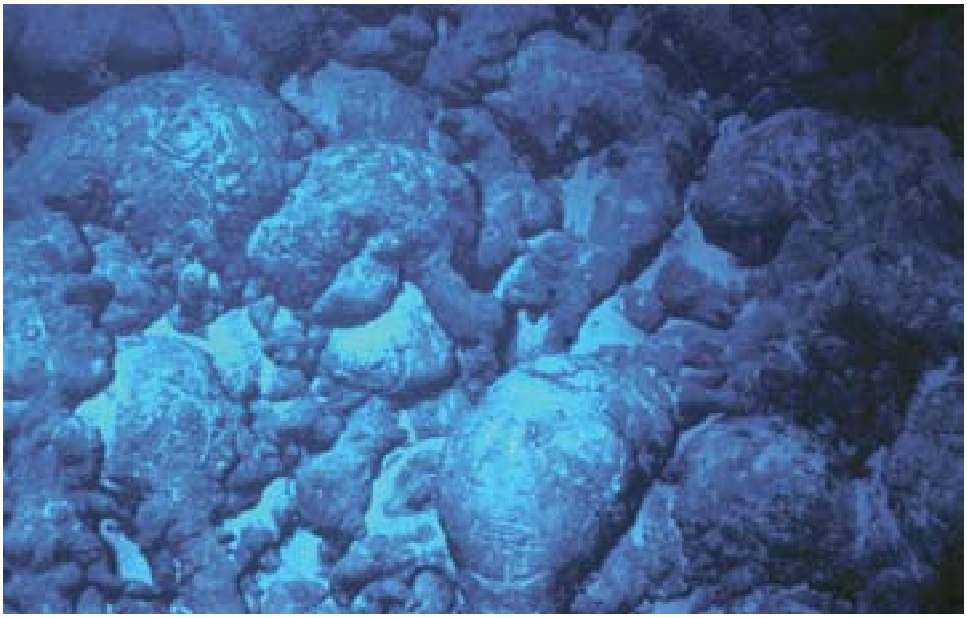


Figure 5.7 Pillow lavas on the Mid-Atlantic Ridge, 3150 m below sea level. Each large 'pillow' is about 0.5 to 1 m in diameter.

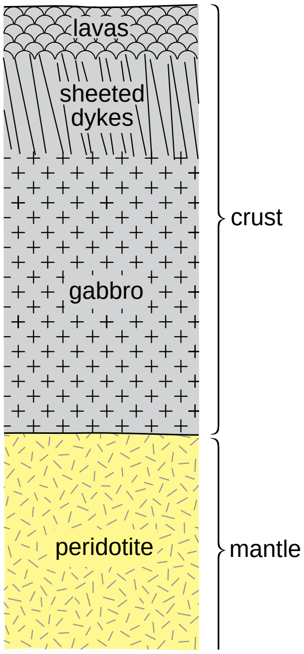


Figure 5.8 The three-fold division of igneous oceanic crust into basalt lavas, dolerite dykes and gabbro.

The picture that emerges from combining all these lines of study is that lava flows make up roughly the top half kilometre of the oceanic crust, which is itself about 7 km thick on average. The lowest part of the oceanic crust is a coarse-grained igneous rock containing predominantly pyroxene and plagioclase feldspar.

■ What rock name would you give to this rock type?

□ According to Figure 4.13 this is gabbro, the coarse-grained (i.e. plutonic) equivalent of basalt.

The gabbro layer is believed to form by slow solidification of intrusions of basalt magma beneath a mid-ocean ridge (hence its large grain size). Fractional crystallisation of olivine, plagioclase and pyroxene during slow cooling of the gabbro magma chamber can produce relatively evolved basalt magma that may spill out onto the ocean floor and erupt as pillow lava. Lying above the gabbro and below the lavas is a layer about 1 km thick consisting of nothing but parallel dykes of medium-grained dolerite. Each dyke has been intruded up the middle of a previous dyke or up the contact between two previous dykes. This layer is referred to as a **sheeted dyke complex**. The three-layer structure of oceanic crust, summarised in Figure 5.8, can be well displayed where segments of old oceanic crust and upper mantle tens or even hundreds of kilometres across have been thrust up onto continental regions, as found for example in Oman. Such slabs of ocean floor are known as an **ophiolite** (pronounced ‘oaf-ee-oh-lite’) or an ophiolite complex (Box 5.1).

Igneous processes at divergent plate boundaries drive other processes where seawater is drawn down into the hot young crust, becoming heated and then rising back towards the sea floor in a hydrothermal circulation system. The hot water reacts with some of the minerals in the crust and changes their chemistry (in particular, olivine becomes altered to serpentine and much of the pyroxene is replaced by the hydrous mineral amphibole) and veins of metal-rich minerals may be deposited.

Box 5.1 The Troodos Mountains, Cyprus

The geology of Cyprus, in the eastern Mediterranean Sea, is largely a combination of late Mesozoic and Cenozoic sedimentary rocks and slightly older igneous rocks. It is the igneous rocks that form the main mountain range – the Troodos Mountains – and they have been dated at 90–92 Ma. The igneous rock types are basaltic pillow lavas, parallel dolerite dykes packed so tightly that entire tracts of land are composed of metre-scale dykes, gabbro, and coarse-grained olivine–pyroxene rocks. The mineral content of the latter rock type defines them as varieties of the ultramafic rock type peridotite. Some of the peridotites have been heavily altered by reaction with hot water, turning olivine into serpentine; such rocks are known as serpentinite – an allusion to the impression that the rocks look like snake skin. The simplified geological map in Figure 5.9a shows that these are arranged in an almost concentric fashion over an area about 90 km long and 40 km wide. Pillow lavas are exposed almost all the way around the fringe of the outcrop. From the geological map, it is apparent that the pillow lavas are overlain by younger sediments. Copper sulfide mineralisation is widespread within the pillow lavas and has been exploited for more than 4000 years; it is no surprise that Cyprus derives from the Greek word for copper (*cupros*). Beneath the pillow lavas lie the dyke rocks (forming a sheeted dyke complex because this rock unit is composed of sheet-like intrusions; Figure 5.9b) and these pass into gabbro, which is in turn in contact with the peridotites and serpentinites.

The origin of the igneous rocks is evident from their geology: sea-floor eruption of pillow lava, shallow magma transport in dykes, plutonic crystallisation of gabbro, and peridotite typical of upper mantle rocks. Viewed as a whole they represent a cross-section through mafic igneous crust, from the mantle roots up through to the intrusive magma storage regions and transport routes that once generated new oceanic crust at a mid-ocean ridge.

The Troodos ophiolite was the first such example to be interpreted in terms of plate tectonic processes once a major mapping project carried out by Greek and British geologists during the 1960s led to the realisation that the Troodos rocks matched the rock types of the oceanic crust, despite now being well and truly part of continental lithosphere. A key scientist in these discoveries was Professor Ian Gass, the first professor of Earth Sciences at The Open University.

The outcrop pattern on Figure 5.9a and the inclined dip of the sheeted dykes (Figure 5.9b) show that the rocks are no longer in their original horizontal orientation, and this is because of deformation that occurred while the peridotites were being altered to serpentinite. This chemical reaction introduces water into the mineral structure, causing a large expansion of volume (and decrease in density) which causes the serpentinite to swell upwards, shouldering aside overlying rock and intruding into it. The Troodos ophiolite is an example of oceanic crust that has become sufficiently buoyant to avoid subduction.

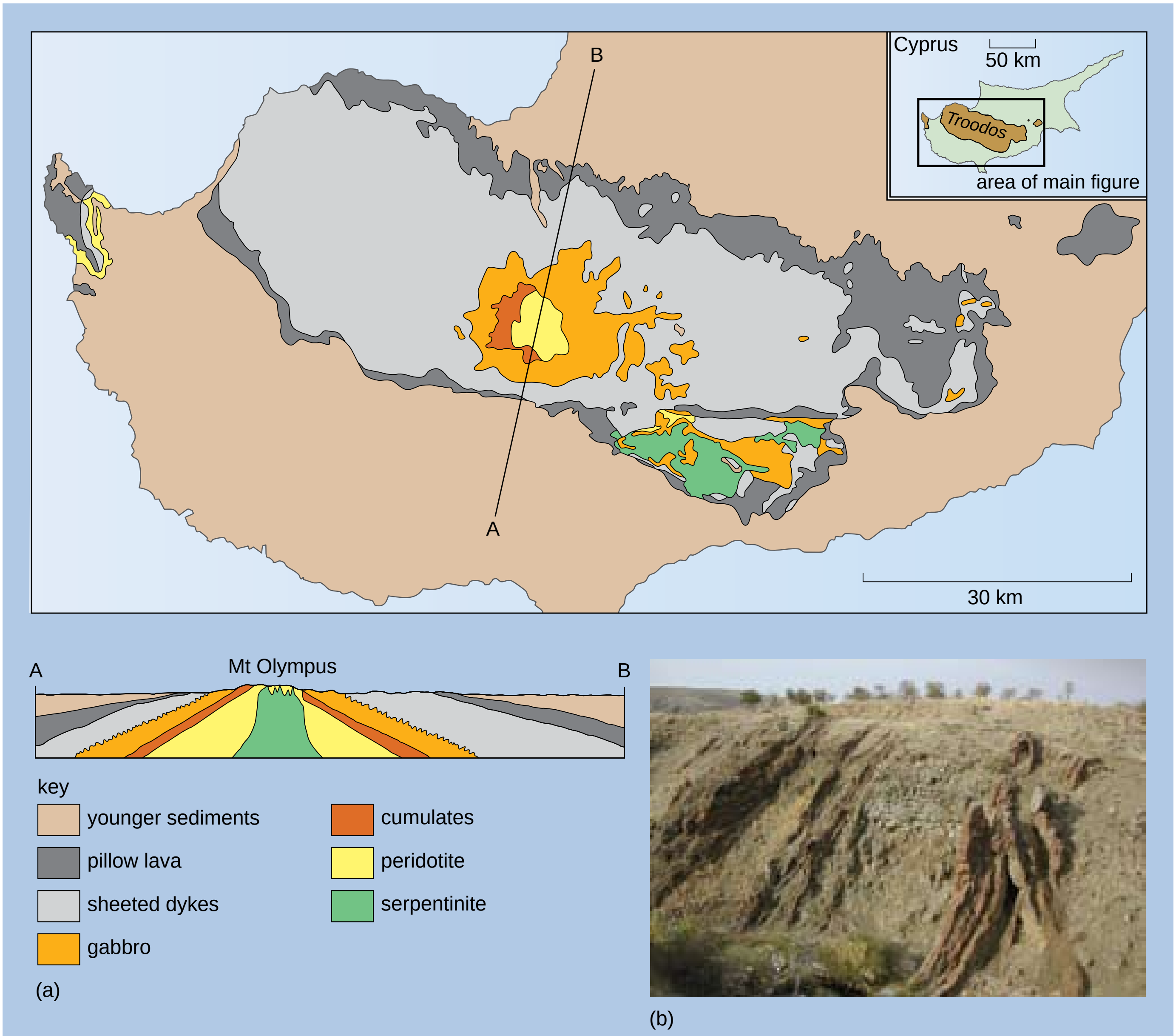


Figure 5.9 (a) Simplified geological map of the Troodos Mountains, with insets showing their location on Cyprus and a north–south cross-section. (b) Sheeted dykes intruding pillow lavas, Klirou, Cyprus. Dykes are 1–2 m wide.

5.4.2 Magmatism at hot spots

Basaltic volcanism abounds on the Big Island of Hawaii, in the middle of the Pacific Ocean, thousands of kilometres from any plate boundary. Indeed, the whole island is composed of basalt. This is a **hot spot**, where large volumes of basalt erupt in isolation from plate boundaries. Volcanism here, and at other **intraplate volcanoes** (or **within-plate volcanoes**) is best explained by decompression melting of plumes of unusually hot mantle rising buoyantly from great depth, possibly from as deep as the core–mantle boundary. On approaching the lithosphere, the material in the plume is at a sufficiently low pressure (and high temperature) to partially melt, releasing basaltic magma that erupts through the overlying lithosphere. In most cases the lithosphere is itself moving, but in a horizontal direction as part of plate tectonics. As the lithosphere moves over the hot spot, a chain of extinct volcanoes is left behind on the surface. Active volcanoes lie above the hot spot but progressively older extinct volcanoes lie at increasing distances away, as classically illustrated in the 6000 km-long Hawaii–Emperor chain of islands and seamounts (sub-sea volcanoes) (Figure 5.10).

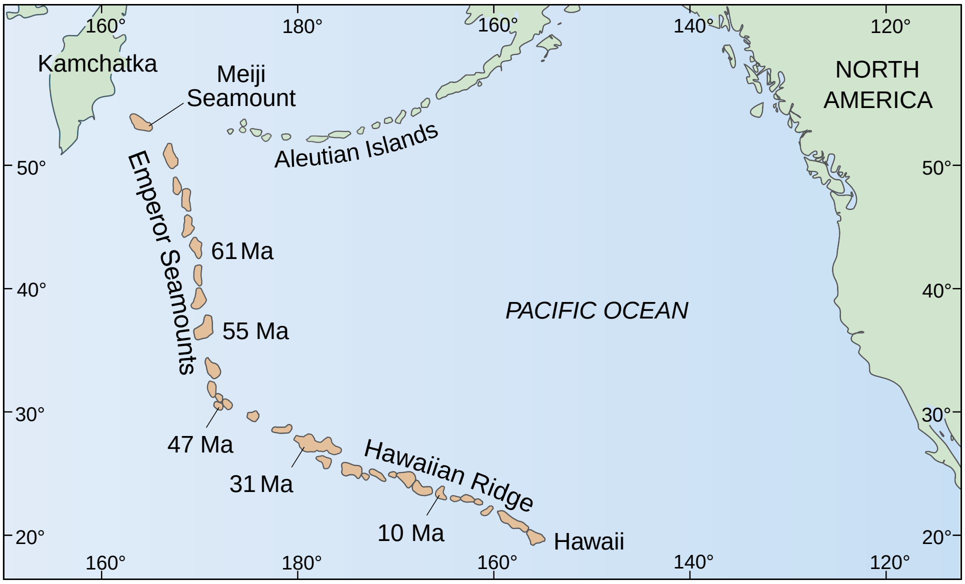


Figure 5.10 The islands and seamounts of the Hawaii–Emperor chain (brown), showing an age progression (in millions of years) away from the active hot spot beneath Hawaii.

If mantle plumes are initiated deep in the mantle, then they can potentially appear at the surface under any region of the Earth, such as in the middle of an oceanic plate in the case of Hawaii. Unusually intense volcanism in Iceland has built this part of the Mid-Atlantic Ridge up above sea level and requires unusually productive partial melting, which is attributed to a hot spot coinciding with a divergent plate boundary. Other examples of hot spots occur within continental lithosphere, such as in the southern Sahara, where about two dozen volcanic centres, some with summit calderas, make up the Tibesti Mountains, rising to 3400 m. There is no obvious linear chain of extinct volcanoes here because the African Plate is hardly moving relative to the underlying mantle plume. Similarly, the dominantly basaltic Cape Verde and Canary Islands (Box 5.2) overlie hot spots on the oceanic portion of the African Plate. A continental hot spot track is, however, found on the faster-moving North American Plate (Figure 5.11), where a series of large rhyolitic ignimbrite calderas are interpreted as resulting from assimilation and melting of continental crust as it passes over a hot spot. This hot spot currently underlies the Yellowstone caldera. The older calderas lie to the southwest and become younger towards Yellowstone, indicating that the North American Plate is moving southwest relative to the hot spot.

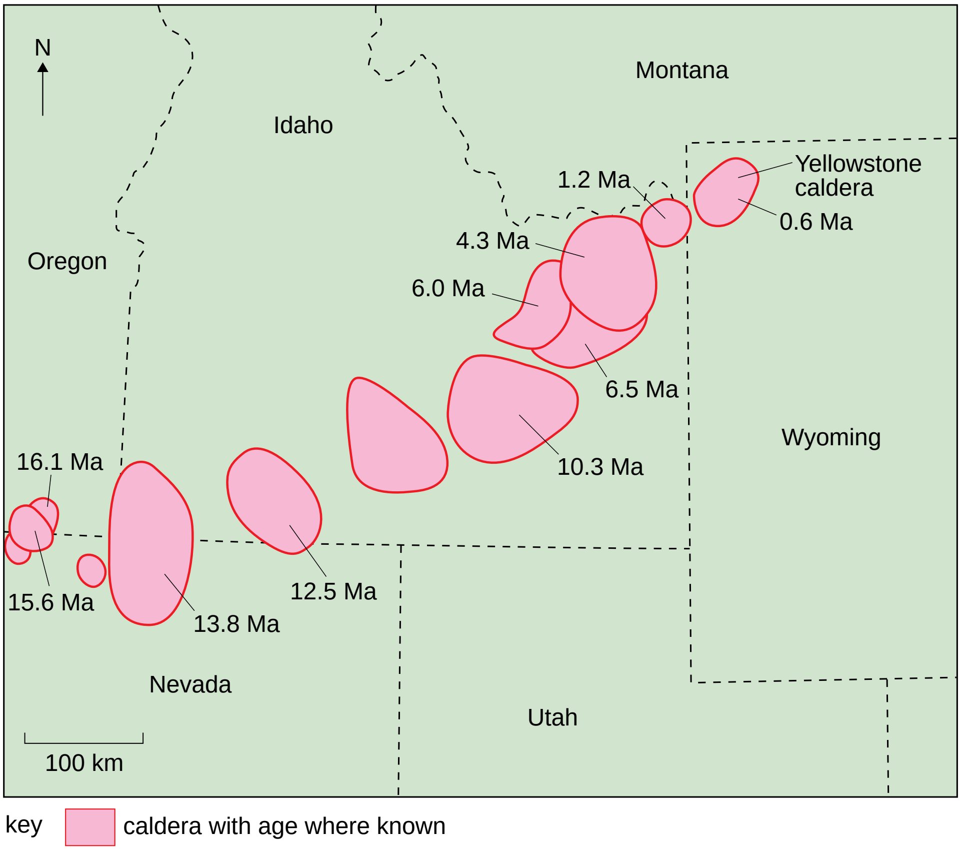


Figure 5.11 Map of the northwestern USA, showing the locations and ages of rhyolite calderas that formed as the North American Plate moved over the Yellowstone hot spot.

Box 5.2 Volcanoes of Tenerife, Canary Islands

Tenerife, the largest of the Canary Islands, is the third largest volcanic ocean island on Earth by volume (Figure 5.12a). It is dominated by El Teide (or Mount Teide), a World Heritage Site, which forms part of the Parque Naçional del Teide. At 3718 m above sea level, and approximately 7500 m above the floor of the Atlantic Ocean, Teide is the highest mountain in Spain, and the highest point in the Atlantic Ocean. The active, but dormant stratovolcanoes of Teide and its sister peak Pico Viejo (‘Old Peak’, even though it’s actually younger!) form the most recent central volcanic complex on Tenerife.

Teide and Pico Viejo are located in a 16 km × 19 km depression known as the Las Cañadas caldera that was formed about 200 000 years ago, although the mechanism (catastrophic eruption or collapse) is still a matter of contention among researchers. The southern part of the caldera wall provides a cross-section through the older volcano (Figure 5.12b). Much younger eruptions of felsic magmas produced pyroclastic flows and fall deposits, including those known as Bandas del Sur, that are well exposed along the south coast of the island.

Eruptions from the northwest and northeast rift zones are thought to be fed from dykes extending laterally from beneath the summit. Eruptions on Tenerife in historical times include lava flows from near the summit of Pico Viejo in 1798 and the 1909 eruption of a small strombolian mafic cone with associated lava flows and black scoria fall from vents on the northwest rift zone (Montaña Chinyero). Teide currently has active fumaroles at the summit, which release hydrogen sulfide and other gases.

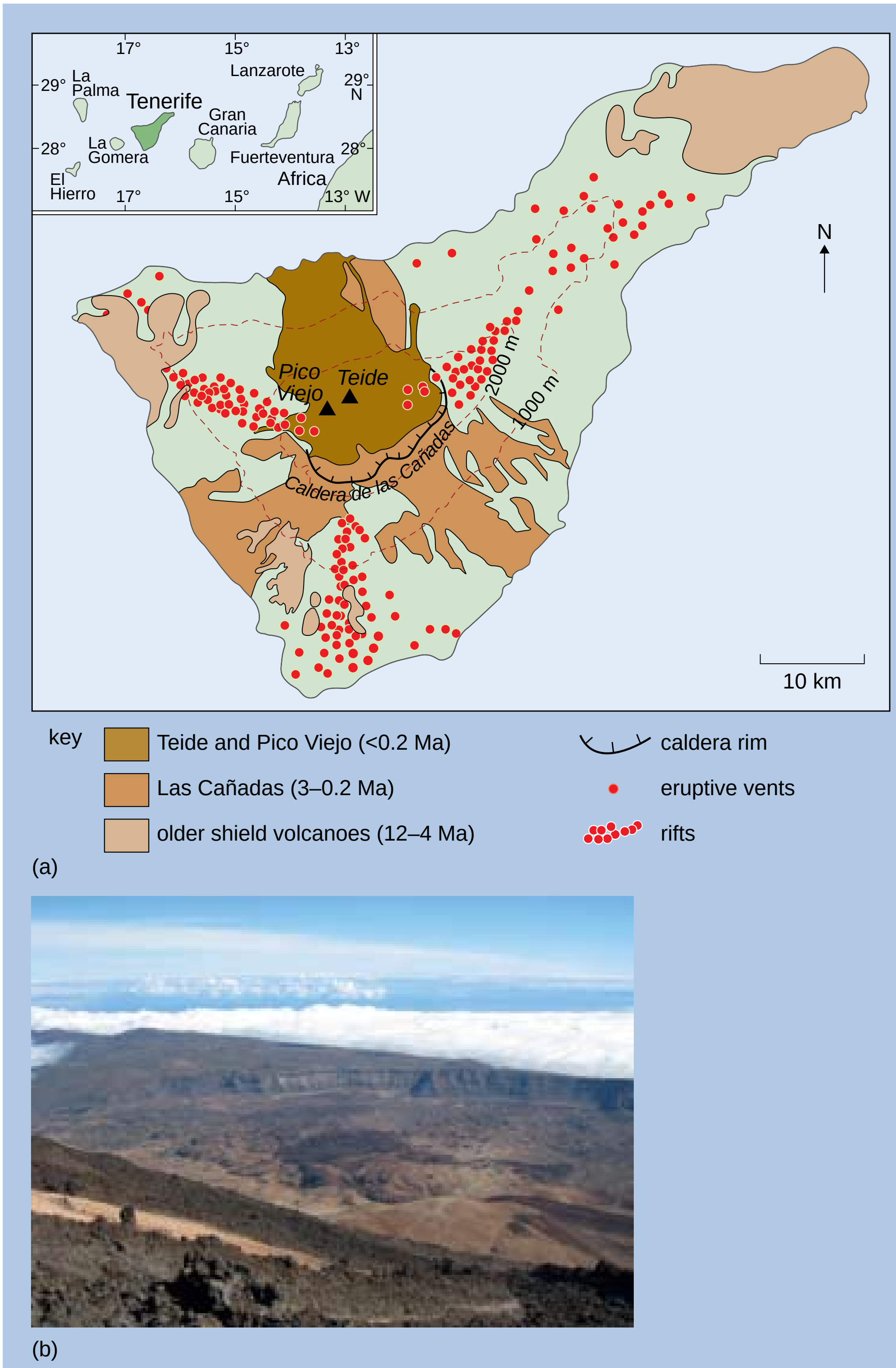


Figure 5.12 (a) Map of Tenerife, showing the wall of the Caldera de Las Cañadas, the outcrop of shield and stratovolcanoes, and the distribution of vents along the northeastern, northwestern and southern rift zones and their associated lavas (green). (b) Photograph from the flanks of Teide, showing a’a lava flows in the foreground and the floor and southeastern part of the Caldera de Las Cañadas. Note the layers of lava exposed within the caldera wall. Lava and pyroclastic deposits from the 2000-year-old Montaña Blanca eruption occur on the caldera floor.

5.4.3 Flood basalts

At around the time that the oldest rhyolitic caldera associated with the Yellowstone hot spot was forming (about 16 million years ago), a huge burst of basaltic volcanism occurred to the north, in the states of Oregon and Washington. The most recent investigations and mapping indicate that some 230 000 km^3 of basaltic lava inundated 200 000 km^2 in a series of enormous eruptions fed from a dyke swarm in the space of about half a million years. It built a volcanic pile about 1 km thick, known as the Columbia River Basalt Group (Figure 5.13a). It is the youngest example on Earth of a **flood basalt province** – characteristically large by the scale of present-day volcanic systems and typically involving about one million cubic kilometres of rather monotonous basalt lavas emplaced in a timespan of about one million years or less (a long-term output rate ten times that of Hawaii). Many flood basalt provinces are to be found on the continents, but they can also be erupted onto the ocean floor. This is how the Ontong–Java Plateau in the western Pacific (about 120 million years ago) was formed. The Columbia River Basalt Group is actually on the 'small' side in comparison with the Deccan Traps basalts that cover a large part of western India (Figure 5.13b). It took only half a million years, about 65 million years ago, for more than one million km^3 of basalt magma to be erupted while India and the Seychelles were starting to split apart.

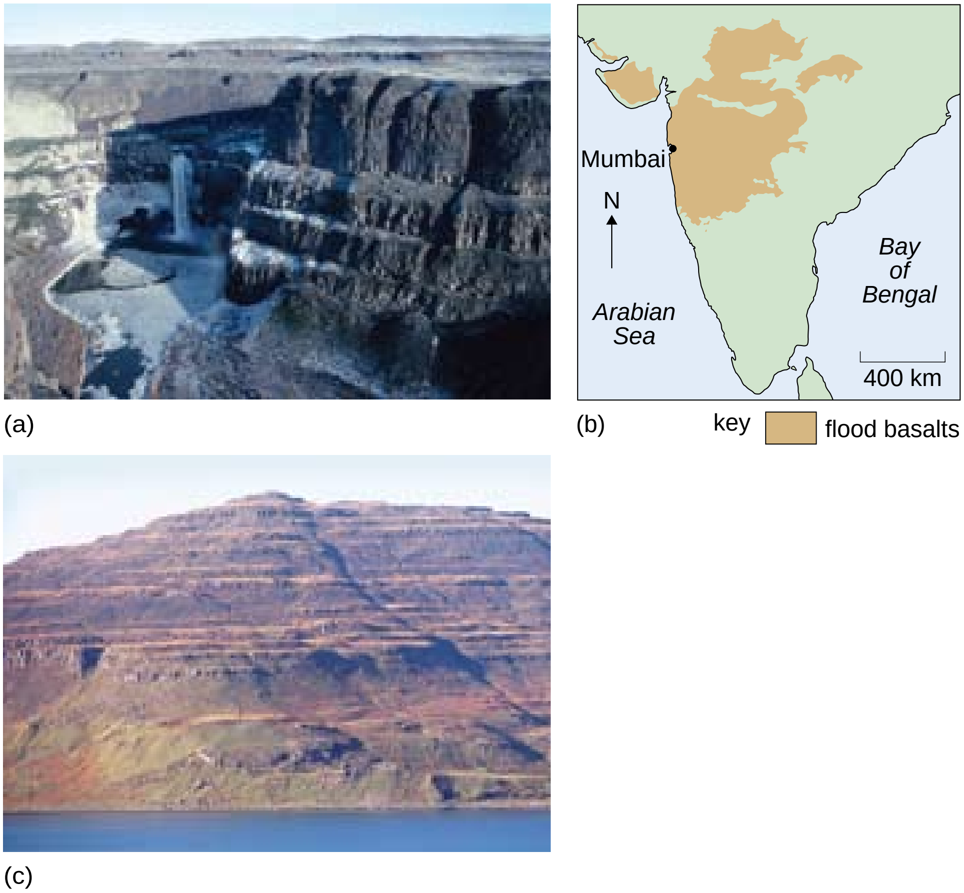


Figure 5.13 Examples of flood basalts. (a) Basalts of the Columbia River Province exposed in a canyon of the Columbia River at Palouse Falls, Washington, USA. Total thickness of lavas here is 115 m. Note vertical jointing in lavas. (b) Map showing the present-day extent of the Deccan Traps continental flood basalts in India. The average total thickness is about 1 km. They cover an area of 0.5×10^6 km^2 today, but prior to erosion, and taking into account what is below sea level, their original extent may have been 1.5×10^6 km^2. (c) Flat-lying lavas on western Mull, dating from about 60 million years ago, in which individual flow units have eroded to produce a stepped topography. Reproduced with the permission of the British Geological Survey © NERC. All rights Reserved.

Continental flood basalts were also erupting during the rifting of the British Isles from Greenland in the early Palaeogene, heralding the opening of the North Atlantic Ocean along the Mid-Atlantic Ridge and vigorous magmatism at the Iceland hot spot. The remains of these flood basalts (the North Atlantic Igneous Province) are to be found in Greenland, the Faeroes, and on the fringes of western Scotland and Northern Ireland (Figure 5.13c). Included within this province are volcanic and plutonic centres on the islands of Mull, Rum and Skye, some of whose shallow and plutonic parts were examined in Chapter 4 (e.g. Figure 4.16).

Flood basalts seem to owe their origin to the arrival of anomalously hot mantle near the Earth's surface, sometimes coupled with stretching and splitting of continental lithosphere. Both processes aid the process of decompression melting that is required to yield exceptionally large quantities of basalt. A useful working hypothesis is that the first burst of activity of the plume is the most volcanically productive because the top of the plume carries the hottest material whereas later plume material, whilst still anomalously hot, is somewhat cooler and therefore produces less melt. This explanation accounts for a long-lived period (tens of millions of years) of hot spot volcanism that produces a hot spot trail (such as the Hawaii–Emperor chain, the Yellowstone chain, and the Iceland hot spot) following the rapid emplacement of a flood basalt province. The exact mechanisms of these processes are not yet resolved, and the details may vary between provinces. Some researchers argue that the available geological evidence is inconsistent with anomalously high temperatures in the mantle being the cause of flood basalt provinces and other voluminous centres of basaltic volcanism that lie far from plate boundaries. Instead, they argue that localised compositional variations in the upper mantle are responsible for regions of anomalously high degrees of partial melting.

Whatever their exact cause(s), flood basalts seem to be erupted at random intervals at random places on the globe, and are among the clearest demonstrations that the Earth's interior is not in a monotonously uniform steady state, but undergoes episodes of enhanced activity. The lava erupted in a large flood basalt province would clearly have devastating local effects, but the amount of volcanic gas liberated over a comparatively short period could also have significant effects on global climate and other environmental systems. The environmental disturbances caused by the eruption of at least two flood basalt provinces – the Deccan Traps (65 Ma) and the Siberian Traps (250 Ma) – may have contributed to the mass extinction events at the end of the Cretaceous and Permian Periods, respectively. Fortunately for us and many other species, there are no indications that eruption of another suite of flood basalts is imminent.

5.4.4 Magmatism at subduction zones

Most of the world's on-land volcanoes occur at convergent plate boundaries on the edges of continents and in oceanic island arcs where andesitic stratovolcanoes of the kind described in Section 3.6.2 predominate. In this subsection, we shall examine how subduction leads to volcanism. Once again, phase diagrams hold the key.

Look back at Figure 2.5. This shows oceanic lithosphere being subducted, and volcanoes situated above the subduction zone. Clearly, melt must be being generated somewhere below the volcanoes. Consider what happens to the downgoing plate, usually referred to as the **slab**, during subduction.

■ What will happen to the pressure and temperature in the slab?

□ The pressure will increase according to depth. As it gets deeper, the slab comes into contact with progressively deeper and warmer parts of the mantle. Heat will be conducted into the slab, causing it to warm up.

Pressure is transmitted instantaneously, so the pressure in the slab must always be the equilibrium pressure for the depth, but the rate at which heat can be conducted into the slab is limited by its thermal conductivity. Therefore, the slab will warm up slowly, at a rate that will depend on how quickly the slab is subducting and how cold it was when it entered the subduction zone. These parameters certainly vary between subduction zones because not all plates move at the same speed and old oceanic lithosphere is colder than younger oceanic lithosphere.

■ What combination of relative age and plate speed will favour the slab reaching high temperatures at shallow depth?

□ Young (warm) lithosphere subducting slowly will have been heated to a higher temperature by the time it reaches a given depth than will old (cold) lithosphere.

Mathematical models of the temperatures within subduction zones suggest that only in the case of the youngest lithosphere will the subducted slab reach its solidus because of heat conducted into it from the hot mantle it encounters. This limitation may explain why lavas with specific intermediate compositions of the sort expected from the partial melting of basalt at high pressure are rare. They are found in the few arcs (such as parts of the Aleutians and southern Chile) where the plate age and speed would be appropriate for the slab to melt. However, the simple geological observation that basalt magmas are erupted from volcanoes in all subduction zones means that partial melting of the basaltic slab cannot be the sole, or even dominant, process for generating magmas in this tectonic setting. To produce basalt, ultramafic peridotite must be partially melted, and this is the rock type that makes up the **mantle wedge** lying above the subducted slab. But the solidus of peridotite is at even higher temperatures than that of basalt, and the wedge is cooled by the cold slab plunging into it. This does not seem to be a situation that could cause the mantle wedge to melt. To understand how basaltic magmas are generated within arcs requires closer consideration of the nature and fate of the slab.

Although oceanic crust is created by partial melting from an anhydrous source (as you saw in Section 5.4.1), by the time it is subducted anhydrous conditions no longer apply.

■ What are the reasons for this?

☐ The most obvious reasons are that a subduction zone begins under water, so the upper part of the slab must be wet. Seawater will permeate cracks and fissures in the oceanic crust, and the veneer of sediments overlying the lavas (which may be subducted with the slab) will also be wet. As you saw in Section 5.4.1, much of the original pyroxene in the crust of the slab will have been hydrothermally replaced by amphibole, which is a hydrous mineral.

The descending slab is made largely of hydrous rocks, as a result of reactions between anhydrous oceanic lithosphere, produced at a divergent plate boundary, and seawater at low pressures. Subduction has the effect of placing these rocks in a high-pressure environment with the result that they experience metamorphic reactions that change a hydrous mineral assemblage to an anhydrous mineral assemblage. These reactions are known as **dehydration reactions** and they release water from the slab. For example amphibole-rich rock (amphibolite) dehydrates to **eclogite**, a rock dominated by pyroxene and garnet. The water leaves the slab and enters the mantle wedge, turning initially dry peridotite into wet peridotite.

The significance of this stems from the fact that water lowers the melting temperature (solidus) of rock (Figures 5.2, 5.3 and 5.6).

Question 5.5

The mantle wedge beneath volcanic arcs lies between the Moho, at around 35 km depth (1000 MPa, or 1 GPa), and the top of the slab at about 100 km depth (3000 MPa, or 3 GPa). If the temperature in this region ranges from 1000 to 1300 °C, what can you deduce from Figure 5.6 about the ability of this region to yield partial melts?

Water not only plays a critical role in the genesis of magmas at subduction zones, but it also becomes an important component of the magma, with arc magmas containing several weight per cent water. This is the driving force behind the explosivity of many subduction zone volcanoes.

Magmatism in arcs (and other tectonic settings) stems from the production of basalt by partial melting of mantle peridotite. The crust of the overriding plate has no direct role in the process. However, magmas of intermediate and felsic composition are also intruded and erupted at subduction zones, granites and felsic ignimbrites being particularly notable at continental arcs. This is because fractional crystallisation and assimilation (Section 5.3) can cause the silica content of magma to increase during ascent. The full interplay of processes is complex, and varies between volcanoes and through the lifetime of a single volcano, but the essential story is summarised in Figure 5.14.

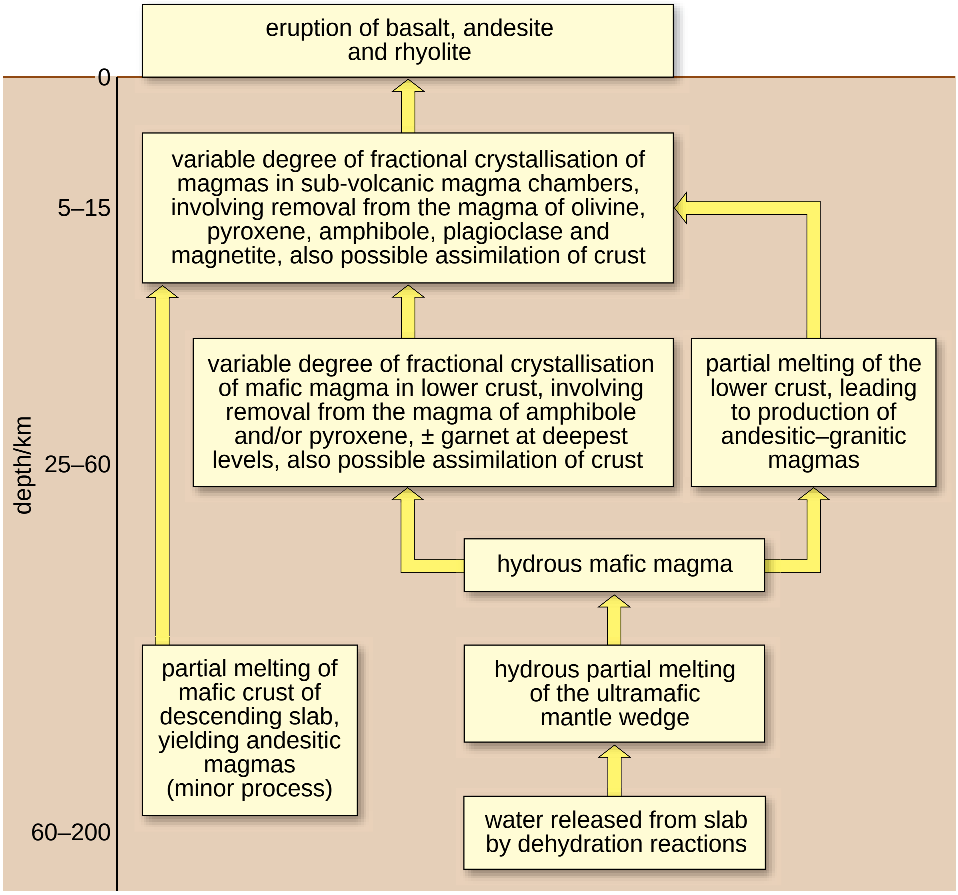


Figure 5.14 Possible processes involved in the generation and evolution of magmas at subduction zones.

Question 5.6

Summarise the difference between magma generation and evolution at subduction zones (Figure 5.14) and at mid-ocean ridges, paying attention to: (a) depth and process of magma generation by partial melting; (b) depth and degree of fractional crystallisation and crustal assimilation; and (c) the composition of the erupted magmas.

5.4.5 Magmatism associated with continental collision

There remains only continental collision to complete our survey of the settings for igneous processes. Figure 2.6b showed that subduction stops after two continents have collided, at which time the magma-generating processes described in Figure 5.14 must also terminate. However, it is found that plutons, especially granites, are emplaced in continental collision zones for several tens of millions of years after a collision. The Lake District batholith is an example of this type, having been intruded in the aftermath of a collision that united England and Scotland in the Silurian.

Geochemical arguments can be used to demonstrate that there is a substantial contribution from the mantle in some post-collision granite magmas, which is not very well understood. Other granite magmas can be explained more simply by considering what collision does to the crust. Compressional deformation of the leading edge of the overriding continental crust combined with partial subduction of the leading edge of the continental crust on the subducted plate means that the crust near the suture zone is much thicker than before collision (Section 2.3.1). For example, the continental crust below the Himalaya (the site of a suture zone between India and the main mass of Asia) is about 70–90 km thick, or about twice the average continental thickness.

■ What are the relative contributions of continental crust and mantle to radiogenic heating within the Earth?

☐ Volume for volume, continental crust generates heat at a rate orders of magnitude greater than the mantle (see Section 2.4).

Deep in the thickened continental crust of a collision zone radiogenic heating will slowly raise the temperature, insulated by a thick layer of crust that is itself producing radiogenic heat by the decay of K, U and Th. Gradually, the temperature may become hot enough for the deep crustal rocks to melt.

■ Which compositional attributes will lead to a rock having a low solidus temperature at a given pressure?

☐ Rocks that are rich in felsic components, that are made from several minerals, and which contain hydrous minerals such as amphibole and mica, have lower solidus temperatures than rocks without these attributes.

The crustal rock types with the lowest solidus temperatures are mudstones and their metamorphic equivalents. Reaction of muscovite and quartz in these rocks will yield a liquid with the composition of granite and an unmelted solid residue. Together these form a migmatite and, if the liquid can segregate from its residue, give rise to granite intrusions (Figure 4.9a). Many of the granites in collision zones were formed by the melting of muddy sediments and other less silica-rich crustal rocks (average continental crust is intermediate in composition, see Section 2.2.1) as a result of the high rates of radiogenic heating within the thickened crust.

You will take a closer look at how heat flow is affected by crustal thickening in Chapter 6, but for now perhaps you will be relieved to have at last found a melting mechanism that relies on heat rather than pressure changes! You have also reached the interface between igneous and metamorphic processes.

Activity 5.1 Igneous rocks and igneous processes

In this activity, you will relate igneous rock specimens included in the Digital Kit and Virtual Microscope to the processes that formed them.

5.5 Summary of Chapter 5

1 Any silicate rock consisting of more than one mineral melts over a range of temperatures and pressures. In a *P–T* phase diagram, the solidus and liquidus separate stability fields of solid only, solid (crystals) + liquid, and liquid only.

2 Under hydrous conditions, or in the presence of any other volatile, both the solidus and liquidus plot at lower temperatures for a given pressure (except for zero pressure).

3 Partial melting of silicate rocks yields a melt that is more felsic than the starting material. This can become separated from the remaining crystals, leaving these behind as a residue that is less felsic (more mafic) than the starting material.

4 Magmas tend to rise because they are less dense than solid rock.

5 A magma can evolve to a more felsic composition by fractional crystallisation (if the crystals are removed) as it cools within a magma chamber, or by assimilating crustal rocks of more felsic composition.

6 At divergent plate boundaries and intraplate volcanoes, mafic magma is generated by decompression melting of anhydrous upwelling mantle.

7 At subduction zones, partial melting occurs in hydrous conditions. Water escaping from the slab hydrates the mantle wedge to produce partial melts of mafic (basalt) composition. Melts may become more felsic during ascent, mainly as a result of fractional crystallisation but perhaps also through crustal assimilation.

8 The rate of radiogenic heat production in continental crust that has been thickened as a result of continent–continent collision can cause sufficiently high temperatures for magmas of mainly felsic composition to be produced by partial melting within the crust.

5.6 Objectives for Chapter 5

Now you have completed this chapter, you should be able to:

5.1 Use *P–T* phase diagrams to explain how melting and crystallisation may occur as a result of changes in pressure, temperature or water content.

5.2 Explain how the composition of igneous rocks depends on a combination of melting, crystallisation and assimilation of country rocks.

5.3 Describe the most common types of igneous activity at divergent and convergent plate boundaries and in intraplate settings, and explain how the magma is generated and how its composition may evolve during ascent.

5.4 State the igneous rock types most likely to form in each plate tectonic setting, but be aware of possible exceptions.

Now try the following question to test your understanding of Chapter 5.

Question 5.7

In continental areas, temperature usually rises with depth at a rate of 20–40 °C km^{-1}. Suppose the rate in a particular region is 30 °C km^{-1}.

(a) What would you expect the temperature to be at a depth of 30 km? (Assume that the surface temperature is 10 °C.)

(b) Plot your answer to part (a) on Figure 5.3, using the depth scale on the right of the diagram, and label it point A. If conditions are anhydrous, explain whether the physical state of felsic material will be solid, liquid, or both.

(c) Suppose that sufficient water is introduced to saturate the system. Explain what the physical state of the felsic material will be now.

(d) Suppose the resulting water-saturated magma rises towards the surface, and that it loses heat at an average rate of 5 °C km^{-1}. Plot a line starting from point A representing the changing *P* and *T* followed by magma cooling at this rate as it rises, and with reference to this line describe what will happen. Speculate on whether or not any magma will reach the surface.

Chapter 6 Metamorphism

6.1 Introduction

As described in Section 2.4, the Earth's crust is heated both by primordial heat and by radiogenic heat from decay of the heat-producing isotopes of K, U and Th. This is why temperatures increase with depth inside the Earth. Granites in continental collision zones are formed by melting crustal rocks, but heating the Earth's crust does not always result in melting. As rocks become hotter, their constituent minerals recrystallise and form different minerals that are stable at higher temperatures. Deeper in the Earth's crust (or mantle), the mass of the overlying rocks results in higher pressures imposed on the rocks and new minerals form that are stable at these pressures. Moreover, the shapes and alignments of minerals change in response to increasing pressures as rocks reorganise themselves to take up smaller volumes or different shapes. The changes in the mineralogy and in the orientation of minerals that make up a rock when subjected to changes in temperature and/or pressure are known as metamorphism.

Broadly, the study of metamorphic rocks comprises three main aspects, which should be considered together wherever possible:

1 Evaluating the conditions of temperature and pressure to which the rocks were subjected during metamorphism

There is direct evidence from many laboratory experiments performed over the past 50 years that new minerals form at given temperatures and pressures. Rocks of different compositions are heated and squeezed in a laboratory under specified pressures and temperatures, and the resulting minerals are compared with those observed in naturally occurring metamorphic rocks. Pressures of up to 100 GPa can be obtained in the laboratory using a diamond-anvil cell (Figure 6.1). Note that a pressure of 1000 MPa (1 GPa) is equivalent to a depth of about 35 km beneath the Earth's surface.

The conditions of pressure and temperature that produced the naturally occurring metamorphic rocks are assumed to be the same as those imposed during the experiment. However, the assumption that rocks in the laboratory will behave similarly to rocks in the Earth's crust or mantle may not always be valid. Natural rocks are rather complex chemical systems. In addition to the major elements that make up the more abundant minerals, rocks contain a large number of minor elements that, although present in minute concentrations, can have a strong effect on the stability of metamorphic minerals. Equally important, minerals in natural rocks normally form in the presence of a fluid phase (usually a mixture of H_2O and CO_2), and the composition of this fluid influences which minerals are stable under the particular P–T conditions. Hydrothermal fluids of appropriate composition can be introduced into diamond-anvil cell experiments. Unfortunately, only rarely can the fluid phase that was present during the metamorphism of rocks be sampled, and the fluid compositions usually have to be inferred from the numerous minerals observed. There is also the problem of time. Natural rocks usually have millions of years in which to react to new conditions; but in the laboratory, the available timescale is restricted to days

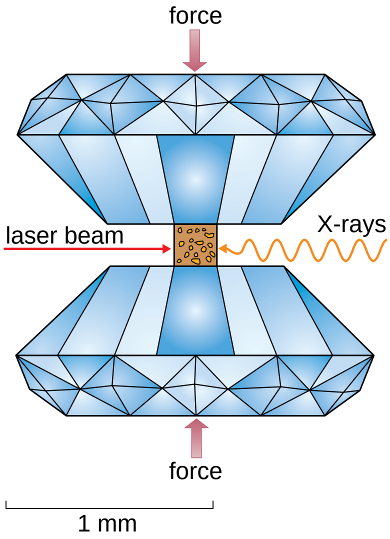


Figure 6.1 Diamond-anvil cells used for high-pressure experiments. A mineral sample is placed in a small hole (0.25 mm diameter) in metal foil inserted between two cut diamonds, each with 16 facets and weighing 0.125 carats. Diamonds are used because of their hardness and transparency to the X-rays needed to identify the resulting minerals. Temperatures are raised by laser beams impacting on the sample, which is inserted into a screw vice that squeezes the diamond tips together and imposes the required pressure on the sample.

or perhaps weeks. This is most important at low temperatures, because under such conditions the rates of reaction are very slow, and it is difficult to achieve equilibrium in the laboratory experiments. At higher temperatures, reactions proceed more rapidly, and so equilibrium can be achieved in a comparatively short space of time.

2 Thermal evolution and its relationship with the tectonic environment

In trying to interpret the significance of the *P*–*T* information gained from studying metamorphic minerals, the geologist must make sure it fits in with all the other geological evidence available. Heat for metamorphism comes from both internal sources (largely radiogenic heat), and from external sources such as the intrusion of hot magma. The precise combination of these sources determines the way that temperature increases with depth. This is recorded by the distribution of metamorphic minerals, and reflects the tectonic setting in which the rocks have formed. When metamorphic studies are combined with structural studies, the tectonic setting can often be identified, not only as a result of the highest metamorphic temperatures that the rocks have experienced, but also whilst the rocks were cooling during their ascent towards the surface of the Earth as a result of uplift and subsequent erosion. This process is called **exhumation**.

3 The relationship between metamorphism, as seen in the growth of new minerals, and deformation events recorded in large-scale and small-scale structures

A careful study of the mineral textures seen in thin sections and hand specimens provides the bridge between an estimate of metamorphic conditions and an understanding of the geological events that caused them: without it, we should learn little of the history of a metamorphic terrain. Such studies are based on textural evidence as discussed in Chapter 9.

In practice, the subject of metamorphism is limited to those transformations that take place between the surface zones of sedimentation and zones deep in the Earth's crust and mantle where partial melting begins. Such boundaries are gradational, and so inevitably there are areas of overlap in the study of the mineralogical and chemical changes that occur during the compaction and lithification of sediment, and the study of igneous petrology, where rocks start to melt. These two examples are of very low **metamorphic grade** and very high metamorphic grade, respectively.

The simplest definition of metamorphic grade is that higher grades of metamorphism reflect conditions of higher temperatures. It is a rather loose definition, but it will be defined more precisely later.

Activity 6.1 Metamorphic assemblages

In this activity, you will re-examine metamorphic rocks using the Virtual Microscope.

6.2 Metamorphic reactions

Rocks are solid objects and do not appear to react, at least at room temperature and atmospheric pressure. However, minerals, like most other chemical compounds, will react if temperatures and pressures are changed sufficiently. In fact, all mineral assemblages in metamorphic rocks result from chemical reactions that take place as the rock undergoes metamorphism. These reactions can be expressed in terms of either mineral names or chemical formulae.

For example, consider the reaction:

$$\text{muscovite} + \text{quartz} \rightleftharpoons \text{alkali feldspar} + \text{sillimanite} + H_2O \qquad (6.1)$$

H_2O will be a gas or a liquid depending on the *P*–*T* conditions of the reaction. The minerals muscovite and quartz are the solid **reactants** and alkali feldspar and **sillimanite** (a metamorphic mineral) are the mineral **products**. The same reaction can be expressed by an equation using the chemical formulae of the four minerals, and the fluid phase:

$$\underset{\text{muscovite}}{KAl_3Si_3O_{10}(OH)_2} + \underset{\text{quartz}}{SiO_2} \rightleftharpoons \underset{\substack{\text{alkali}\\\text{feldspar}}}{KAlSi_3O_8} + \underset{\text{sillimanite}}{Al_2SiO_5} + H_2O \qquad (6.2)$$

There are two properties of chemical reactions we now need to introduce in order to understand metamorphic reactions better.

First, any chemical reaction involves a change in **entropy** (represented by the symbols ΔS, pronounced 'delta S') between reactants and products (i.e. the sum of the entropies of the products minus the sum of the entropies of the reactants). Entropy reflects the ordering of the atomic structure of matter. A system with zero entropy is perfectly ordered, whereas the same system with high entropy is more disordered. At any given pressure, a liquid or a gas is more disordered than a crystalline solid. Consider the chemical reaction involved in dissolving sugar in a cup of coffee. The reactants are sugar (a crystalline solid) and coffee (a liquid); the product is entirely liquid (sweet coffee), so the system has increased its atomic disorder and therefore increased its entropy.

■ To take another example, which has the lowest entropy: ice, water or steam?

☐ Ice, with a crystal structure, is the most highly ordered and so has the lowest entropy.

Second, any chemical reaction involves a change in **molar volume** as represented by the symbols ΔV ('delta V'). The molar volume is the volume occupied by one molecular weight of the mineral, gas or liquid. ΔV is defined as the change in volume between the sum of the molar volumes of the reactants and the sum of those of the products. A substance with a small molar volume has a closely packed atomic structure and is dense; conversely, a substance with a large molar volume is less dense. So a gas has a larger molar volume than a liquid, which in turn has a larger molar volume than a solid. When graphite is transformed into diamond at high pressures, there is a net increase in density and so a decrease in the molar volume.

■ Returning to Reaction 6.1, is the value of ΔS positive or negative? (You can assume that the four minerals involved have similar values of entropy.)

□ H_2O is the only fluid (liquid or gas) involved. Fluids have less-ordered structures than any solid and so have higher entropies than any solid. Consequently, entropy of the products is greater than the entropy of the reactants. In other words, entropy increases as the reaction proceeds to the right and so ΔS is positive.

■ Is the value of ΔV positive or negative in Reaction 6.1?

□ It is positive, because two solids have reacted to form two other solids and a fluid, which is generally much less dense (and so has a large molar volume) than any solid.

Entropy may also be used to derive a more satisfactory definition of metamorphic grade. Earlier (Section 6.1), we stated that higher grade is usually taken to mean higher temperature. We may now define metamorphic grade in terms of the entropy change (ΔS) of the metamorphic reactions concerned. *Increasing grade involves an increase in entropy of the metamorphic system* (including, of course, any gas phase).

■ In Reaction 6.1, which assemblage will be present at higher metamorphic grade: (i) muscovite and quartz, or (ii) sillimanite, alkali feldspar and H_2O?

□ Assemblage (ii), sillimanite, alkali feldspar and H_2O, as this assemblage is less ordered, so has the higher entropy.

Most metamorphic reactions take place as the metamorphic grade increases, and are called **prograde reactions**; they tend to decrease the degree of ordering in the mineral system concerned.

Question 6.1

Another common metamorphic reaction is:

$$\text{tremolite} + \text{calcite} + \text{quartz} \rightleftharpoons \text{diopside} + H_2O + CO_2 \qquad (6.3)$$

where tremolite (a type of an amphibole) and diopside (a type of pyroxene) are both rich in Ca and Mg.

Determine: (a) which assemblage has the lower entropy, and (b) which assemblage would be stable at a higher grade of metamorphism.

For the muscovite + quartz reactions (Reactions 6.1 and 6.2), at a pressure of 150 MPa, temperatures of more than 550 °C are needed for muscovite and quartz to react to form alkali feldspar, sillimanite and H_2O. If alkali feldspar, sillimanite and H_2O are taken below 550 °C again, they will react together and revert to muscovite and quartz. How can it be then that alkali feldspar and sillimanite are often found in the same rock at room temperatures?

There are two reasons:

1. One of the products of the prograde reaction is a gas, H_2O, and the short answer is that it is not likely to stay around and wait for the rock to cool. Thus, if alkali feldspar and sillimanite are unaccompanied by H_2O at temperatures below 550 °C, they obviously cannot revert to muscovite and quartz.

2 The rates at which reactions take place vary enormously. In particular, they increase exponentially with increasing temperature, so that many reactions are extremely slow at low temperatures. Thus, if an assemblage is cooled rapidly, it may not have enough time to react and form low-temperature minerals.

In general, if conditions are such that reactions take place during cooling, they are called **retrograde reactions**; these occur as rocks readjust to the lower grades of metamorphism. The best textural evidence for retrograde metamorphism is provided by a relatively high-grade mineral surrounded, and partially replaced, by a mineral that is stable at lower metamorphic grades. For example, Figure 6.2 shows garnet partly being replaced by chlorite. The rock shows **textural disequilibrium** between the garnet and the chlorite because the reaction has not gone to completion. Retrograde reactions tend to occur during slow cooling in rocks that contain an H_2O-rich fluid phase. The reason that some garnet is preserved in Figure 6.2 is because either the reaction ran out of fluid, or the rock cooled too fast for the reaction to go to completion.

Figure 6.2 Garnet crystals (high relief) partly replaced by pale-green chlorite as a result of retrograde metamorphism (viewed in plane-polarised light; width of image = 1 mm).

Unfortunately, because the experimental problems are acute, very little is known about the actual rates of specific metamorphic reactions between coexisting solids. An exception is the change of aragonite to calcite. These two minerals are polymorphs: both have the same composition ($CaCO_3$) but different crystal structures, and the change from one to the other takes place in less than ten years at temperatures above 400 °C. The reaction has been studied experimentally at these temperatures and the results extrapolated to lower temperatures by plotting a graph of temperature against the time taken for all the aragonite to change to calcite (Figure 6.3). Note that such extrapolations can introduce comparatively large errors.

The experiments used to construct Figure 6.3 were done in the absence of water, but if water is present the times are greatly reduced. Even at 50 °C, all the aragonite would change to calcite in a mere million years, a consequence of which is that aragonite is very rarely found in fossils or as cement in ancient rocks.

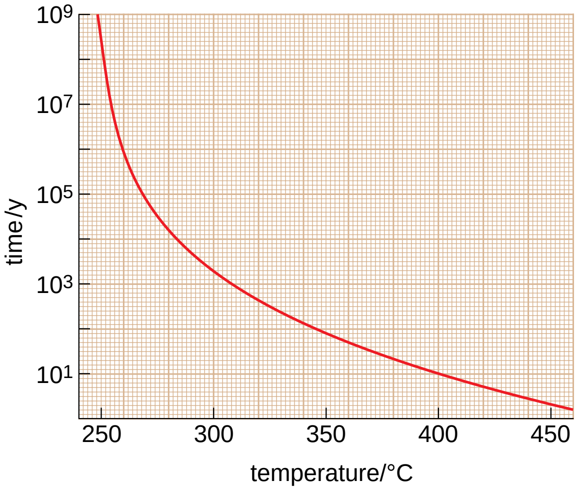


Figure 6.3 Time in years for complete change from aragonite to calcite as a function of temperature (anhydrous conditions).

Why should water speed up the change – bearing in mind that it cannot contribute chemically in a reaction in which one polymorph of $CaCO_3$ goes to a second polymorph of $CaCO_3$?

Water is an excellent **catalyst**, and is particularly important in speeding up metamorphic reactions that otherwise might not take place in the time available. In completely dry rocks, chemical changes can only occur by the slow process of ionic diffusion (the passage of ions through a solid crystal lattice). When water is present, it tends to spread itself along grain boundaries as an intergranular film, providing a network of chemical 'arteries' in which ions may move rapidly in solution and so speed up the metamorphic reactions.

This discussion of the muscovite + quartz reaction and the structural change from aragonite to calcite highlights three important general points:

1. Many prograde metamorphic reactions release H_2O and/or CO_2. Because these products are mobile and tend to move away from the rocks in which they were generated, they are often unavailable for retrograde reactions that might otherwise take place as the temperature falls.
2. Rates of reactions tend to be slow, but they increase exponentially with increasing temperature and are also speeded up by the presence of water.
3. Conversely, since many prograde metamorphic reactions release H_2O and CO_2, both tend to be driven off as metamorphism continues. Thus, by the time the highest grades are reached, almost all the fluid will have left the system, and recrystallisation and the growth of new minerals will have sealed off all the intergranular spaces, making the rock almost impermeable. Not only are H_2O and CO_2 not present to take part in reverse reactions, they are also unavailable as catalysts. Thus, such 'dry' rocks will then survive cooling more or less unaltered, because the rates of reaction are limited by the very slow rates of ionic diffusion. So high-grade mineral assemblages become 'frozen in', and preserved at low temperatures.

High-temperature minerals can be present at low temperatures, but they do not lie within their stability field when plotted on a phase diagram. Minerals that exist outside their stability fields are described as being **metastable**, which means that while they are not in a state of change, they can be made to change by some sort of impetus. For example, water can remain a liquid at temperatures below its freezing point, particularly if it is flowing. Freezing can be induced by halting its flow, or by adding a chunk of ice. For metamorphic assemblages, the impetus either comes from further metamorphism, when the rock is reheated and/or squeezed under a new set of *P*–*T* conditions, or from weathering, when the rock is attacked by atmospheric precipitation and its minerals are transformed into those that are in equilibrium at the temperatures and pressures present at the surface of the Earth – for example clay minerals in sediments.

■ How do we know that metamorphic rocks were in fact formed at high temperatures and pressures?

□ From experiments designed to reproduce naturally occurring mineral assemblages under conditions of known pressure and temperature (Figure 6.1).

The results of such experiments are most simply illustrated using a phase diagram. In the experiments on which phase diagrams are based, there are three things that can be varied easily – the chemical composition of the sample, the temperature and the pressure. The minerals that form will depend on all three factors. It is obviously difficult to illustrate how they all vary on a two-dimensional diagram, so for convenience we usually keep one constant and show what happens as the other two vary. In most metamorphic experiments, it makes sense to work with a fixed composition, and to plot the results on a graph of pressure against temperature. Always remember, however, that such a *P*–*T* phase diagram only illustrates results for a particular composition; if the composition is altered, the details of the phase diagram will change.

Phase diagrams such as Figure 6.4a illustrate the stability ranges of three minerals all of which share the same composition, in this case Al_2SiO_5. These minerals are quite different in appearance. Kyanite forms blue blades (Figure 6.4b), sillimanite forms bundles of white fibres (Figure 6.4c) and andalusite forms prismatic needles with diamond-shaped cross-sections (Figure 6.4d).

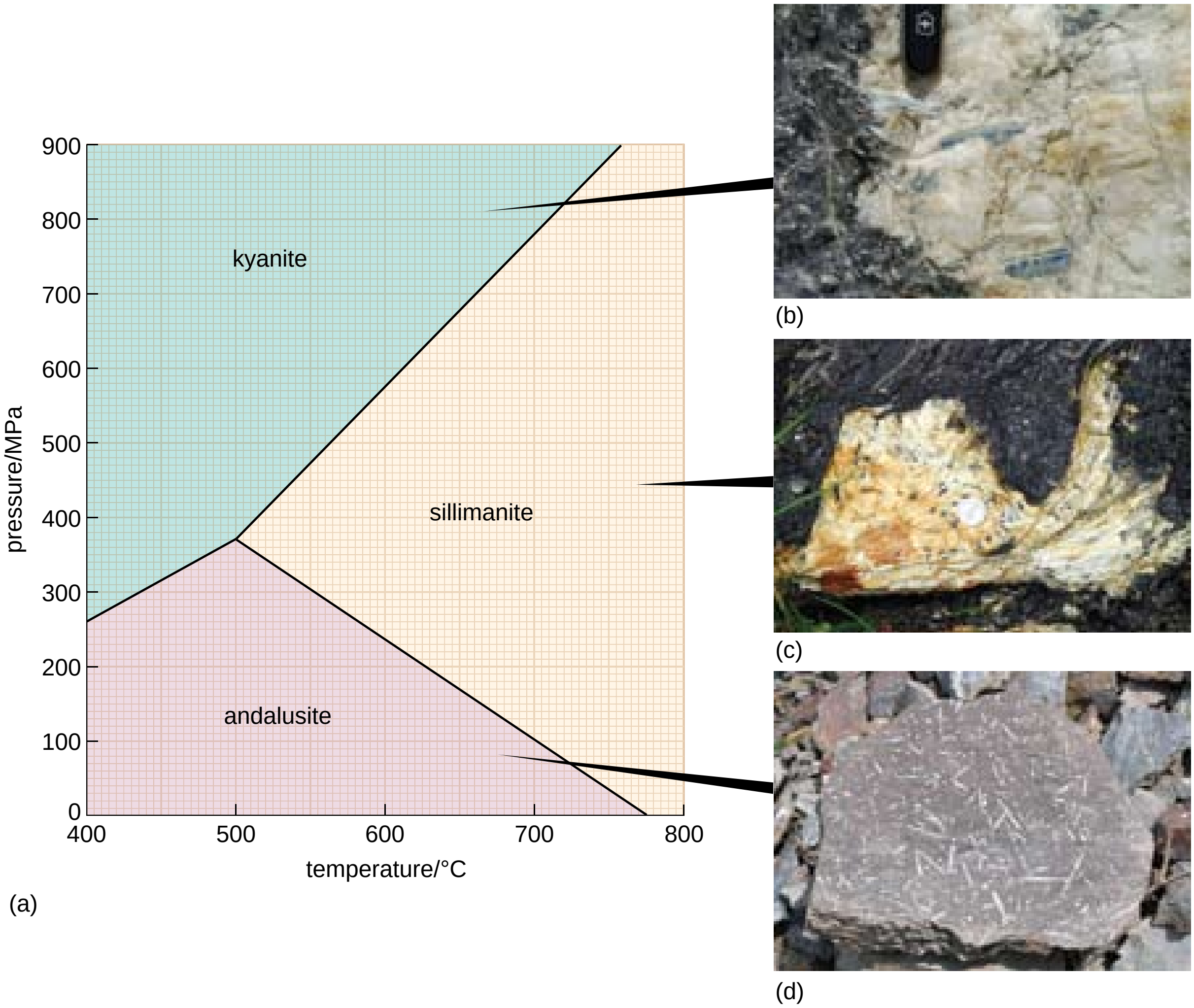


Figure 6.4 (a) Phase diagram illustrating the stability fields of three aluminium silicates: andalusite, kyanite and sillimanite. (b) Blades of blue kyanite in quartzite from the Himalaya (NW India). See also kyanite in the Digital Kit. (c) Sheaves of white fibres of sillimanite, locally discoloured by iron staining, from the Himalaya (NE India) (the coin is about 2 cm across). (d) White needles of andalusite in a spotted hornfels from the aureole of the Skiddaw granite, Cumbria, England. Andalusite is diamond-shaped in cross-section (width of image is 20 cm). See also andalusite in the Digital Kit.

Several points need to be emphasised in interpreting the phase diagram.

1. You may recall that a *phase* differs chemically and/or physically from the rest of the system being considered. In this case, kyanite, sillimanite and andalusite are three different phases because, although they share the same chemical composition, they have different crystal structures (i.e. they are polymorphs).
2. At equilibrium, the minerals kyanite, sillimanite and andalusite each exist over a range of different pressures and temperatures (Figure 6.4a). These are the conditions under which each mineral is stable, and the area depicting these conditions on a phase diagram is called its stability field. (Note also that to describe a mineral as 'stable' is synonymous with saying it is 'at equilibrium'.)
3. If a phase is present under conditions outside its stability field, it is said to be metastable.
4. The lines that mark the boundaries between the stability fields of kyanite, sillimanite and andalusite are phase boundaries. These lines mark the only places on this phase diagram where two of these minerals can exist together *at equilibrium*.

If either the pressure or temperature is changed so that conditions move off the phase boundary, only one of the minerals becomes stable. You have seen how such changes can take a very long time to complete (Figure 6.3).

Question 6.2

Use the Al_2SiO_5 phase diagram (Figure 6.4a) to answer the following questions.

(a) Is andalusite stable at 300 MPa and 700 °C?

(b) Which phase is stable at 800 MPa and 500 °C?

(c) At what temperature do kyanite and sillimanite coexist in equilibrium at a pressure of 700 MPa?

(d) Kyanite and andalusite were reported together in the same sample at 400 °C at atmospheric pressure (0.1 MPa). Which of these two phases is metastable (i.e. lies outside its stability field)?

Your answers to Question 6.2 demonstrate that in the case of the kyanite–andalusite–sillimanite system, it is comparatively easy to make very general statements about the conditions of pressure and temperature, depending on which mineral was present at equilibrium; however, it is difficult to be specific. This can be improved by using a sample in which two of the minerals occur together in equilibrium, since this indicates that conditions were somewhere along the line of the phase boundary. But can the estimates of pressure and temperature be made even more precise? Can we say precisely where along the phase boundary the minerals from a sample were in equilibrium? The short answer is no – unless help is available from elsewhere. What we need is some other mineral with a different stability field to be present in the same rock.

■ What other reaction have you looked at that also involves sillimanite?

☐ Reactions 6.1 and 6.2, in which muscovite and quartz react to form sillimanite, alkali feldspar and H_2O.

This reaction is plotted in Figure 6.5, and in this case, the phase boundary marks the boundary between the stability field of muscovite and quartz and that of Al_2SiO_5, alkali feldspar and H_2O. Note that the term sillimanite has been replaced with the formula Al_2SiO_5 since other polymorphs of this composition are stable within the *P*–*T* conditions shown. The boundary defines the conditions at which the five phases can coexist together in equilibrium. Because the gradient of this slope is fairly steep, we can say that, for normal crustal pressures ($P < 1000$ MPa), all five phases will coexist over a temperature range of about 400–800 °C. But how can a more precise estimate be obtained?

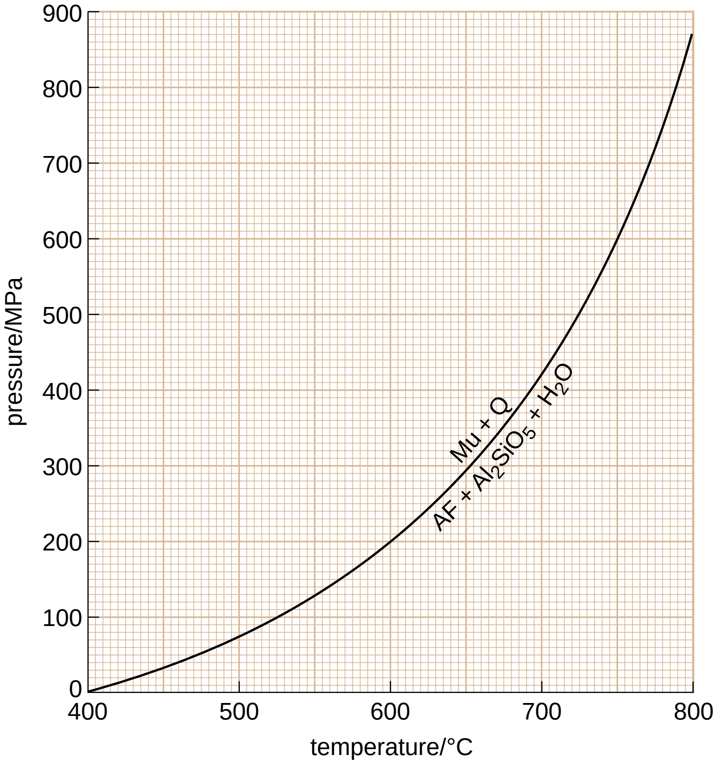


Figure 6.5 Pressure–temperature plot for the reaction muscovite (Mu) + quartz (Q) $\rightleftharpoons$ alkali feldspar (AF) + $A1_2SiO_5 + H_2O$.

Figure 6.6 combines the phase diagrams for sillimanite–kyanite–andalusite and for muscovite–quartz–Al_2SiO_5–alkali feldspar–H_2O. By using all four phase boundaries, the diagram becomes subdivided into smaller areas of pressure and temperature (A, B, C, D and E) in which particular combinations of minerals are stable. You can see, for example, that muscovite and quartz only coexist in equilibrium with andalusite in area A.

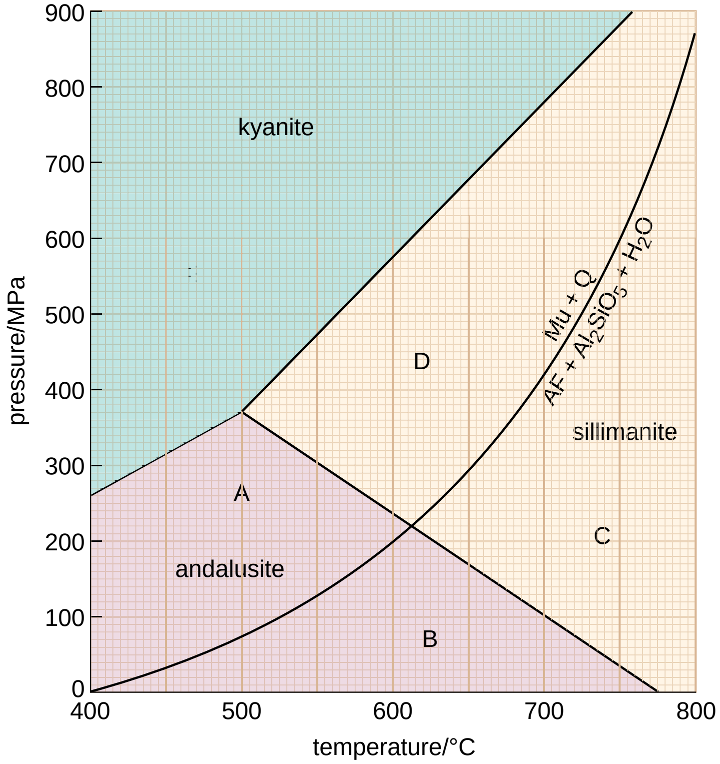


Figure 6.6 Phase diagram combining information from Figures 6.4a and 6.5.

There is one point on Figure 6.6 where six phases can occur together. So, if muscovite, quartz, alkali feldspar, andalusite, sillimanite and H_2O are found in equilibrium in one sample, it can only have crystallised at a particular pressure and temperature.

■ What would that pressure and temperature be?

□ These six phases only coexist at about 610 °C and 220 MPa (where fields A, B, C and D meet at a point on Figure 6.6).

In summary, it should be apparent that:

1 It is not possible to obtain precise estimates of pressure and temperature on every sample of metamorphic rock. Usually, all you can say is that the rock crystallised somewhere in the stability fields of its particular minerals.

2 To obtain a reasonable estimate of metamorphic conditions, it is necessary to seek out rock specimens that contain certain key metamorphic minerals, and so may reflect conditions on a particular phase boundary. In some areas, suitable samples may be scarce or non-existent.

In general, it is necessary to combine information from different metamorphic reactions in an attempt to limit the possible range of pressure and temperature in which the rock could have crystallised. The grid-like pattern that is formed as more and more boundaries are added to the phase diagram is called a **petrogenetic grid**, and Figure 6.6 is a simple example.

There is another way of determining pressure and temperature of which you should be aware. In contrast to the minerals discussed so far in this section, many minerals have a range of compositions – the plagioclase feldspars, for example, range in composition between albite and anorthite by the interchange of the elements Na and Si with the elements Ca and Al. Reactions involving such minerals will not occur along a line on a phase diagram but be spread out across a range of *P*–*T* conditions.

For example, there is a reaction involving the minerals garnet, **cordierite**, quartz and any one of the three aluminosilicates (Al_2SiO_5):

$$\text{garnet} + \text{quartz} + Al_2SiO_5 \rightleftharpoons \text{cordierite} \qquad (6.4)$$

Cordierite (Figure 6.7a) is a silicate mineral found in some aluminous metamorphic rocks (Figure 6.7b). Its composition varies from $Mg_2Al_4Si_5O_{18}$ to $Fe_2Al_4Si_5O_{18}$. Similarly, garnet in such rocks has a composition that varies from $Mg_3Al_2(SiO_4)_3$ to $Fe_3Al_2(SiO_4)_3$. The position of the phase boundary of the reaction described by Reaction 6.4 depends on the precise Fe/Mg ratio of the rock. In Figure 6.8, this phase boundary is plotted for both the iron-rich compositions and the magnesium-rich compositions, determined from experiments containing only Fe and only Mg compositions, respectively. For naturally occurring rocks containing both Fe and Mg, the reaction will lie somewhere between these boundaries, depending on the Fe/Mg ratio of the rock. For example, Figure 6.8 also shows the phase boundary for a natural cordierite with Fe/Mg = 1.5 (a typical value). For reactions involving such minerals, the geologist must not only identify which minerals are present but also analyse the precise composition of the cordierite and garnet to find the appropriate position of the phase boundary at which the two minerals coexist in equilibrium with an aluminosilicate and quartz.

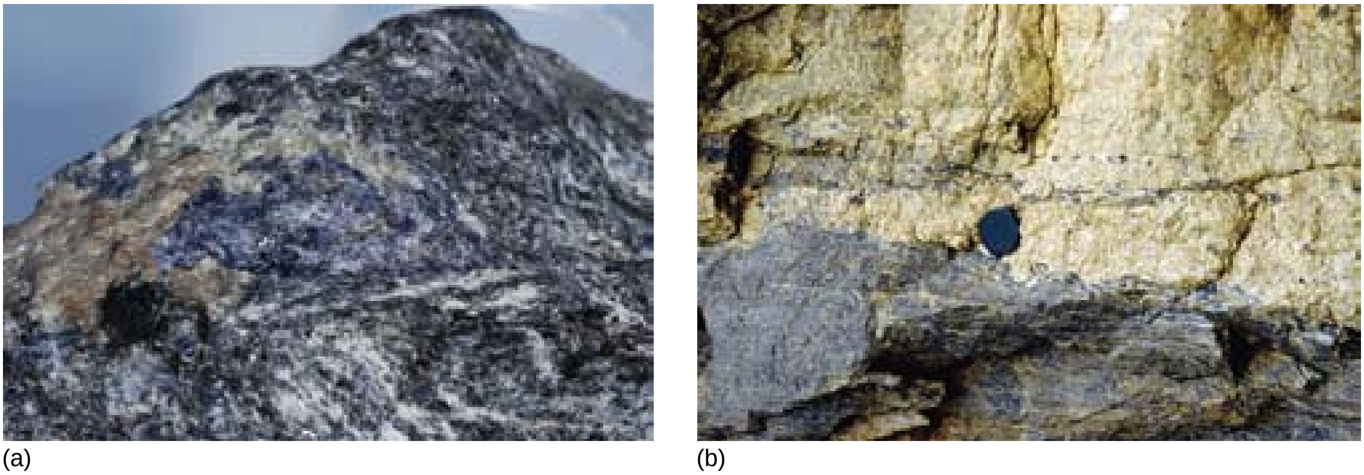

(a) (b)

Figure 6.7 (a) Cordierite (blue mass of crystals in quartz-rich patch about 2 cm across) and biotite (black) crystals from (b) migmatite outcrop, central Madagascar (lens cap for scale).

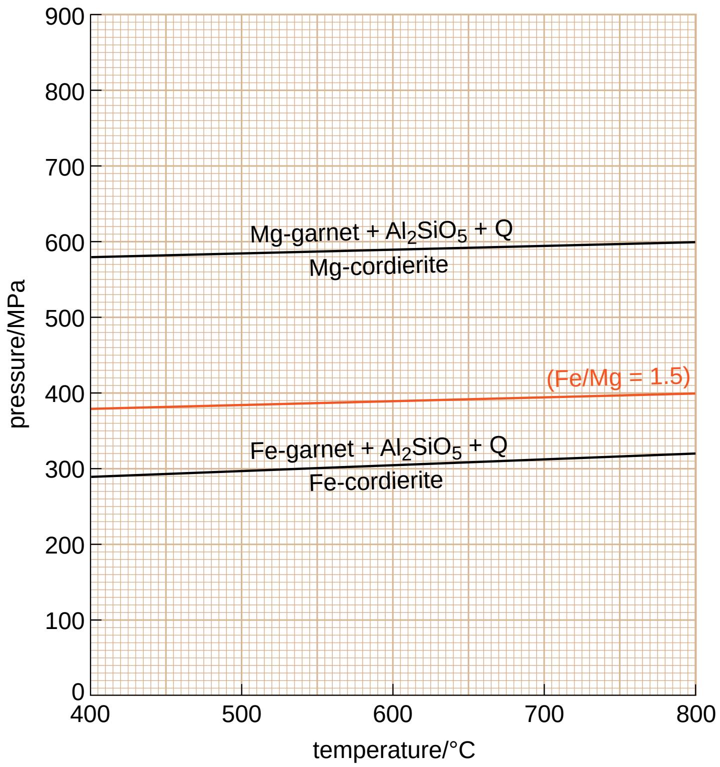


Figure 6.8 Phase boundaries for Reaction 6.4 using Fe and Mg end members. Also shown is the phase boundary for a cordierite with the Fe/Mg ratio = 1.5 (the middle (red) line).

■ What is the main difference between the phase boundaries shown on Figure 6.8 and that on Figure 6.5?

☐ The phase boundary in Figure 6.5 is much steeper.

■ Which reaction gives a better idea of the pressure at which the phases on either side of the boundary can coexist?

☐ The reaction involving cordierite, garnet, sillimanite and quartz, as it gives a narrow range of pressures for a given Fe/Mg ratio.

■ If a rock contains negligible Mg, what is the pressure range over which the mineral cordierite, garnet, sillimanite and quartz can coexist?

☐ From the lower phase boundary in Figure 6.8, a range of 290–320 MPa is indicated.

The reason that this reaction is nearly parallel to the temperature axis is that it involves a large change in molar volume (ΔV) but only a small change in entropy (ΔS).

The large change in volume is because garnet is a denser phase (density = 3.58×10^3–4.32×10^3 kg m^{-3}) than cordierite (2.53×10^3–2.78×10^3 kg m^{-3}). Reactions with a gentle slope, such as Reaction 6.4, make good natural barometers and those with a steep slope, such as Reaction 6.1, make good natural thermometers of metamorphic conditions. Good natural barometers are provided by reactions with a large change in molar volume (ΔV) (but should not also involve a large change in entropy), and good natural thermometers are provided by reactions with a large change in entropy (ΔS).

So far we have stressed the importance of pressure and temperature in determining which minerals are likely to be stable.

■ What is the third, and arguably most important, factor that governs whether a particular metamorphic mineral is stable?

☐ The bulk chemical *composition* of the rock being metamorphosed: rocks of different composition tend to contain different minerals, even if they have been at equilibrium under the same conditions of temperature and pressure (Book 1, Section 8.3).

The reason for the importance of chemical composition is simple. If a rock does not contain the appropriate elements for a particular mineral, then that mineral cannot be formed.

Question 6.3

Which of the minerals garnet, calcite, muscovite and feldspar are likely to be present in the metamorphosed equivalents of: (a) basalt; (b) limestone; and (c) mudstone? Explain your answer in each case.

The important point to remember is that if metamorphic rocks contain different minerals but have been exposed to the same temperatures and pressures, then they must have different chemical compositions.

6.3 Types of metamorphism

At the beginning of this chapter, we described metamorphic processes as those that are associated with mineralogical and structural changes in rocks. You have seen a few of the sorts of reactions that take place, the factors that control them, and the methods used to estimate the conditions of temperature and pressure. You will now look a little more closely at the kinds of rocks that metamorphism produces, and the three main types of metamorphism.

6.3.1 Dynamic metamorphism

Dynamic metamorphism takes place in areas of intense *local* deformation, such as in fault zones or **shear zones**. Temperatures and pressures may be low if dynamic metamorphism occurs at high levels in the Earth's crust where rocks are brittle, but at deeper levels, higher temperatures result in more 'ductile' structures in which flow predominates over fracture. The differences between brittle and ductile deformation will be examined more fully in Chapter 7.

In a major fault zone, there is mechanical movement of one rock over another so that any rocks within the fault zone may be ground down in a process known as **cataclasis** (from the Greek meaning 'break down'). The resulting rocks are called cataclastic rocks. Almost all brittle faults contain a zone in which there are cataclastically broken or crushed rocks known as a **fault breccia** (Figure 6.9).

Figure 6.9 Fault breccia from Barra, Outer Hebrides.

More ductile deformation causes slippage between layers within the rocks, and between planes of atoms within minerals. Recrystallisation under these conditions results in a fine-grained foliated rock with a streaked-out texture indicating the direction of shearing, which is known as a **mylonite** (Figure 6.10a). During the formation of mylonites, quartz is broken down into fine-grained aggregates that form elongated ribbons (Figure 6.10b). During mylonite formation, or indeed any other kind of dynamic metamorphism, the grain size is reduced.

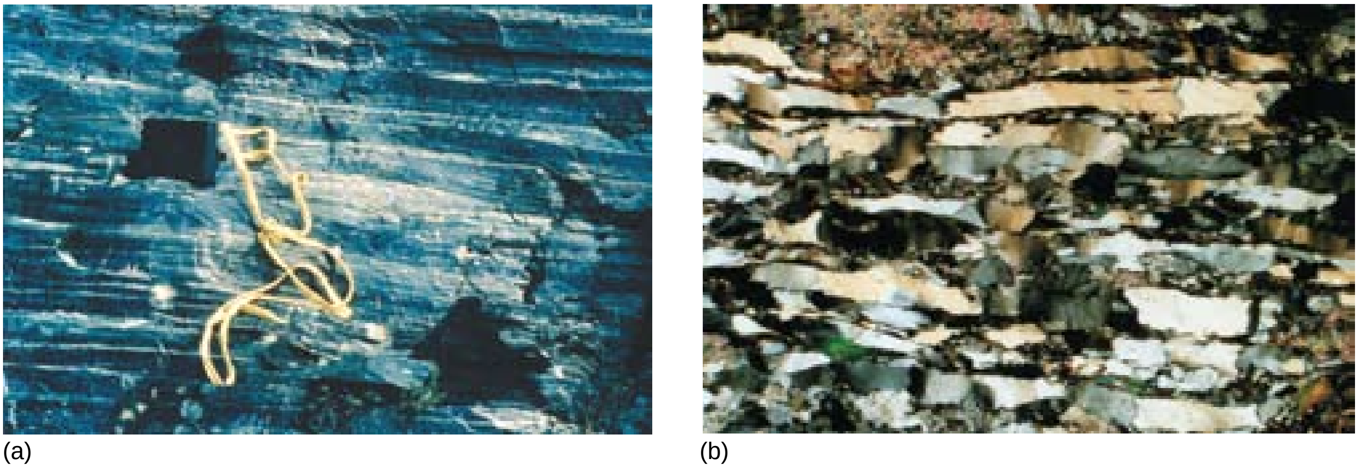


Figure 6.10 (a) Mylonite from the Moine Thrust, Scotland. (b) Ribbon texture from recrystallised quartz in a mylonite (viewed between crossed polars; width of image = 5.5 mm).

Movements along fault planes expend considerable amounts of energy to overcome friction and most of this energy is released as *heat*. Although the size of the resultant increase in temperature depends on (i) frictional heating along the fault plane, and (ii) the thermal conductivities of rocks on either side, it appears to be most sensitive to (iii) the rate of movement of the fault. Earthquakes are produced by rapid movements on faults and it can be shown that movements of this kind, although of very short duration, can produce a temperature rise sufficient to cause melting in the deformed rock – about 5 cm per second has been suggested as the sort of speed at which rock melting might be expected.

Rocks melted by frictional heat are rare, but they do turn up in small quantities all over the world. Such melts form black, fine-grained rocks that look like basaltic glass and because they intrude the country rocks adjacent to the fault zone in small irregular dykes and veins (Figure 6.11), they have often been identified wrongly as igneous rocks of more conventional origin, hence their name **pseudotachylites** (tachylite being an old term for basaltic glass).

Figure 6.11 Pseudotachylite (cross-cutting dark band) within a biotite migmatite gneiss from Uist, Outer Hebrides.

6.3.2 Contact metamorphism

Contact metamorphism (also known as thermal metamorphism) refers simply to the metamorphic changes taking place in response to the heat associated with igneous bodies. It is most obvious around intrusive rocks because, unlike their extrusive equivalents, most of their heat is not lost to the atmosphere but is dissipated into the surrounding country rocks. Naturally, the temperatures are highest close to the igneous body itself, so there is therefore a very marked increase in metamorphic grade near the contact. The zone of metamorphic rocks around the intrusion is termed a metamorphic aureole.

In a metamorphic aureole, high-temperature minerals will grow nearest to the contact and low-temperature minerals are found furthest away. We can plot on a map where these metamorphic minerals occur in just the same way as we can record any other features, and in so doing would map **metamorphic zones**. Each zone is known by the name of a particular characteristic mineral (termed an **index mineral**) present in that zone; for example, the biotite zone or the garnet zone. Provided rocks of similar composition are being examined, the index mineral is a qualitative indication of the metamorphic grade. This approach is extremely useful where rocks of similar compositions are exposed over a large enough area, because the metamorphic zones provide an indication of the relative metamorphic grades, even when precise estimates of pressure and temperature are not available.

For example, the aluminosilicate polymorph that occurs within metamorphosed mudstones very close to their contact with the granites of southwest England is andalusite (Figure 6.4d). The distribution of andalusite around a granite defines the boundaries of the *andalusite zone*. Farther away from the granite, where the temperature of the rocks reached during or after intrusion was lower, andalusite is not found.

In this way, a series of metamorphic zones can be mapped out around an intrusion, each zone representing successively lower temperatures away from the body as, for example, has been done around the Ardara granodiorite in Donegal

(Figure 6.12). In this example, the three aluminosilicates kyanite, andalusite and sillimanite are particularly useful.

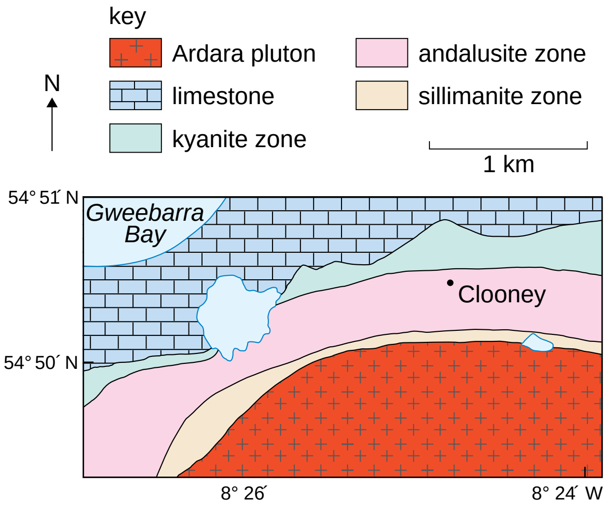


Figure 6.12 Metamorphic aureole around the Ardara pluton, Donegal, Ireland.

■ With reference to Figure 6.4a, determine whether, for any particular pressure, sillimanite is stable at higher or lower temperatures than andalusite. Is that consistent with their relative positions in the contact aureole in Figure 6.12?

☐ Sillimanite is always stable at higher temperatures than andalusite at a fixed pressure (Figure 6.4a), which explains why it occurs nearer to the granite contact than andalusite does in Figure 6.12.

Question 6.4

(a) Since andalusite rather than kyanite reacts to form sillimanite in the contact aureole of the Ardara pluton, what is the maximum pressure at which metamorphism could have taken place?

(b) The pressure at which the granodiorite was emplaced was 200 MPa. Use the phase diagram (Figure 6.4a) to estimate the temperature at the outer margin of the sillimanite zone.

(c) Estimate the depth at which the intrusion was emplaced.

In general, large-scale changes in the bulk chemical composition of the original rocks do not occur during metamorphism. This applies to contact metamorphism, but with some important exceptions. Right up against the igneous contact, elements can be introduced from the crystallising magma. For example, carbonate-rich rocks are sometimes replaced by calcium and magnesium silicates, iron oxides and sulfides.

■ What is the name of the process that changes the bulk composition of rocks during the influx of hydrothermal fluids?

☐ Metasomatism (see Section 4.2).

In some metamorphic aureoles, metasomatism has resulted in the concentration of economic amounts of ores. One such example occurs around the Skiddaw granite in Cumbria, northwest England. Extensive hydrothermal activity has resulted in the alteration of the granite near its margins and the development of mineralised veins that have been exploited by tungsten mining. The granite itself forms a flat-topped intrusion of biotite granite of Devonian age (400 Ma) intruded into Ordovician slates and siltstones (Skiddaw Group, Figure 6.13). An elliptical aureole has developed within the sediments around the granite. Hard hornfels is developed near to the granite, with spotted slates developed in the outer parts of the aureole. The minerals formed in these zones carry important information about the conditions at which the granite has formed.

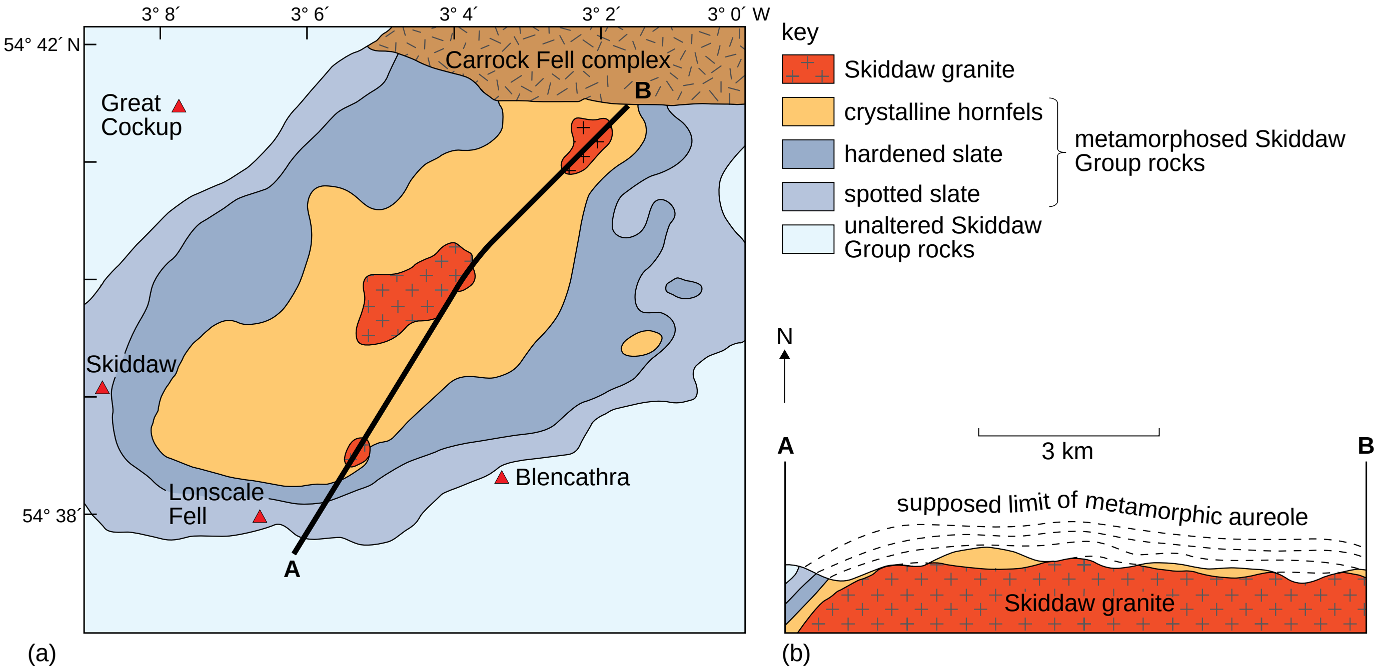


Figure 6.13 (a) Geological sketch map of the Skiddaw aureole. (b) Cross-section along the line A–B in (a).

Finally, the textures of rocks in a thermal aureole can be distinctive. Since the emplacement of magma may not cause a preferred alignment of recrystallised minerals (at least at shallow crustal levels), the main characteristic of rocks that have undergone contact metamorphism is the formation of a hard, compact and splintery rock called a hornfels. Frequently, the first sign of contact metamorphism in the metamorphic aureole is a distinctive 'spotting' caused by the growth of clusters of new metamorphic minerals. Such a rock is called spotted hornfels, as shown in Figure 6.4d.

Activity 6.2 The Skiddaw granite and its metamorphic aureole

This activity brings together the field relations, thin sections and pressure–temperature diagram from a metamorphic aureole in Cumbria, using the Digital Kit.

6.3.3 Regional metamorphism

Regional metamorphism is a very much larger-scale phenomenon than either dynamic or contact metamorphism. Regional metamorphism is usually associated with continental collision or subduction, and therefore with deformation. At the highest grades of metamorphism, rocks begin to melt and ultimately granites are formed.

The patterns of changing metamorphic grade across an area differ between contact and regional metamorphism. In contact metamorphism, the zones tend to be concentric around a particular intrusion, whereas in areas of regional metamorphism, the zones are often linear – as you might expect if they represent deeply eroded mountain belts, formed along some ancient convergent plate boundary or collision zone.

The third and perhaps most important distinction between contact and regional metamorphism is that the sources of heat are different, which has important implications for the rate at which metamorphism occurs. During contact metamorphism, the heat source is a cooling igneous body. This is an effective heat source only for as long as the magma is crystallising. Even for a large granite body, this will be less than 100 000 years. For a dyke, heat may be exhausted after 100 years or less. During regional metamorphism, the heat source comes largely from radiogenic heat from the decay of heat-producing elements. This is a much slower process. For example, garnets from metamorphosed sedimentary rock in the Himalaya have grown during regional metamorphism following continental collision (Figure 6.2). Highly precise dating of the mineral garnet has shown that the core of a crystal is 30 Ma old, and that successively younger overgrowths were formed towards the outer parts of the garnet which are only 25 Ma old. This study indicates that during regional metamorphism rocks can be heated for at least five million years.

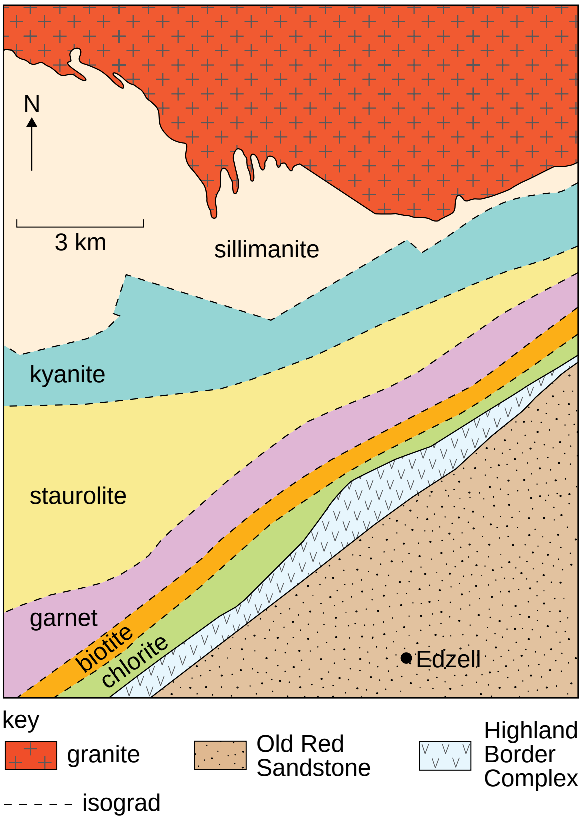


Figure 6.14 Geological sketch-map of the Barrow zones of Glen Esk (Angus), southeast Highlands of Scotland. Note that the Highland Border Complex and the Old Red Sandstone are part of younger sequences, largely unaffected by the regional metamorphism.

Our present understanding of the metamorphism at continental collision zones is possible because of the foundations laid by a handful of 19th century geologists. In 1893, George Barrow made a classic study of the regionally metamorphosed rocks of the Scottish Highlands. He discovered a 10–20 km thick sequence of sedimentary and volcanic rocks that had been highly deformed and metamorphosed about 470 million years ago during what is now known to have been a period of subduction and collision. Barrow mapped these rocks and studied their thin sections. By considering the sequence of metamorphosed mudstones and shales, often called **pelitic rocks** (or pelites), he was able to establish a series of zones, each characterised by a distinctive mineral (Figure 6.14).

Barrow was the first person to recognise metamorphic zones and to use index minerals to identify successive grades of metamorphism from the appearance of distinctive new minerals in a traverse across rocks of successively higher grade. Each new mineral represents a further step towards higher temperatures and/or pressures, and so becomes an index mineral indicative of metamorphic grade. He recognised six metamorphic zones in the metamorphic rocks of the southeast Highlands. The index minerals he based them on were (in order of increasing metamorphic grade) chlorite, biotite, garnet, staurolite (a brown iron–aluminium silicate, Figure 6.15), kyanite and sillimanite; of these minerals, kyanite is the most spectacular (Figure 6.4b).

Figure 6.15 Muscovite schist with large red garnet crystals (centre of image) and several dark staurolite porphyroblasts from Karakoram, Pakistan (field of view 35 mm across). Note twinned staurolite in lower left of image – staurolite takes its name from such cross-shaped twins.

What can we say about the conditions of pressure and temperature reflected in the appearance of these index minerals? Once again, the approach is very simple in principle – experiments are carried out at known pressures and temperatures and the reactions that take place are then observed. The change in *P*–*T* conditions between the kyanite and sillimanite zones, for example, can be interpreted by considering the experimentally determined phase diagram in Figure 6.4a.

■ How do the aluminosilicates recognised by Barrow in the southeast Highlands compare with those present in the metamorphic aureole in Figure 6.12?

☐ Andalusite is absent from the Barrow zones.

■ Do the minerals in the Scottish examples reflect higher or lower pressures than those in the metamorphic aureole?

☐ In Figure 6.4a, the phase boundary between kyanite and sillimanite exists only at pressures greater than 370 MPa, whereas the phase boundary between andalusite and sillimanite is at less than 370 MPa. Thus, the transition from kyanite directly to sillimanite in the Scottish rocks indicates that they were formed at higher pressures (>370 MPa) and therefore greater depths.

Question 6.5

Figure 6.16 is the pressure–temperature diagram for a reaction in which garnet is formed. Having argued that the pressure in this area was greater than 370 MPa, use Figure 6.16 to infer the approximate temperature at which garnet first appears, that is, at the beginning of the garnet zone.

In the case of contact metamorphism, the near-concentric mineral zones around an igneous body essentially represent zones of successively higher temperatures. Although the same principles apply, regional metamorphism is more complicated, since both pressure and temperature vary. In both types of metamorphism, however, the points of first appearance of an index mineral can be mapped and linked by a line. Provided such minerals are *in rocks of similar composition*, this line marks the position of rocks with ‘equal metamorphic grade’ and so is known as an **isograd**. Conventionally, an isograd is named after the index mineral on the side of *higher* metamorphic grade. Thus, the border between the kyanite and sillimanite zones in Figure 6.14 is marked by the appearance of sillimanite as an index mineral and is called the sillimanite isograd.

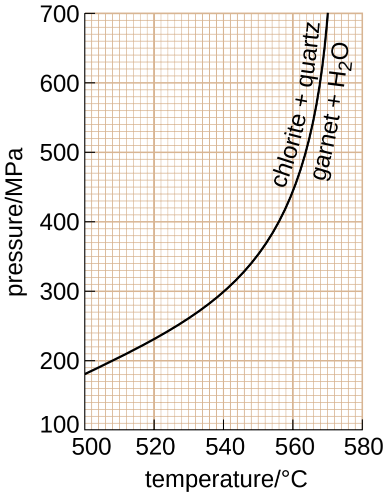


Figure 6.16 Pressure–temperature diagram for the reaction chlorite + quartz $\rightleftharpoons$ garnet + H_2O.

It could be argued from Figure 6.4a that the transition from kyanite to sillimanite does not *necessarily* involve an increase in temperature, but could reflect a decrease in pressure. This possibility can be excluded by a more detailed examination of the minerals on either side of the isograd – and, in general, sequences of metamorphic zones are determined by changes of temperature rather than by changes of pressure.

Question 6.6

Figure 6.17 is a geological sketch map illustrating some folded metamorphosed sediments (metasediments) and a gabbro intrusion.

(a) Sketch in and label the isograds between the different mineral zones.

(b) Is the metamorphism older or younger than the folding?

(c) Does the metamorphism appear to be related to the gabbro intrusion (i.e. is it an example of contact metamorphism)?

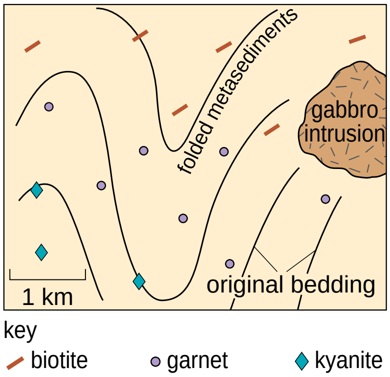


Figure 6.17 Geological sketch map for use with Question 6.6. Different symbols identify the index mineral found at various exposures.

The isograds you drew on Figure 6.17 may look convincing, but do not assume that they can be drawn as precisely as, for example, a topographic contour line. Rocks have complex chemical compositions involving many elements and thus represent messy chemical systems. None the less, several experienced geologists working in the same area would usually put an isograd at roughly the same place on the map.

The significance of isograds for the rate of temperature increase with depth, or geothermal gradient, during metamorphism may not be clear cut. In going from south to north across Barrow's zones, you are going from an area of low-grade metamorphism to one where both the temperatures and pressures were higher. Whether the two areas record metamorphism at the same time is a complex question beyond the scope of this course.

Finally in this section, we must reiterate the importance of rock composition in determining which metamorphic minerals will crystallise. In a sequence of pelites, calcic and mafic rocks, different minerals will grow in each rock type and so there will be different index minerals in each rock type. You have seen that garnet, staurolite, kyanite and sillimanite can be used as important index minerals to establish metamorphic zones; they are all rich in aluminium, and they occur in Al-rich metamorphosed sediments (pelites). Fewer mineralogical changes occur in carbonates and mafic igneous rocks and so pelites are generally more useful in establishing metamorphic zones. However, it is important to realise that if, as is common, metamorphic zones are based on index minerals found in pelites, the zones may well include rocks of other compositions. So there is nothing wrong with talking about mafic igneous rocks being in the staurolite zone, even though they contain no staurolite.

Question 6.7

Figure 6.18 illustrates a series of isograds based on minerals in pelitic rocks. The isograds cut across a series of calcic, pelitic and mafic igneous rocks. Use Table 6.1 to indicate what mineral assemblages you would expect to find at points A, B, C and D.

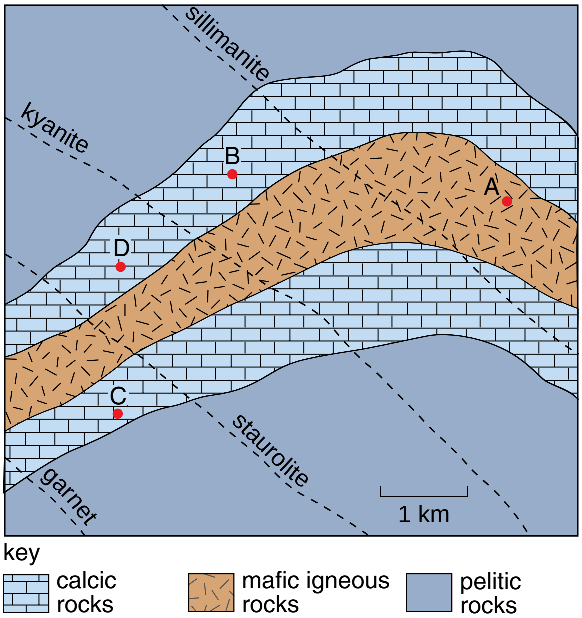


Figure 6.18 Isograds (based on pelitic rock minerals) cross-cutting a sequence of calcic, pelitic and mafic igneous rock types. (For use with Question 6.7.)

Table 6.1 Metamorphic assemblages produced in pelitic, calcic (originally carbonate-rich) and mafic igneous rocks.

	Assemblage produced (index mineral in italics)		
Metamorphic zone (based on pelites)	**Pelites**	**Calcic rocks (calcite may also be present)**	**Mafic igneous rocks**
garnet	*garnet*, mica, quartz, Na-rich plagioclase	garnet, epidote, amphibole	chlorite, albite, epidote
staurolite	*staurolite*, garnet, quartz, mica, Na-rich plagioclase	garnet, anorthite, amphibole	amphibole, plagioclase
kyanite	*kyanite*, garnet, quartz, mica, Na-rich plagioclase	garnet, anorthite, amphibole	amphibole, plagioclase
sillimanite	*sillimanite*, garnet, quartz, mica, Na-rich plagioclase	garnet, pyroxene, anorthite	amphibole, plagioclase, garnet

6.4 Metamorphic facies

The mapping of metamorphic zones using index minerals is very useful for describing metamorphism from a particular area. However, it is difficult to compare the metamorphic grades of different areas on the basis of index minerals because of the possibility of different bulk compositions in the two areas, which will result in different metamorphic index minerals even under similar conditions of pressure and temperature. **Metamorphic facies** is a term coined by the Finnish geologist Pentti Eskola (1883–1964) to embrace *all* the possible metamorphic mineral assemblages produced in rocks of different composition at similar temperatures and pressures. The assignment of a rock to a metamorphic facies is based on the observed mineral assemblage, which corresponds to a particular P–T range of metamorphism.

This definition requires amplification:

1 The composition of a metamorphic rock determines its mineral assemblage under any particular conditions of temperature, pressure and fluid composition. Thus, given a chemical analysis of a rock, it should be possible to predict the mineral assemblage that is likely to be stable under a given set of physical conditions (Table 6.1).

2 A metamorphic facies is determined from the *observed mineral assemblages* in associated rocks: it does *not* require knowledge of the precise conditions of temperature and pressure, which are estimated subsequently from a petrogenetic grid.

The first point you have met before, but the second is important because it often causes confusion. You have seen how metamorphic conditions may be estimated from experimental results on particular metamorphic reactions. Moreover, when data on a number of reactions that occur under different conditions are plotted together on a P–T diagram, a petrogenetic grid is formed, which is a criss-cross of different phase boundaries that allows likely metamorphic conditions to be inferred from particular mineral assemblages. But such a petrogenetic grid is obviously experimentally determined, and thus, although its general form is unlikely to change significantly, it will be continually modified as experimental techniques develop and more experiments are carried out, particularly in the presence of fluid phases of different compositions. A metamorphic facies, by contrast, is assigned to the rock from the observed mineral assemblages. Provisional limits of temperature and pressure may then be assigned to the particular metamorphic facies from the petrogenetic grid as currently understood.

Let us take an example that you have already met. In Barrow's area of the Scottish Highlands, pelitic rocks in the higher grades of metamorphism contain sillimanite, whereas the calcic rocks contain garnet and pyroxene. Although the minerals found in the two types of rock are different because the rocks have different chemical compositions, both assemblages belong to the same metamorphic facies. In areas of lower-grade metamorphism, pelites contain chlorite and biotite, and the calcic rocks contain calcite, epidote (a Ca-rich chain silicate) and an amphibole, both rock types belonging to the same metamorphic facies.

■ What metamorphic zone would the mineral garnet indicate if found in a calcic rock?

□ From Table 6.1, garnet is present in all metamorphic zones in calcic rocks. It does not therefore help to define the particular zone.

■ Will a garnet-bearing rock of pelitic composition always be from the same metamorphic facies as a garnet-bearing rock of a calcic composition?

□ No. Because they can be from different metamorphic zones, they can also be formed at different pressures and temperatures. If so, the metamorphic facies may be different as well.

Today, geologists recognise eight principal metamorphic facies that cover temperatures from 100 °C to over 800 °C, and pressures from atmospheric pressure to over 1400 MPa (Table 6.2). One of these (eclogite) you have met before in

Section 5.4.4. There, it was defined as a particular rock type; a metamorphic rock of mafic composition with garnet and pyroxene present (Figure 6.19). Here, we are using the term **eclogite facies** to indicate the *P*–*T* field over which such an assemblage is stable (i.e. a metamorphic facies). Although it is not necessary to define each facies here, the facies names generally result from the stable assemblage found in rocks of mafic composition. For example, **blueschist facies** rocks contain blue minerals, such as the amphibole, glaucophane, only if the bulk composition is mafic (Figure 6.20). A pelite, when metamorphosed under blueschist-facies conditions, will be made up largely of white mica and quartz, but neither of these is blue! None the less, if we know that it formed under these conditions it would be correct to say it is a blueschist-facies rock.

Table 6.2 Summary of metamorphic facies and their likely environments of formation. Zeolite, prehnite and pumpellyite are all low-temperature calcic aluminosilicates.

Facies	Environment
zeolite	buried sediments
prehnite–pumpellyite	more deeply buried sediments
blueschist	relatively high pressure but very low temperature: subduction zones
greenschist	extremely widespread facies, low-grade regional metamorphism in orogenic areas
amphibolite	widespread facies, high-grade regional metamorphism in orogenic areas
granulite	highest grade of regional metamorphism, occurs in the lower crust and within some Precambrian cratons
hornfels	metamorphic aureoles: moderate to high temperatures but at low pressures
eclogite	very high pressures and moderate to high temperatures

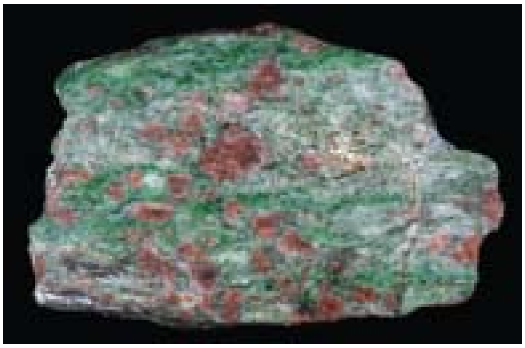
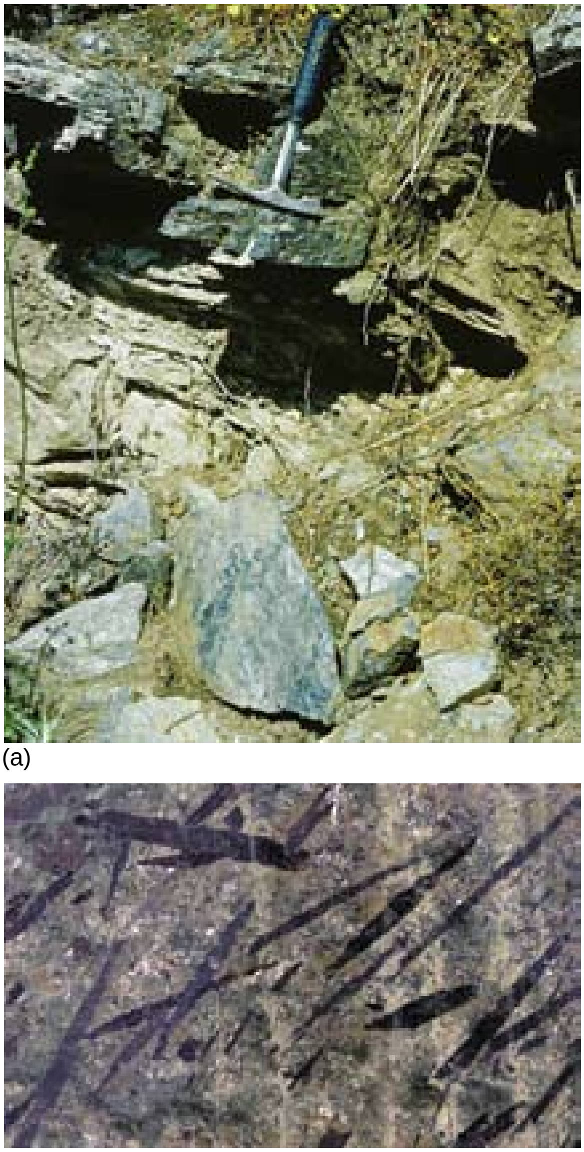


Figure 6.19 Eclogite, a spectacular, high-density rock from Norway. The principal minerals are garnet (pink) and a sodium-rich pyroxene (green) called jadeite (width of image = 150 mm).



(a)

(b)

Figure 6.20 (a) A blueschist-facies locality from western Turkey. The blue tinge results from the presence of glaucophane. (b) Polished block (width = 5 cm) of blueschist with glaucophane porphyroblasts; also present are muscovite, garnet and epidote. Glaucophane can be observed using the Virtual Microscope by viewing the blue amphibole present in the metamorphic rock specimen ‘Amphibolite’.

Our present estimates of the *P*–*T* fields that characterise the eight metamorphic facies are reproduced in Figure 6.21. Broadly, metamorphic facies may be linked to estimates of pressure and temperature – and so to the change in temperature with depth known as the geothermal gradient. For example, hornfels results from contact metamorphism, which is characterised by high but localised geothermal gradients due to the associated magmatic activity. Very low geothermal gradients occur when cold material from the Earth's surface is taken down to lower crust or upper mantle depths. In between, there are more 'normal' geothermal gradients, which characterise much of the Earth's continental crust.

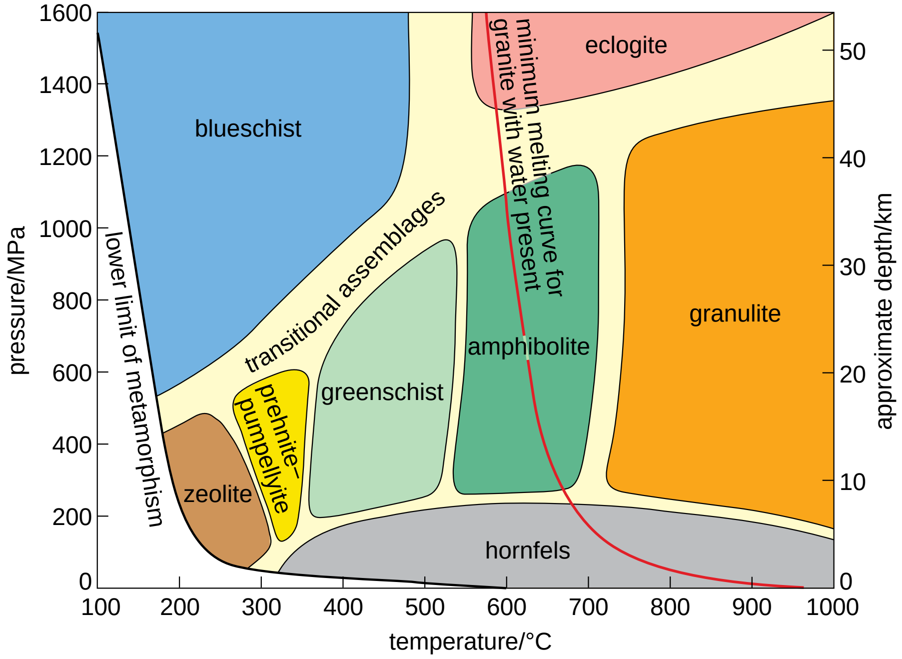


Figure 6.21 The *P*–*T* fields of the major metamorphic facies.

■ If you plot a typical continental geothermal gradient of 30 °C km^{-1} on Figure 6.21, what sequence of metamorphic facies would you expect to find with increasing depth in the Earth's crust? Assume a surface temperature of 0 °C.

☐ The geotherm starts at 0 °C at the surface and increases by 300 °C for every 10 km of depth. Near the surface, it passes through **zeolite** and **prehnite–pumpellyite facies**, then successively **greenschist** and **amphibolite facies**, reaching **granulite facies** at a depth of about 25 km.

This observation raises two important issues. First, the occurrence of rocks indicative of metamorphic facies not in this sequence suggests that the Earth's geothermal gradient can be significantly disturbed. Second, how and why are rocks formed in higher pressure metamorphic facies subsequently brought back up to the surface? One explanation is that their exhumation results from isostasy in response to unusual thicknesses of continental crust. As an example, take Barrow's sillimanite isograd, which was formed at a pressure of 680 MPa, equivalent to a depth of ~25 km during metamorphism. At present, these rocks are exposed on top of a 35 km-thick crust. The simplest interpretation is that during the metamorphism that piece of crust was at least 25 + 35 km, or 60 km, thick – compared to an average thickness for continental crust of 30–35 km. Since then, the thick crust has been uplifted, the top 25 km has been eroded, and the sillimanite isograd exhumed and exposed at the surface.

■ The metamorphic facies diagram covers a very wide range of temperatures and pressures. Other than metamorphism, what other effects might occur at very high temperatures?

☐ Clearly, if rocks are heated to a high enough temperature, they are going to melt. On Figure 6.21, the red line represents the melting curve for rocks of granitic composition when water is present (it is the water-saturated solidus, see Figure 5.3). At pressures greater than 400 MPa, melting takes place at about 600 °C, in the presence of H_2O.

Melts formed during regional metamorphism may rise to shallower levels and even cause their own metamorphic aureole. Thus, the effects of thermal and regional metamorphism may cause complex changes in metamorphic grade both horizontally and vertically.

6.5 Plate tectonics and metamorphism

The sequence of metamorphic facies recorded during deformation of the Earth's crust reflects the complex interaction between available heat sources and structural evolution. Some associations of metamorphic facies are indicative of particular tectonic settings and we shall examine two of these in a little detail.

6.5.1 Metamorphism and subduction

In general, the dominant heat source for regional metamorphism is supplied internally by the decay of heat-producing elements (Section 2.4). The geothermal gradient $\frac{\Delta T}{d}$ varies from one tectonic environment to another, but is related to the **heat flow** (q) by the equation:

$$q = \frac{K \Delta T}{d} \qquad (6.5)$$

where K is the **thermal conductivity** of the rocks and ΔT is the change in temperature that occurs over a depth range d. Heat flow is a measure of the amount of energy passing through unit surface area in unit time, usually expressed in milliwatts per square metre, mW m^{-2}.

Over much of the Earth's surface, the heat flow is surprisingly uniform at about 60 mW m^{-2}. Striking exceptions occur near mid-ocean ridges and island arcs where the heat flow is high, and above oceanic trenches at convergent plate boundaries where it is low. This is illustrated in a profile of the heat flow observed across the western Pacific in Figure 6.22.

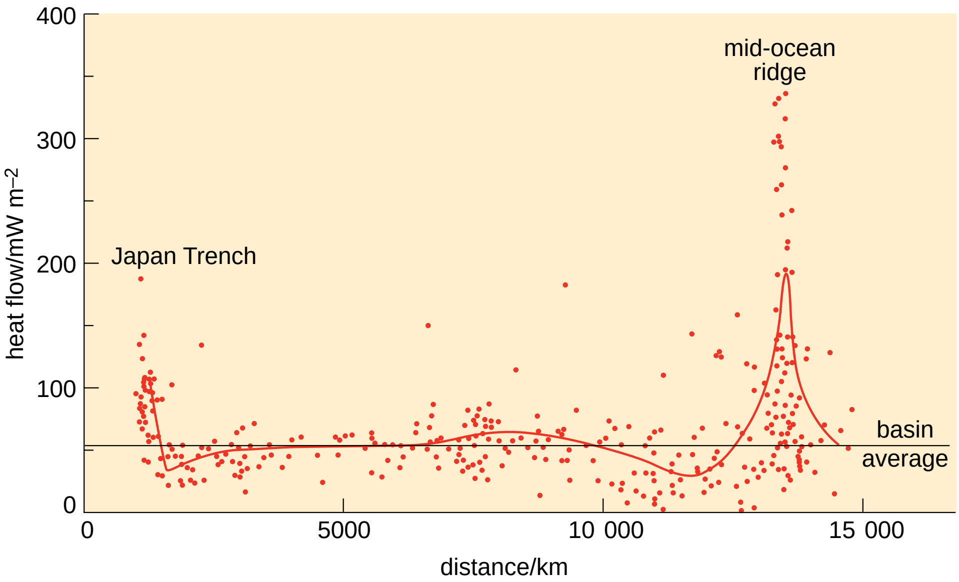


Figure 6.22 Variation in surface heat flow, measured across the western Pacific Ocean. The curved red line represents the average local heat flow, and the horizontal line illustrates the average heat flow from all ocean basins.

The low heat flow values on the flanks of the mid-ocean ridge in Figure 6.22 are due to cooling caused by the circulation of considerable volumes of seawater through the top few kilometres of the newly formed ocean crust.

■ Why are the heat flow values so much higher in Japan and on the mid-ocean ridge?

☐ These are both areas where considerable volumes of magma are intruded. Magmas are hot, and as they move up towards the surface they bring heat with them, thereby increasing the heat flow.

■ Assuming that the thermal conductivity does not change significantly, will the geothermal gradient be higher or lower in areas of higher heat flow?

☐ Higher, since the geothermal gradient is proportional to the heat flow (Equation 6.5).

■ What might be the cause of the low heat flow values in the Japan Trench?

☐ Such trenches mark the site of active subduction (see Figure 2.4). The ocean crust has been at or near the surface. It is therefore relatively cold, and as oceanic lithosphere is subducted into the mantle the isotherms are also dragged down resulting in a low geothermal gradient and thus low heat flow to the surface (Figure 6.23).

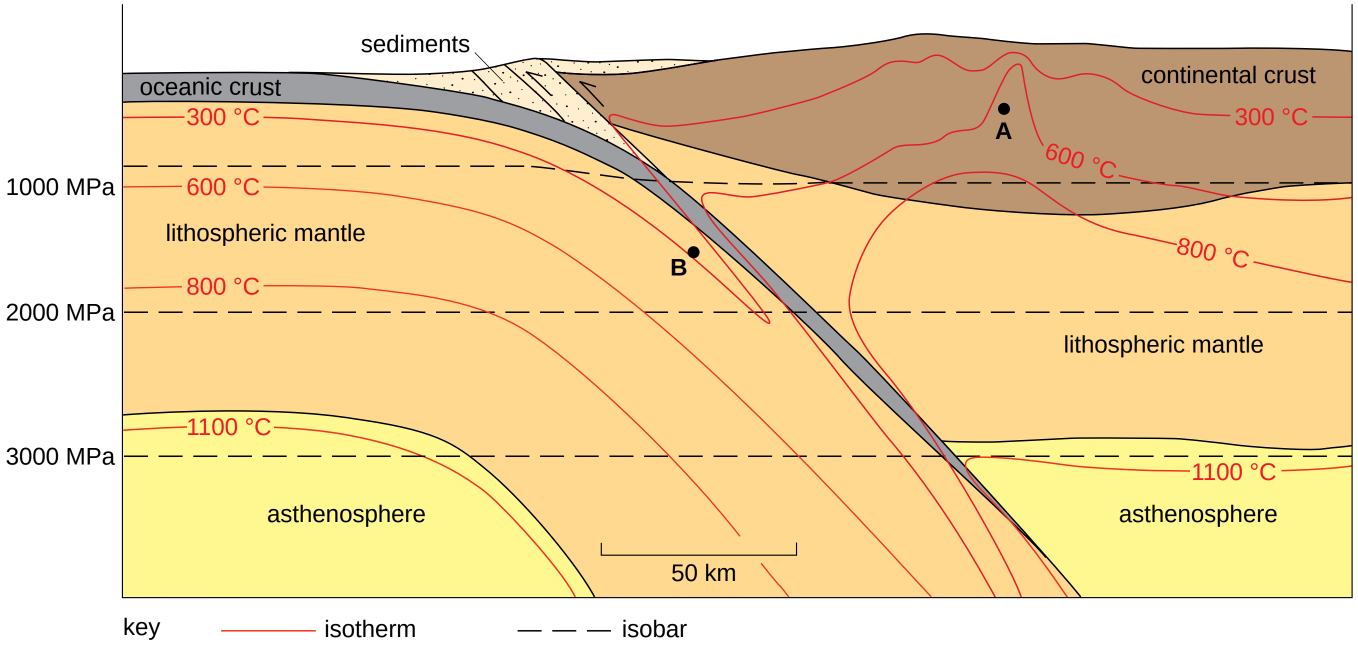


Figure 6.23 Cross-section through a convergent plate boundary illustrating high temperature–low pressure metamorphism (at point A) and high pressure–low temperature metamorphism (at point B). Together these make up a paired metamorphic belt, which runs parallel to the plate boundary.

In general, magmatic provinces are characterised by high heat flow and high geothermal gradients. Moreover, convergent plate boundaries would appear to be unique in that they contain a zone of high geothermal gradients (where the magmas are intruding) adjacent to one where the temperatures have been reduced by subduction (Figure 6.23). The net result is that in addition to the distinctive igneous rocks (Section 5.4) there should also be two roughly parallel suites of metamorphic rocks, one containing relatively high-temperature and low-pressure mineral assemblages, associated with magmatism in the overriding plate, and the other typical of relatively high pressures and low temperatures, associated with subduction of the downgoing slab. This is called a **paired metamorphic belt** and is characteristic of convergent plate boundaries.

Question 6.8

Using Figures 6.21 and 6.23, what sort of geothermal gradient and what metamorphic facies will be present at points A and B on Figure 6.23?

Paired metamorphic belts identified from around the Pacific were first explained simply in terms of the varying geothermal gradients that are associated with subduction zones (Figure 6.23). The presence of blueschist assemblages (at point B) requires not only that very high pressures occurred at low temperatures during active subduction, but also that the rocks were then uplifted rapidly to the surface before they had time to warm up. However, blueschist-facies rocks are not always preserved in ancient subduction zones. In some examples from subduction zones, the initial low-temperature/high-pressure minerals warm and recrystallise as the rocks are brought to the surface, resulting in the loss of blueschist-facies minerals and the preservation of only higher-grade metamorphic facies. Throughout the metamorphic history of any rock in a subduction (or collision) zone, there is a competition between the rate of heating imposed largely by radioactive decay of heat-producing elements and the rate of cooling imposed by the rock being brought towards the surface during exhumation. In the case of blueschists, their high *P*–low *T* assemblage is preserved only when the rate of cooling due to exhumation (controlled largely by isostasy) exceeds the rate of heating (controlled by the thermal conductivity of the rocks).

6.5.2 Metamorphism and continental collision

The destruction of an oceanic basin by complete subduction of the oceanic crust that formerly separated two continents is likely to be followed by continental collision and crustal thickening. Such momentous events lead to metamorphism on a regional scale.

■ What is the heat source responsible for regional metamorphism during crustal thickening?

☐ Most of the heat within the Earth is generated by the radioactive decay of U, Th and K, and these elements are most abundant in the rocks of the continental crust, particularly the uppermost crust. So a thicker crust means a steeper geothermal gradient.

The changing geothermal gradient during continental collision can best be understood by considering the effects of thrusting one crustal slice over another, hence thickening the crust. Initially the rocks will heat up, followed by cooling as the rocks are brought to the surface. For a simple model in which a segment of upper continental crust is thrust onto another with a similar geothermal gradient, the immediate result following thrusting is a distinctive ‘saw-tooth’ variation of temperature with depth (Figure 6.24a; Figure 6.24b, curve 1). At this instant, the pressure in the lower segment has increased, but the temperatures remain as yet unchanged. This clearly cannot last since temperatures of near 0 °C cannot be maintained in what is now well within a thickened segment of continental crust overlain by hot rocks. Moreover, the continuing radioactive decay of U, Th and K, which are now present in larger quantities since there is more upper crustal material in the vertical section, gradually increases the temperature, until, after about 30 Ma, its variation with depth is illustrated by curve 2 of Figure 6.24b. Predictably enough, a rock at a depth of 17 km on Figure 6.24b (near the top of the underlying piece of crust) has had its temperature increased considerably from 200 °C to 800 °C. What is perhaps more surprising is that even in the upper segment of crust the geothermal gradient has also been increased by the emplacement of radioactive crust beneath it.

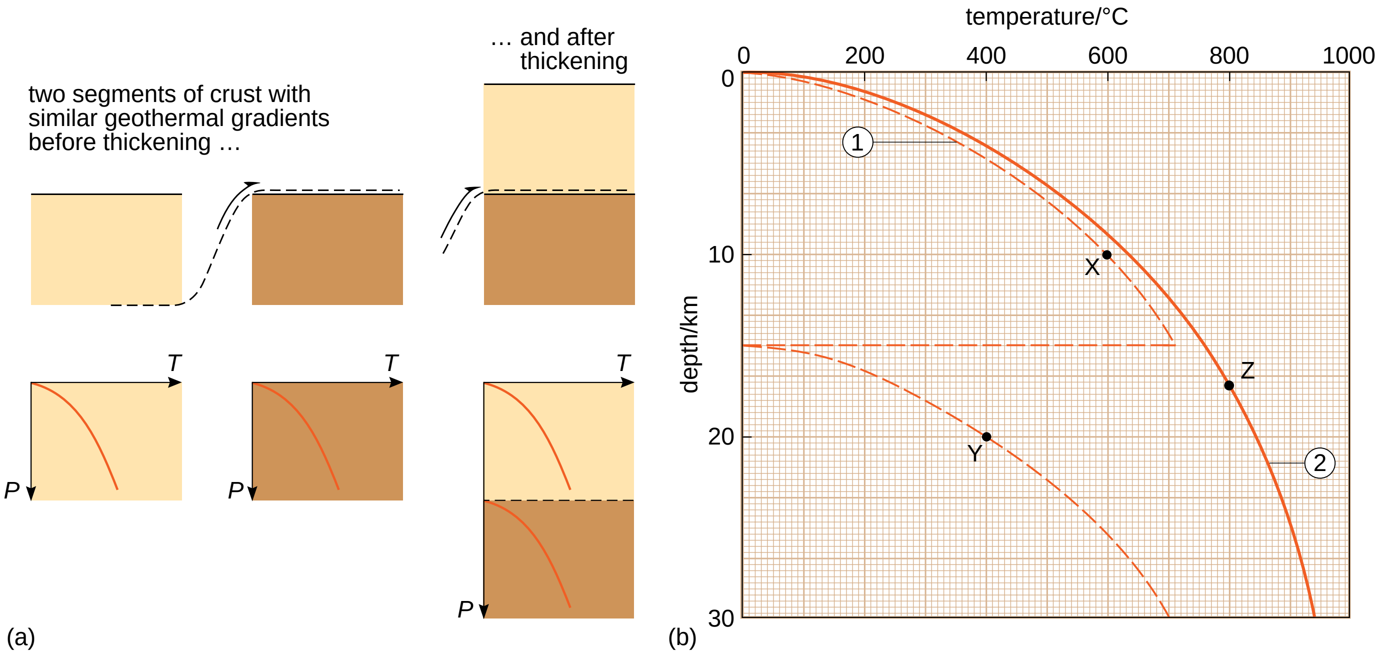


Figure 6.24 (a) Two segments of crust with similar geothermal gradients (shown as curved red lines) before and immediately after crustal thickening. (b) Temperature variation with depth: curve 1 (dashed), immediately after thrusting; curve 2 (solid line) after 30 Ma. Points X, Y and Z are discussed in the text.

Now we shall consider what happens as rocks are brought to the surface. Because of isostasy the thickened crust will be susceptible to exhumation, and rocks must cool as they approach the surface. In some cases, this cooling effect is rapid enough to prevent the geothermal gradient from reaching the equilibrium temperature gradient indicated as curve 2 in Figure 6.24b. If the rocks are exhumed very rapidly after thrusting, then the metamorphic facies appropriate to the *P*–*T* conditions at points X and Y along curve 1 of Figure 6.24b will be preserved.

■ From Figure 6.21, what are the metamorphic facies of rocks metamorphosed at X and Y?

☐ X lies in the amphibolite facies (600 °C at 10 km) and Y in the greenschist facies (400 °C at 20 km).

Both greenschist and amphibolite facies are represented in the Barrow zones, but what is unusual in Figure 6.24 is that the colder rocks (at Y) occur at greater depths than the hotter rocks (at X). Such a temperature inversion has been observed in metamorphic rocks of the Himalaya and its preservation has been interpreted as evidence of very rapid exhumation, of the order of 5–10 mm per year.

If, as is more usual, the geothermal gradient reaches equilibrium along curve 2 of Figure 6.24b, rocks deep in the thickened crust will reach temperatures of 800 °C or more (at Z), well within the granulite facies. In this case, the diagnostic feature of the collision zones is that the temperatures will increase more rapidly with depth than in crust of normal thickness. Many of the granulite-facies rocks exposed today result from the heating of a crust thickened by collision, and subsequent exhumation of deep crustal levels by isostasy. In the UK, granulites are best exposed in the Lewisian terrane of northwest Scotland, where they formed 2800 Ma ago.

This chapter has mainly been concerned with changes in mineralogy that result from heating rocks at different pressures. However, there are also aspects of metamorphic rocks that relate to deformation, particularly where rocks are folded or faulted. Such mechanical processes form the topic of the next three chapters.

6.6 Summary of Chapter 6

1. Metamorphism is the mineralogical response of a rock to imposed conditions of temperature and pressure that differ markedly from those at which the original rock formed.
2. Broadly, the study of metamorphic rocks comprises: (a) evaluation of the conditions of temperature and pressure to which the rocks were subjected during metamorphism, in the light of experimental data; (b) consideration of the thermal evolution of an area and how that may be explained in terms of its tectonic environment.
3. Metamorphic grade is defined in terms of the entropy change (ΔS) of particular metamorphic reactions: increasing metamorphic grade involves an increase in entropy of the metamorphic system. In general, higher-grade rocks reflect higher temperatures. Prograde reactions are those that take place with increasing metamorphic grade, and retrograde reactions are those that take place with decreasing metamorphic grade. Rates of reaction increase exponentially with temperature, and so equilibrium is more likely to be achieved at higher grades of metamorphism. The presence of a fluid phase will also greatly increase the rates of reaction and its absence will inhibit retrograde reactions.
4. The petrogenetic grid is the pattern of experimentally determined phase boundaries plotted on a phase diagram of temperature against pressure. In principle, once the mineral assemblage in a metamorphic rock has been identified, the petrogenetic grid may be used to estimate the conditions of pressure and temperature under which it crystallised.
5. Dynamic metamorphism occurs within localised areas of intense deformation, such as along fault or shear zones. Contact metamorphism refers to changes in response to heat in the vicinity of igneous rocks. Typically it is not associated with deformation, and the metamorphic grade decreases systematically away from the contact with igneous rocks. Regional metamorphism tends to occur over large areas, and as it is usually associated with collision or subduction zones, the rocks will have been deformed.

6 The particular minerals observed in a metamorphic rock reflect its composition, in addition to the conditions of pressure and temperature and the nature of the fluid phase. Thus different minerals occur in rocks of the same metamorphic grade but of different compositions. The term metamorphic facies is used to embrace all possible mineral assemblages in rocks of different composition that are thought to have crystallised under similar P–T conditions. A metamorphic facies is identified from the observed mineral assemblages, and not from the inferred conditions of pressure and temperature, which are estimated from the petrogenetic grid.

7 The mineral assemblages in metamorphic rocks from different areas can reflect different variations in temperature and pressure and these vary depending on the thermal, and hence tectonic, environment. Geothermal gradients are steep in areas of magmatic activity (e.g. island arcs and mid-ocean ridges), and low at subduction zones. Continental collision and crustal thickening also result in an increased gradient, due primarily to the radioactive decay of U, Th and K within the crust. However, the preserved mineral assemblages depend not only on the geothermal gradient during metamorphism, but also on the rate at which the rocks are brought to the surface.

6.7 Objectives for Chapter 6

Now you have completed this chapter, you should be able to:

6.1 Carry out observations on hand specimens of metamorphic rocks, identify metamorphic minerals and textures and interpret these in terms of the processes involved.

6.2 Interpret simple phase diagrams to infer the conditions of temperature and/or pressure at which a particular metamorphic mineral, or assemblage, was at equilibrium.

6.3 Map out isograds and use them to assess the relationship of metamorphism to both the intrusion of an igneous magma and a deformation event.

6.4 Evaluate estimates of metamorphic pressure and temperature from metamorphic assemblages to suggest possible tectonic environments in which the metamorphism could have taken place.

Now try the following questions to test your understanding of Chapter 6.

Question 6.9

A fairly common metamorphic reaction is that of:

$$\text{staurolite} + \text{quartz} \rightleftharpoons \text{garnet} + \text{sillimanite} + H_2O$$

Predict which assemblage has the lower entropy, which is stable at higher grades of metamorphism, and whether this reaction takes place at a staurolite or sillimanite isograd.

Question 6.10

A metamorphosed pelite contains the mineral staurolite in equilibrium with kyanite. Use the following information to estimate the approximate conditions of pressure and temperature during metamorphism:

(a) Staurolite does not form until the temperature exceeds 580 °C in pelitic rocks at pressures greater than 200 MPa.

(b) No granite rocks or migmatites were formed nearby; pelite rocks melt at about 600–650 °C under moderate pressures. You can assume H_2O was present during metamorphism.

(c) The phase relations are as illustrated in Figures 5.3 and 6.4a.

Question 6.11

The inferred metamorphic conditions of temperature and pressure of two rocks from different areas are 200 °C at 800 MPa, and 550 °C at 200 MPa. Using Figure 6.21, for each rock, identify which metamorphic facies it is from, calculate the geothermal gradient (assuming that, for each 1 km in depth, pressure increases by 30 MPa) and suggest in what thermal or tectonic environment it might have been metamorphosed.

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Entries in **bold** represent glossary terms. Page numbers in *italics* refer to figures and tables..