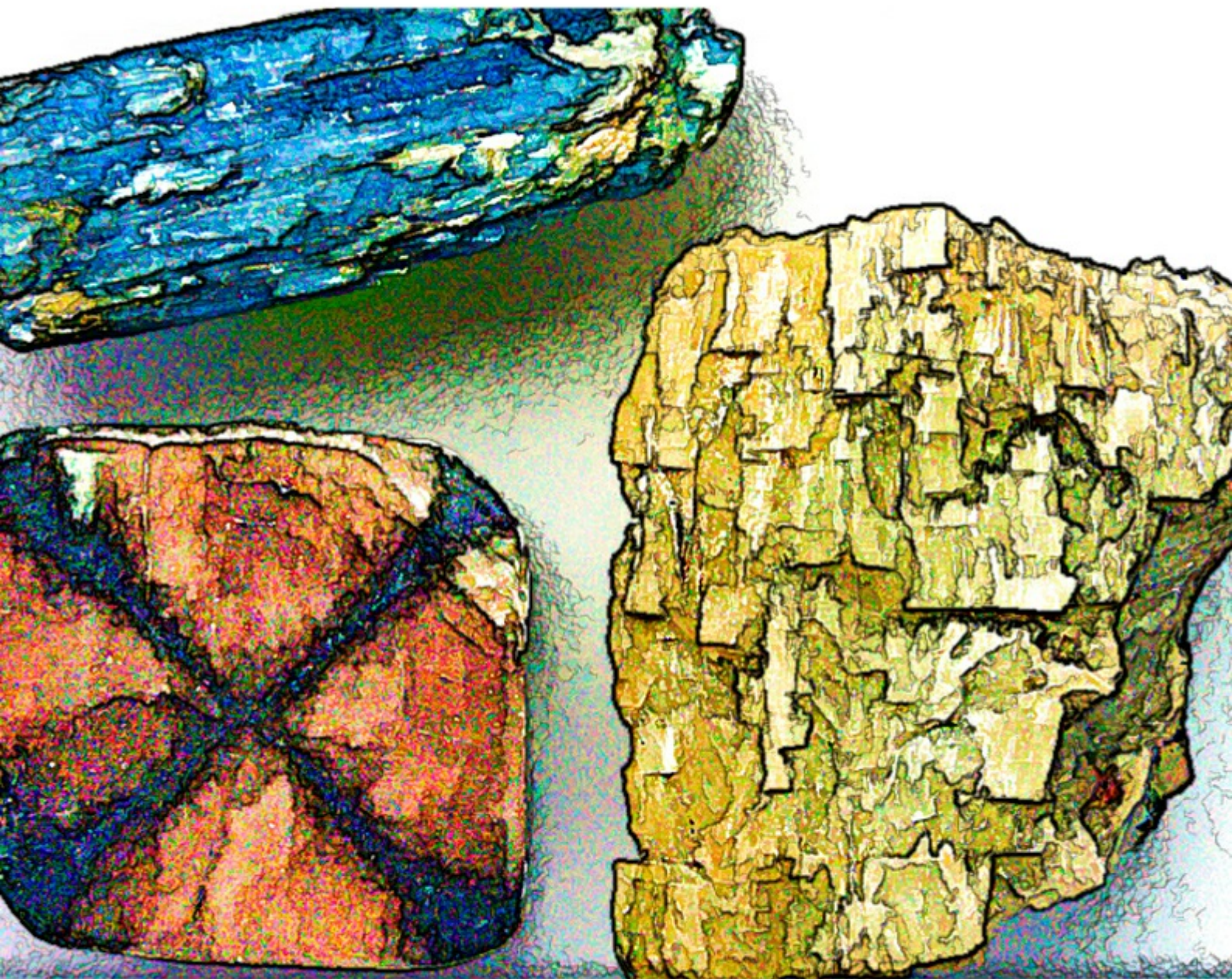


Minerals and Rocks

J. Richard Wilson



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Minerals and Rocks

1st edition

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Preface

The first version of this text was written to serve as lecture notes for a first term geology course in “Minerals and Rocks” at Aarhus University in Denmark in 2003. In Aarhus this course is accompanied by a general “Introduction to Geology” course that presents, for example, the structure of the Earth, plate tectonics and paleontology. These topics are therefore not treated here, and some knowledge of the Earth’s structure and plate tectonics is assumed.

After a brief introduction to the Rock Cycle, this text presents the physical properties of minerals and an introduction to crystallography. The most important rock-forming minerals are then dealt with in a systematic way, followed by the three main rock groups – igneous, sedimentary and metamorphic.

A lecture course in Minerals and Rocks must be accompanied by a parallel sequence of practical classes where, for example, crystal structures, minerals and rock types dealt with in theory are demonstrated in practice.

J. Richard Wilson

June 2014

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1 Introduction

The solid part of the Earth is made up of rocks. Rocks are made up of minerals. A mineral is a naturally occurring inorganic solid. It has a specific chemical composition and a characteristic crystal structure. Quartz is a very common mineral. Most beach sand is composed of quartz. It has the composition SiO_2 and forms elongate 6-sided crystals that terminate at a point (Picture 1.1).



Picture 1.1: Quartz crystals.

Pyrite is also a naturally occurring mineral that forms cubic crystals (Picture 1.2). It is also known as iron pyrites and has the composition FeS_2 .



Picture 1.2: Pyrite cube.

Rocks are naturally occurring, coherent solids consisting of an aggregate of minerals. Glass may be present in some volcanic rocks. There are three main groups of rocks, classified on the basis of how they formed:

- IGNEOUS
- SEDIMENTARY
- METAMORPHIC

1.1 Igneous rocks

Igneous rocks form by the solidification of melts. Molten rock is called *magma*. The most common magma (basaltic magma) forms as a result of partial melting of the Earth's mantle. Magma formed in the mantle rises towards the surface because it has a lower density than the surrounding mantle rocks. If the magma reaches the surface, a volcano is formed. Magma at the surface of the earth is called lava. Volcanic eruptions can produce mainly lava, but some eruptions are explosive and produce large volumes of ash and other fragmentary volcanic rocks, such as pumice. Igneous rocks that form at the surface of the earth are *volcanic rocks*. Because volcanic rocks cool rapidly they are fine-grained. They may contain some large crystals but the matrix is always fine-grained. Lava may cool so fast that crystals do not have time to nucleate and grow and glass forms. Magma below the surface of the earth contains dissolved fluids – mostly water. As magma rises to the surface the pressure decreases and some of the fluids escape as gas. Evidence of escaping gas is common in volcanic rocks in the form of bubbles that are called vesicles.

Magma formed in the mantle may not reach the surface but accumulate in a magma chamber in the crust. Magma here cools slowly and the resulting rock is entirely crystalline and (relatively) coarse-grained. Igneous rocks that form below the surface are called *plutonic rocks*.

1.2 Sedimentary rocks

All rocks exposed at the surface of the earth are subject to *weathering*. Rocks break into fragments that are transported by wind, ice and water and can be deposited to form a sediment. Small fragments are transported further than large fragments. During weathering many of the original minerals break down to produce clay minerals. An important mineral that does not break down is quartz. As they become buried, loose sediments (sand, silt, clay) become *consolidated* and form compact rocks – sedimentary rocks. An example is consolidated clay that is called *shale*. Other sedimentary rocks form as a result of the *precipitation* of minerals from water; rock salt is an example. A wide variety of life forms exist in sedimentary environments and sedimentary rocks often contain evidence of life in the form of *fossils*. Fossils and fragments of fossils can accumulate to form *limestone*. One of the most obvious features of sedimentary rocks is layering.

1.3 Metamorphic rocks

All rock types can be subjected to elevated temperature and/or pressure conditions. For example, sedimentary rocks near a magma chamber will be heated. The magma may have a temperature in the vicinity of 1200°C. Sedimentary rocks close to the chamber will be heated more than those further away. If shale is heated the clay minerals break down to form new minerals that are stable at higher temperatures. The original sedimentary rock changes its mineralogy and structure because it has been heated as a result of its proximity to an igneous rock – it has become *thermally* or *contact metamorphosed*. Rocks below the surface of the earth may be subjected to deformation at the same time as they are heated in what is known as *regional metamorphism*. Regionally metamorphosed rocks develop a *foliation* – a new layered structure – which is evident in, for example, schists and gneisses.

1.4 The Rock Cycle

Sedimentary and volcanic processes take place at (or very near) the surface of the earth. Plutonic and metamorphic processes take place below the surface. The three groups of rocks: igneous, sedimentary and metamorphic, are related to each other by the Rock Cycle (Fig. 1.1). Igneous rocks at the surface of the earth are subjected to weathering and erosion. Material derived from this is deposited to form sediments. The loose sediment consolidates to form a sedimentary rock. This sedimentary rock becomes buried and subjected to heating and/or deformation – it becomes metamorphosed. Metamorphism can be so intense that the rock begins to melt and form an igneous rock. This can be exposed at the surface and the process continues.

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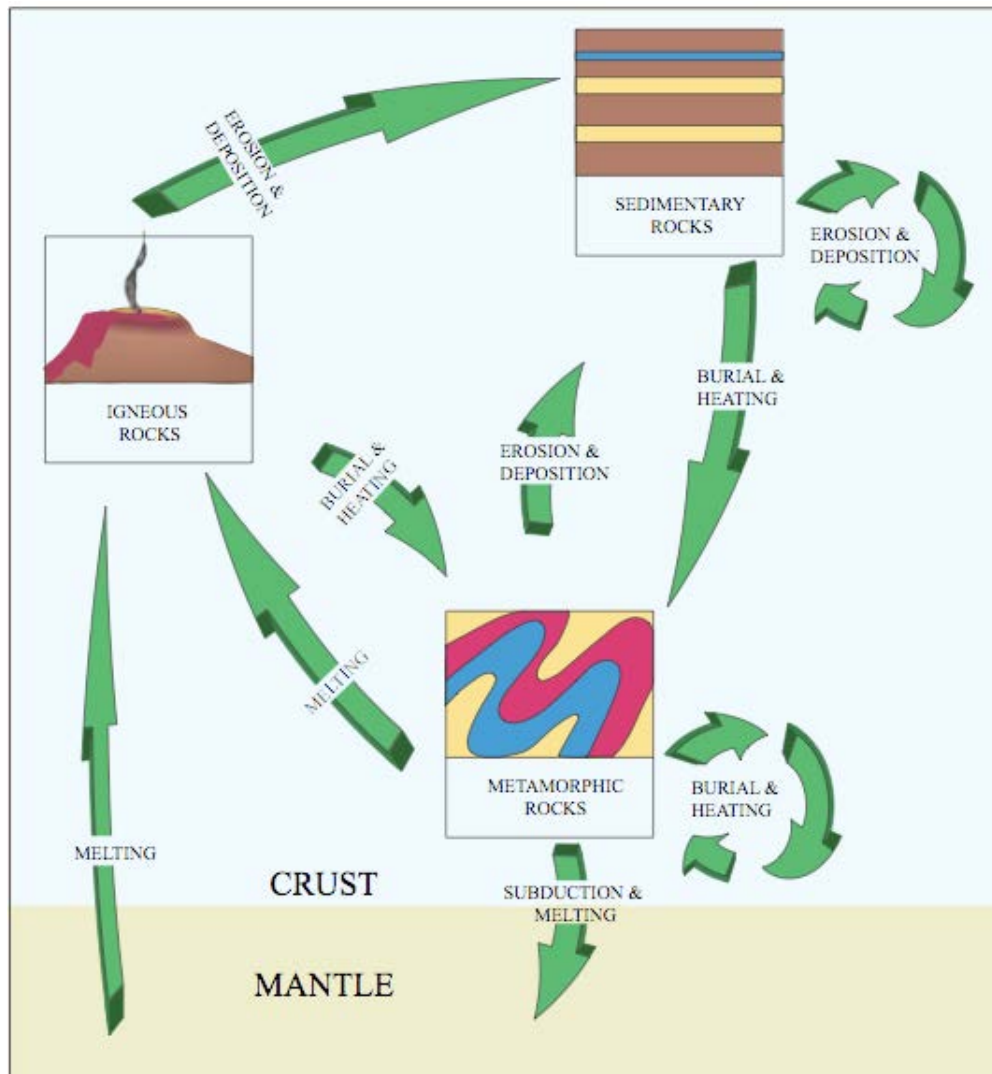


Fig. 1.1: The Rock Cycle

The cycle does not always follow the central circular course. Sedimentary rocks can be weathered and eroded and form new sediments. Metamorphic rocks can likewise form sedimentary rocks. Rocks can be metamorphosed more than once. Igneous rocks can be metamorphosed. While some igneous rocks are formed as a result of the solidification of melts produced by high-grade metamorphism, most are formed as a result of the partial melting of the mantle.

2 Minerals – an Introduction

2.1 Definition of a mineral

A *mineral* is a naturally formed solid; liquids and gases are therefore excluded. A mineral has a *specific composition*. It may be a naturally occurring element e.g. gold (Au), copper (Cu) or diamond (C). Minerals are, however, usually compounds e.g. quartz (SiO_2), pyrite (FeS_2) or olivine ($\text{Mg,Fe}_2[\text{SiO}_4]$) in which the ratio of metal (Fe + Mg) to silicate group [SiO_4] is 2:1. Minerals possess a characteristic *crystal structure* and therefore exclude non-crystalline materials such as glass. For example, *quartz* is a mineral. It is a common naturally formed phase with a simple composition: SiO_2 . It forms elongate, 6-sided crystals that terminate at a point (Picture 1.1). Quartz is an essential component of, for example, *granite* which is a rock. Granite consists of several minerals. It is dominated by feldspar and quartz (light coloured minerals) but also contains a minor amount of dark minerals.

ROCKS are composed of **MINERALS**

To learn about rocks we have to start by understanding minerals.

2.1.1 Polymorphs

Each mineral has a unique crystal structure. The same chemical composition can, however, in some cases develop more than one crystal structure. For example *carbon* (C) occurs in nature in two forms:

GRAPHITE – hexagonal crystals, very soft (marks paper), density 2.3 g/cm^3

DIAMOND – cubic crystals, hardest known mineral, density 3.51 g/cm^3

Diamonds occur in nature in rocks that were formed at extremely high pressure. To convert graphite to diamond in the laboratory requires a pressure of ca. 25 kilobars (25.000 times atmospheric pressure) at low temperature, increasing to 100 kb at about 2500°C (Fig. 2.1). To form diamonds in nature therefore requires a pressure of $>30 \text{ kb}$ which is equivalent to a depth of about 100 km below the surface of the Earth.

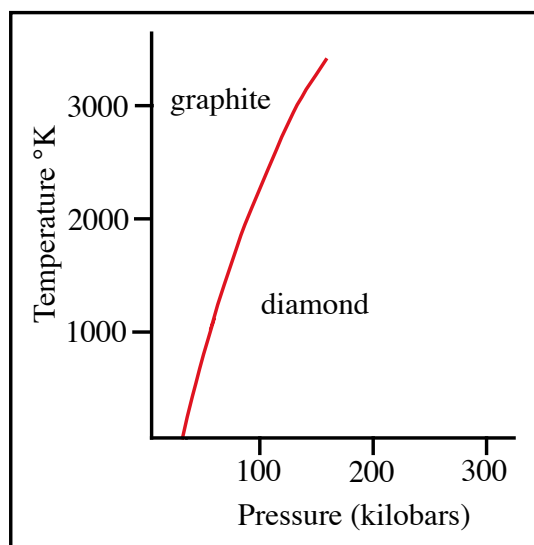


Fig. 2.1: Pressure – temperature phase diagram for carbon

The stable phase of carbon at pressures below ~30 kb is graphite, but (luckily for the diamond industry) the conversion from diamond to graphite at low temperatures is extremely slow.

2.2 Properties of minerals

These are determined by the composition and crystal structure of the mineral. Relevant mineral physical properties include external shape, cleavage, lustre, colour, hardness and density. Some minerals have a characteristic taste; others are magnetic. The ability of a mineral to react with dilute HCl is a commonly used chemical property.

2.2.1 Crystal form, growth habit and twinning

As you will learn in *crystallography*, there are seven *crystal systems*. These are:

Crystal system	mineral examples
cubic	garnet, pyrite, halite (rock salt), magnetite
tetragonal	zircon
hexagonal	beryl
trigonal	calcite
orthorhombic	olivine, topaz
monoclinic	gypsum, orthoclase (K-feldspar)
triclinic	plagioclase (Ca-, Na-feldspar)

The *3-dimensional* appearance of a mineral is referred to as its “habit”. This is, of course, largely influenced by the crystal structure. For example, a hexagonal mineral can form long 6-sided needles or short “stubby” crystals. Both are hexagonal but they have different “habits”. Common descriptive terms include:

flakes	thin plates (typical of mica minerals)
needle-like	thin, long crystals
tabular	flattened but not “flaky”
prismatic	elongate – but not needle-like
fibrous	typical of e.g. asbestos minerals

Crystals of a mineral often occur together in an *aggregate*. These crystal aggregates can develop in a wide variety of forms:

radiating	e.g. pyrite
skeletal	e.g. native copper
reniform (kidney-shaped)	e.g. hematite
rosette	e.g. corundum

A feature shown by many crystals is a symmetrical intergrowth referred to as *twinning*. Twins can be of a variety of types including:

- contact twins* where it looks as if part of a crystal has been rotated (e.g. *swallow tail* twins in gypsum). The two individuals are separated by a definite surface (Fig. 2.2a).
- penetration twins* consist of two interpenetrating individuals, such as is common in staurolite (where a cross-like structure is formed) (Fig. 2.2b) and in K-feldspar (where it is called Carlsbad twinning)
- repeated* (also called multiple or polysynthetic) *twins* are made up of three or more parts twinned in the same fashion. This is very common in plagioclase feldspar.

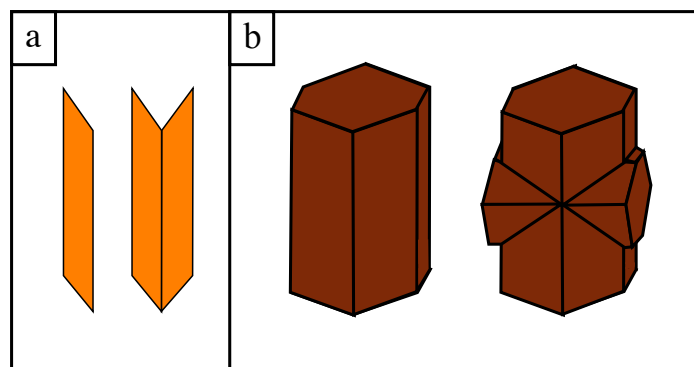


Fig. 2.2: Some common examples of twinning

a) A single (monoclinic) crystal of gypsum is shown on the left. The two “halves” of a gypsum twin are joined along a plane in a contact twin. b) Staurolite commonly develops penetration twins where it appears that two crystals have “grown into” each other.

2.2.2 Cleavage and fracture

This is the tendency for a mineral to break in preferred directions along planar surfaces.

Cleavage planes are breakage surfaces that are controlled by crystal structure but are not to be confused with crystal faces which are growth surfaces. Some examples are:

Mica minerals have a perfect basal cleavage – they break into thin flakes which reflect their layered crystal structure

Calcite (CaCO_3) breaks into rhombohedral fragments

K-feldspar ($\text{K}[\text{AlSi}_3\text{O}_8]$) breaks along two perpendicular surfaces to give near-rectangular fragments

Some minerals do not have well-developed cleavage and break in other characteristic ways. Quartz, for example, is brittle and develops characteristic curved or conchoidal fractures.



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2.2.3 Lustre

The quality and intensity of light reflected from a mineral is referred to as lustre. Two very important groups of minerals are those with *metallic* and those with *non-metallic* lustre. Quartz (Picture 1.1) has non-metallic lustre whereas pyrite (Picture 1.2) has metallic luster. Various types of non-metallic lustre are:

adamantine	sparkling, like diamond (and sometimes zinc blende)
vitreous	glassy, e.g. quartz
pearly	as shown by the very soft mineral, talc
silky	shown by many fibrous minerals
greasy	commonly as a result of surface alteration. Characteristic for nepheline ($\text{Na}[\text{AlSiO}_4]$)

2.2.4 Colour

This is commonly the most striking feature of a mineral, but can be misleading. For example, quartz (SiO_2) is ideally colourless and transparent. Varieties of quartz are, however:

milky quartz	white, full of minute fluid inclusions
bluish quartz	minute inclusions of ?rutile
amethyst	transparent purple (traces of iron)
rose quartz	pink (TiO_2 impurity)
citrine	yellow (colloidal $\text{Fe}_2(\text{OH})_3$)
smoky quartz	brown or grey (Al substituting for Si)

These are all trigonal and consist of >>99% SiO_2 . For a mineralogist these are all “impure quartz” – but they make valuable semi-precious gem stones. The mineral corundum (Al_2O_3) occurs in the gem varieties, ruby (red) and sapphire (blue). The colour of granite largely depends on the colour of the alkali feldspar which comprises 30–50% of the rock. This can be white, pink, red, beige, pale green etc.

2.2.5 Streak

A mineral in powder form may have a different colour from individual crystals. This can be an important feature for identification. It is not practical to crush minerals to identify them, so we use a colourless porcelain plate which we scratch with the mineral. The colour of the streak is then observed. This method is only valid for minerals that are softer than the porcelain plate.

The mineral *pyrite* (FeS_2) (Picture 1.2) has a metallic lustre and is brassy yellow in colour. It has a density of about 5 gm/cm^3 and a hardness of about 6 (see below). It has, however, a black streak. Pyrite is also known as “fools gold” because it was commonly mistaken for this naturally occurring precious metal. But gold has a density of 19.3 gm/cm^3 , is much softer (hardness = 3) and has a yellow streak.

2.2.6 Hardness

This is a mineral's relative resistance to scratching. *Mohs' scale of relative hardness* (Table 2.1) is divided into 10 steps, each marked by a single mineral. Any mineral on the scale will scratch all minerals below it. The softest mineral with hardness = 1 is *talc* (used in talcum powder); the hardest with hardness = 10 is *diamond*. Common minerals on the scale include calcite with H = 3, K-feldspar with H = 6 and quartz with H = 7. For practical purposes common objects have the following hardness values: finger nail 2.5; copper coin 3.5; pocket knife, glass 5.5.

Mohs' Scale of Relative Hardness		
Number in scale	Mineral name	Hardness of some common objects
10	Diamond	
9	Corundum	
8	Topaz	
7	Quartz	
6	K-feldspar	
		Pocket knife; glass
5	Apatite	
4	Fluorite	
		Copper coin
3	Calcite	
		Finger nail
2	Gypsum	
1	Talc	

Table 2.1: Mohs' scale of relative hardness of common minerals

2.2.7 Tenacity

This is the resistance that a mineral offers to breaking, crushing, bending or tearing i.e. its cohesiveness. Commonly used terms are:

brittle	breaks or powders easily (e.g. quartz)
malleable	can be hammered out into thin sheets
ductile	can be drawn into a wire
flexible	bends but does not resume its original shape on pressure release
elastic	resumes its original shape after bending (e.g. mica)

The terms malleable, ductile and flexible are mostly relevant for metallic minerals.

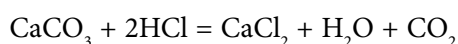
2.2.8 Density

Density, mass per unit volume, is measured in gm/cm^3 . We use the term specific gravity (G) which is the ratio of the weight of a substance to the same volume of water. Since this is a ratio there are no units for G. This is often a very useful criterion for mineral identification. Some useful reference values are:

Gold (Au)	19.3	Galena (PbS)	7.6
Pyrite (FeS_2)	5.2	Halite (NaCl)	2.2
Calcite (CaCO_3)	2.7	K-feldspar ($\text{K[AlSi}_3\text{O}_8]$)	2.6
Olivine ($\text{Mg,Fe}_2[\text{SiO}_4]$)	3.2–4.4	Quartz (SiO_2)	2.6

2.2.9 Other properties

Other useful properties can be relevant in special cases. For example, halite, also called rock salt (NaCl), has a characteristic *taste* and magnetite (Fe_3O_4) is strongly *magnetic*. Calcite (CaCO_3) reacts readily with cold hydrochloric acid (HCl) to release bubbles of carbon dioxide:



Many minerals are birefringent. This is usually only observable with the aid of a microscope, but in calcite double refraction is visible in hand specimen – you can see “double” through transparent crystals.

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3 Crystallography

Crystallography is the science related to crystals. The word is derived from the Greek “crystallus” which means ice. The term was used in ancient Greece for colourless quartz which was believed to be “fossilised” ice. The most obvious feature of a quartz crystal (Picture 1.1) is that it is defined by a series of flat surfaces formed during growth and must, therefore, reflect the internal structure of the material. This feature is shared by many materials and the term used for them is *crystals*.

A vital feature of crystals of the same type is that the *angle* between the crystal faces is identical, no matter how large or small the crystal is. This law of the “constancy of interfacial angles” was demonstrated by Nicolaus Steno in 1669. Minerals are crystalline. When a mineral grows from a melt (magma) it forms crystals with perfect crystal faces. As more crystals grow and the amount of melt decreases the crystals begin to collide with each other. When all the melt has crystallised the solid rock consists of the minerals that formed but their crystal faces will seldom be preserved. Even though the crystalline nature of minerals may not be obvious in a solid rock, it can usually be confirmed with the aid of a microscope and always using X-ray diffraction.

3.1 Symmetry

The shapes illustrated in Fig. 3.1 have *symmetry planes*. The two sides of the face in Fig. 3.1a, are identical. They are the mirror image of each other. The face has one symmetry plane. The diamond in Fig. 3.1b has two symmetry planes, one horizontal and one vertical. The shape in Fig. 3.1c has four symmetry planes.

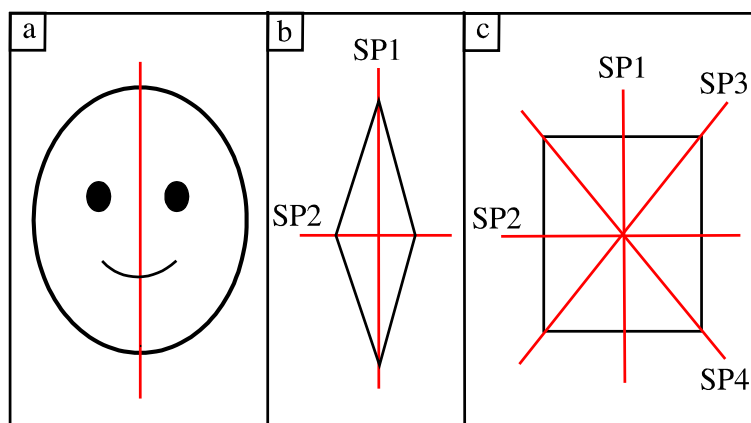


Fig. 3.1: Planes of symmetry in two-dimensional objects
a) contains one symmetry plane; b) contains two; c) has four (SP1 to SP4).

In addition to two symmetry planes, the diamond shape in Fig. 3.1b also has a *symmetry axis*. When it is rotated 180° about an axis through the centre (perpendicular to the plane of the paper) the same image is achieved as before rotation (Fig. 3.2). The axis is called “two-fold” because rotation through 360° repeats the same shape twice. The “distorted diamond” in Fig. 3.3 has a two-fold axis of symmetry but no symmetry planes.

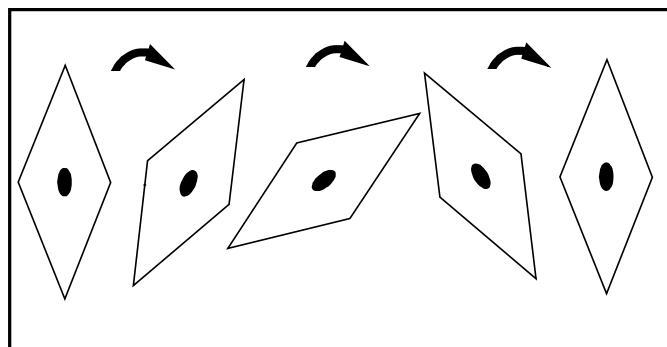


Fig. 3.2: Axis of symmetry

Two-fold symmetry axis in the diamond shape from Fig. 3.1b.

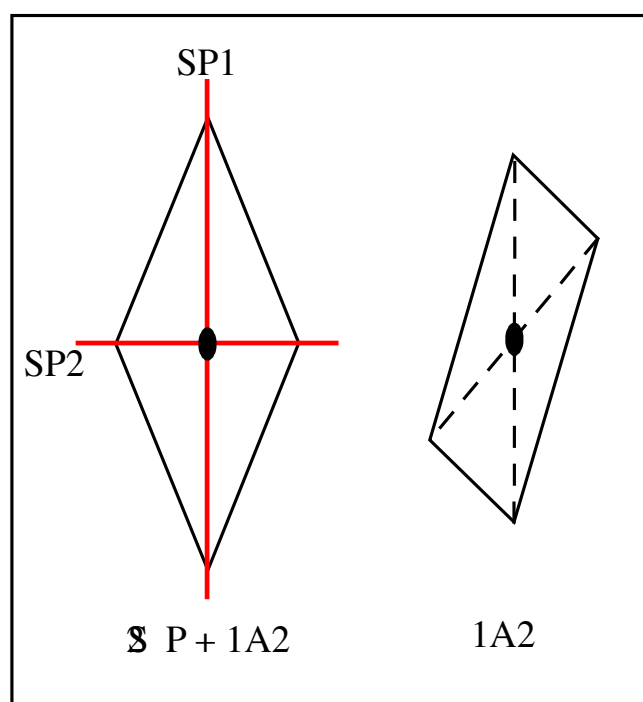


Fig. 3.3: Symmetry planes and axes

The diamond shape (left) has two symmetry planes and a two-fold symmetry axis. The “distorted diamond” (right) has a two-fold symmetry axis but no planes of symmetry.

The rectangle in Fig. 3.4a has a two-fold axis (A_2), the triangle (Fig. 3.4b) has a three-fold axis (A_3), the square (Fig. 3.4c) has a four-fold axis (A_4), and the hexagon (Fig. 3.4d) a six-fold axis (A_6). It is easy to construct figures with 5-fold, 7-fold etc. axes, but these do not occur in crystals. Only 2-, 3-, 4, and 6-fold axes are present in crystals (or none at all).

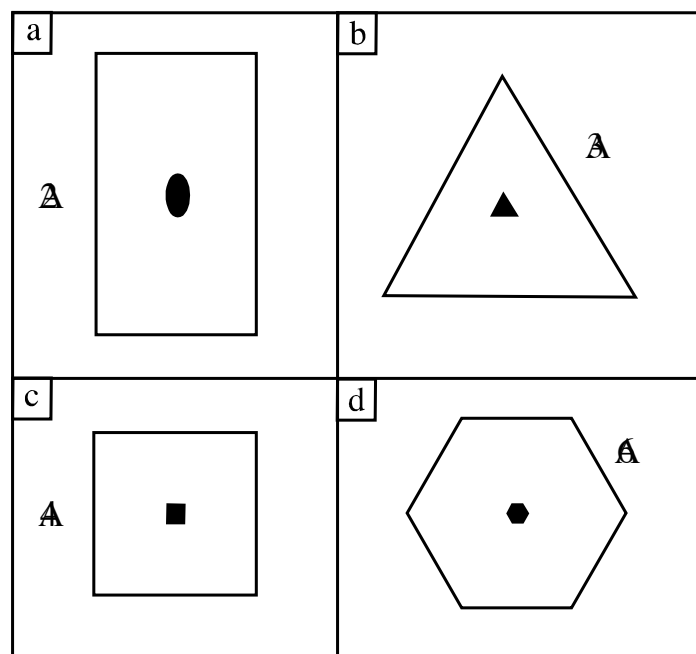


Fig. 3.4: Symmetry axes in crystals

These are restricted to: a) two-fold symmetry axis; b) three-fold symmetry axis; c) four-fold symmetry axis; d) six-fold symmetry axis.

A third form of symmetry is called a *centre of symmetry*. A crystal has a centre of symmetry when like faces and edges are arranged in pairs on opposite sides of a central point. It is obvious that the “brick” shape in Fig. 3.5a has a centre of symmetry, but the shape to the right (Fig. 3.5b) does not (neither does a pyramid, for example).

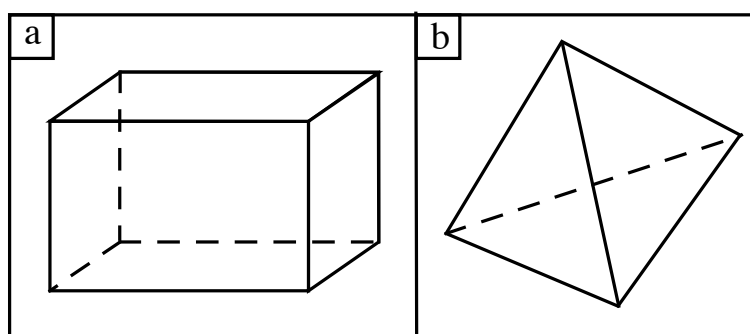


Fig. 3.5: Centre of symmetry

a) Each face of the “brick” on the left has a parallel partner on the opposite side; it has a centre of symmetry (C). b) The figure on the right does not have a centre of symmetry.

Symmetry elements are therefore:

SYMMETRY PLANES	SP
SYMMETRY AXES	A2 A3 A4 A6
CENTRE OF SYMMETRY	C

Several symmetry elements are often combined in crystals. The shape in Fig. 3.6 has a four-fold symmetry axis (A_4) and four two-fold symmetry axes ($4A_2$). There is also a total of 5 symmetry planes and a centre of symmetry.

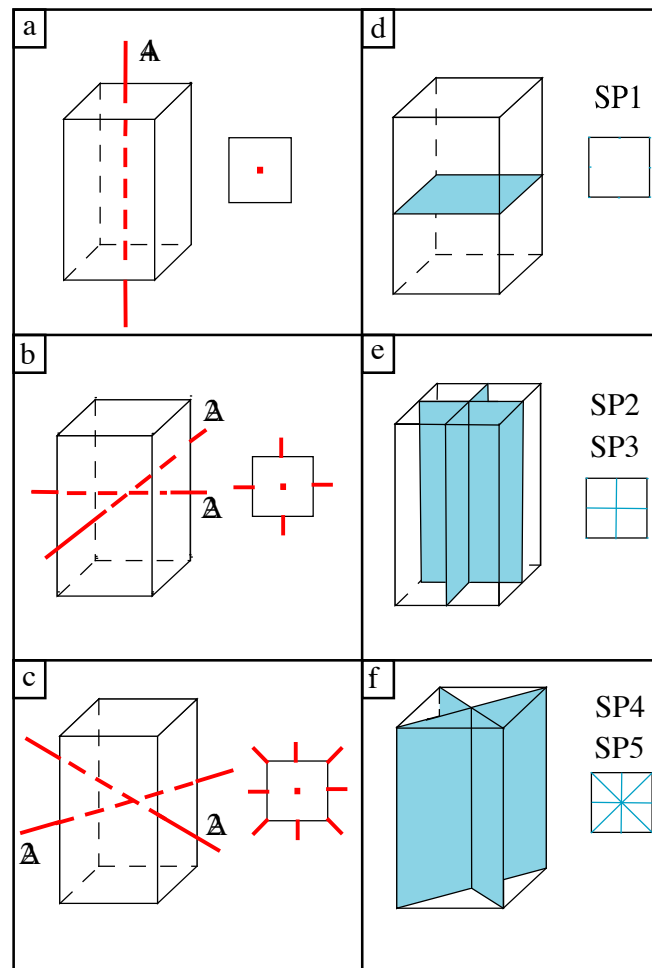


Fig. 3.6: Symmetry elements for a tetragon

a) There is one A_4 axis of symmetry. b) There are two A_2 axes passing through the centres of the large sides. c) There are two A_2 axes passing through the centres of the long edges. d) There are five planes of symmetry (SP1 to SP5). SP1 is parallel with the base and top of the tetragon and is therefore not visible in the two-dimensional form to the right in d). There is also a centre of symmetry.

The total symmetry for the shape in Fig. 3.6 is therefore: **1A4 4A2 5SP C**

This is the maximum symmetry in the *tetragonal crystal system*. Tetragonal minerals include zircon ($Zr[SiO_4]$).

3.2 Crystal systems

There are 7 crystal systems, each of which has particular minimum symmetry elements (Table 3.1).

CRYSTAL SYSTEM	MINIMUM SYMMETRY
CUBIC	4 A ₃
TETRAGONAL	1 A ₄
HEXAGONAL	1 A ₆
TRIGONAL	1 A ₃
ORTHORHOMBIC	3 A ₂
MONOCLINIC	1 A ₂
TRICLINIC	NO SYMMETRY

Table 3.1: Minimum symmetry elements developed in the 7 crystal systems

To aid the description of crystals a coordinate system is used whose “0”-point (origin) is located at the centre of the crystal. The three axes are called the a-, b- and c-axes.

In the *orthorhombic* system the three axes are perpendicular to each other (i.e. all at 90° to each other) and a, b and c are of different lengths (Fig. 3.7).



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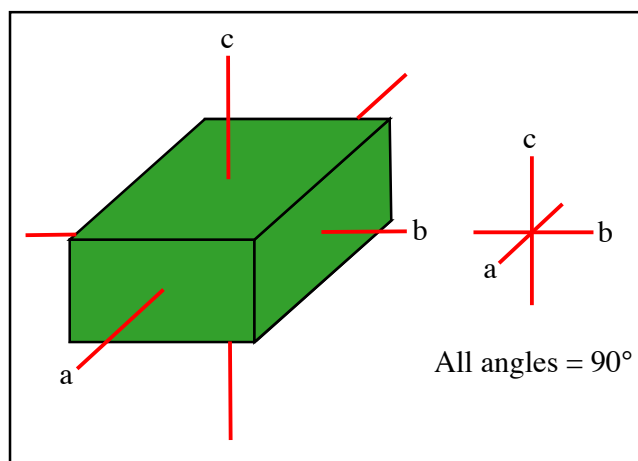


Fig. 3.7: Crystallographic axes in the orthorhombic system
There are three axes at 90° to each other and of unequal lengths (a , b and c) in the orthorhombic crystal system.

In the *tetragonal* system the three axes are again perpendicular to each other, but two of them have the same length (Fig. 3.8). The convention is to place the c -axis as the unique vertical axis (the A_4 symmetry axis) and the two others are identical with each other (and may be called a_1 - and a_2 -axes rather than the a - and b -axes).

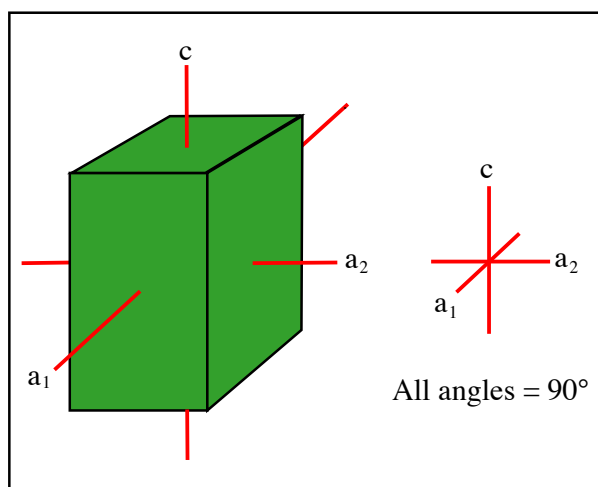


Fig. 3.8: Crystallographic axes in the tetragonal system
The c -axis is the A_4 symmetry axis. The other two axes are at 90° to c and each other and of equal lengths (a_1 and a_2).

In the *cubic* system (Fig. 3.9) the three axes are mutually identical; each one represents an A_4 symmetry axis. They can be called a_1 -, a_2 - and a_3 -axes rather than the a -, b - and c -axes.

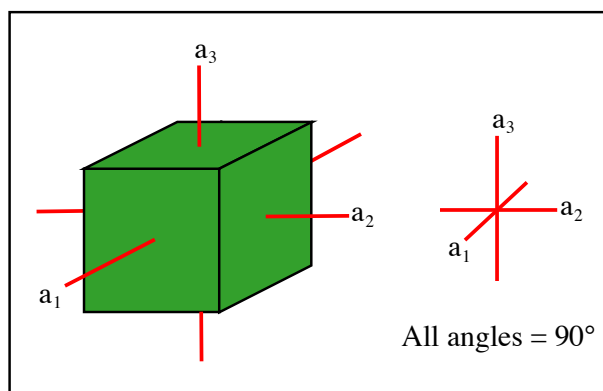


Fig.3.9: Crystallographic axes in the cubic system

The three axes are interchangeable – of equal lengths and at 90° to each other (a_1 , a_2 and a_3).

The *monoclinic* system can be obtained by “distorting” the orthorhombic symmetry in one plane (Fig.3.10). The angle between the a- and c-axes is no longer 90° . The obtuse angle between the a- and c-axes is referred to as β . In Fig.3.10, $\beta = 120^\circ$. The value of β of course varies from mineral to mineral. The angles between the a- and b-axes, and between the c- and b-axes, remain 90° , as in the orthorhombic system.

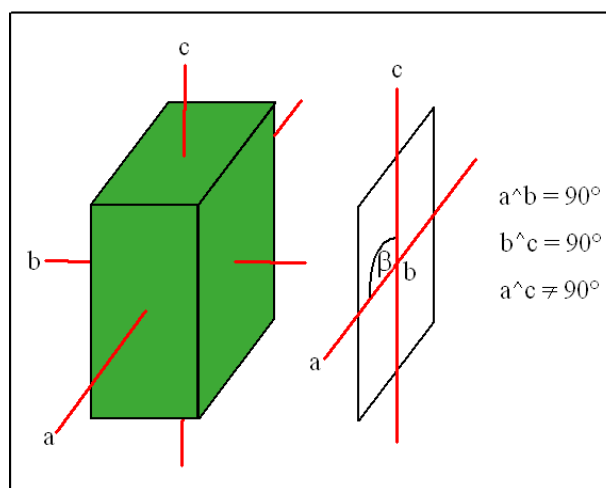


Fig.3.10: Crystallographic axes in the monoclinic system

The b-axis is identical with the A2 symmetry axis. The c-axis is at 90° to the b-axis. The a-axis is at 90° to the b-axis but not at 90° to the c-axis. On the right it is clear that the a-axis is parallel with top and bottom sides of the shape that consequently are only cut by the c-axis. The size of β varies from mineral to mineral.

In the *triclinic* system a monoclinic crystal is distorted (Fig.3.11). This results in none of the angles between the a-, b- and c-axes being 90° .

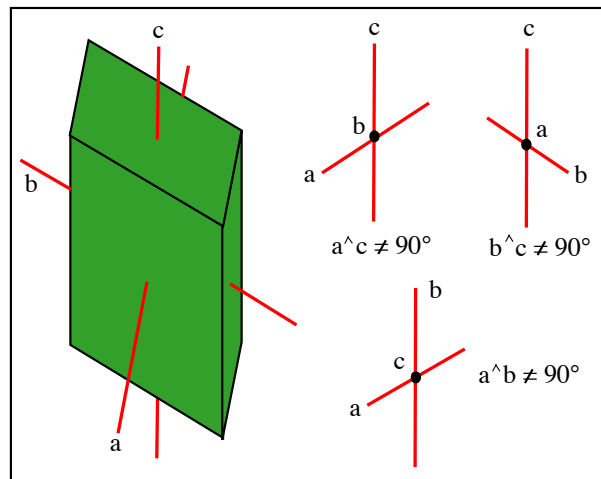


Fig.3.11: Crystallographic axes in the triclinic system

None of the angles between the three axes (a, b and c) are 90° in the triclinic system.

In the *trigonal* system, which has an A_3 axis of symmetry, it is more convenient to use four crystallographic axes rather than three. The unique A_3 axis of symmetry is the c-axis. The other three axes are identical and arranged at 60° to each other; they are called the a_1 -, a_2 - and a_3 -axes (Fig.3.12).

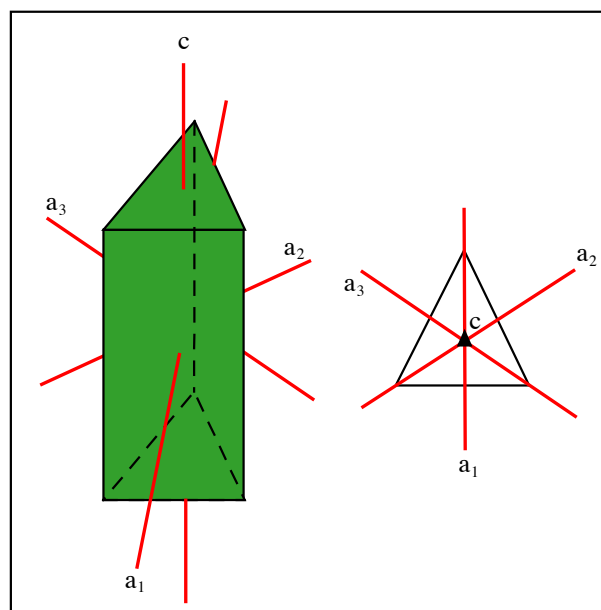


Fig.3.12: Crystallographic axes in the trigonal system

The c-axis is the A_3 symmetry axis. The three identical a-axes (a_1 , a_2 and a_3) are at 90° to the c-axis and at 60° to each other.

In the *hexagonal* system the A_6 symmetry axis is the c-axis. Otherwise the convention is the same as for the trigonal system with three identical a-axes arranged at 60° to each other (Fig.3.13).

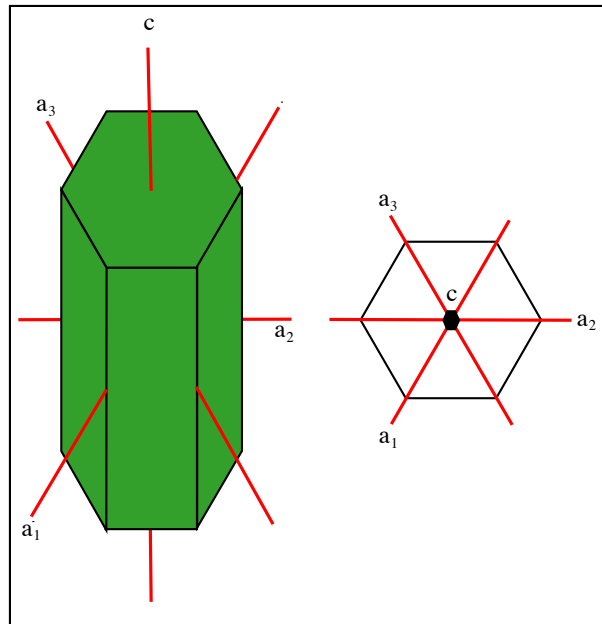


Fig.3.13: Crystallographic axes in the hexagonal system
These are arranged in the same way as in the trigonal system.


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3.3 Crystal classes

We have seen that there are 7 crystal systems. Within each system there are several three-dimensional shapes that can show the minimum symmetry required to define the symmetry. This can be illustrated by two very different shapes that both belong to the cubic crystal system.

The total symmetry of a cube (Fig.3.9) is $3A_4$, $4A_3$, $6A_2$, $9SP$, C .

The total symmetry of a tetrahedron (Fig.3.5b) is $4A_3$, $3A_2$, $6SP$.

Both cube and tetrahedron have $4A_3$ axes of symmetry and therefore belong to the cubic crystal system (Table 3.1). These are two classes of the cubic crystal system. There are a total of 32 crystal classes, but these are beyond the scope of this text.

3.4 Indices of crystal faces

Different types of crystal faces are defined according to their relationship to the crystallographic axes. In Fig.3.14 there are 3 crystallographic axes OX , OY and OZ which meet at O (the origin) and are perpendicular to each other (i.e. 90°). ABC is a crystal face which intersects all three axes. In Fig.3.14 there is an additional crystal face (DEF) that also cuts all three crystallographic axes. One of the vital features of crystal faces is that they cut the crystallographic axes at distances that have a simple, whole number ratio to each other. In Fig.3.14 the ratios of the lengths along the axes defined by the two faces are: $OD = OA$, $OE = 2OB$ and $OF = \frac{1}{2}OC$.

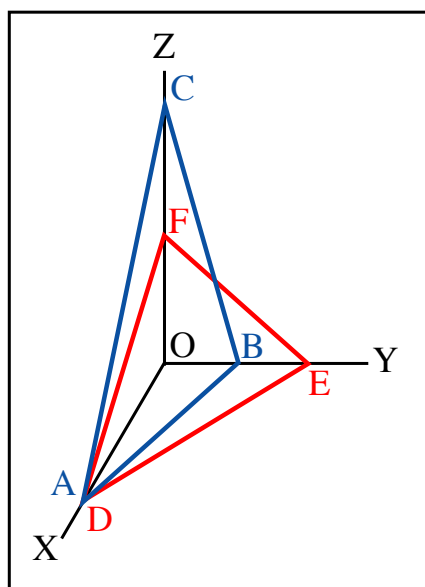


Fig.3.14: Crystallographic parameters
See text for explanation.

The parameters of face DEF relative to ABC are therefore $\frac{1}{1} \frac{2}{1} \frac{1}{2}$

The form of the crystal face ABC (to which other faces are related) varies from crystal to crystal and is called the unit form. There is a system for the notation of crystal faces that was invented by W.F. Miller in 1836. The notations are called Miller indices. Miller indices are the reciprocal of the parameters defined above. The unit form (which cuts all 3 axes) is called (111), as illustrated in Fig.3.15a.

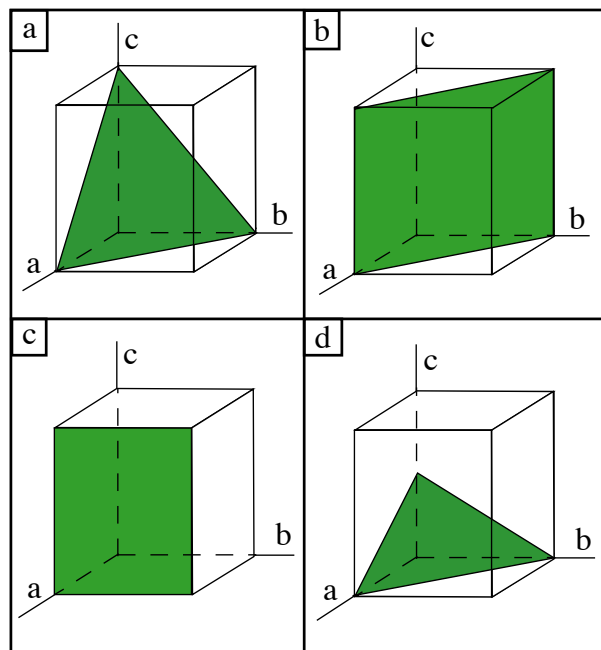


Fig.3.15: Miller indices

a) The unit form (111) cuts all three axes. b) This face cuts a and b at the same distance as (111) and is parallel with c. Its parameters are (11 ∞) and Miller indices (110). c) Parameters (1 $\infty\infty$); Miller indices (100). d) Parameters (11 $\frac{1}{2}$); Miller indices (112)

The crystal face in Fig.3.15b cuts the a- and b-axes at the same distance as (111), but is parallel with the c-axis. The parameters for this face are therefore (11 ∞). It is, however, not practical to use ∞ . Miller indices use the reciprocal: (11 ∞) \rightarrow ($\frac{1}{1} \frac{1}{1} \frac{1}{\infty}$) \rightarrow (110)

In Fig.3.15c the crystal face cuts only the a-axis and is parallel with both the b- and c-axes. The parameters of this face are therefore (1 $\infty\infty$) and its Miller indices are (100). The crystal face in Fig.3.15d has the parameters (11 $\frac{1}{2}$) so that its Miller indices are (112). Note that it cuts the c-axis at half the distance of the unit form (111) in Fig.3.15a, but its Miller indices are (112).

A cube has 6 identical crystal faces, all of which cut one crystallographic axis and are parallel with the other two. Since it is useful to be able to identify any individual face of a crystal, a convention is used in which each crystallographic axis has a positive and a negative end (Fig.3.16). The face that cuts the positive end of the a-axis and is parallel with the b- and c-axes is referred to as (100); the face opposite and parallel to this cuts the negative end of the a-axis and is referred to as ($\bar{1}$ 00).

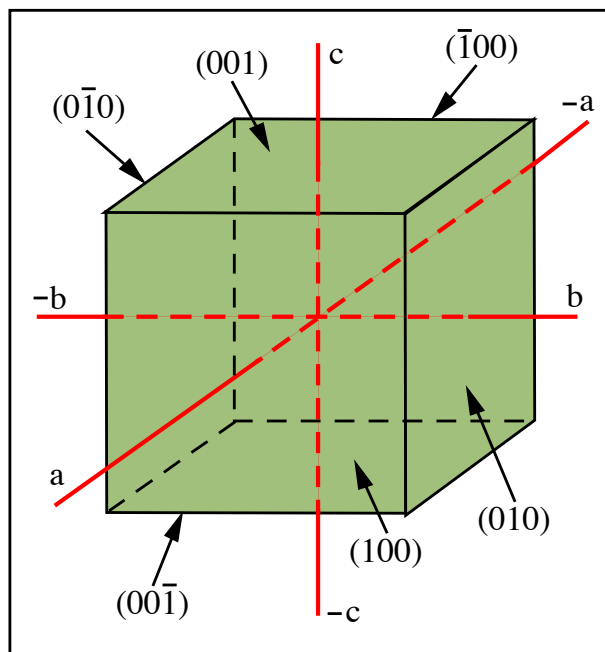



Fig.3.16: Miller indices of the form $\{100\}_6$ in the cubic system

Cubes are a commonly developed crystal form in, for example, the minerals pyrite, halite and fluorite.

The six faces that combine to form a cube are (100) , $(\bar{1}00)$, (010) , $(0\bar{1}0)$, (001) , and $(00\bar{1})$. These 6 identical crystal faces define the cube form which is conventionally shown using “curly brackets” $\{.....\}$. $\{100\}_6$ in the cubic system means all these 6 faces.



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In the cubic crystal system there are 12 faces of the type (110) , as shown in Fig.3.17. These 12 identical faces $\{110\}_{12}$ combine to form a dodecahedron.

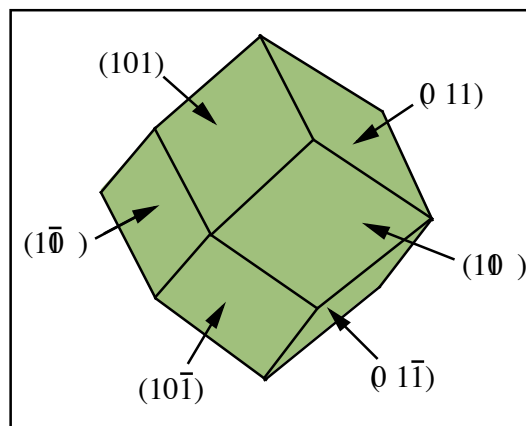


Fig.3.17: There are 12 faces in the form $\{110\}$ in the cubic system which combine to give a dodecahedron. The $\{110\}_{12}$ form is commonly developed in garnet crystals.

There are 8 faces of the type (111) in the cubic system (Fig.3.18) that combine to form an octahedron.

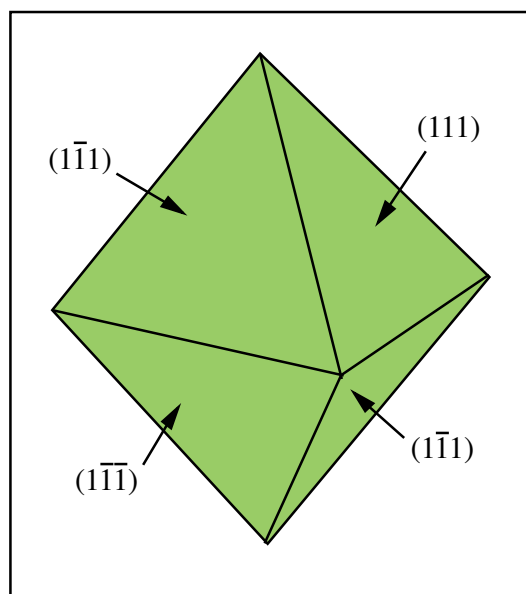


Fig.3.18: There are 8 faces in the form $\{111\}$ in the cubic system. These combine to form an octahedron. The $\{111\}_8$ form is commonly developed in magnetite crystals and is a common cleavage in fluorite.

There are 12 faces of the type (210) in the cubic system (Fig.3.19). These combine to form a pyritohedron, so-called because it is commonly developed in pyrite (FeS_2).

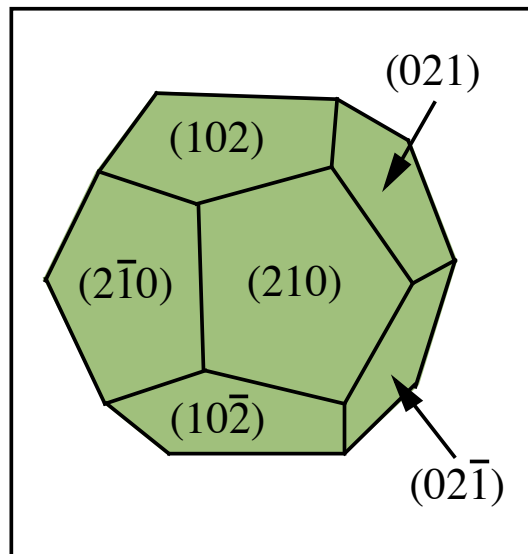


Fig.3.19: In the cubic system, twelve identical faces of the type $\{210\}$ combine to form a pentagonal dodecahedron
 The $\{210\}_{12}$ form is commonly developed in the mineral pyrite and is also known as a pyritohedron.

There are 24 faces of the type $\{112\}$ in the cubic system (Fig.3.20). These combine to form a trapezohedron.

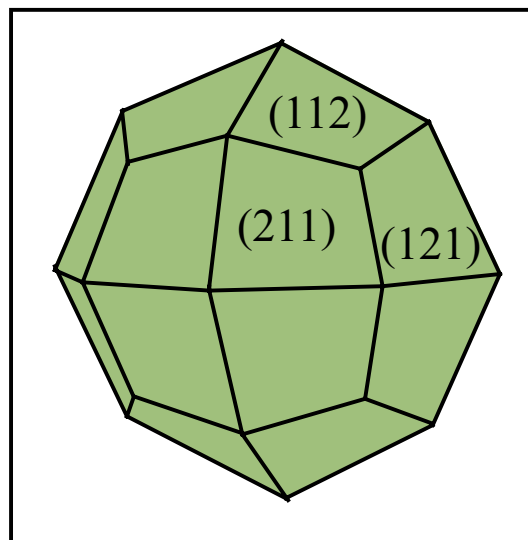


Fig.3.20: There are 24 identical faces in the form $\{112\}$ in the cubic system which combine to give a trapezohedron
 The $\{112\}_{24}$ form is commonly developed in the minerals garnet and leucite.

In the tetragonal system there are two important types of crystal faces, those parallel with the *c*-axis (*prism* faces) and those that cut both the *a*- and *c*-axes (*pyramid* faces) (Fig.3.21). It is apparent that there are 4 prism faces $\{110\}$ and/or 8 pyramid faces $\{111\}$ that must combine to make a tetragonal crystal. The form $\{001\}$ (if it is developed) consists of two faces, one at either end of the crystal. This type of face is called a *pinacoid*.

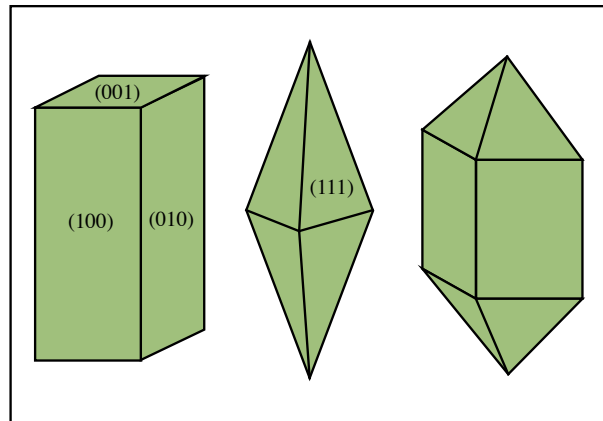


Fig.3.21: Prism, pyramid and pinacoid faces in the tetragonal system
A combination of prism and pyramid faces are commonly present in the mineral zircon.

There are three pinacoid forms in the orthorhombic system $\{100\}$, $\{010\}$ and $\{001\}$ (Fig.3.22). Faces of the type $\{110\}$ are prism-faces and $\{111\}$ -type are pyramids, as in the tetragonal system.

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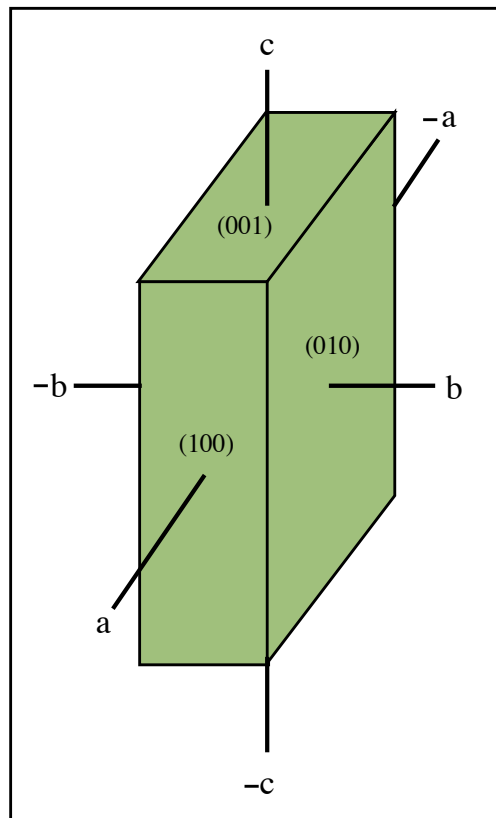


Fig.3.22: Pinacoid faces in the orthorhombic system

In the monoclinic crystal system the crystallographic a- and c-axes are not at 90° to each other. The convention is that the b-axis is identical with the A_2 axis of symmetry (Fig.3.23).

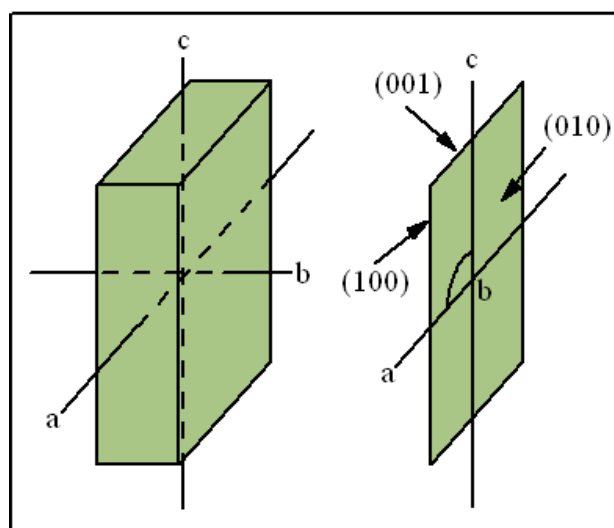


Fig.3.23: Crystallographic axes in the monoclinic system

The a- and c-axes are not perpendicular to each other in the monoclinic system. The angle between the a and c axes (referred to as the obtuse ($>90^\circ$) angle β) varies from mineral to mineral. In the monoclinic mineral gypsum $\beta \neq 114^\circ 12'$. In hornblende $\beta \neq 105^\circ 44'$.

In the triclinic system, where none of the angles between the three crystallographic axes are 90° , there is a free choice as to how the axes are located relative to the faces.

In the trigonal and hexagonal crystal systems it is convenient to use 4 crystallographic axes. The c-axis is unique (it is the A3 or A6 axis of symmetry) and the three identical a-axes at 60° to each other are called a_1 , a_2 and a_3 with positive and negative ends (Fig.3.24). The positive and negative ends are located alternately. This results in the sum $a_1 + a_2 + a_3 = 0$.

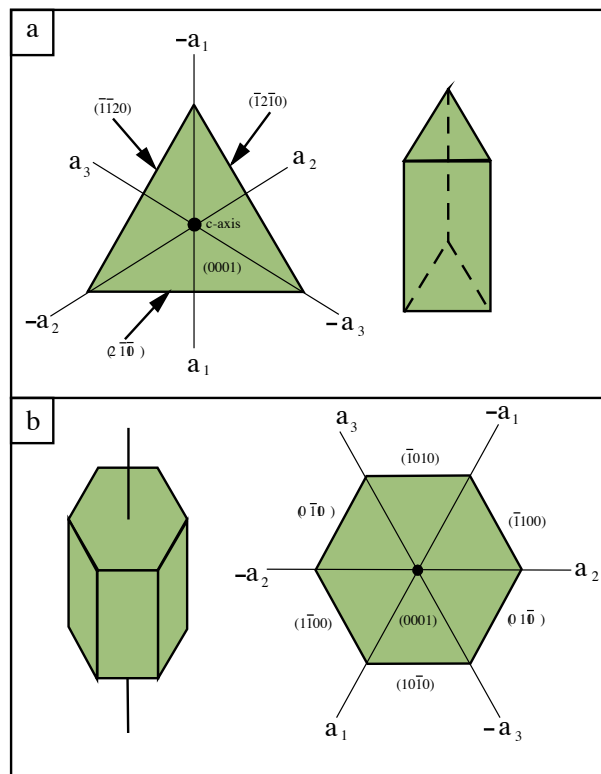


Fig.3.24. Convention for the location of crystallographic axes in the (a) trigonal and (b) hexagonal systems
A face parallel with the c-axis that cuts a_1 and $-a_2$ at equal distances and is parallel with a_3 is called (1100) etc.

A form that is common in the trigonal system is that of a rhomb (Fig.3.25). The most commonly developed crystallographic form is $\{10\bar{1}1\}$. This is the form of the excellent cleavage in the trigonal mineral calcite. Calcite rhombs are often used to demonstrate double refraction (birefringence).

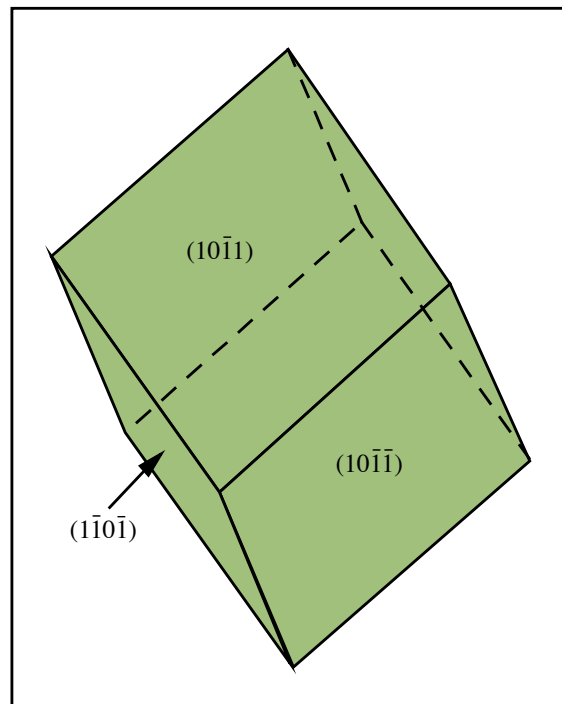


Fig.3.25. A rhombic form is commonly developed in the trigonal system (e.g. calcite cleavage).

We have, in general, only dealt with the most simple types of crystal faces: (100), (110) and (111). Many other faces can be developed in crystals. The Miller indices notation for faces of the type that cut a, b and c is {hkl}. This includes forms {111}, {112}, {113}, {123} etc. The notation {hk0} includes {110}, {120}, {130}, {230} etc.

4 Systematic Mineralogy

4.1 Silicate minerals

About 3600 minerals have been identified. Most of these occur in the Earth's crust. So far about 30 minerals have been mentioned in this text. Some minerals are, of course, more common than others. The crust is dominated by the elements oxygen and silicon. Oxygen forms O^{2-} anions and compounds that contain O^{2-} are called oxides (hematite Fe_2O_3 and quartz SiO_2 are oxides). Silicon forms Si^{4+} cations. Silicon and oxygen together form an extremely strong complex ion: the *silicate ion* $[SiO_4]^{4-}$. Minerals that contain the silicate ion are *silicate minerals*; these dominate the crust.

O^{2-} has an ionic radius of 1.32 Å whereas Si^{4+} is a relatively small cation with an ionic radius of 0.42 Å. Considering the ions as spheres, 4 large oxygen ions can be packed around one small silicon ion giving a tetrahedral structure. This structure has 4 positive and 8 negative charges, giving a net charge of 4-. Silicate minerals are dominated by the $[SiO_4]^{4-}$ silicate tetrahedron. $[SiO_4]^{4-}$ tetrahedra exist independently in some minerals but can share one, two, three or all four oxygen anions in other minerals. This possibility gives a range of *silicate structures* (Fig. 4.1).

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	Type of Linkage	Formula + Valency	Si:O	• Si ⁴⁺ • O ²⁻	Example	General trends
NESO-	Separate tetrahedra	(SiO ₄) ⁴⁻	1:4		Olivine Mg ₂ [SiO ₄]	<div>Increasingly open structures</div> <div>Decreasing density</div> <div>Lower crystallisation temperature</div> <div>(OH)-commonly present</div> <div>Increasing room for large cations eg. K</div>
SORO-	Double tetrahedra	(Si ₂ O ₇) ⁶⁻	1:3.5		(Uncommon)	
CYCLO-	(3-, 4- & 6-) tetrahedron-rings	(Si ₆ O ₁₈) ¹²⁻	1:3		Beryl Be ₃ Al ₂ [Si ₆ O ₁₈]	
INO-	Single chains	(SiO ₃) ²⁻	1:3		Pyroxene var. Diopside Ca Mg[SiO ₃] ₂	
INO-	Double chains	(Si ₄ O ₁₁) ⁶⁻	1:2.75		Amphibole var. Tremolite Ca ₂ Mg ₅ [Si ₄ O ₁₁] ₂ (OH) ₂	
PHYLLO-	Sheets	(Si ₂ O ₅) ²⁻	1:2.5		Mica var. Muscovite KAl[AlSi ₃ O ₁₀](OH) ₂ Biotite K(Mg,Fe) ₃ Al[AlSi ₃ O ₁₀](OH) ₂	
TECTO-	Three-dimensional networks	SiO ₂	1:2	Not readily depicted in 2 dimensions	Feldspars eg. Orthoclase K[AlSi ₃ O ₈] Silicon dioxide Quartz SiO ₂	

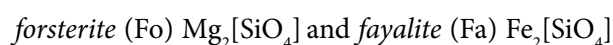
Fig. 4.1: Types of SiO₄ linkage in silicate minerals

The most simple silicate structure involves individual SiO₄-units (nesosilicates). The structure becomes more complex as the number of oxygens shared by adjacent SiO₄-units increases. One oxygen is shared in sorosilicates and two are shared in a ring-structure in cyclosilicates. Two oxygens are shared to form a chain structure in single chain inosilicates, and these chains are linked by shared oxygens in double chain inosilicates. Three of the four oxygens in the SiO₄-units are shared to form a sheet structure in phyllosilicates, and all four are shared to form a complex three-dimensional structure in the tectosilicates. As the amount of oxygen-sharing increases and the structure becomes more open the density of silicate minerals generally decreases and there is more room for large ions (e.g. cations Na⁺ and K⁺; (OH)⁻ anions).

4.1.1 Nesosilicates

4.1.1.1 Olivine group

The olivine group consists of two end members:



in which Mg^{2+} and Fe^{2+} readily substitute for each other. All intermediate compositions exist. The composition of the olivine group can be written as $(\text{Mg,Fe})_2[\text{SiO}_4]$. Note that the silicate part of the composition $[\text{SiO}_4]$ is in square brackets so that it is straightforward to identify the type of silicate (Fig. 4.1), whereas the cations that substitute for each other (Mg,Fe) are in round brackets.

The density of olivine increases with Fe-content; forsterite $G = 3.2$; fayalite $G = 4.4$. Olivine is orthorhombic, but well-formed crystals are seldom observed. Fresh crystals are green; the mineral is named after its olive-green colour (Picture 4.1). Mg-rich olivine is a major component of the Earth's mantle. The mantle consists of rocks called peridotite since they are dominated by olivine, the gem variety of which is known as peridot. Olivine is an important, early crystallizing mineral from high temperature basaltic magma. It is commonly altered to a reddish alteration product or to serpentine $(\text{Mg}_3[\text{Si}_2\text{O}_5](\text{OH})_4)$. Note that serpentine has a composition close to forsterite + water.



Picture 4.1: The green nodule is enclosed in solidified basaltic lava. The nodule, that comes from the mantle, consists of the ultramafic rock type called peridotite. The dominant minerals in this peridotite are pale green magnesium-rich olivine and lesser amounts of darker green chromium-rich clinopyroxene. The sample comes from Lanzarote, one of the Canary Islands.

Pure forsterite melts at 1890°C and pure fayalite at 1205°C and there is complete solid solution between them. A temperature – composition diagram for olivine is a “cigar-diagram” (Fig. 4.2).

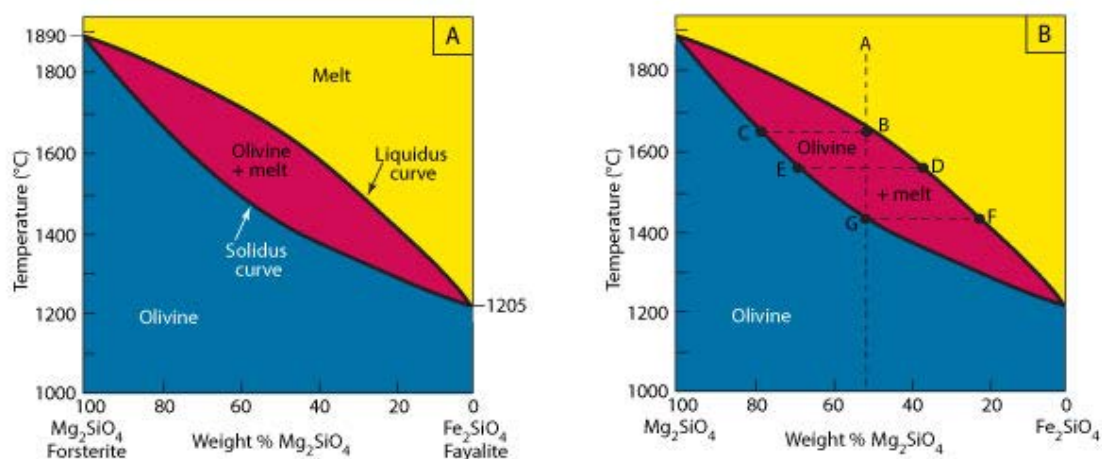


Fig. 4.2: Temperature-composition diagram (at atmospheric pressure) for the olivine system (forsterite $\text{Mg}_2[\text{SiO}_4]$ – fayalite $\text{Fe}_2[\text{SiO}_4]$).

A. The diagram is divided into three fields by two curves. A "melt" field occurs above the upper (liquidus) curve. A "solid" field occurs below the lower (solidus) curve. The two curves outline a cigar-shaped field where solid (olivine crystals) and melt occur in equilibrium. B. The crystallization process illustrated here is explained in the text.

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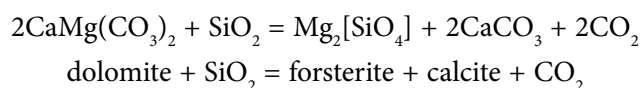
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Consider a melt with composition A (melt with the composition of Fo_{50}) at a temperature of $>1800^\circ\text{C}$. On cooling this will intersect the *liquidus* at B at a temperature of 1660°C . Crystals of Mg-rich olivine with composition C (Fo_{77}) will begin to form. Crystals with composition C are in equilibrium with melt with composition B and they are, of course, at the same temperature. With further cooling new crystals will grow and the crystals that have already formed will gradually change composition along the *solidus* curve while the melt changes composition along the *liquidus* curve. The crystals will always remain in equilibrium with the cooling melt e.g. crystals at E will be in equilibrium with melt D. The amount of crystals relative to melt will gradually increase and the final drop of melt will have composition F (melt with a composition of Fo_{21}) which is in equilibrium with crystals at G. Note that G (Fo_{50}) has the same composition as the initial melt A.

Note that *the composition of the melt will always be more iron-rich than that of the coexisting crystals at the same temperature*. This crystallization path assumes perfect *equilibrium* – the melt is always able to react with the crystals. This is seldom achieved in nature. For example, early crystals may be removed from the melt to form a rock composed of Mg-rich olivine (called dunite). In our diagram the final melt can therefore reach more Fe-rich compositions than F and the final crystals can therefore be more Fe-rich than G (Fo_{50}) even though the starting melt had a composition of Fo_{50} . A fraction of the crystallizing assemblage has been removed from the system and this process, called *fractional crystallization*, is very important in igneous petrology.

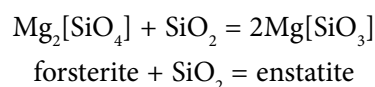
Alternatively, crystallization can take place so rapidly that individual olivine crystals cannot change their composition by reaction with the melt. Olivine crystals can therefore become *zoned* with Mg-rich cores and Fe-rich rims. Exchange between Mg^{2+} and Fe^{2+} in olivines, however, takes place very easily, and zoning is seldom observed. Other mineral groups that show solid solution are well known for zoned crystals, in particular the plagioclase feldspars, as we will see later.

Forsterite occurs in some metamorphosed carbonate rocks. To form forsterite requires both Mg and Si. Pure limestone is composed of calcite (CaCO_3) – so no olivine can be formed from this. Many limestones are, however, impure and contain both quartz (SiO_2) and dolomite ($\text{CaMg}(\text{CO}_3)_2$). At a temperature of $\sim 500^\circ\text{C}$ these can react together to form forsterite:



Metamorphosed limestone is marble and the rock produced would be forsterite marble.

Mg-rich olivine cannot exist in equilibrium with quartz because they react together to form a new mineral – Mg-rich orthopyroxene called enstatite (section 4.1.4.1.1):



4.1.1.2 Garnet group

Another mineral group which contains independent $[\text{SiO}_4]^{4-}$ tetrahedra is the garnet group. Garnets have a much wider compositional range than olivine and have the general formula $\text{R}^{2+}_3\text{R}^{3+}_2[\text{SiO}_4]_3$ where $\text{R}^{2+} = \text{Mg}^{2+}, \text{Fe}^{2+}, \text{Ca}^{2+}$ or Mn^{2+} (divalent cations) and $\text{R}^{3+} = \text{Al}^{3+}, \text{Fe}^{3+}$ or Cr^{3+} (trivalent cations). The most common garnet, which occurs in metamorphic rocks and has a reddish brown colour, is called almandine ($\text{Fe}_3\text{Al}_2[\text{SiO}_4]_3$). This occurs in metamorphosed Al – rich rocks, the most usual of which are clay – rich sediments which have been heated to over $\sim 500^\circ\text{C}$. The green garnet in Picture 4.2 is rich in the grossularite ($\text{Ca}_3\text{Al}_2[\text{SiO}_4]_3$) end-member. Garnets are cubic and commonly form 12-sided (dodecahedra) or 24-sided crystals (trapezohedra). As garnets are quite hard ($H \approx 7$), have no cleavage and are not uncommon they are used as abrasive material for grinding and polishing. Most varieties of garnet are cut as gemstones. Its name comes from the Latin *granatus*, meaning grain-like.



Picture 4.2: Green grossularite-rich garnet showing the dodecahedral form illustrated in Fig. 3.17.

4.1.1.3 Zircon

Zircon ($\text{Zr}[\text{SiO}_4]$) is the main mineral that contains the element zirconium and occurs in small amounts in a wide range of rock types – it is a common accessory mineral. Zircon, which forms brownish tetragonal crystals (Picture 4.3), is a very important mineral for age determinations. This is because zircon is an extremely stable mineral and uranium can enter the zircon structure. The radioactive isotope of uranium (e.g. ^{238}U) decays to radiogenic lead (^{206}Pb) at an extremely slow rate (half life = 4.47×10^9 years). Measurement of the amounts of these isotopes allows determination of the time at which the zircon crystallized.

The zircon age determination method is the most reliable technique for very old rocks and has provided information on the oldest rocks in the world. Zircon is the main source for the element zirconium that is used in nuclear reactors. ZrO_2 is extremely refractory and is used to make crucibles for melting platinum at 1755°C .



Picture 4.3: Zircon crystal illustrating its tetragonal crystal structure with prism and two types of pyramid faces developed (see also Fig. 3.21).



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4.1.1.4 Sphene

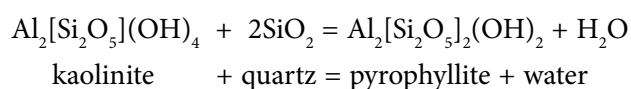
Sphene ($\text{CaTi}[\text{SiO}_4](\text{O},\text{OH},\text{F})$) (alternatively called titanite) is also a widely developed accessory mineral, especially in granites. It forms brownish wedge-shaped crystals and is the most important Ti-bearing silicate mineral. There are other Ti-bearing minerals, most notably ilmenite (FeTiO_3) and rutile (TiO_2) but these are oxides and not silicates.

4.1.1.5 Aluminium silicate polymorphs

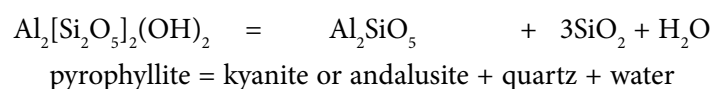
There are three minerals with the composition Al_2SiO_5 . Writing the composition in this way masks the fact that they contain independent silicate tetrahedra; this becomes clear when the formula is written $\text{Al}_2\text{O}[\text{SiO}_4]$. The three aluminium silicate polymorphs are:

sillimanite	–	orthorhombic
andalusite	–	orthorhombic
kyanite	–	triclinic

Sillimanite forms white, andalusite brown and kyanite pale blue prismatic crystals (Picture 4.4). These occur in clay-rich sedimentary rocks (most clay minerals are Al-rich; for example kaolinite ($\text{Al}_2[\text{Si}_2\text{O}_5](\text{OH})_4$)) that have been subjected to high temperature and/or pressure (i.e. metamorphism). Clay-rich rocks are collectively called pelites; after metamorphism they are metapelites. It is clear from the PT diagram (Fig. 4.3) that andalusite is only stable at pressure below 4 kbar (equivalent to a depth of 12–14 km), sillimanite only forms at temperatures above 525°C, and kyanite is the high pressure phase. During metamorphism kaolinite reacts with quartz to form the mineral pyrophyllite ($\text{Al}_2[\text{Si}_2\text{O}_5]_2(\text{OH})_2$).



With increasing temperature pyrophyllite breaks down to form andalusite or kyanite (depending on the pressure).



Note that both these reactions release water. This is a typical feature of metamorphic reactions involving hydrous minerals. The Al_2SiO_5 polymorphs are widely used to determine pressure/temperature conditions during the metamorphism of clay-rich sedimentary rocks.



Picture 4.4: Top left: blue kyanite; right: white sillimanite; bottom left: andalusite with a cross-structure (chastolite).

Andalusite sometimes develops a black cross (Picture 4.4) formed by tiny carbonaceous inclusions. This special variety of andalusite is called chastolite.

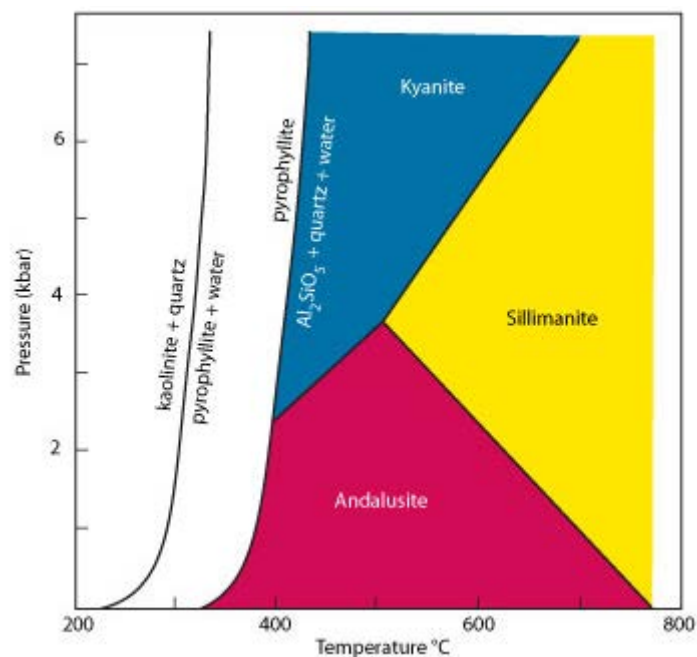


Fig. 4.3: Pressure-temperature diagram for the Al_2SiO_5 polymorphs (kyanite, andalusite and sillimanite) and the breakdown curves for kaolinite and pyrophyllite.

Clay-rich rocks (pelites) containing kaolinite commonly develop one of the Al_2SiO_5 polymorphs during metamorphism, via pyrophyllite. Either andalusite or kyanite will be the first polymorph to form, depending on the pressure.

4.1.1.6 Staurolite

The mineral staurolite also occurs in metamorphosed Al-rich sedimentary rocks (metapelites) at $>500^{\circ}\text{C}$ (Fig. 4.4). It has a brownish colour and commonly occurs as well-formed monoclinic (actually pseudo-orthorhombic – i.e. they appear to be orthorhombic) crystals. Cruciform twins are a characteristic feature of staurolite (Fig. 2.2). It is named after the Greek word *stauros* meaning cross, referring to its cruciform twins.

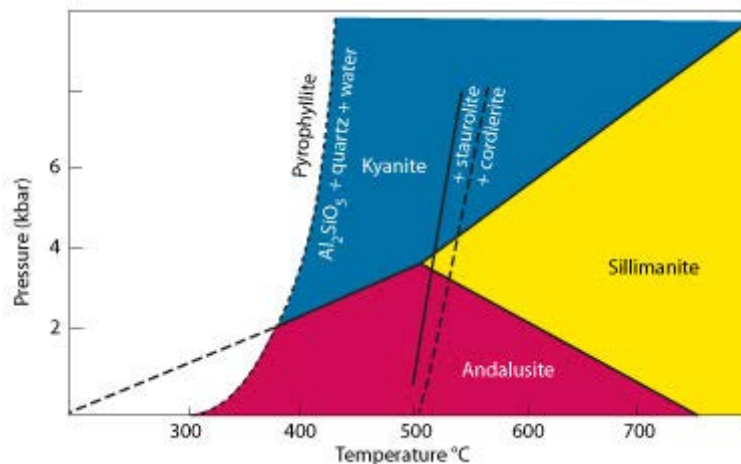


Fig. 4.4: Formation of staurolite and cordierite relative to the Al_2SiO_5 polymorphs during metamorphism. These minerals are very useful in the determination of the pressure-temperature conditions during the metamorphism of clay-rich sediments (metapelites). Cordierite is in section 4.1.3.3.

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4.1.1.7 Topaz

Topaz ($\text{Al}_2[\text{SiO}_4](\text{OH},\text{F})_2$) is an orthorhombic mineral that is often used as a gem stone. It occurs in granitic rocks. The best crystals are found in granitic pegmatites. A pegmatite is a very coarse grained igneous rock.

4.1.2 Sorosilicates (epidote)

There are not many important sorosilicates (at least not important for us at present). The *epidote* group of minerals, however, contains both $[\text{SiO}_4]^{4-}$ and $[\text{Si}_2\text{O}_7]^{6-}$ units and has a complicated chemical composition in which Ca^{2+} , Al^{3+} and Fe^{3+} are involved. Epidote (with the composition $\text{Ca}_2(\text{Al},\text{Fe}^{3+})_3\text{O}[\text{SiO}_4][\text{Si}_2\text{O}_7](\text{OH})$) is a greenish mineral (Picture 4.5) which occurs as a relatively low temperature (typically 200–400°C) alteration product in many rocks types (where it commonly occurs in cracks and veins) and as a metamorphic mineral. It is an essential mineral in “greenschists” which are metamorphosed basalts.



Picture 4.5: An aggregate of green prismatic epidote crystals.

4.1.3 Cyclosilicates

4.1.3.1 Beryl

Beryl ($\text{Be}_3\text{Al}_2[\text{Si}_6\text{O}_{18}]$) is the “type” ring silicate in which the hexagonal symmetry reflects the six-membered rings of SiO_4 tetrahedra in the structure (Fig. 3.13). It is a green mineral (bright green varieties are known as emerald; bluish green as aquamarine). Beryllium forms a very small cation (Be^{2+}) which does not enter the structure of the common rock-forming minerals during fractional crystallization of magmas. Be^{2+} is therefore concentrated in the residual magma that will also be enriched in other small cations (e.g. boron, lithium).

The same applies to very large cations (e.g. thorium, uranium) and large anions (e.g. fluorine, hydroxyl ((OH)-groups), chlorine). Late stage magmatic fluids are therefore volatile-rich and can crystallize relatively rare minerals that contain small (or large) cations. The mineral beryl therefore occurs in late stage granitic rocks, commonly in very coarse-grained rocks called pegmatites. These can contain very large individual crystals (in extreme cases up to several meters long). Beryl is typically associated with quartz, K-feldspar, Li-mica, tourmaline (a boron-mineral) and other “exotic” minerals.

4.1.3.2 Tourmaline

This is another ring-structured silicate mineral with $[\text{Si}_6\text{O}_{18}]$ -units. The structure contains boron in BO_3 -groups as well as Na, Mg, Al, Fe, Li and (OH)-groups. The colour of tourmaline varies with its composition and many varieties are used as gemstones. The most common, Fe-rich varieties, are black; Li-rich ones are green; brown, green and red types also occur. Colour zoning is common. Tourmaline forms trigonal crystals that commonly have a triangular cross-section (Picture 4.6). Since tourmaline contains the small boron cation (B^{3+}) it occurs (like beryl) in late stage granitic rocks, especially pegmatites. Tourmaline crystals are commonly striated parallel with the c-axis.



Picture 4.6: Black tourmaline crystal where the trigonal form is obvious.

4.1.3.3 Cordierite

Cordierite ($(\text{Mg,Fe})_2\text{Al}_3[\text{Si}_5\text{AlO}_{18}]$) has the same structure as beryl at high temperature but changes (or “inverts”) to an orthorhombic symmetry on cooling. Its external form is therefore hexagonal but its internal structure is orthorhombic. Comparison with the composition of beryl shows that one of the 6 Si^{4+} -cations is replaced by Al^{3+} which has a similar ionic radius. The Al^{3+} replacing Si^{4+} has a coordination number of 4 (like Si^{4+}). The other Al present in the structure helps to link the $[\text{Si}_5\text{AlO}_{18}]$ -rings together, and has a coordination number of 6. Al^{3+} can therefore occur in two different sites in silicate minerals with 4- or 6-fold coordination. This is very important in some minerals, as we shall see later. Cordierite occurs in Al-rich sedimentary rocks which have been metamorphosed to temperatures $>500^\circ\text{C}$ (metapelites) (Fig. 4.4).

4.1.4 Inosilicates

There are two types of chain silicates, those with single chains ($[\text{SiO}_3]^{2-}$ -units) and those with double chains ($[\text{Si}_4\text{O}_{11}]^{6-}$ -units). The pyroxene group of minerals are single chain silicates; the amphiboles are double chain silicates.

4.1.4.1 Pyroxenes

Some pyroxenes are orthorhombic, others are monoclinic (Fig. 4.5). The name pyroxene is a mistake! It comes from a Greek word meaning “stranger to fire” since it was erroneously believed that it did not occur in igneous rocks.

4.1.4.1.1 Orthopyroxenes

Orthopyroxenes, like the olivine group, form a solid solution series between $\text{Mg}[\text{SiO}_3]$ (*enstatite*) and $\text{Fe}[\text{SiO}_3]$ (*ferrosilite*). The most common orthopyroxene, with a composition intermediate between these two end-members, is called hypersthene. The modern pyroxene nomenclature does not include “hypersthene”, but it is common in the older literature. Orthopyroxenes, which are dark brown to black, occur in igneous rocks where they are essential components in some basalts and their coarse grained equivalents (e.g. plutonic rocks composed of orthopyroxene and plagioclase feldspar are called norites).



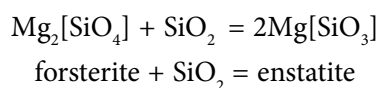
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Orthopyroxenes are sometimes formed when Mg-rich olivine reacts with the SiO_2 -component in a melt:



Orthopyroxene is also common in plutonic rocks that contain few or no light minerals, the so-called ultramafic rocks. It is an important mineral in mantle peridotites. In metamorphic rocks the presence of orthopyroxene is evidence of high temperatures.

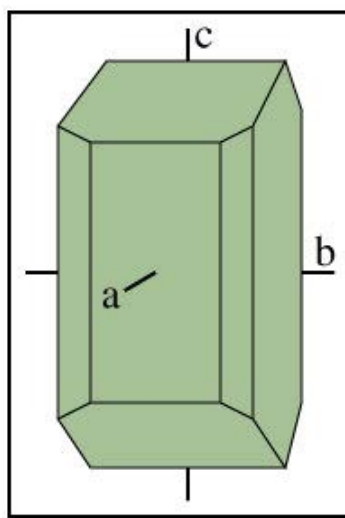


Fig. 4.5: Pyroxenes are either orthorhombic (orthopyroxenes) or monoclinic (clinopyroxenes). The drawing illustrates a typical orthopyroxene crystal. In clinopyroxenes the angle between the **a** and **c** axes is not 90° .

4.1.4.1.2 Clinopyroxenes

Clinopyroxenes cover a wide range of compositions. The most common ones have compositions intermediate between the end members $\text{CaMg}[\text{Si}_2\text{O}_6]$ (*diopside*), $\text{CaFe}[\text{Si}_2\text{O}_6]$ (*hedenbergite*), enstatite ($\text{Mg}[\text{SiO}_3]$) and ferrosilite ($\text{Fe}[\text{SiO}_3]$) (Fig. 4.6). The most widespread clinopyroxene, which has an intermediate composition in the “pyroxene quadrilateral”, is called *augite*. Augite is a black mineral which occurs in basalts and their plutonic equivalents, gabbros. It also occurs in some ultramafic rocks and is a very important phase in mantle peridotites. Augite also occurs in some high-temperature metamorphic rocks.

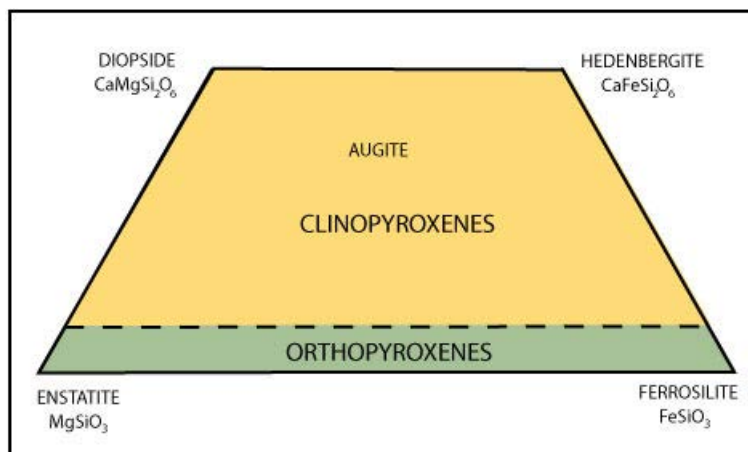


Fig. 4.6: Classification of the Ca-Mg-Fe pyroxenes in the “pyroxene quadrilateral”. Orthopyroxenes are orthorhombic whereas clinopyroxenes are monoclinic.

Other clinopyroxenes include *aegirine* which has the composition $\text{NaFe}[\text{Si}_2\text{O}_6]$, where Fe here is Fe^{3+} to achieve valency balance with Na^+ and $[\text{Si}_2\text{O}_6]^{4-}$; it is dark green and occurs in Na-rich igneous rocks. Compositions intermediate between aegirine and augite occur; these are (logically enough) called *aegirine-augite*. Another end-member is *jadeite* ($\text{NaAl}[\text{Si}_2\text{O}_6]$) which forms under high pressure. Compositions intermediate between jadeite and augite are called *omphacite* and are noteworthy for their occurrence in eclogites, which are the metamorphic equivalents of basalts formed under very high pressures and relatively low temperatures (in subduction zones). Eclogites consist dominantly of two minerals: green omphacite and red-brown garnet. These very attractive, dense, rocks are quite rare.

Pyroxenes can often be distinguished from amphiboles by their cleavage (Fig. 4.7) or their crystal outlines (Picture 4.7). Both groups of minerals have prismatic cleavage, but pyroxenes break into fragments with square or rectangular cross sections (i.e. 90° between the cleavage planes) whereas amphiboles break into diamond-shaped cross sections (60° between the cleavage planes).

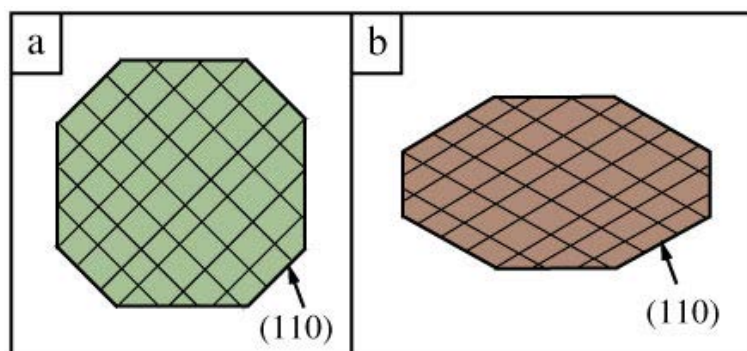


Fig. 4.7: Pyroxenes and amphiboles can often be distinguished by their cleavage. The angle between cleavage surfaces in pyroxenes is close to 90° whereas in amphiboles the angle is close to $60^\circ/120^\circ$.



Picture 4.7: Crystal shape can sometimes be used to distinguish between pyroxenes and amphiboles. Individual crystals of pyroxene (left) and amphibole (right) have been cut through to illustrate their cross-sectional shapes. The angles between prismatic crystal faces in pyroxenes are close to 135° whereas in amphiboles they are close to 60° or 120°.

4.1.4.2 Amphiboles

Like the pyroxenes, some amphiboles are orthorhombic but most are monoclinic. The compositional variation of amphiboles is expressed by the general formula:



A = Na, K; B = Ca, Na, Mg, Fe²⁺; C = Mg, Fe²⁺, Fe³⁺, Al; T = Si, Al

The A-site is for relatively large cations; this site is empty in some amphiboles. The B-site is slightly larger than the C-site. Note three points: a) large cations (K⁺) may be present; b) Al³⁺ substitutes quite extensively for Si⁴⁺; c) The presence of (OH)⁻ groups. These features reflect the fact that the amphibole structure is fairly “open”, in contrast to more compact structures like those of olivine and garnet (silicates with independent SiO₄-tetrahedra), as is shown in Fig. 4.1. Distinction between amphiboles and pyroxenes can commonly be achieved using the angle between cleavage surfaces (Fig. 4.7) or crystal outlines (Picture 4.7).

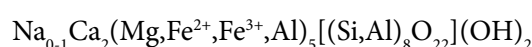
4.1.4.2.1 Orthorhombic amphiboles

The most common orthorhombic amphibole is *anthophyllite* (Mg₇[Si₈O₂₂](OH)₂) in which the A-site is empty and both B- and C-sites are occupied by Mg. Anthophyllite is a grey-brown mineral that forms elongate prisms that occur in aggregates. It occurs in metamorphosed, olivine-rich ultramafic rocks.

4.1.4.2.2 Monoclinic amphiboles

Two end-members in the monoclinic amphiboles are *tremolite* ($\text{Ca}_2\text{Mg}_5[\text{Si}_8\text{O}_{22}](\text{OH})_2$) and *actinolite* ($\text{Ca}_2\text{Fe}_5[\text{Si}_8\text{O}_{22}](\text{OH})_2$). Tremolite is colourless to pale green and typically occurs as a result of the metamorphism of Ca- and Mg-bearing carbonate sediments (dolomites). Compositions intermediate between tremolite and actinolite occur as a result of the metamorphism of the most common pyroxene, augite. They also occur as one of the green minerals in metamorphosed basaltic rocks known as greenschists. The other green minerals in greenschists are epidote and a mineral with a layered structure called chlorite. The Na-feldspar, albite, is also present in greenschists.

The most widespread amphibole is called *hornblende* that has a complex and variable composition:



Hornblende (Fig. 4.8) is a dark green to black mineral that occurs in many different rock types. In igneous rocks it may be the only hydrous mineral in, for example, basalts and gabbros, but is more common in intermediate and acidic (SiO_2 -rich) types like granites. Hornblende is an important mineral in metamorphic rocks, especially in the metamorphosed equivalents of basaltic rocks known as amphibolites. These consist dominantly of hornblende and a feldspar mineral (plagioclase) and form at higher temperatures than greenschists.

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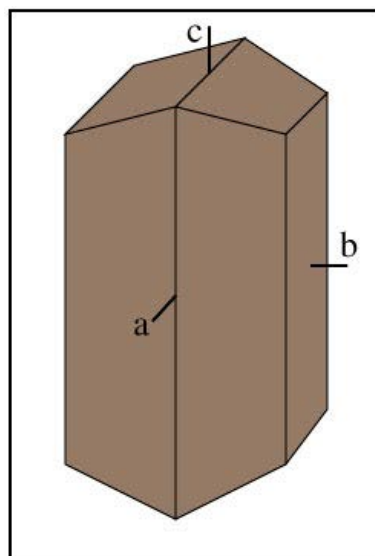


Fig. 4.8: Illustration of a typical hornblende crystal.
The angle between the **a** and **c** axes is not 90°.

There are many other amphiboles, but the only one we will consider here is called *glaucofan* ($\text{Na}_2\text{Mg}_3\text{Al}_2[\text{Si}_8\text{O}_{22}](\text{OH})_2$). This amphibole is blue and occurs in basaltic rocks that have been metamorphosed at relatively high pressures and low temperatures – called blueschists.

4.1.5 Phyllosilicates

Most phyllosilicates – minerals with layered silicate structures (i.e. containing Si_2O_5 -units) – have a platy habit and one prominent cleavage. They are generally soft, with relatively low density, and are commonly flexible.

4.1.5.1 Serpentine

Serpentine is a hydrated Mg-silicate mineral with a layered structure ($\text{Mg}_3[\text{Si}_2\text{O}_5](\text{OH})_4$). The most common occurrence of serpentine is as a metamorphic alteration product of olivine ($\text{Mg,Fe})_2[\text{SiO}_4]$. Olivine-rich rocks that have been extensively altered to serpentine are called serpentinites. Olivine contains both Mg and Fe, whereas the serpentine structure can contain only very little Fe. The excess Fe released from olivine during alteration to serpentine usually occurs as small grains of magnetite (Fe_3O_4) so that serpentinites are weakly magnetic. Serpentine is a relatively soft mineral ($H = 3-5$) with a greasy lustre when massive and silky in fibrous varieties. Many, but not all, serpentines are green. Serpentine can occur with a fibrous habit and is one of the minerals used commercially as asbestos.

4.1.5.2 Talc

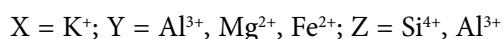
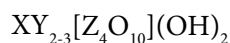
Talc ($\text{Mg}_3[\text{Si}_4\text{O}_{10}](\text{OH})_2$) is the first mineral with which you ever came into contact – it is the main component in talcum powder. Talc is a very soft mineral and defines hardness = 1 on Mohs' scale. It has a layered structure but commonly occurs in foliated masses; a rock composed of talc is called soapstone because of its greasy feel. The composition of talc is similar to that of serpentine and the two minerals often occur together. The mineral *pyrophyllite* ($\text{Al}_2[\text{Si}_4\text{O}_{10}](\text{OH})_2$) is compositionally related to talc with Al^{3+} instead of Mg^{2+} . It is usually very fine grained and forms during the low grade metamorphism of clay-rich sediments by reaction between kaolinite (section 4.1.5.3) and quartz. Pyrophyllite breaks down in turn to give andalusite or kyanite (Fig. 4.3).

4.1.5.3 Kaolinite

Kaolinite is a clay mineral with the composition $\text{Al}_2[\text{Si}_2\text{O}_5](\text{OH})_4$. Kaolinite, like other clay minerals, usually forms tiny (submicroscopic) flakes that occur in aggregates which are soft ($H = \sim 2$). It is mostly formed by the alteration of feldspars and is an important component of many soils. Kaolinite is just one of many clay minerals that are beyond the scope of this text.

4.1.5.4 Micas

The mica minerals form an important group of layer silicates which are all characterised by perfect basal cleavage. The layered structure is quite “open” so that the large K^+ cation can be accommodated. They have the general formula:



Micas are hydrous minerals i.e. they contain (OH)-groups, and there is quite extensive substitution of Al for Si. Each SiO_4 -tetrahedron shares 3 oxygens with its neighbour to form 6-sided units in a two dimensional sheet-like structure. The layers are, however, not stacked directly one on top of each other so that micas are monoclinic (Fig. 4.9). Pseudo-hexagonal forms are, however, commonly developed; stacks of mica flakes are often referred to as “books”. The name mica was probably derived from the Latin *micare* meaning to shine.

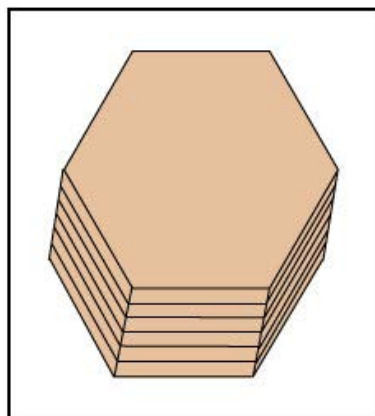


Fig. 4.9: Micas commonly form 6-sided crystals where the basal cleavage is a dominant feature.

The 6-sided flakes are not located directly on top of each other so that the crystal symmetry is not hexagonal but monoclinic.

4.1.5.4.1 Muscovite

Muscovite is the most common colourless mica (Picture 4.8) with the composition $\text{KAl}_2[\text{AlSi}_3\text{O}_{10}](\text{OH})_2$. Note that Al occurs in two sites – one replacing Si (with coordination number (CN) = 4) and one slightly larger site, linking the $[\text{AlSi}_3\text{O}_{10}]$ layers together (CN = 6). The large site occupied by K has CN = 12. Fluorine (F) can occur replacing some of the (OH) – groups.

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Picture 4.8: Flakes of brownish-black biotite (top) and colourless muscovite (bottom) illustrating their perfect basal cleavage.

Muscovite is a soft mineral ($H = \sim 2$) which forms elastic flakes with $G = \sim 2.9$. It occurs in some granites and granitic pegmatite where it can form meter-sized plates. It is common in metamorphosed clay-rich sediments called mica schists. Here the muscovite defines a foliation that develops in response to pressure during metamorphism. Large muscovite flakes used to be used instead of glass in Russia when it became known as “Muscovy-glass”, which is probably how it got its name. A Li-bearing variety of mica which is pink to purple in colour is known as lepidolite.

4.1.5.4.2 Biotite

The composition of biotite (dark mica) is similar to muscovite except that the Al with CN = 6 is replaced by Mg and Fe. This gives $K(Mg,Fe)_3[AlSi_3O_{10}](OH,F)_2$. The structure is similar to that of muscovite, forming pseudohexagonal flakes. It is slightly harder ($H = 2.5-3$) and denser ($G = 2.9-3.4$, increasing with Fe-content) than muscovite. Mg-rich varieties are called *phlogopite* and are brown, whereas Fe-rich varieties (biotite) are black (Picture 4.8).

Biotite occurs in small quantities in many plutonic rocks i.e. in rocks that crystallized slowly from magma at some depth below the surface of the Earth. Its composition shows that (OH)-groups are essential for its formation i.e. the magma must be hydrous. The amount of H_2O that can be dissolved in magma depends to some extent on the confining pressure. Magma at the surface of the Earth (= lava) cannot contain much water in solution. Hydrous minerals are therefore not common in volcanic rocks but can form in plutonic rocks. This explains why, for example, micas are common in granite but rare in its volcanic equivalent called rhyolite.

Biotite is also very common in many metamorphic rock types and is a major component of mica schists. Books of biotite are common in granitic pegmatites. It is named after a French physicist, J.B. Biot.

4.1.5.5 Chlorite

The chlorite group of minerals has the composition $(\text{Mg,Fe,Al})_6[(\text{Si,Al})_4\text{O}_{10}](\text{OH})_8$. Like other layer silicates, chlorite has a perfect basal cleavage and is fairly soft ($H = 2-2.5$). Chlorites are usually green (it is named from the Greek chloros meaning green) and are formed by the alteration of other silicate minerals that contain Mg and Fe (e.g. olivine, augite, hornblende, biotite). These mineral reactions, which take place at temperatures in the range $\sim 100-500^\circ\text{C}$, require the presence of a hydrous phase.

Chlorite is a widespread “late-stage” mineral. For example, it commonly fills “holes” in volcanic rocks called vesicles that are formed as a result of the escape of a gas phase from the magma at low pressure. It is a common vein-filling mineral in many rock types. It is a major component of greenschists (basalts metamorphosed at $300-500^\circ\text{C}$), together with, amongst other minerals, epidote.

4.1.6 Tectosilicates

The silicate structures in this group are based on a 3-dimensional framework of SiO_4 -tetrahedra in which all the four corner O^{2-} anions are shared with neighbouring tetrahedra. When all tetrahedra have Si^{4+} at their centres, the O^{2-} anions are all valency-satisfied and the SiO_2 unit is therefore electrically neutral. SiO_2 is, of course, the composition of quartz. No other compositions would be possible if it were not for the fact that some of the Si^{4+} cations can be replaced by Al^{3+} . This gives rise to a wide variety of minerals, including most importantly the feldspars. Framework silicates (mostly feldspars and quartz) make up about 64% of the continental crust and are therefore very important minerals in geology. The proportions of quartz and feldspar are used to classify most igneous rocks, as we will see later.



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4.1.6.1 Quartz

Quartz, SiO_2 , is trigonal and forms prismatic 6-sided crystals (Picture 1.1). It defines hardness = 7 on Mohs' scale and $G = 2.65$. It has a vitreous (glassy) lustre. Colourless crystals are the most usual but many coloured varieties occur. Quartz has a conchoidal fracture i.e. curved fracture surfaces. The names commonly given to some of the coloured, coarsely crystalline varieties, have been mentioned in section 2.2.4.

Quartz also occurs in microcrystalline varieties that appear to be amorphous. Their crystalline nature is only revealed by powerful microscopes or X-ray studies. The general term for microcrystalline varieties of quartz is *chalcedony*. It is commonly deposited from aqueous solutions and is frequently found lining or filling cavities in rocks. Colour and banding give many varieties:

<i>carnelian</i>	–	red chalcedony
<i>chrysoprase</i>	–	green chalcedony
<i>agate</i>	–	layered with different colours. Many agates sold commercially have been artificially coloured. Moss agate has moss-like patterns.
<i>onyx</i>	–	a layered variety in which the layers are planar and parallel
<i>flint and chert</i>	–	grey to black compact varieties
<i>fossilised wood</i>	–	has commonly been silicified (replaced by microcrystalline quartz)
<i>opal</i>	–	has the composition $\text{SiO}_2 \cdot n\text{H}_2\text{O}$ and is of the few amorphous minerals

Quartz is in fact only one of several polymorphs of SiO_2 . Two naturally occurring high temperature polymorphs are tridymite and cristobalite. Two high pressure polymorphs are coesite and stishovite. Coesite is formed from quartz by, for example, meteorite impact. Stishovite has been formed as a result of the extremely high local pressure produced by underground atomic explosions.

Quartz is an extremely widespread mineral in the continental crust. It is the main component of yellow beach sand. It is an essential component of many igneous and metamorphic rocks. Under their breakdown by weathering processes, quartz survives and is therefore a major component of many sedimentary deposits.

4.1.6.2 Feldspars

The feldspars form an extremely important group of minerals. Their compositions can be expressed in terms of three end members involving the cations K^+ , Na^+ and Ca^{2+} :

$\text{K}[\text{AlSi}_3\text{O}_8]$	–	ORTHOCLASE (Or)
$\text{Na}[\text{AlSi}_3\text{O}_8]$	–	ALBITE (Ab)
$\text{Ca}[\text{Al}_2\text{Si}_2\text{O}_8]$	–	ANORTHITE (An)

Feldspars with compositions intermediate between $\text{K}[\text{AlSi}_3\text{O}_8]$ and $\text{Na}[\text{AlSi}_3\text{O}_8]$ are known as the **ALKALI FELDSPARS**.

Those between $\text{Na}[\text{AlSi}_3\text{O}_8]$ and $\text{Ca}[\text{Al}_2\text{Si}_2\text{O}_8]$ are the **PLAGIOCLASE FELDSPARS**.

Note that albite is therefore an end member in both feldspar series. The compositional variation in the feldspars can be expressed in terms of a triangular diagram (Fig. 4.10). There is more solid solution at high than at low temperature.

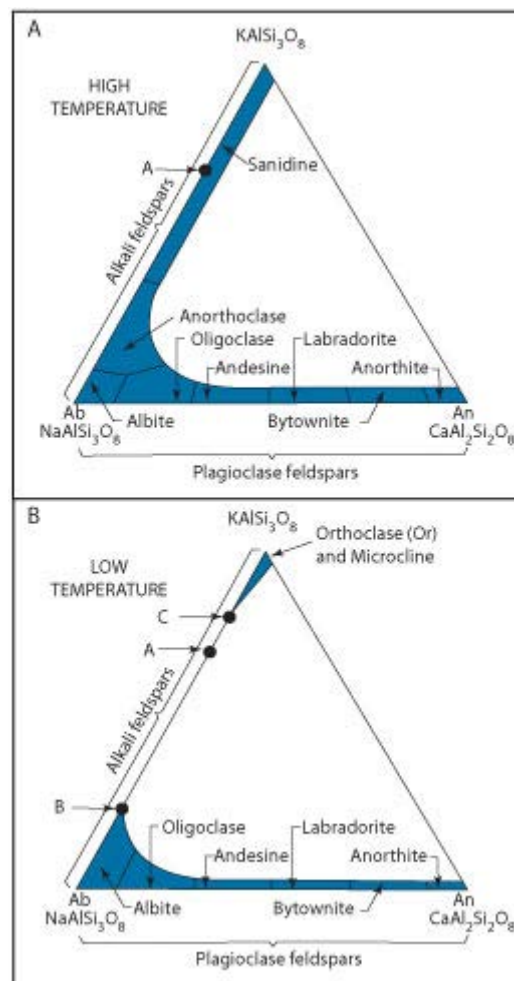


Fig. 4.10: Triangular diagrams showing the compositions of the feldspar minerals at high and low temperature. The amount of solid solution in the feldspars is much greater at high temperature (A) than at low temperature (B).

4.1.6.2.1 Alkali feldspar

The K-rich end-member of the alkali feldspar series occurs in three polymorphs, *orthoclase* (monoclinic) (Fig. 4.11), *microcline* (triclinic) and *sanidine* (monoclinic). Orthoclase and microcline are low temperature forms; sanidine is the high temperature polymorph. Na-rich alkali feldspars (with >10% Or) are called *anorthoclase*.

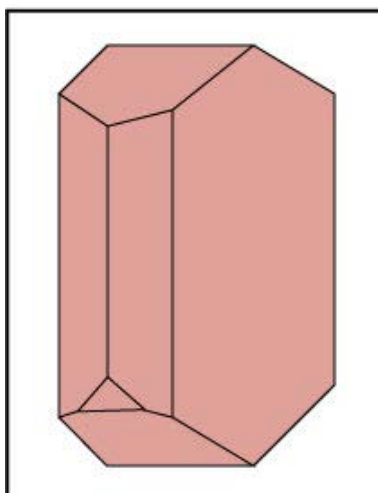


Fig. 4.11: Illustration of a typical orthoclase crystal.

At high temperatures there is complete solid solution in the alkali feldspars (i.e. Na^+ and K^+ are completely interchangeable in the feldspar structure) but on cooling they split into two separate phases – one K-rich and one Na-rich. The K-rich phase orthoclase is normally dominant and veins or patches of albite are *exsolved* from the original homogeneous feldspar (Figs. 4.12 & 4.13). The coexistence of two *intergrown* phases is commonly visible in hand specimens of slowly cooled alkali feldspars; such intergrowths are known as *perthite* (Picture 4.9).

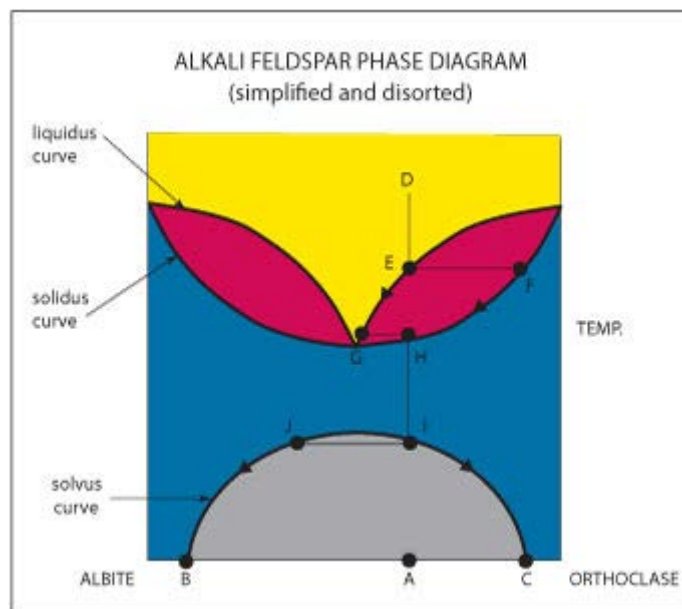


Fig. 4.12: Phase relations in the alkali feldspar system.

Melt D begins to crystallize Or-rich feldspar F when it reaches the liquidus at E. On cooling melt E \rightarrow G and crystals F \rightarrow H. Crystals H on fast cooling will remain homogeneous (sanidine). On slow cooling they will split into two (I & J) when they reach the solvus curve. On further cooling they will change in composition J \rightarrow B and I \rightarrow C. The bulk composition will be A (I=H=D) but the final product will be PERTHITE consisting of exsolved albite – rich feldspar (B) in an orthoclase – rich host (C) as illustrated in Fig. 4.13.

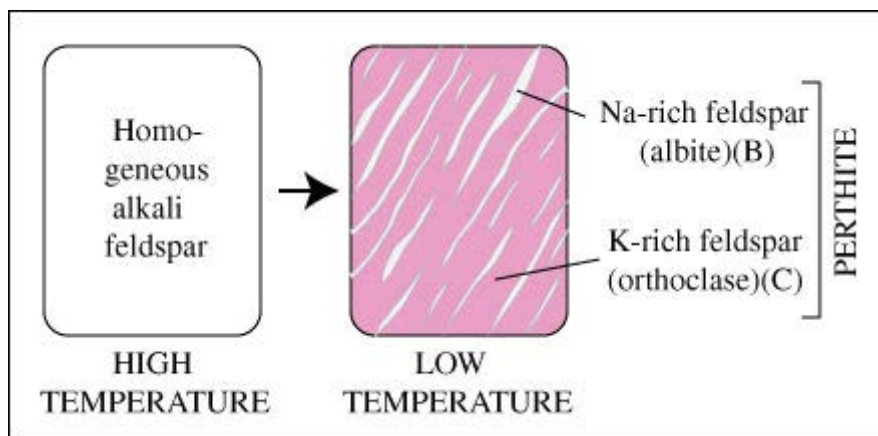


Fig. 4.13: Development of perthite in alkali feldspars.

Homogeneous alkali feldspar at high temperature (A in Fig. 4.10A) splits into two phases during cooling (B and C in Fig. 4.10B and Fig. 12).



Picture 4.9: Alkali feldspar with well-developed perthite structure. The pinkish host phase is rich in orthoclase (C in Figs. 4.12 & 4.13) whereas the whitish veins are rich in albite (B in Figs. 4.12. & 4.13).

Orthoclase and microcline have perfect (001) and good prismatic (010) cleavage, and readily form roughly rectangular-shaped cleavage fragments. They define hardness = 6 on Mohs scale and have $G \sim 2.57$. Their colour is usually white to pale yellow or grey. Pink to red varieties are due to the presence of minute flakes of hematite (Fe_2O_3). Green microcline is known as amazonite. Twinning is frequently developed; the most common type is called Carlsbad twinning.

Microcline and/or orthoclase are essential components of many igneous rocks such as the plutonic rock types granite and syenite. Their volcanic equivalents (rhyolite and trachyte respectively) contain the high temperature polymorph sanidine. Microcline and/or orthoclase are also important in many metamorphic rocks, particularly in gneisses. K-feldspar is widely used as a component in the manufacture of ceramics.

4.1.6.2.2 Plagioclase feldspar

There is complete solid solution in the plagioclase series, like in olivine (section 4.1.1.1). Anorthite is the high-temperature end member (melts at $\sim 1560^{\circ}\text{C}$) and albite the low-temperature (melts at 1118°C) one in a cigar-shaped phase diagram. Individual plagioclase feldspars are given specific names (Fig. 4.10). Expressed in terms of the % anorthite (An) end member these are:

An ₀₋₁₀	ALBITE
An ₁₀₋₃₀	OLIGOCLASE
An ₃₀₋₅₀	ANDESINE
An ₅₀₋₇₀	LABRADORITE
An ₇₀₋₉₀	BYTOWNITE
An ₉₀₋₁₀₀	ANORTHITE

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One of the consequences of fractional crystallization (i.e. when crystals and melt do not keep in equilibrium during the crystallization of magma) is that zoning can develop. This is uncommon in olivines but is very common in plagioclase feldspars. This is because when plagioclase changes composition during reaction with the melt it requires a coupled reaction involving $\text{Na}^+ + \text{Si}^{4+} = \text{Ca}^{2+} + \text{Al}^{3+}$. Not only does some calcium become replaced by sodium in a site with coordination number 6-8, but in order to maintain electronic neutrality, some aluminium has to be replaced by silicon in the tetrahedral site (CN = 4). This is a slow process, and often results in incomplete reaction so that compositional zoning of plagioclase crystals results with Ca-rich cores and Na-rich margins.

All members of the plagioclase series are triclinic (Fig. 4.14). Crystals are commonly tabular parallel to (010). Repeated twinning (also called multiple or polysynthetic twinning) parallel with (010) is extremely common and is sometimes visible in hand specimen. The hardness of plagioclase feldspars is close to 6; the density increases with Ca-content from 2.62 for albite to 2.76 gm/cm^3 for anorthite. Plagioclases can be colourless, white or grey. A beautiful play of colours (called labradorescence) is seen in some plagioclase crystals (Picture 4.10).

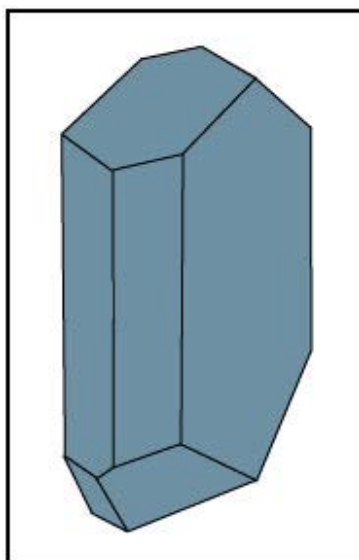


Fig. 4.14: Illustration of a perfect crystal of albite.
All members of the plagioclase series are triclinic.

Plagioclase feldspars are even more widely distributed than alkali feldspars. The classification of igneous rocks is to a large extent based on the proportions of plagioclase to alkali feldspar. Amongst volcanic rocks plagioclase is essential in, for example, basalt that is the most common rock type of all (ocean floors are formed of basalt, usually below a thin layer of sediments). Plagioclase is therefore also a major component of the plutonic equivalent of basalt, called gabbro. The composition of plagioclase in igneous rocks varies with the temperature of formation. In keeping with the phase diagram, Ca-rich plagioclases form at higher temperatures than Na-rich ones. For example, gabbros typically contain labradorite-andesine whereas granites typically contain oligoclase. Plagioclase feldspars are also important in many metamorphic rocks.



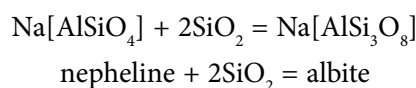
Picture 4.10. Play of colours (labradorescence) in plagioclase. The blue area is slightly more albite-rich than the rest. Repeated twinning is visible on either side of the blue area.

4.1.6.3 Feldspathoids

The feldspathoids are anhydrous framework silicates that are compositionally related to the alkali feldspars. The main difference is their SiO_2 content; feldspathoids contain less SiO_2 than the feldspars. The two most important feldspathoid minerals are nepheline and leucite.

4.1.6.3.1 Nepheline

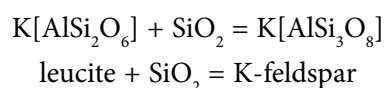
Nepheline has the composition $\text{Na}[\text{AlSiO}_4]$ and is related to albite thus:



Nepheline cannot coexist in equilibrium with quartz. It is hexagonal, usually colourless to grey, hardness = 5.5 – 6 and with a relatively low density ($\sim 2.63 \text{ gm/cm}^3$). It has a greasy lustre. Nepheline occurs in plutonic and volcanic silica-poor igneous rocks and (like leucite) is used for the classification of these.

4.1.6.3.2 Leucite

Leucite ($\text{K}[\text{AlSi}_2\text{O}_6]$) is related to K-feldspar thus:



Leucite forms white crystals with cubic symmetry from K-rich, SiO_2 -poor lavas. On cooling the symmetry inverts to tetragonal, but the external cubic form (typically trapezoidal i.e. with 24 faces; Fig. 3.20) is preserved. Leucite-bearing volcanic rocks are quite rare on a global scale. One famous location is Mt. Vesuvius, the volcano that erupted to destroy Pompeii in A.D. 79.

4.1.6.3.3 Other feldspathoids

There are several other feldspathoid minerals which are used in the classification of igneous rocks. Their compositions are related to nepheline but contain e.g. chlorine or sulphur in the structure. Most of these are cubic and they can be brightly coloured. Two of the most common are called sodalite and lazurite. The latter is deep blue. When lazurite occurs together with (typically) calcite and pyrite it forms a precious stone known as lapis lazuli.

4.1.6.3.4 Zeolites

Zeolites are aluminosilicates with a framework structure enclosing cavities occupied by large cations (Ca^{2+} , Na^+ , K^+) and water molecules. There are about 45 naturally occurring zeolites. A common occurrence is as vesicle-filling in volcanic rocks.

4.2 Non-silicate minerals

Most rocks are dominantly composed of silicate minerals. Some non-silicates are, however, also important rock-forming minerals. Limestone, for example, is mainly made of the mineral calcite (CaCO_3). The vast majority of economically important minerals are also non-silicates. Here we will briefly consider some of the commonest or most important non-silicates.



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4.2.1 Native elements

Gold, silver, copper and *platinum* all occur naturally in the form of native metal elements, but most of them are rare! Native non-metallic minerals include the two polymorphs of carbon, *diamond* and *graphite* (section 2.1.1), and the element *sulphur*. Native sulphur forms soft ($H \approx 2$), yellow, orthorhombic crystals where volcanic gases have been active.

4.2.2 Sulphides

The sulphides form an important group of minerals that include the majority of the ore minerals. They (nearly) all have metallic lustre and high densities.

4.2.2.1 Galena

Galena (PbS) forms cubic crystals, has perfect cubic cleavage, is fairly soft ($H = 2.5$) and has high G (7.5). Its colour and streak are dark grey. Galena often contains some silver in its structure and is an important ore for both Pb and Ag .

4.2.2.2 Sphalerite

Sphalerite (also known as zinc blende) (ZnS) is cubic and forms tetrahedral crystals with perfect cleavage. $H = 3.5-4$ and $G \sim 4$. Its lustre is non-metallic (it is sometimes adamantine) and it is commonly yellowish brown to black. It is often found together with galena.

4.2.2.3 Pyrite

Pyrite (FeS_2) is cubic and generally forms cubes or 12-sided crystals called pyritohedra (Fig. 3.19). Cube faces are commonly striated (Fig. 4.15). Cubes can grow together in an interpenetrating fashion. It is quite hard for a sulphide mineral ($H = 6-6.5$). $G = 5$. It has a very metallic lustre and is pale brassy yellow. The streak is black. It is sometimes referred to as “fools gold” (section 2.2.5). It is the most widespread sulphide mineral and occurs as an accessory mineral in many rock types.

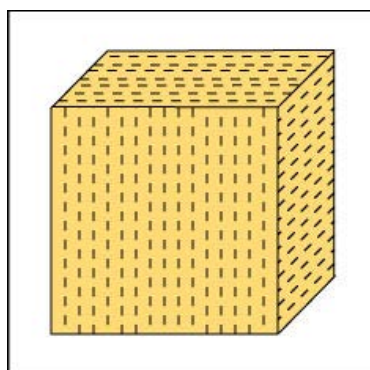


Fig. 4.15: Pyrite crystals commonly form perfect cubes. The cube-faces may show characteristic striations. The striations on opposite faces have identical orientations.

4.2.2.4 Chalcopyrite

Chalcopyrite (CuFeS_2) is tetragonal but usually occurs in a massive form. $H = 3.5\text{--}4$; $G = 4.2$. It has a metallic lustre and is brass-yellow, often with a tarnished appearance. It has a greenish-black streak. It is one of the most important ores of copper. Like pyrite it is sometimes called “fools gold”.

4.2.3 Oxides

4.2.3.1 Corundum

Corundum (Al_2O_3) is hexagonal and defines $H = 9$ on Mohs' scale (Picture 4.11). Its colour varies widely. Gem varieties include ruby (red) and sapphire (blue). Powdered corundum is used as an abrasive. It occurs in silica-poor igneous and metamorphic rocks.



Picture 4.11. Corundum forms 6-sided (hexagonal) crystals.
The reddish variety illustrated here is used commercially as ruby.

4.2.3.2 Hematite

Hematite (Fe_2O_3) is hexagonal and typically forms thin tabular crystals. Two common habits of hematite are kidney-like (reniform) (Picture 4.12) and micaceous or platy (specular). It is reddish-brown to black with $H = \sim 5$ and $G = 5.26$. Crystals have a metallic luster, but other varieties may be dull. It has a reddish-brown streak. Hematite is formed by the oxidation of other Fe-bearing minerals (note that hematite contains exclusively Fe^{3+}) and commonly imparts a reddish colouring to rocks. It is the most important iron ore for the manufacture of steel.



Picture 4.12. Hematite sometimes forms kidney-shaped aggregates.

4.2.3.3 Ilmenite

Ilmenite (FeTiO_3) is hexagonal but usually occurs in massive aggregates. It is black with a metallic to sub-metallic lustre, $H = 5.5-6$ and $G = 4.7$ and a black streak. It may be weakly magnetic. It is a very common accessory mineral in many rocks. It is the main sources of Ti metal and TiO_2 that is widely used as a white pigment in, for example, paper and paint.

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4.2.3.4 Magnetite

Magnetite and *chromite* are two minerals that belong to the *spinel* group. The general formula for spinels is XY_2O_4 in which X is a divalent and Y a trivalent cation. In the mineral spinel, $X = Mg$ and $Y = Al$ giving $MgAl_2O_4$. All spinels are cubic and form octahedral crystals (Fig. 3.18). In magnetite (Fe_3O_4) $X = Fe^{2+}$ and $Y = Fe^{3+}$. It forms black crystals with a metallic lustre. $H = 6$ and $G = 5.18$. It is strongly magnetic and occurs as an accessory mineral in many rock types.

4.2.3.5 Chromite

Chromite ($FeCr_2O_4$), like all the spinel minerals, is cubic. It forms small octahedra but massive occurrences are more usual. It is black with a dark brown streak. $H = 5.5$; $G = 4.6$. Chromite is one of the first minerals to crystallize from high temperature magmas and is common in, for example, ultramafic rocks. It is the only ore of chromium that is very important in steel production.

4.2.4 Chlorides and fluorides

4.2.4.1 Halite

Halite ($NaCl$) is a cubic mineral that has perfect cubic cleavage. Crystalline masses of halite are called rock salt. It has $H = 2.5$ and $G = 2.16$. The salty taste is unmistakable. Salt is soluble in water and handling of samples with sweaty hands gives the samples a smooth surface and imparts a salty taste to the fingers. Salt is formed by the evaporation of seawater and is usually accompanied in evaporite deposits by, for example, gypsum and calcite.

4.2.4.2 Fluorite

Fluorite (CaF_2) is a cubic mineral. A (111) cleavage is perfect and fluorite octahedra are common (Picture 4.13). $H = 4$; $G = 3.18$. Fluorite crystals are transparent to translucent, but the colour varies very widely (green, blue and yellow varieties are common). Colour banding can occur in individual crystals and in massive aggregates. Fluorite is commonly associated with ore minerals in veins.



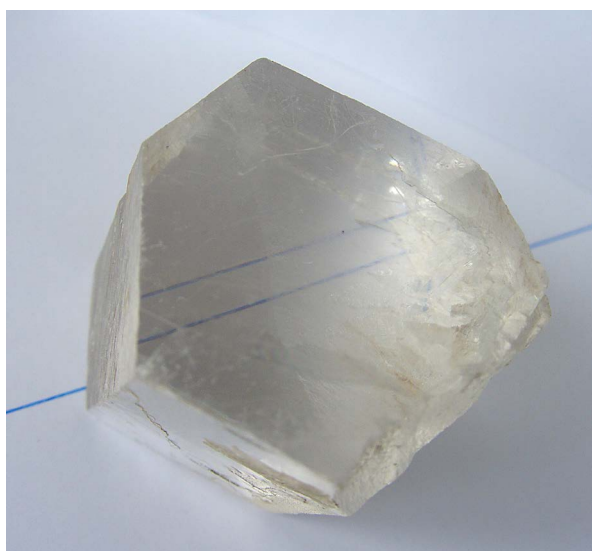
Picture 4.13. Octahedral purple fluorite. The octahedral faces are cleavage surfaces.

4.2.5 Carbonates

4.2.5.1 Calcite

Calcite (CaCO_3) is a trigonal mineral with a wide range of forms (more than 300 different varieties have been described). Rhombohedral cleavage is perfect, as illustrated in Fig. 3.25. Calcite defines $H = 3$; $G = 2.71$. Calcite is usually white, but many coloured varieties occur. It is readily identified by its crystal form, hardness, birefringence and reaction with dilute HCl (section 2.2.9.). Calcite is an extremely widespread mineral. It is the main component of limestone. Metamorphosed limestone (marble) is also composed of finely crystalline calcite. It occurs in many igneous and metamorphic rocks as an accessory mineral, and is found together with many ore minerals. The vast majority of igneous rocks are dominated by silicate minerals, but one very rare type is dominated by carbonates – these rocks are known as carbonatites. A very important property of calcite is its very high double refraction (birefringence) (Picture 4.14).

The minerals magnesite (MgCO_3) and siderite (FeCO_3) are related to calcite.



Picture 4.14. *Calcite has extremely high birefringence. This is evident here in that a single line appears to be double when viewed through a transparent calcite crystal.*

4.2.5.2 Aragonite

Calcite is in fact one of two minerals with the composition CaCO_3 . The other polymorph is *aragonite* which is orthorhombic. The shells of molluscs are largely formed of aragonite even though calcite is the stable phase at low temperatures and pressures.

4.2.5.3 Dolomite

Dolomite $\text{CaMg}(\text{CO}_3)_2$ is compositionally related to calcite and is also trigonal. With $H = 3.5\text{--}4$ it is somewhat harder than calcite and only reacts very slowly with cold HCl . It occurs in sedimentary rocks, particularly in dolomitic limestones. The metamorphic equivalents of these can contain a wide variety of Mg- and/or Ca-bearing minerals like forsterite, diopside, tremolite and talc.

4.2.6 Sulphates

4.2.6.1 Gypsum

Gypsum ($\text{CaSO}_4 \cdot 2\text{H}_2\text{O}$) forms monoclinic crystals which are usually tabular on (010). Swallowtail twins are common, as illustrated in Fig. 2.2a. It has a perfect (010) cleavage and defines $H = 2$. $G \sim 2.3$. Gypsum is colourless, white or grey; it can be transparent. Crystalline gypsum is also known as selenite. Satin spar is a fibrous variety; alabaster is a fine-grained massive variety. Gypsum is a widely distributed mineral in sedimentary rocks and is formed by the evaporation of seawater, usually together with halite – rock salt. The attractive aggregates known as “desert roses” are composed of gypsum. Gypsum also occurs together with many ore minerals in veins.

4.2.6.2 Barytes

Barytes (BaSO_4) usually forms colourless or white tabular orthorhombic crystals with perfect (001) cleavage. It has a notably high density for a non-metallic mineral ($G = 4.5$) and $H = 3\text{--}3.5$. Barytes occurs together with a variety of ore minerals in veins.

4.2.7 Phosphates

4.2.7.1 Apatite

Apatite ($\text{Ca}_5(\text{PO}_4)_3(\text{F,Cl,OH})$) is a hexagonal mineral which commonly forms long prismatic crystals. It is usually greenish in colour. It has hardness = 5 and can just be scratched with a knife. $G \sim 3.18$. Apatite is a very widely distributed accessory mineral but also occurs in large deposits, sometimes in pegmatites where it can form large green hexagons.

5 Igneous rocks

Petrology is the study of rocks that are naturally occurring aggregates of minerals. As we have seen, there are three main types of rocks: igneous, sedimentary and metamorphic. Here we are concerned with the study of igneous rocks.

5.1 Classification of igneous rocks

Igneous rocks can be classified according to many criteria, such as texture, crystal or grain size, colour, mineralogy, chemical composition, mode of occurrence, and genesis. Since it must be possible to give a rock a name in the field, a first order classification cannot be based on chemical composition since this first requires chemical analysis. Genetic classifications should be avoided as far as possible, but this is not always easy. For example, the main subdivision of rocks into sedimentary, igneous and metamorphic is genetic but cannot be avoided.

Igneous rocks are classified by their content of essential minerals i.e. those that make up the bulk of the rock – this is known as the *mode* of the rock. Minor amounts of accessory minerals are not considered. As we will see later, this is fairly straightforward for coarse grained rocks but can be very difficult for very fine grained ones – and impossible for glassy rocks.



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Two terms used in connection with grain size are:

phaneritic: individual crystals can be distinguished with the naked eye.

aphanitic: most of the individual crystals cannot be distinguished with the naked eye.

Igneous rocks are divided into two main groups on the basis of their field relations or on their grain size.

PLUTONIC – crystallized at depth. Phaneritic. Average crystal or grain size > 5mm (coarse); 1–5mm (medium); 0.5–1mm (fine grained).

VOLCANIC – extruded at the surface of the Earth. Aphanitic. Grain size < 0.5mm (very fine grained).

It is not unusual for igneous rocks to have large crystals of one or more mineral(s) in a finer grained *groundmass* (also called the *matrix*). In this case it is the grain size of the matrix that is used. For example, a rock with a black, fine grained matrix may contain cm-sized pyroxene crystals. The rock was extruded from a volcano as lava and is volcanic, despite the presence of some large crystals which crystallized in a magma chamber before eruption. Large crystals in a finer grained matrix are called *phenocrysts*. The presence of phenocrysts in a finer grained matrix is referred to as a *porphyritic* texture.

5.1.1 Plutonic rocks

Plutonic rocks are classified according to their *modal* mineral content i.e. actual mineral content in volume %. The classification (from 1976) is known as the “Streckeisen” system after the Austrian professor who was chairman of an international committee to systematise the naming of igneous rocks. Before this there was considerable international confusion and disagreement as to how igneous rocks should be named.

For the naming of plutonic rocks, five groups of minerals are used. These are:

- Q quartz
- A alkali feldspar – orthoclase, microcline, perthite, anorthoclase and albite (An_{00-05})
- P plagioclase (An_{05-100})
- F feldspathoid minerals or FOIDS – nepheline, leucite, sodalite etc.
- M mafic minerals. “Mafic” stands for Mg-Fe minerals and includes olivine, pyroxenes, amphiboles, micas, garnets, oxide and sulphide minerals etc.

Q, A, P and F are light coloured minerals. M are dark minerals. The percentage of dark minerals (%M) is often referred to as the COLOUR INDEX of a rock. The light coloured minerals are referred to as *felsic* minerals from **f**eldspar and **s**ilica. The word *mafic* is constructed from the **m**agnesium – **f**errous (= iron) nature of the relevant minerals.

The primary division of plutonic rocks is based on the amount of mafic minerals. When dark minerals make up more than 90% (i.e. $M > 90\%$) the rocks are ULTRAMAFIC and are classified according to the mafic minerals present. But the majority of igneous rocks have $M < 90\%$ and we will start with these.

5.1.1.1 $M < 90\%$ (QAPF double triangle)

The light coloured (felsic) minerals $Q + A + P + F$ are calculated to 100%, ignoring the dark minerals. Since quartz and feldspathoid minerals cannot coexist, the light minerals will in fact be:

$$Q + A + P = 100 \text{ or } F + A + P = 100$$

These are then plotted in the QAPF double triangle (Fig. 5.1). The QAPF double triangle is divided into 15 fields with different rock names. The fact that the QAPF double triangle is used for all plutonic rocks with $M < 90\%$ means that even if a rock consists of, for example, 89% olivine and 11% plagioclase, it is named according to the 11% plagioclase.

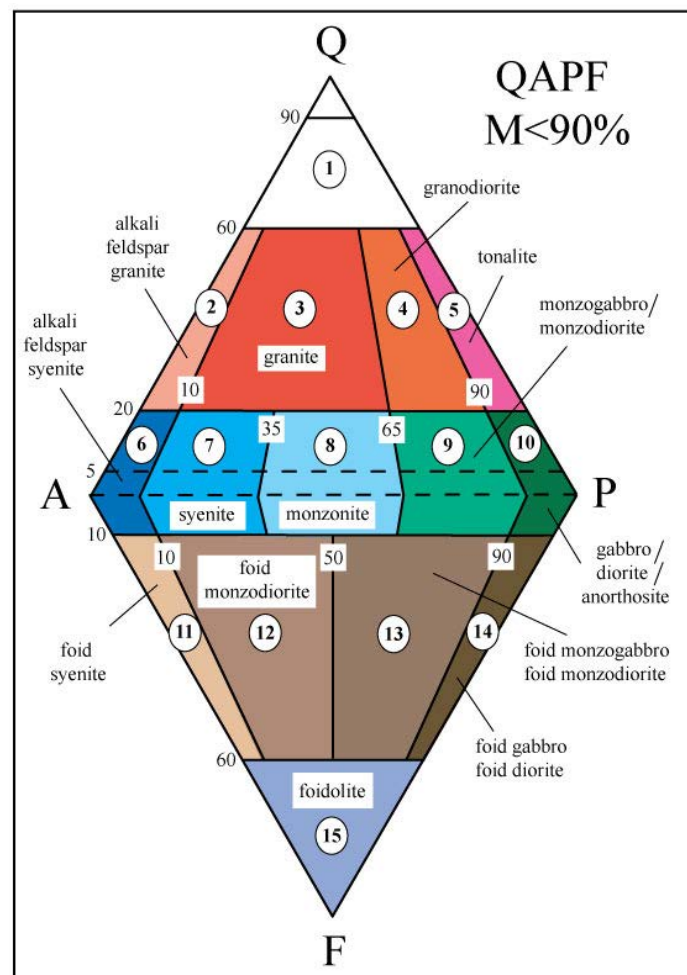


Fig. 5.1: The QAPF double triangle for the classification of plutonic rocks

This is known as the Streckeisen system of classification and is used for plutonic rocks with less than 90 vol. % dark (mafic) minerals.

The following notes refer to the numbered fields in Fig. 5.1. The most important names are in *italics*.

1. Igneous rocks with > 60% quartz amongst the light minerals are very rare. Rocks with > 90% are quartzolites. With 60–90% quartz the term quartz-rich is used as a prefix to the appropriate name from field 2, 3, 4 or 5. e.g. quartz-rich granodiorite.
2. Alkali feldspar granites are, as the name implies, granitic rocks which are rich in alkali feldspar and poor in plagioclase.
3. The *granite* field is the largest one in the QAP triangle and granites are very common in the continental crust. Note that granites can contain 20–60% quartz (ignoring dark minerals) and that the ratio of alkali feldspar to plagioclase can vary from 9:1 to ~1:2 (actually 35:65). Fields 2, 3 and 4 are all “granitic rocks”. The term “granite” *sensu strictu* is exclusively used for rocks that lie in field 3.
4. *Granodiorite* is the name given to rocks that are more plagioclase-rich than true granites. They can be considered as having a composition essentially intermediate between granites (field 3) and diorites (field 10)
5. Rocks dominated by quartz and plagioclase are tonalites.
6. Rocks that are dominantly composed of alkali feldspar are alkali feldspar syenites. If there is 5–20% quartz the term quartz alkali feldspar syenite is used. With up to 10% feldspathoid mineral (e.g. nepheline) the rock is e.g. a nepheline-bearing alkali-feldspar syenite.
7. *Syenite* (see point 6 for quartz syenite and nepheline-bearing syenite).
8. *Monzonite* (see point 6 for quartz and nepheline-bearing types).
9. *Monzodiorite* or *monzogabbro* (see point 6 for quartz and nepheline-bearing types). For distinction between monzodiorite and monzogabbro see point 10.
10. This field is used for three rock types – *diorite*, *gabbro* and *anorthosite*. Anorthosites consist of > 90% plagioclase. Distinction between diorite and gabbro is usually based on the composition of the plagioclase in the rock. Gabbro contains plagioclase with $An_{>50}$; diorite has $An_{<50}$. This is, of course, difficult to determine in hand specimen! Diorites are usually more felsic than gabbros. Diorites typically contain andesine (plagioclase in the range An_{30-50}) + hornblende ± biotite ± clinopyroxene. Gabbros contain, for example, labradorite (plagioclase An_{50-70}) + clinopyroxene ± orthopyroxene ± olivine. Gabbroic rocks are further classified according to their dark minerals (section 5.1.1.1.1).
11. Rocks in this field are foid syenites. If the foid (= feldspathoid) mineral is, for example, nepheline, the rock is called a *nepheline syenite*.
12. Foid monzosyenite e.g. nepheline monzosyenite.
13. Foid monzodiorite or foid monzogabbro. e.g. nepheline monzodiorite.
14. Foid gabbro or foid diorite e.g. nepheline gabbro.
15. Rocks with > 60% foid minerals (amongst the light minerals) are rare and are called foidolites (e.g. nephelinolite). The term nephelinite seems more obvious but is used for equivalent volcanic rocks.

Not all these rock types are equally common. The most important types are rocks of field 10 (particularly gabbros and diorites), granitic rocks (fields 2, 3 and 4) and various syenitic rocks (fields 6, 7 and 11). The intermediate rock types (monzo-) and F-rich types occur less frequently.

Quartz-bearing rocks lie in the upper (QAP) triangle and are silica-oversaturated. Rocks that lie on the AP boundary are silica-saturated. Those that lie in the APF triangle contain a foid mineral and are silica-undersaturated.

5.1.1.1.1 Gabbroic rocks

Gabbroic rocks, which lie in field 10 in the QAPF classification, consist dominantly of plagioclase, clinopyroxene (typically augite), olivine and orthopyroxene (commonly hypersthene). Some special names are used for rocks belonging to the gabbro family (Table 5.1).

Main minerals	Rock name
clinopyroxene + plagioclase	gabbro
orthopyroxene + plagioclase	norite
olivine + plagioclase	troctolite

Table 5.1. Rocks belonging to the gabbro family



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Intermediate types are quite common. For example, a rock consisting of clinopyroxene + orthopyroxene + plagioclase is called a gabbro. One consisting of olivine + clinopyroxene + plagioclase is an olivine gabbro, and so on.

Prefixes are commonly used depending on the colour index (M). These are *mela-* (short for “melanocratic”) when M is in the range 65–90 e.g. melatroctolite; *meso-* when M is 35–65 e.g. mesonorite; and *leuco-* when M is 10–35 e.g. leucogabbro.

The “meso-” prefix is not often used since it refers to the “normal” modal composition. Note that if $M > 90$ the rock classifies as *ultramafic*, and if $M < 10$ in field 10 (i.e. plagioclase > 90%) the rock is an *anorthosite*.

The prefixes *mela-* and *leuco-* are used for all rock types which are relatively enriched in dark or light minerals respectively. e.g. melasyenite, leucogranite.

5.1.1.1.2 Charnockitic rocks

A special nomenclature is used for plutonic rocks that contain orthopyroxene. This is because these rocks have a special mode of occurrence and tend to be found related to one another. We have already encountered one rock type that belongs to the charnockitic family – *norite*. The term *charnockite* is used for hypersthene granite. Other charnockitic rocks types include mangerite (= hypersthene monzonite) and jotunite (= hypersthene monzonorite). Charnockitic rocks typically occur associated with large bodies of anorthosite in which the dominant mafic mineral is orthopyroxene.

5.1.1.2 M > 90% (Ultramafic rocks)

Ultramafic rocks are usually dominated by one or more of the minerals olivine, orthopyroxene and clinopyroxene (Fig. 5.2). Ultramafic rocks with > 40% olivine are *peridotites*. Various types of peridotite are *dunite* (> 90% olivine), *harzburgite* (olivine + orthopyroxene), *wehrlite* (olivine and clinopyroxene), and *lherzolite* (olivine + two pyroxenes).

Rocks with > 60% pyroxene are *pyroxenites*. Pyroxenites are divided into 6 types: orthopyroxenite (> 90% orthopyroxene), clinopyroxenite (> 90% clinopyroxene), websterite (two pyroxenes), olivine orthopyroxenite, olivine clinopyroxenite and olivine websterite.

Common accessory minerals in ultramafic rocks are garnet or spinel.

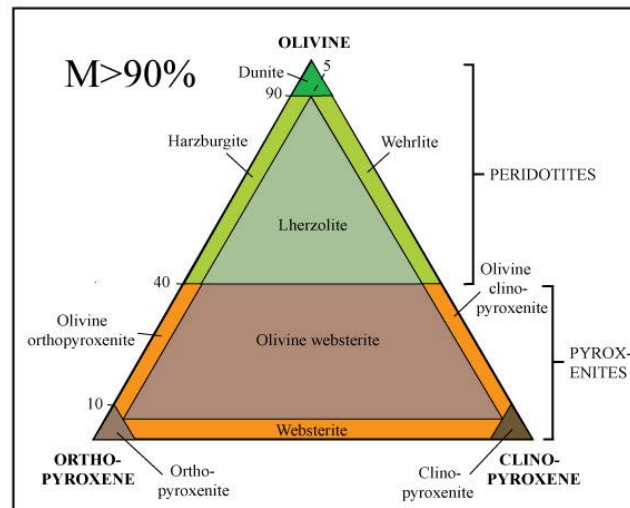


Fig. 5.2: Classification of ultramafic rocks composed of olivine, orthopyroxene and clinopyroxene. Ultramafic rocks are defined as containing more than 90 vol. % dark (mafic) minerals.

Hornblende may be important to give various types of hornblendites and e.g. hornblende peridotites. We will not consider these further.

The upper mantle is composed of peridotite. Mafic minerals usually crystallize before plagioclase from basaltic magma. Ultramafic rocks can therefore form as the early crystallization products. It is important to note that ultramafic magmas are extremely rare so that there is no classification system for volcanic rocks equivalent to the ultramafic classification for plutonic rocks.

5.1.1.3 Very coarse grained plutonic rocks

Very coarse grained igneous rocks are called *PEGMATITES*. The most common are granitic pegmatites, but all plutonic rocks can occur in pegmatitic varieties e.g. gabbro pegmatite, nepheline syenite pegmatite etc. Pegmatites are commonly very attractive rocks because of the large grain size of the crystals (up to several metres long in extreme cases). They can also be of economic interest because of the presence of exotic minerals containing rare elements.

5.1.1.4 Rocks intermediate between plutonic and volcanic

The main subdivision of igneous rocks is into those which have crystallized slowly at depth (plutonic) and those which cooled rapidly at the surface (volcanic). Lava flows are, of course, volcanic, but the centre of a thick lava flow may cool so slowly that a phaneritic rock results. A hand specimen of the centre of a thick basaltic lava flow may therefore be identified as a gabbro. Some magmas cool at shallow depths and form rocks which can be considered as texturally intermediate between plutonic and volcanic. For example, feeders to lava flows may have glassy margins (called a chilled margin because it is formed by rapid cooling against cold wall rocks) but phaneritic centres.

Dykes and *sills* (these are dealt with later) are examples of minor intrusions that cooled near the surface of the Earth. The igneous rock types in minor intrusions are compositionally the same as those that cooled more slowly at depth, but will typically be relatively fine grained (e.g. 0.3–0.8 mm). This can be expressed by using the prefix “micro-” which refers to grain sizes intermediate between those of typical plutonic and volcanic rocks. Examples are *microgabbro* and *microgranite*.

There are some terms that are specifically used for rocks that are texturally intermediate between plutonic and volcanic equivalents. They typically occur in minor intrusions (most commonly in dykes and sills), near the margins of larger intrusions (which cooled relatively quickly) and near the centres of thick lava flows (which cooled relatively slowly). Some of these terms are:

dolerite and *diabase* synonyms (originally English and American respectively) for *microgabbro*. These terms are very commonly used for dykes and sills of appropriate composition.

granophyre a granitic rock in which the groundmass quartz and alkali feldspar occur in a micrographic intergrowth.

aplite fine grained granitic rocks which typically occur as dykes, veins or patches in or associated with granitic intrusions.

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lamprophyre	this porphyritic, usually melanocratic, rock type usually occurs in minor intrusions. Feldspar/feldspathoids only occur in the groundmass; biotite \pm amphibole are common; hydrothermal alteration of mafic minerals is widespread; calcite is a common primary mineral. Lamprophyres are geochemically distinct.
lamproite	similar to lamprophyre but with more extreme chemical composition.
<i>kimberlite</i>	ultramafic minor intrusions which typically occur as pipes. Important minerals are olivine (largely serpentinised), Mg-rich mica (phlogopite), pyroxenes, garnet and carbonates. Kimberlites are formed at very high pressure and are the main source of <i>diamonds</i> .

5.1.1.5 Carbonatite

The overwhelming majority of igneous rocks are composed of silicate minerals. A tiny minority of igneous rocks (that occur both as plutonic and volcanic varieties) consist dominantly of carbonate minerals. The carbonate mineral can be calcite (CaCO_3), dolomite ($\text{CaMg}(\text{CO}_3)_2$) or siderite (FeCO_3). Different names are used depending on the composition of the carbonate mineral.

5.1.2 Volcanic rocks

Volcanic rocks have, by definition, an aphanitic groundmass but their classification is more complicated than that of plutonic rocks. As we have seen, the latter are classified according to the minerals present. This is often difficult or impossible to determine for volcanic rocks. When it is possible a QAPF double triangle is used, but in many cases (e.g. when a rock is extremely fine grained) a chemical analysis is required and the so-called *TAS classification* is used (TA = total alkalies $\text{Na}_2\text{O} + \text{K}_2\text{O}$; S = silica). Many volcanic rocks are formed by explosive activity, resulting in the formation of fragmentary rocks – the so-called pyroclastic rocks (pyro = heat; clast = fragment). This is also taken into consideration in the naming of volcanic rocks.

Ultramafic volcanic rocks are extremely uncommon. Ultramafic plutonic rocks are formed by the separation of mafic minerals (commonly olivine and/or pyroxenes) from magmas. The magma itself is not ultramafic.

5.1.2.1 QAPF double triangle for volcanic rocks

This classification (Fig. 5.3) can only be used if the mineral mode can be estimated. In many cases it is based exclusively on the phenocryst phases present. The numbers of the fields are the same as those for the plutonic rock classification (Fig. 5.1).

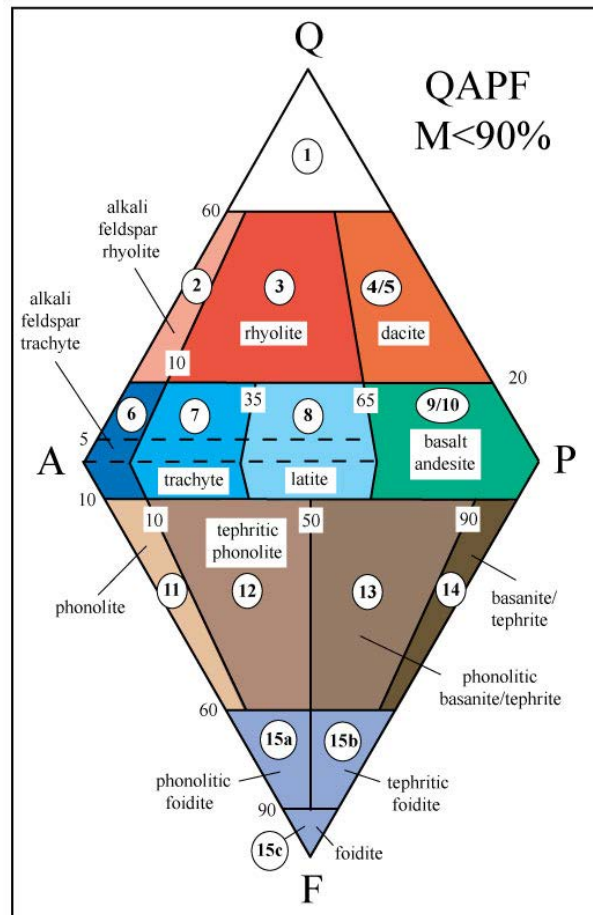


Fig. 5.3: The QAPF double triangle for the classification of volcanic rocks.

This classification cannot be used if the rock is too fine grained to allow mineral identification. It may be based on phenocrysts (if present). When this classification cannot be used it is necessary to use the TAS diagram (Fig. 5.4).

Comments on the individual fields in Fig. 5.3 are given below. The most important rock names are in *italics*.

1. There are no volcanic rocks in this field.
2. The term alkali feldspar rhyolite corresponds to alkali feldspar granite.
3. *Rhyolite* is the volcanic equivalent of granite.
4. & 5. Fields 4 and 5 are covered by *dacite* which is the volcanic equivalent of both granodiorite and tonalite.
6. Alkali feldspar trachyte. With minor (5–20%) quartz the term quartz alkali feldspar trachyte is used; with minor foid (0–10%) the term e.g. nepheline-bearing alkali feldspar trachyte is used.
7. *Trachyte* is the volcanic equivalent of syenite. See point 6 for quartz- and foid-bearing varieties.
8. *Latite* is the volcanic equivalent of monzonite. See point 6 for subtypes.

9. & 10. These fields contain the large majority of volcanic rocks – *basalts* and *andesites*. These are broadly the volcanic equivalents of gabbros and diorites respectively. Distinction can be made using the colour index (basalt has $M > 35$; andesite $M < 35$). The plagioclase composition is difficult to apply because andesites commonly contain phenocrysts of calcic plagioclase (labradorite or bytownite) and compositional zoning is commonly developed in the plagioclase in volcanic rocks. The TAS classification is very widely used for dark volcanic rocks (which includes basalts and andesites).
11. *Phonolites* are the volcanic equivalents of foid syenites e.g. nepheline syenites. In volcanic rocks leucite ($K[AlSi_2O_6]$) may be an important foid mineral.
12. Tephritic phonolites are relatively rare.
13. Phonolitic basanites and phonolitic tephrites are relatively rare. For distinction between them see point 14.
14. Basanites and tephrites belong to this field. They are separated by the amount of olivine present. Basanite has $> 10\%$ olivine; tephrite has $< 10\%$ olivine. Because these rocks are very fine grained it is not possible to determine the amount of olivine optically (even with a microscope) and the amount of olivine is estimated on the basis of the chemical composition of the rock which is recalculated in terms of simple mineral proportions – this is called the normative composition of a rock (hypothetical mineral composition of a rock calculated from its chemical composition) and is not to be confused with its modal composition (volume % minerals actually present).
15. This field is subdivided into three subfields:
 - 15a: Phonolitic foidite e.g. phonolitic nephelinite
 - 15b: Tephritic foidite e.g. tephritic leucitite
 - 15c: Foidite e.g. nephelinite

Volcanic rocks commonly have holes formed by escaping expanding gas during solidification. These are called *vesicles* and the rock is vesicular. Vesicles are commonly filled by later material; they are then referred to as *amygdales* and the rock is amygdaloidal.

Some special terms are used for glassy rocks:

OBSIDIAN	Massive glass. Commonly black. The most common type of obsidian is rhyolitic.
PERLITE	Pearly grey glass with concentric curved fractures called a perlite texture.
PUMICE	Lightweight, highly vesicular, frothy, glassy rock. The most common types of pumice are rhyolitic or phonolitic. Pumice can usually float on water.

5.1.2.2 Chemical classification using the TAS diagram

Volcanic rocks show a great range of variation in chemical compositions. This variation is, however, far from random and shows a series of rational patterns. Two of the most important and useful chemical parameters in volcanic rocks are SiO_2 and $\text{Na}_2\text{O} + \text{K}_2\text{O}$ (total alkalis). The vast majority of volcanic rocks have between 35 and 75% SiO_2 and 1–16% total alkalis. It is clear that a fine grained basalt consisting essentially of augite ($\text{Ca}(\text{Mg,Fe})[\text{Si}_2\text{O}_6]$), calcic plagioclase ($\text{Ca}[\text{Al}_2\text{Si}_2\text{O}_8] - \text{Na}[\text{AlSi}_3\text{O}_8]$) and magnesian olivine ($(\text{Mg,Fe})_2[\text{SiO}_4]$) will be relatively poor in SiO_2 and $\text{Na}_2\text{O} + \text{K}_2\text{O}$. A fine grained rhyolite, on the other hand, consisting essentially of quartz (SiO_2), potassium feldspar ($\text{K}[\text{AlSi}_3\text{O}_8]$) and sodic plagioclase ($\text{Na}[\text{AlSi}_3\text{O}_8] - \text{Ca}[\text{Al}_2\text{Si}_2\text{O}_8]$) will be relatively enriched in SiO_2 , K_2O and Na_2O . The chemical compositions of rocks are therefore closely related to their mineralogy.

The TAS system in Fig. 5.4 is widely used for the classification of volcanic rocks when a mineralogical classification is difficult to apply. In addition to volcanic rock names, most of which are also used in Fig. 5.3, some important terms used in the description of volcanic rocks are shown in Fig. 5.4. These are based solely on the *SiO₂-content of the rock*. This is the total amount of the chemical component silicon dioxide (silica) and is NOT the amount of quartz that is a mineral with the composition SiO_2 . A basalt does not contain quartz but has 45–52% SiO_2 (which is mainly present in the silicate minerals augite, olivine and plagioclase or in basaltic glass). Volcanic rocks are divided into four groups on the basis of their SiO_2 -content (Table 5.2).



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%SiO ₂	
< 45	ultrabasic
45-52	basic
52-63	intermediate
> 63	acid

Table 5.2. Subdivision of igneous rocks on the basis of their SiO₂-content.

Note in Fig. 5.4 that basalts are basic, andesites are intermediate and rhyolites are acidic. The terms acid, intermediate, basic and ultrabasic are used for plutonic and some metamorphic rocks as well as volcanics. Gabbros are therefore basic, granites are acidic.

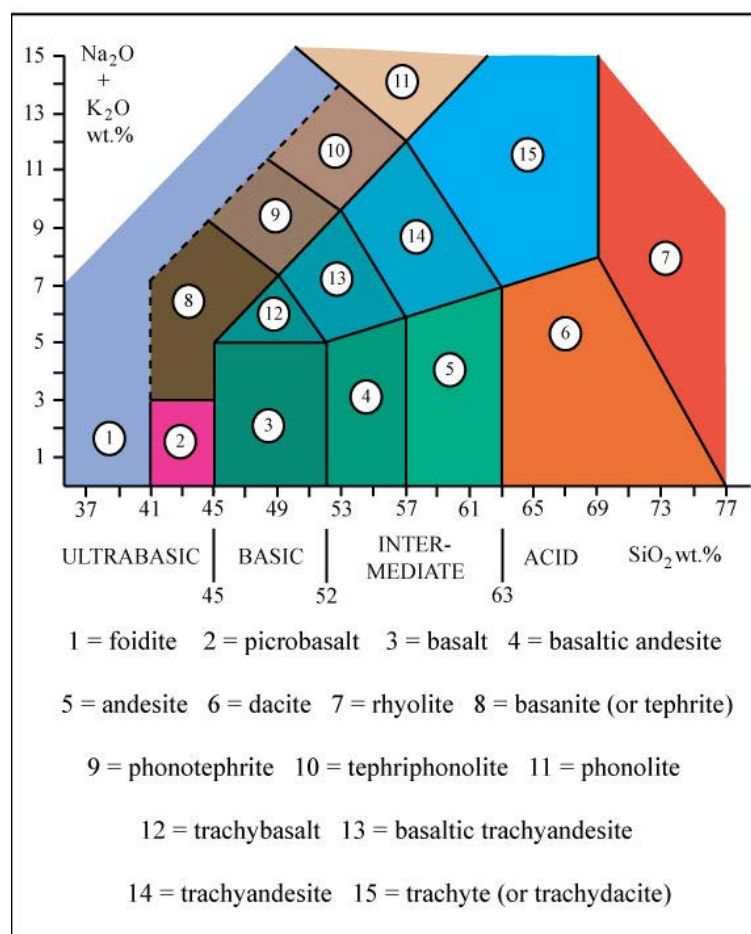


Fig. 5.4: The TAS (Total Alkalies vs. Silica) classification for volcanic rocks.

This classification is based on chemical analysis of the rock for SiO₂, K₂O and Na₂O.

Most ultramafic rocks are also ultrabasic e.g. peridotite. An exception is, however, the rock eclogite. This is a metamorphic basaltic rock consisting of green clinopyroxene and red-brown garnet. Eclogites are formed under very high pressures – so high that plagioclase is no longer stable. The rock is basic in terms of its chemical composition (chemically = basalt) but it consists entirely of mafic minerals and is therefore ultramafic.

The TAS diagram (Fig. 5.4) is divided into 15 fields which are commented on here. The most important names are in *italics*.

1. Foidite e.g. nephelinite. These rocks, which are relatively rare, can contain a wide range of total alkalies and cover a large area in the SiO₂-poor part of the diagram.
2. Volcanic rocks with 41–45% SiO₂ and < 3% total alkalies are called picobasalts.
3. *Basalts* are basic rocks with 45–52% SiO₂ and < 5% total alkalies.
4. Basaltic andesites (52–57% SiO₂) and
5. *Andesites* (57–63% SiO₂) are intermediate volcanic rocks with relatively low total alkali contents.
6. *Dacites*, which have > 63% SiO₂, are acidic volcanics which grade into
7. *Rhyolites* with 69–77% SiO₂ depending on the alkali content.
Note that fields 3–7 (basalt, basaltic andesite, andesite, dacite, rhyolite) lie on the margins of or within the upper triangle in the QAPF system (Fig. 5.3). They are silica-saturated or silica-oversaturated.
8. This field defines tephrites and basanites. Distinction is based on the normative olivine content (basanite > 10% olivine; tephrite < 10% olivine) (see point 14 in section 5.1.2.1.).
With increasing SiO₂ and total alkalies these pass into
9. Phonotephrite,
10. Tephriphonolite, and
11. *Phonolite*.
Note that fields 1 and 8–11 lie within the lower triangle in the QAPF system (Fig. 5.3).
12. This small triangular field defines the composition of trachybasalt.
13. Basaltic trachyandesite.
14. Trachyandesite.
15. *Trachyte* (< 20% normative quartz) and trachydacite (> 20% normative quartz).

5.1.2.3 Volcaniclastic rocks

During the eruption of a volcano lava may be extruded, gas is vented, and fragments of volcanic rocks are ejected into the air (or into water if the eruption takes place subaqueously). The lavas are igneous rocks and have been considered in sections 5.1.2.1. and 5.1.2.2. The gas is dispersed and leaves no record. The ejected material, collectively known as *tephra*, will be deposited around the volcano as a subaerial or subaqueous deposit. Clearly such deposits are intermediate between true igneous and true sedimentary rocks. Sediments composed of fragments of volcanic material are called *volcaniclastic* deposits. The size of volcanic fragments is fundamental (Table 5.3).

	< 2mm	2–64mm	> 64mm
volcanic	ash	lapilli	block, bomb
sedimentary	clay, silt, sand	granule, pebble	cobble, boulder

Table 5.3. Names of volcaniclastic rocks are based on particle size. Comparison is made with sedimentary deposits.

Blocks are angular clasts (fragments) produced by the breaking-up of solid rock. *Bombs* are fragments that were partly molten when they were ejected and were streamlined into smooth forms as they solidified during flight through the air.



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Volcanic fragments may also be classified according to their composition:

<i>vitric</i>	– glassy
<i>lithic</i>	– rock fragments
<i>crystalline</i>	– consists of crystals

Ash-sized clasts are usually vitric (glass shards) or crystal, whereas blocks are usually lithic. Clasts of volcanic origin may still be extremely hot when they are deposited so that they become *welded* together. *Ignimbrites* (“fiery cloud” rocks) are deposited from one of the most violent types of volcanic eruption – *pyroclastic flows*. These fragmental volcanic rocks are usually SiO₂-rich (dacitic or rhyolitic) and commonly contain flattened glassy fragments (called “flames”). Ignimbrites are commonly welded.

Volcanic deposits that have become consolidated are classified according to the size of the fragments.

<i>ash tuff</i>	rock composed entirely of ash-sized fragments
<i>lapilli tuff</i>	rock composed entirely of lapilli-sized fragments
<i>volcanic breccia</i>	rock composed entirely of blocks

Intermediate types occur.

Agglomerate is a term sometimes used for unsorted volcanoclastic deposits with a mixture of blocks, bombs, lapilli and ash that accumulated near the volcanic vent.

It is common practice to combine the different type of terms e.g. lithic lapilli tuff, vitric tuff breccia, crystal tuff, andesite breccia, rhyolitic pumice etc.

5.1.3 Mineral assemblages

As we have seen above, most igneous rocks are classified according to their mineralogy. Rocks that consist entirely of glass cannot, of course, be classified on a mineralogical basis. Ultramafic rocks are classified using mafic minerals whereas the QAPF double triangle uses felsic minerals. Apart from ultramafic rocks, plutonic and volcanic rocks only differ in terms of grain size and texture; their mineralogical variation is essentially similar. The relationship between mineral assemblage (i.e. all the important minerals in a rock) and rock composition is illustrated in Fig. 5.5.

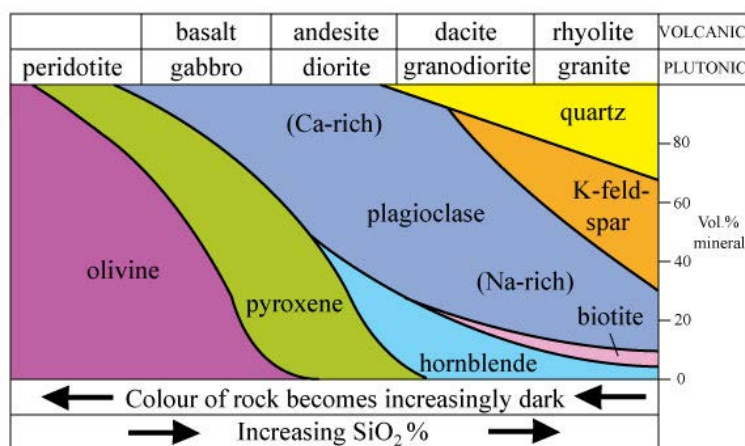


Fig. 5.5: Variations in mineral assemblages in plutonic and volcanic rocks.
This diagram only applies to rocks in the QAP part of the QAPF double triangle.

Several main features of Fig. 5.5. are:

a) The names of equivalent plutonic and volcanic rocks are:

<i>volcanic</i>		basalt	andesite	dacite	rhyolite
<i>plutonic</i>	peridotite	gabbro	diorite	granodiorite	granite


Note that there is no volcanic equivalent to peridotite. Such a rock does exist (it is called komatiite) but it is so rare that we will not consider it further in this introductory text.

- The SiO_2 content increases from left to right from $< 45\%$ for the ultramafic (and ultrabasic) peridotite to $> 63\%$ for the felsic, acidic granite/rhyolite.
- The colour index (% mafic minerals, M) decreases from left to right.
- Peridotite consists of $> 90\%$ olivine \pm pyroxene and up to 10% Ca-rich plagioclase feldspar.
- Gabbros/basalts consist dominantly of Ca-rich plagioclase and pyroxene \pm olivine.
- Diorites/andesites are dominated by plagioclase and hornblende \pm pyroxene.
- Granodiorites/dacites contain quartz, Na-rich plagioclase and hornblende \pm K-feldspar \pm biotite.
- Granites/rhyolites are dominated by quartz and K-feldspar with Na-rich plagioclase and minor mica (biotite \pm muscovite) and hornblende. Micas are, however, very rare in volcanic rocks.
- Plagioclase becomes progressively poorer in the anorthite ($\text{Ca}[\text{Al}_2\text{Si}_2\text{O}_8]$) component and richer in the albite ($\text{Na}[\text{AlSi}_3\text{O}_8]$) component from left to right.
- All the mafic minerals (olivine, pyroxene, amphibole, biotite) become progressively more Fe-rich and Mg-poor from left to right.

This systematic relationship is not coincidental. Basaltic magma, which is by far the most common, changes in composition during crystallization in magma chambers. This process, which is called *fractional crystallization*, will be dealt with later.

5.2 Magma

All igneous rocks are formed by the solidification of molten rock matter – *magma*. Magmas are generated by the partial melting of deep crustal or upper mantle rocks. Magma reaches the surface of the Earth through, for example, a volcano. Molten rock coming out of a volcano is called *lava*. Magmas that reach the surface form *volcanic* (or *extrusive*) rocks; those which crystallize at depth are known as *plutonic* (or *intrusive*) rocks.



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5.2.1 Where does magma come from?

The commonest and most important type of magma is basaltic in composition. Basaltic magma is formed by the partial melting of mantle peridotite (an ultramafic rock consisting mostly of olivine and pyroxenes). Rocks melt over a temperature interval. The temperature at which the first melt begins to form is the solidus temperature. It is entirely molten (liquid) at the liquidus temperature. Basalt forms when the temperature of mantle peridotite is above its solidus. The solidus temperature of mantle peridotite increases with depth (Fig. 5.6). The temperature of the mantle also increases with depth, but for it to exceed its solidus temperature requires special circumstances. This can occur in two ways; by pressure decrease or by the introduction of water. Mantle material is subject to large-scale convection. The convection cells rise beneath mid-ocean ridges where hot mantle peridotite is moved up to a depth where it begins to melt. The melt produced by about 10–25% partial melting of mantle peridotite has a basaltic composition.

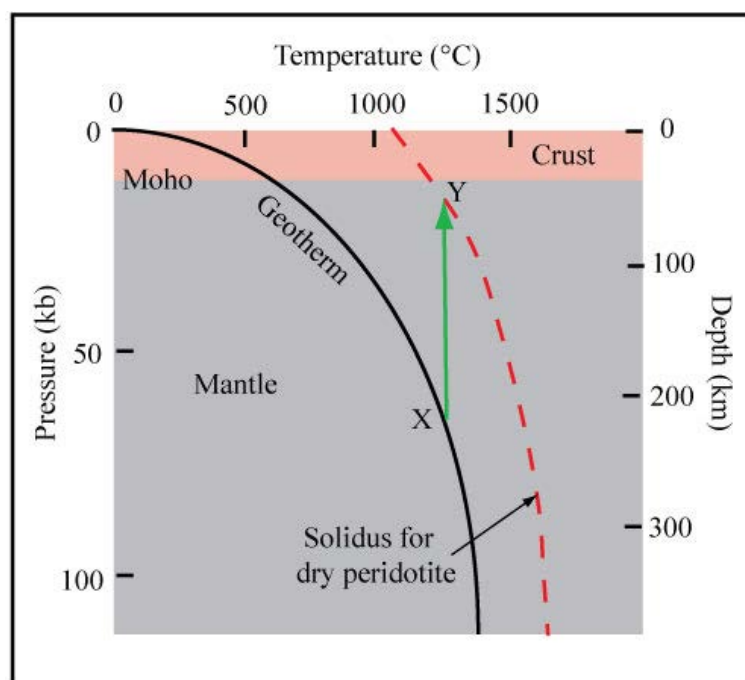


Fig. 5.6: Pressure – temperature diagram showing the melting curve (solidus) for dry mantle peridotite and the geothermal gradient.

Mantle peridotite will begin to melt if it is moved from X to Y. This does not involve heating but a decrease in pressure. This is achieved under mid-ocean ridges as a result of mantle convection. About 10–25% melting of peridotite produces the basaltic magma that is erupted along mid-ocean ridges.

The magma that is erupted along mid-ocean ridges (Fig. 5.7) is therefore basaltic. The partial melting of mantle peridotite also takes place at “hot spots”. These are the surface expression of mantle plumes – columns of mantle material that rise towards the surface and partially melt to produce locally huge volumes of basaltic magma. There are hot spots under, for example, Hawaii, the Canary Islands, and Iceland. Most hot spots are located far from mid-ocean ridges or subduction zones.

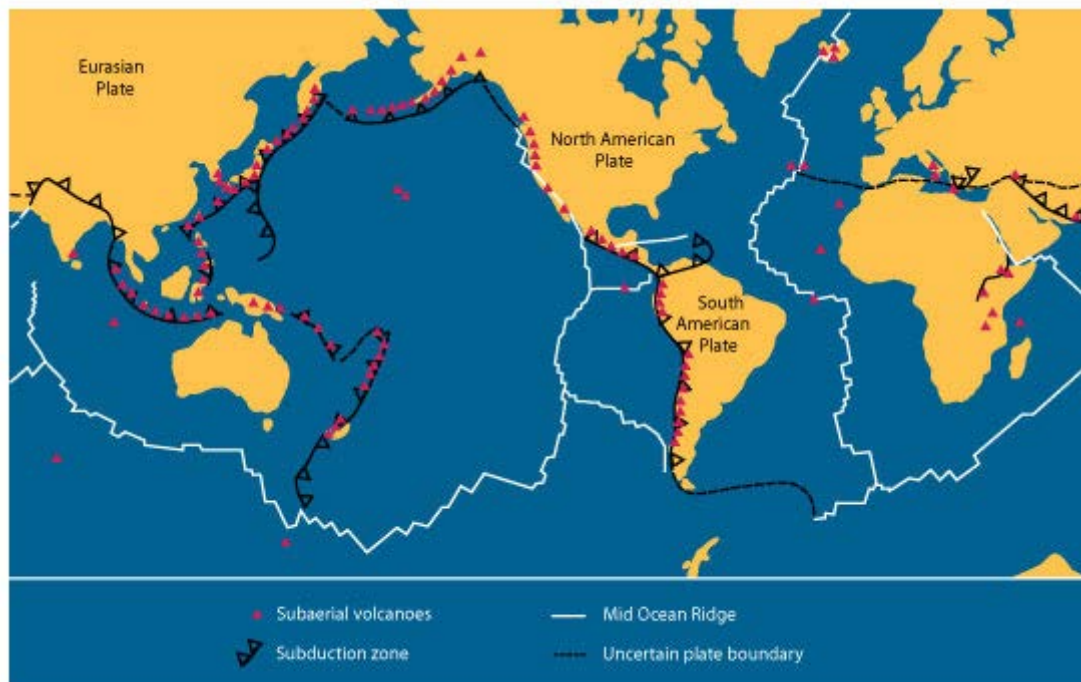


Fig. 5.7: Global distribution of tectonic plate boundaries and volcanic activity.

Basaltic magma comes to the surface along a) mid-ocean ridges because of decompressional partial melting of the mantle as a result of mantle convection b) in connection with hot spots because of decompressional partial melting of the mantle as a result of rising plumes c) in connection with subduction zones as a result of the sinking of the solidus temperature of peridotite by the introduction of water.

The third setting where large volumes of magma are produced is in connection with subduction zones (Fig. 5.8). The oceanic crust that is subducted contains water (in metamorphic minerals; this will be dealt with later when we consider metamorphism). During subduction the oceanic crust is heated and water is released into the overlying wedge of mantle peridotite. The presence of water lowers the solidus temperature of the peridotite and it starts to melt (Fig. 5.9). This magma rises towards the surface and may form a magma chamber on the way and/or reach the surface to form a volcano. Subduction zones are therefore volcanically active. The magma produced is generally andesitic rather than basaltic.

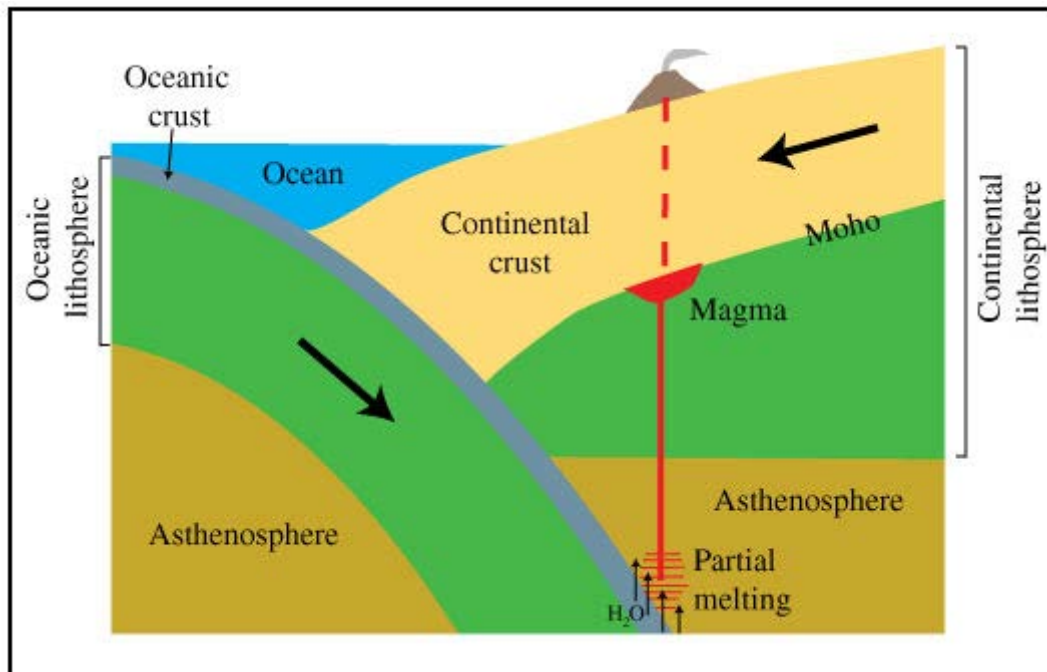


Fig. 5.8: Subduction zone at the margin between oceanic and continental lithospheric plates. Basalt formed at mid-oceanic ridges becomes metamorphosed by circulating seawater. Hydrous minerals like chlorite, hornblende and serpentine form. During subduction these minerals release water (and turn into other, dry minerals). The water enters the overlying mantle wedge. The solidus temperature of the peridotite is lowered by the water, and it starts to melt.

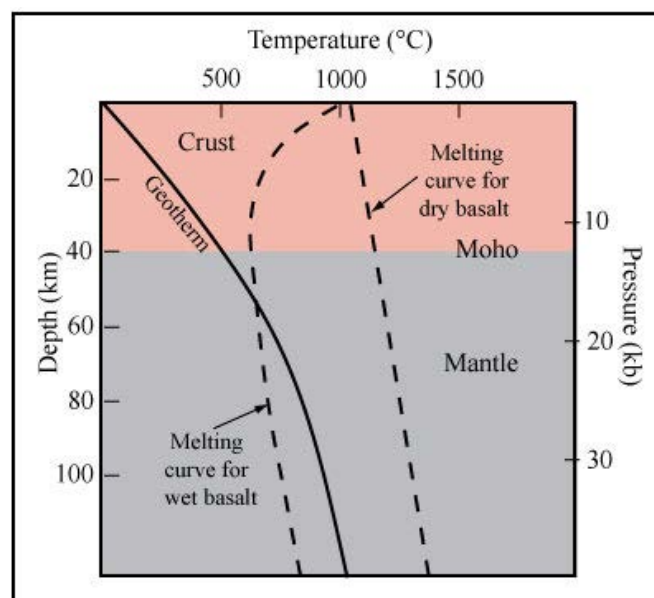


Fig. 5.9: Pressure – temperature diagram showing the influence of water on the melting curve (solidus) for basalt and the geothermal gradient. The presence of water lowers the solidus temperature of rocks.

5.2.2 The composition of magma

As we have seen when dealing with the classification of igneous rocks, three common types of volcanic rock are basalt, andesite, and rhyolite. Dacite has a composition intermediate between andesite and rhyolite. These are formed from basaltic, andesitic, dacitic and rhyolitic magmas. In the TAS diagram (Fig. 5.4) it is apparent that basalt represents a basic (45–52% SiO_2) magma, andesite is intermediate (52–63% SiO_2) and dacite and rhyolite are both acidic (> 63% SiO_2) magmas. The chemical compositions of magmas (and rocks) are expressed in terms of major (> 0.1% oxide) and trace (< 0.1% or 1000 ppm = parts per million) elements. Major elements are expressed as % oxide; trace elements as ppm element. Some elements are “major” in one magma/rock and only occur in trace quantities in another (e.g. phosphorus is “major” in basalt, andesite and dacite and “trace” in rhyolite in Table 5.4).

Table 5.4 illustrates that magmas and igneous rocks consist mainly of only a few (eight) major oxides ($\text{SiO}_2 + \text{Al}_2\text{O}_3 + \text{Fe}_2\text{O}_3 + \text{FeO} + \text{MgO} + \text{CaO} + \text{Na}_2\text{O} + \text{K}_2\text{O}$). The remaining oxides (TiO_2 , MnO , P_2O_5 and H_2O) are usually present in only small amounts. The total should, of course, be close to 100%. This is a check that the analysis is reliable. Values between 98.5 and 100.5 are acceptable. This \pm is accepted because of analytical error giving some \pm on all the elements. Some trace elements may be present in higher concentrations than normal, and there are other elements/oxides that can be present that are not normally determined (sulphur, carbon dioxide etc.).

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	<i>BASALT</i>	<i>ANDESITE</i>	<i>DACITE</i>	<i>RHYOLITE</i>
SiO ₂	49.60	57.94	65.01	72.82
TiO ₂	1.84	0.95	0.58	0.28
Al ₂ O ₃	15.84	16.67	15.91	13.47
Fe ₂ O ₃	3.79	2.50	2.43	1.48
FeO	7.13	4.92	2.70	1.11
MnO	0.20	0.12	0.09	0.06
MgO	6.99	3.91	1.58	0.39
CaO	9.70	6.78	4.32	1.14
Na ₂ O	2.91	3.54	3.79	3.65
K ₂ O	0.51	1.76	2.17	4.50
P ₂ O ₅	0.95	0.29	0.15	0.07
H ₂ O	0.35	1.15	1.20	1.10
Total	99.81	99.94	99.93	100.07
MgO/FeO	0.98	0.79	0.59	0.35
CaO/Na ₂ O	3.33	1.92	1.14	0.31

Table 5.4. Chemical compositions of typical basalt, andesite, dacite and rhyolite

From basalt to rhyolite there are several important geochemical trends:

- SiO₂ increases
- mafic oxides (TiO₂, MgO, FeO, Fe₂O₃, MnO) decrease
- CaO decreases
- alkalies (Na₂O and K₂O) increase
- P₂O₅ decreases
- the MgO/FeO ratio decreases
- the CaO/Na₂O ratio decreases

The mafic oxides are those that enter Mg-Fe minerals like olivine, pyroxenes, amphiboles, magnetite and ilmenite. The MgO/FeO ratio decreases because early-crystallizing mafic minerals are Mg-rich. The CaO/Na₂O ratio decreases because early-crystallizing plagioclase is Ca-rich and Ca enters early-formed clinopyroxene. We will see why these compositions are related in this way later on.

Basaltic, andesitic, dacitic and rhyolitic magmas are not formed in equal abundance. Basaltic magma is by far the most common. As we have seen, basaltic lava is erupted on Iceland and Hawaii and beneath the sea along mid-ocean ridges. The ocean floor (under a thin sheet of marine sediments) consists of basalts. Andesites and dacites are subduction-related and erupted along active mountain chains (e.g. Andes, Rocky Mountains). Rhyolites are erupted in continental areas e.g. Yellowstone Park. It is clear that the distribution of these different types of magmas is closely related to plate tectonics.

5.2.2.1 Magmatic gases

Anyone who has opened a bottle of beer (which probably includes you, the reader) is aware that gas is released. This gas (CO_2) is dissolved in the beer which is above atmospheric pressure in the unopened bottle. The solubility of CO_2 in beer increases with pressure, and when the pressure drops suddenly the gas forms bubbles that are released. Magmas behave in the same way. The most important gas phase in magmas is water vapour. The amount of H_2O that can be dissolved in magma depends on the magma composition and the pressure (i.e. the depth below the surface of the Earth), but usually lies in the range 0.2–3% by weight.

As magma crystallizes, the first minerals to crystallize (olivine, pyroxenes, plagioclase) do not contain any H_2O in their structure and its concentration in the melt therefore increases. As magma rises towards the surface the pressure decreases and the magma may not be able to retain all the gas in solution. Gas is therefore released. This can be in a catastrophic way in the form of an explosive eruption. Water vapour is usually the most abundant gas in magmas but others can be important, including carbon dioxide, nitrogen, hydrogen chloride, rare gases and sulphur compounds. The latter give the characteristic sulphurous smell associated with many active volcanic areas.

5.2.3 Temperature

The temperature of magma depends largely on its composition and the volatile (water) content which increases with pressure. If we consider dry lava at the surface of the Earth, molten basalt typically has a temperature of $\sim 1200^\circ\text{C}$ whereas rhyolitic lava is $\sim 900^\circ\text{C}$. Lava can, of course, be fluid after it has begun to cool and crystallize. There are two important terms used in this connection: the *liquidus* temperature of a melt is that temperature when the first crystals begin to form; the *solidus* temperature is when the last drop of melt crystallizes. (In section 5.2.1 we considered these terms from the “opposite” perspective – melting). For dry basalt the liquidus temperature is $\sim 1200^\circ\text{C}$ and the solidus $\sim 990^\circ\text{C}$. This means that basalt crystallizes completely over a temperature interval of $>200^\circ\text{C}$. For dry rhyolite at the surface of the Earth the liquidus is at $\sim 900^\circ\text{C}$ and the solidus is at $\sim 750^\circ\text{C}$.

5.2.4 Viscosity

The ease with which a liquid flows is expressed by its viscosity. Basaltic magma flows fairly rapidly and has a relatively low viscosity (10^2 – 10^4 poises, similar to thick syrup). Rhyolitic magma, on the other hand, has much higher viscosity (10^5 – 10^{15} poises). The viscosity of magma therefore depends on its composition (including the volatile content); it also depends on the temperature.

As we have seen, the most important rock-forming minerals are silicates. These have a variety of structures in which the $[\text{SiO}_4]^{4-}$ units are linked together in different ways. It is clear from Table 5.5 that rhyolite consists of minerals with more complicated structures than basalt.

basalt minerals	silicate structure	rhyolite minerals	silicate structure
Olivine	<i>Independent SiO_4-units (Nesosilicate)</i>	Quartz	<i>Framework silicate (Tectosilicate)</i>
Pyroxene	<i>Chain structure (Inosilicate)</i>	K-feldspar	<i>Tectosilicate</i>
Plagioclase	<i>Tectosilicate</i>	Plagioclase	<i>Tectosilicate</i>

Table 5.5. The silicate structures of basalt minerals compared with rhyolite minerals. The greater degree of polymerisation of rhyolite minerals results in a high viscosity.

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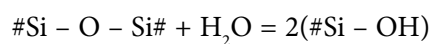
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Silicate melts also contain $[\text{SiO}_4]^{4-}$ units and, as in silicate minerals, they occur with different degrees of oxygen-sharing, or polymerisation. Since all the main minerals that crystallize from rhyolitic magma are framework silicates (tectosilicates), the magma will be highly polymerised before it starts to crystallize. This results in rhyolitic magma being highly viscous. Basaltic magma, on the other hand, is less polymerised because the minerals that eventually crystallize include olivine, which has a simple structure with independent SiO_4 -tetrahedra (nesosilicate), and pyroxene, a chain silicate (inosilicate), in addition to plagioclase feldspar (tectosilicate). The presence of dissolved water has the effect of reducing viscosity because H_2O forms (OH)-groups which break the Si-O bonds:



where Si# denotes 3 Si-O bonds. Temperature also effects viscosity. The greater the temperature above the liquidus, the lower the viscosity. Lava flows, particularly basalts, can solidify to give smooth, “ropy”-surfaced lava called “pahoehoe” on Hawaii. Some basaltic (and other) flows solidify to give a blocky surface (“aa” on Hawaii). These different surface effects reflect small differences in viscosity.

5.2.5 Density

The densities of most magmas range between about 2.5 and 3.0 g/cm³. Mafic magmas are denser than felsic ones. As we will see later, magmas do not always reach the surface of the Earth and may accumulate in the crust in magma chambers. Magma may be repeatedly added to a chamber. As the magma crystallizes its density will normally decrease as “heavy” components are removed first. This means that when new, dense magma enters the base of the chamber it can remain at the floor and the magma in the chamber can become compositionally zoned. It is not uncommon for magma chambers to develop in such a way that there is (dense) basaltic magma at the floor and (buoyant) rhyolitic magma at the roof.

5.3 Eruption of magma

Magmas are formed by the partial melting of rocks in the Earth’s crust or mantle. Magmas are less dense than the rocks from which they form and rise towards the surface. Sometimes they do not reach the surface but accumulate in chambers where they crystallize slowly at depth to form relatively coarse-grained, plutonic rocks. Sometimes they reach the surface where they are erupted. Magmas may reach the surface directly from their source area. Others may reside for some time in a magma chamber and start to crystallize (thereby changing their composition by fractional crystallization, as will be discussed later) before the remaining magma in the chamber makes its way to the surface. The presence of phenocrysts in lavas reflects partial crystallization of magma on the way to the surface, most often in a magma chamber.

The way in which magma is erupted at the surface depends on many factors, the most important of which are its composition (including volatile content) and its viscosity. Explosive eruptions (like Mt. St. Helens, USA, in 1980) involve volatile-rich magmas (i.e. with a lot of dissolved gas) with high viscosities. Non-explosive eruptions involve magmas with low amounts of dissolved gas and low viscosities (like the eruptions on Hawaii).

5.3.1 Non-explosive eruptions

Any volcanic eruption is a violent process in terms of human experience, but some are more violent than others. Relatively non-violent eruptions involve magma with low gas contents and low viscosities, such as those involving basaltic magma. Even non-explosive eruptions can be quite violent in the early stages when dissolved gas is released rapidly. Gas can be released so fast that spectacular lava fountains occur. Bubbles formed in the magma rise to the surface and burst, throwing a shower of fragments of molten, glowing lava into the air. When the lava fragments hit the ground they can pile up to form a spatter cone near to the vent. As the fountaining dies down, lava commonly emerges from the vent and moves down slope as a lava flow.

The upper surface of a lava flow cools rapidly against the air and a solidified crust forms, below which lava continues to flow in well defined channels. The pre-existing topography has a major influence on the course of the flowing lava which will follow valleys. The flow may be concentrated in lava tubes and liquid lava (if its viscosity is low enough) may flow for considerable distances from the vent. Lava near the vent can flow smoothly to give “ropy” lava (pahoehoe; section 5.2.4) but as viscosity increases with cooling it flows less readily and form a “blocky” surface (aa lava; section 5.2.4).

As lava cools its viscosity increases and escaping bubbles become trapped in the solidifying lava. These bubbles, which retain their original sub-spherical shape, are called *vesicles* and the resulting rock would be, for example, vesicular basalt. With time, vesicles can gradually become filled with crystalline material (e.g. calcite, quartz) as fluids circulating through the rocks deposit the material they have in solution. Vesicles that have become filled by such secondary minerals are called *amygdules*, and the resulting rock would be, for example, and amygdaloidal andesite.

The majority of volcanic eruptions take place beneath the sea along mid-oceanic ridges. The basaltic lava that is erupted is immediately cooled by seawater and large blobs of lava are formed with glassy rims (very rapidly chilled lava). The central part of the blob remains hot for some time. These “blobs” solidify to form what is known as *pillow lava* (Fig. 5.10).

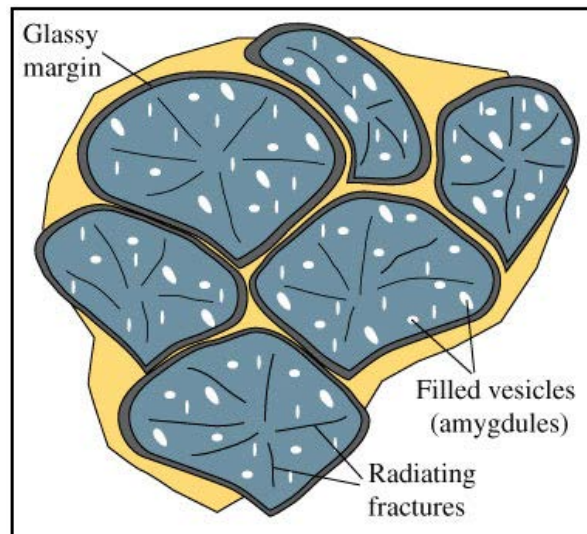


Fig. 5.10: Pillow-shaped bodies are formed when lava comes into contact with water. They are particularly common along mid-ocean ridges. They can sometimes be used as a “way-up” criterion in ancient, deformed rocks.

Pillows lie in a pile that resembles a heap of sand bags. Seawater can react with the hot lava that can change its composition, most importantly involving the introduction of water and sodium (from the salt water) into the pillows. Basaltic lava that is erupted on the sea floor therefore differs from that erupted on land by having pillow structures and by being relatively enriched in H_2O and Na_2O .

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The history of the Earth has been punctuated by several short (a few millions of years) periods involving the eruption of huge quantities of basaltic lava above hotspots (section 5.2.1.). In the north-western USA, the Columbia River basalts in Oregon and Washington originally covered an area of ca. 163.000km² and had a volume of 174.300km³. Some of the individual lava flows had volumes >2000km³. This huge amount of basaltic lava was erupted from fissures (not from central volcanoes) over a period of about 3 million years some 17–14 million years ago (Mid-Miocene). The basaltic lavas inundated and “drowned” the pre-existing topography and formed a huge plateau which was up to 10km thick (Fig. 5.11). These provinces are known as flood basalts or plateau basalts.

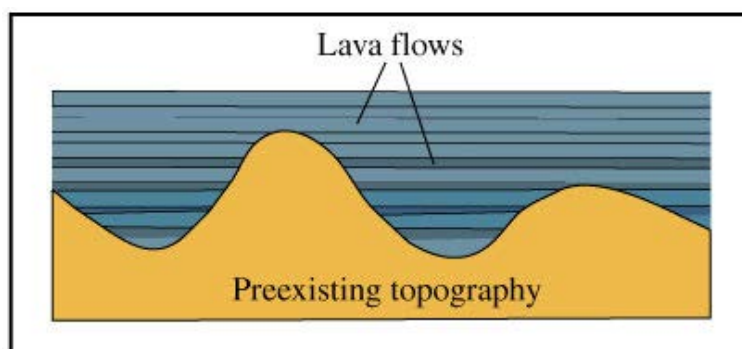


Fig. 5.11: Plateau (or “flood”) lavas drown the pre-existing topography. Enormous volumes of lava are produced in plateau lava provinces.

The Columbia River Province is not the largest of its kind – there are about 10 that are larger elsewhere in the world. Those in Siberia (Permo-Trias), the Karoo area of South Africa (Jurassic) and in the Parana area of Brazil (Cretaceous) covered areas of >2.000.000km².

5.3.2 Explosive eruptions

Magmas with elevated gas contents and high viscosities are liable to erupt extremely violently. This is because the gas released during pressure decrease as magma approaches the surface cannot escape gradually. When the pressure within the magma exceeds the confining pressure, gas escapes in an explosive manner. An extreme example of this causes the formation of *pumice*. This is a rock which consists entirely of very thin glass-walled vesicles; there are so many bubbles (vesicles) relative to glass that pumice can float on water. The escape of gas can be so violent that the magma breaks into tiny glassy fragments called *volcanic ash*.

5.3.2.1 Pyroclasts and tephra

The products of volcanism are commonly hot (Greek = pyro) fragments (clasts) and are known as pyroclastic rocks. Deposits formed of pyroclasts are known as tephra deposits which are named according to their size. Tephra is the name used for the unconsolidated (young) deposit; other names are used for the consolidated rock (Table 5.6).

Average particle diameter (mm)	Tephra (unconsolidated material)	Pyroclastic rock (consolidated material)
> 64	bombs	agglomerate
2-64	lapilli	lapilli tuff
< 2	ash	ash tuff

Table 5.6. Nomenclature for tephra and pyroclastic rocks

5.3.2.2 Eruption columns and tephra falls

The most violent explosive eruptions involve viscous magma with a high gas content. This is particularly the case for rhyolitic magmas which are silica-rich and have a highly polymerised melt structure. Decompression of the rising magma results in the rapid escape of huge volumes of hot gas which rise into the air together with tephra to form an eruption column. These columns can reach as high as 45km up into the atmosphere.

At the level in the atmosphere where the density of the cooling column reaches that of the air it spreads out laterally to form an anvil-shaped cloud. This cloud will drift with the prevailing wind. Particles will fall from this drifting cloud and accumulate as tephra deposits. The climactic eruption of Mt. Mazama 6600 years ago which gave rise to Crater Lake (Oregon, USA) produced a 30cm-thick layer of ash some 130km away from the vent, and mm-thick layers some 1100km away. Fine volcanic dust from such eruptions can become entrained in the upper atmosphere so that dispersal is global and produce multi-coloured sunsets over long time periods.

5.3.2.3 Pyroclastic flows

Probably the most devastating type of volcanic eruption involves highly mobile, hot mixtures of gas and ejecta that move swiftly along the ground from an explosive vent. Pyroclastic flows deposit sheets consisting of a mixture consisting of a matrix of ash with fragments of glass, pumice and crystals, as well as accidental lithic clasts. These flows can cover up to thousands of square kilometers with thicknesses of a few meters up to a few hundred meters. The origin of these sheets was first understood after observation of the eruption on May 8th, 1902, of Mt. Pelée on the island of Martinique, West Indies. Some 29.000 people were killed. Two terms that are commonly used for pyroclastic flows are “*nuée ardente*” (French for “glowing cloud”) that illustrates the nature of the flow, and “*ignimbrite*” (Latin “*ignis*” = fire; “*nimbus*” = cloud) for the poorly sorted deposit.

5.4 Volcanoes

5.4.1 The shapes of volcanoes

The largest volcanoes on Earth are *shield volcanoes* because their shape resembles that of a warrior's shield. The Hawaiian Islands rise as much as 10km above the sea floor and have diameters of over 100km. They are largely built up of successive flows of low-viscous basaltic lava with relatively little pyroclastic (explosive) material. The slopes of young shield volcanoes are usually between 5° and 10° (Fig. 5.12).

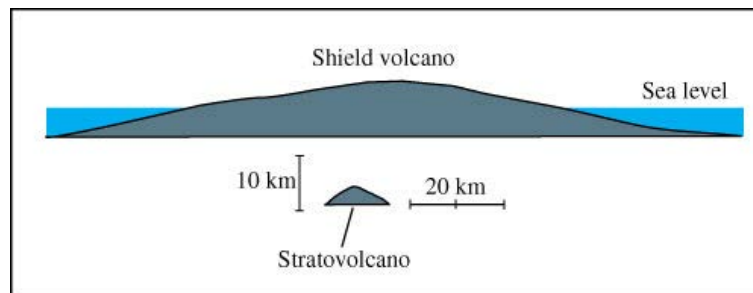


Fig. 5.12: Comparative sizes and shapes of shield (for example Muana Loa, Hawaii) and stratovolcanoes (for example Mt. Fuji, Japan).
Shield volcanoes are by far the largest in the world.

Eruptions commonly involve lava fountains in which tephra is thrown into the air. The tephra is deposited near the vent to form a conical pile called a tephra cone. The size of the tephra is such (in the range of 2–64mm in diameter) that these usually consist dominantly of lapilli. The slopes of *tephra cones* are typically about 30°. Volcanoes that dominantly consist of material more viscous than basalt, such as andesite, typically produce pyroclastic material (i.e. explosive products) in addition to lava flows. These so-called *stratovolcanoes* build up steep conical mounds. This is because the relatively viscous lava cannot flow very far, and the tephra deposits accumulate close to the vent.

5.4.2 Calderas

There is a magma chamber beneath most volcanoes. As magma is withdrawn from the chamber to be extruded via the volcano (or more than one volcano) it can be partially emptied. The roof is then unsupported and can collapse into the chamber. This commonly takes place along ring-shaped fractures. This collapse of a huge block of rocks into the magma chamber is accompanied by a major (usually explosive) eruption. The result is a circular depression called a caldera (Fig. 5.13). Large calderas are developed around Mt. Teide on Tenerife and related to the Greek island of Santorini where most of the caldera is below the sea but some of the steep circular walls are visible. Crater Lake in Oregon, USA, is the result of caldera collapse.

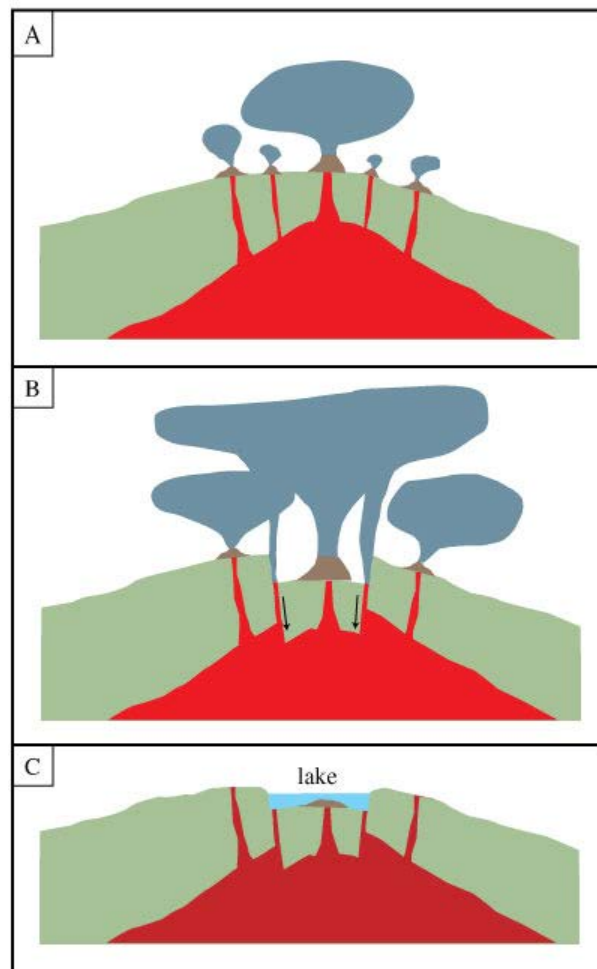


Fig. 5.13: Three stages in the formation of a caldera.

A. Eruptions from volcanoes above a magma chamber results in partial emptying of the chamber. This results in B, collapse of the volcano(es) into the chamber. This is usually associated with an enormous explosive event. Collapse usually occurs along ring-fractures. Subsequent erosion results in (C) a circular-shaped depression that can be occupied by a lake.

5.5 Plutonic rocks

Volcanic rocks are extruded at the surface; plutonic rocks are the crystalline products of magma that has intruded into the crust and solidified beneath the surface. Different types of plutons are named according to their style and size.

5.5.1 Minor intrusions (dykes and sills)

On its way towards the surface, magma is commonly intruded along more or less vertical fractures. The magma, which is under pressure, forces its way upward, opening and filling a fissure. The magma subsequently solidifies to form a tabular, sheet-like body called a *dyke* (Fig. 5.14). Dykes may reach the surface and act as feeders to volcanoes. Dykes are commonly 20cm–20m wide but narrower and wider ones also occur.

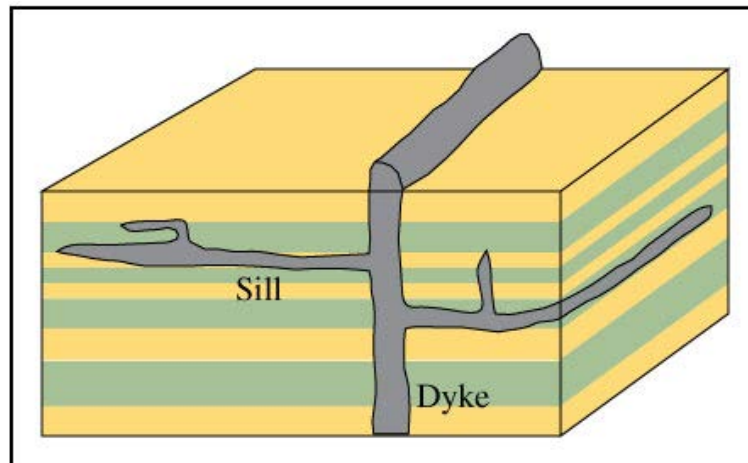



Fig. 5.14: Sills and dykes.

Dykes are more or less vertical sheets of igneous rock. They are commonly feeder channels for volcanoes. Sills are emplaced more or less horizontally, parallel with bedding in sedimentary rocks.

Magma intruded in a dyke commonly cuts across more or less horizontal sedimentary rocks. It may be easier for the magma to spread out laterally between two layers of sediment than to continue upwards to the surface. Tabular, sheet-like bodies that form parallel to the layering are called *sills*. Sills can locally be transgressive to the sedimentary bedding. The Whin Sill in northern England strikes roughly east-west and locally forms a steep escarpment towards the north. The Roman emperor Hadrian built a wall on top of this sill.



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Instead of spreading out laterally to form a sill, the magma may form a bulge by pushing up the overlying rocks to form a dome-shaped feature called a *laccolith*.

Volcanoes may be fed by cylindrical pipes rather than dykes. Magma can solidify in this type of conduit to form a *volcanic pipe*. If the surrounding rocks are eroded away to leave the pipe as a resistant feature it is known as a *volcanic neck*.

A feature that is commonly developed in dykes and sills, and sometimes in lava flows, is *columnar jointing* (Fig. 5.15). As the solidified magma cools it contracts. Cooling is largely through the walls of the intrusion and contraction results in the development of cracks normal to the cooling surface. The cracks usually form polygonal shapes, as is commonly observed in drying out mud. These cracks result in the development of polygonal columns. Almost perfectly hexagonal columns are developed at Giant's Causeway in Northern Ireland.

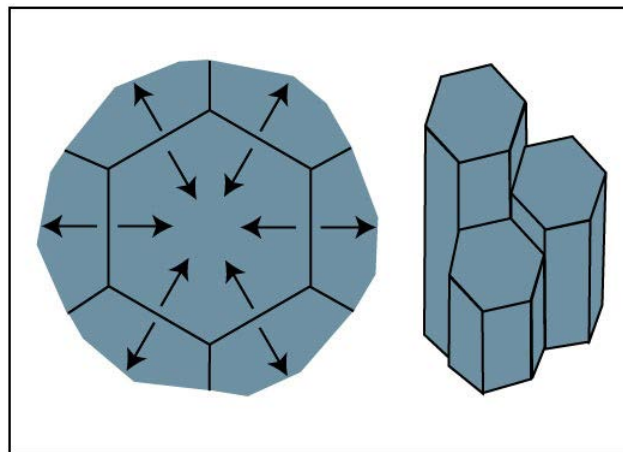


Fig. 5.15: The formation of columnar jointing in cooling volcanic rocks. Contraction during cooling commonly results in the development of polygonal cracks perpendicular to the cooling surface. The resulting columns can be extremely symmetrical.

5.5.2 Major intrusions (plutons)

The largest kind of pluton is called a *batholith*. Most batholiths consist of a large number of smaller plutons, but each of these can have surface areas of several 100km² (Fig. 5.16a). The total area of the largest batholiths (which comprise hundreds of intrusions) is more than 1000 × 250km. Huge batholiths occur in the Andes and in the western USA and Canada. Batholiths, which cut across the enveloping rocks (i.e. they are discordant bodies), are dominantly composed of granitic rocks. Granitic magma, formed by the partial melting of continental crustal rocks (Fig. 5.16b), rose towards the surface in diapiric-like bodies (well illustrated by “lava lamps”) because their density was lower than the enveloping rocks.

At low levels in the crust these country rocks would behave in a plastic manner and could be deformed (Fig. 5.16c); higher up they would be less deformable and would fracture (Fig. 5.16d). Blocks of country rock can be broken off and fragments sink through the rising magma (a process called *stoping*). These fragments of foreign rock (called *xenoliths*) can react with the magma and be assimilated or survive to be visible in the solidified granite. Small, high level plutons may be referred to as stocks. Intrusive diapirs of granitic magma can form successively and intrude previous intrusions to form a complex of intrusions with cross-cutting relationships.

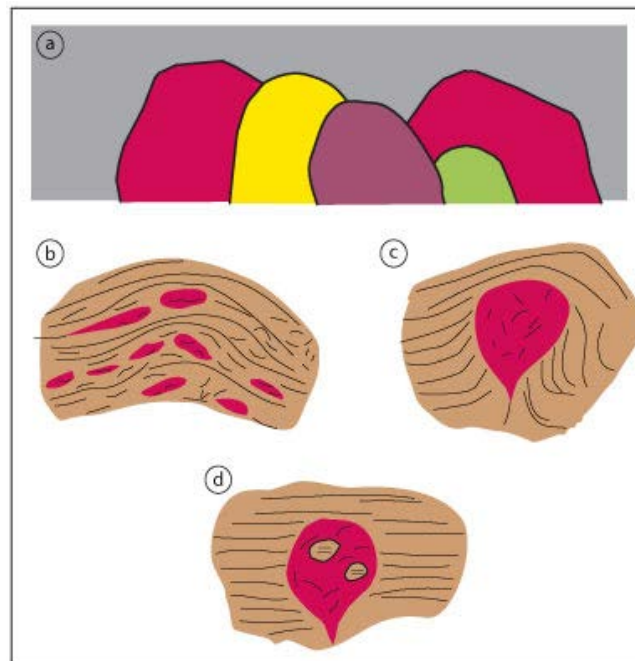


Fig. 5.16: Batholiths and the forms of granitic intrusions at different crustal depths.

a) Batholiths consist of a number of individual plutons. b) Granitic magma forms by the partial melting of the lower continental crust. c) The buoyant magma rises diapirically through the plastic lower crust and deforms the surrounding rocks. d) At higher levels the crust behaves in a brittle fashion and fractures. Blocks of country rock (xenoliths) become enveloped in the magma.

Magma chambers filled by basaltic magma commonly form funnel-shaped bodies with a feeder channel at the base, gradually broadening upwards (Fig. 5.17). Magma chambers of this type are commonly filled by many influxes of magma. The resulting magma mixing can be responsible for the formation of important economic mineral deposits of, for example, chromium and platinum-group elements.

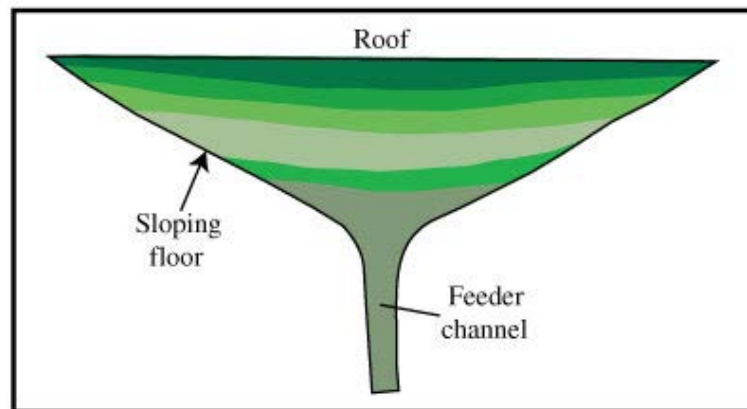


Fig. 5.17: Repeated magma influx is commonly involved in the formation of magma chambers. Funnel-shaped magma chambers (filled here by six major influxes of magma) are common for mafic intrusions. The dimensions are usually in the range 2–10km thick and 5–100km across. This type of intrusion may be layered.

5.5.3 Chilled margins

Intruded magma is always hotter than the surrounding rocks. The magma that comes into contact with these “*country rocks*” will therefore be rapidly cooled and solidify quickly. The interior of the intrusion will cool and crystallize more slowly. The “*chilled margin*” may cool so rapidly as to give a glassy rock that is indistinguishable from lava cooled at the surface of the Earth. The margins of pillow lavas are chilled against seawater (Fig. 5.10). The chilled margin will gradually become more coarsely crystalline away from the margin. The thickness of a chilled margin will depend on the temperature difference between the magma and the country rocks and the size of the intrusion. While the intrusion is cooling down the country rocks will be heated. This can result in the development of new minerals in rocks surrounding intrusions – the rocks become *contact metamorphosed*.

5.6 The origin of magma

As we have seen, there are three main types of magma – basaltic, andesitic and rhyolitic. Of these, basalts are the most voluminous and widespread. The three types of magma occur in different settings in a plate-tectonic context; this must be an expression of their contrasting mode of origin.

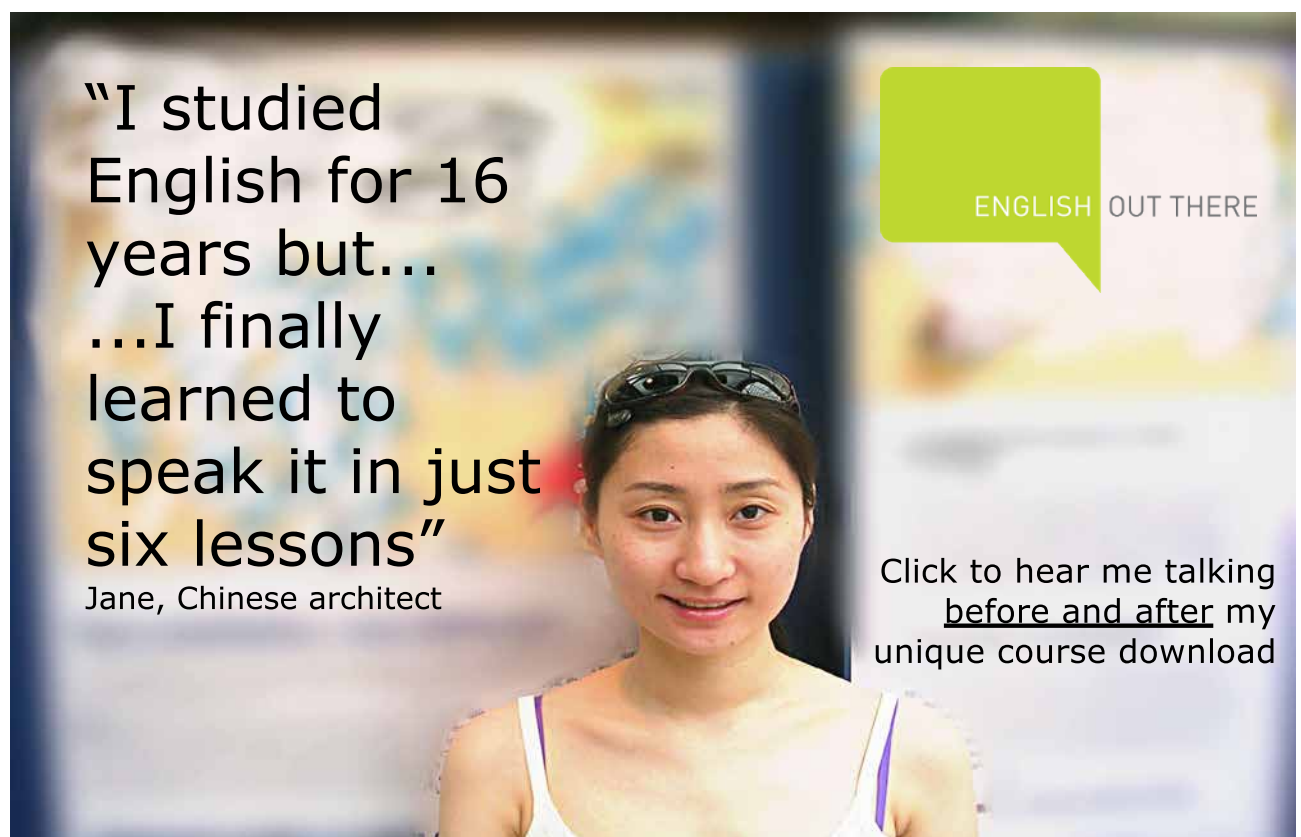
5.6.1 Distribution of volcanoes

We have noted in section 5.2.1 that basaltic magma is erupted more or less continuously along mid-ocean ridges and forms new oceanic crust. This magma must be formed in the mantle because there is nothing else below oceanic crust. Basaltic magma is also erupted from “hotspots” which are areas where large volumes of magma (largely basaltic) are produced, not associated with mid-ocean ridges (Fig. 5.7). An example is Hawaii which is located on oceanic crust far from any plate margin.

Hawaii is in fact the youngest of a long chain of extinct volcanic islands which extend to the northwest. This is because the hotspot from which the basaltic magma is produced is stationary and the oceanic plate is moving towards the northwest so that extinct volcanoes have been removed from their roots and transported with the plate. The oldest extinct volcanoes in the extreme northwest were formed about 70 million years ago. The hotspot has therefore been active and stationary for at least this length of time. Hotspots also occur in continental crust; the plateau basalts of the Columbia River in USA are an example of the products of a continental hotspot. This magma presumably formed in the mantle and passed through the continental crust on its way to the surface. Iceland is an exceptional hotspot because it is located close to a mid-ocean ridge.

Andesitic volcanism occurs both on oceanic and continental crust. It is not, however, related to mid-ocean ridges or hotspots. Andesite magmatism (and some basalt volcanism) is related to convergent zones where oceanic crust is subducted (Fig. 5.8).

Rhyolitic volcanism is restricted to the continental crust and is usually explosive in nature – because of the high viscosity and elevated amount of dissolved gas. Since rhyolites do not occur on oceanic crust it is unlikely that they are derived from the mantle and a continental crustal source seems likely.



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5.6.2 Origin of basaltic magma

Laboratory experiments have demonstrated that basaltic magma can be formed by the partial melting of dry peridotite. The upper mantle is composed of peridotite (an ultramafic rock consisting of olivine and pyroxene(s) together with minor amounts of an Al-bearing phase: garnet, spinel or plagioclase, in sequence of decreasing pressure). As we have seen above, the distribution of basaltic volcanism implies that basaltic magma is derived from the mantle, and this is supported by experimental evidence.

For the sake of simplicity we will consider that a rock which is being melted consists of two minerals A and B which have a so-called *eutectic* relationship (this requires that the two minerals do not react together in any way). In Fig. 5.18 there are four phase fields: melt; melt + crystals of A; melt + crystals of B; and crystals of A + B (i.e. totally solid). Point E is called the eutectic point and is in contact with all four fields. The boundaries separating the liquid-only fields from the liquid + crystals fields are *liquidus* curves. The boundary separating solid-only from crystals + liquid is the *solidus*. Mineral A melts at temperature T_1 and B at T_2 , but mixtures of A and B begin to melt at a temperature lower than pure A or B called the eutectic temperature (T_3).

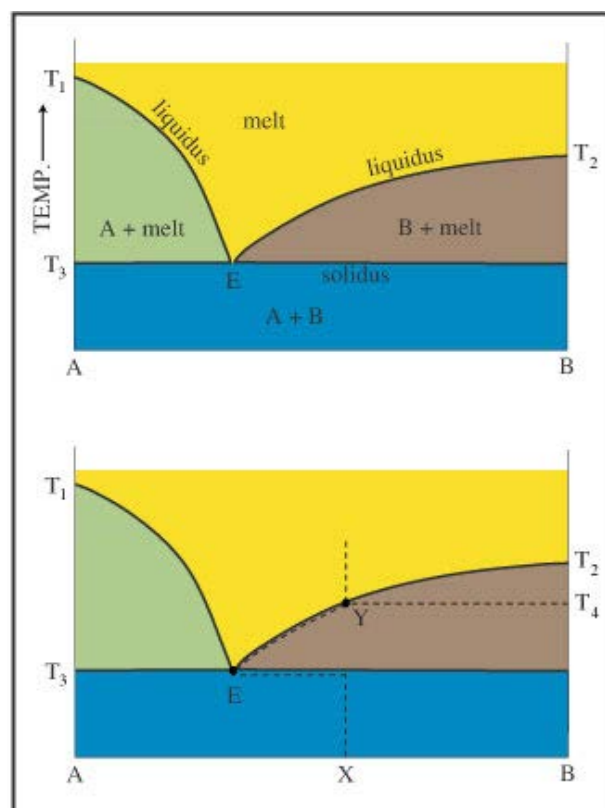


Fig. 5.18: Melting relations in a eutectic system.

In a very simplified way this diagram illustrates why partial melting of mantle peridotite (X) produces a melt with a different, but constant composition (basalt at E).

When a rock with composition X consisting of minerals A and B is heated, nothing happens (apart from it getting hotter!) until temperature T₃. At the eutectic temperature the two minerals will begin to melt in eutectic proportions and produce a liquid with the eutectic composition (E). Melting continues at constant temperature until the last crystals of A disappear at which time there are still crystals of B present. With further heating the temperature rises, further crystals of B melt and the liquid changes in composition along EY. When the last crystal of B disappears at T₄ the liquid has the same composition as the original solid and further heating simply raises the temperature of the liquid.

As we can see, the composition of the first melt is different from the composition of the rock being melted and always has the eutectic composition. This is a very simplified model for mantle melting; the mantle peridotite would be equivalent to the melting assemblage (A + B) and basaltic magma has the equivalent of the eutectic composition. This explains why basaltic magma has essentially the same composition no matter where it occurs. In reality, more than two minerals are involved in mantle melting and the location of the eutectic point varies with pressure, but as a first approximation this system illustrates mantle melting to produce basaltic magma.

A 10–15% partial melt of mantle peridotite is sufficient to produce enough basaltic magma that it can separate from the residual solid and move upwards. The mantle material that remains after it has produced a portion of basaltic magma is called “depleted”, “barren” or “residual” mantle peridotite (it will still be a peridotite); mantle material that has not been partially melted is called “undepleted” or “fertile” peridotite.

5.6.3 Origin of andesitic magma

Andesitic volcanism is related to subduction zones. Laboratory experiments have demonstrated that andesitic magma is produced by the partial melting of mantle peridotite under wet conditions. But how does water get down into the mantle? Basalts are produced at mid-ocean ridges. A circulating system of hot hydrous fluids is established as seawater, which percolates downwards through the basaltic lavas and dykes, becomes heated and rises to the surface. This is a so-called hydrothermal system. The hot basalts react with the water and the dry minerals in basalt become altered to water-bearing minerals (serpentine, chlorite, epidote). When these water-bearing rocks are subducted down into the mantle at a convergent margin, water is released into the overlying mantle. This lowers the solidus temperature of the peridotite which starts to melt to produce andesitic magma.

5.6.4 Origin of rhyolitic magma

The fact that rhyolitic volcanism is restricted to continental areas and has high volatile contents implies that it is derived from the “wet” partial melting of continental crust. This is supported by laboratory experiments which show that the first melt that is produced by the partial melting of wet continental crustal material has a rhyolitic composition. The source of heat is the mantle or rising mafic magma so that partial melting takes place in the lower continental crust. The rhyolitic melt will be less dense than the surrounding rocks and will rise towards the surface. The felsic magma will, however, be very viscous and the surrounding rocks will be relatively cold. The rhyolitic magma will therefore commonly cease to rise and crystallize at depth to form a granitic pluton. This is why granitic plutons are much more common than rhyolitic volcanism. As we will see later, rhyolitic magma can also be produced by the extensive fractional crystallization of basaltic magma in magma chambers.

5.6.5 Crystallization of magmas

We have established that basaltic, andesitic and rhyolitic magmas can be produced by the partial melting of different source materials in different plate tectonic settings. Basaltic magma is more *primitive* (is a higher temperature, more mafic magma) than andesite which is more primitive than rhyolite which is an *evolved* magma (low temperature, felsic). Magmas change composition as they crystallize. This is because the minerals that are formed have different compositions from the magma. This can be illustrated by considering crystallization in a eutectic system.



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5.6.5.1 Eutectic crystallization

Fig. 5.19 is simply the reverse of the melting process dealt with in Fig. 5.18. A melt with composition K will start to crystallize mineral A at temperature T_5 when it has cooled sufficiently to reach the liquidus at L. A monomineralic rock consisting of crystals of A will form. The melt will change in composition along the liquidus curve from L towards E as cooling continues and mineral A crystallizes. When the melt reaches the eutectic point, minerals A and B will crystallize together in their eutectic proportions. The temperature will remain constant (T_3) until the whole system has crystallized and there is no liquid left. If both minerals can sink through the melt the rocks at the floor will comprise a monomineralic rock A overlain by biminerallitic rock AB in their eutectic proportions. If we consider that A is augite and B is plagioclase, the rock sequence will be clinopyroxenite overlain by gabbro. We have considered a simple binary system; in nature magmas are more complex and involve the successive crystallization of several minerals, but it is clear that magma become more evolved as it crystallizes and minerals are removed.

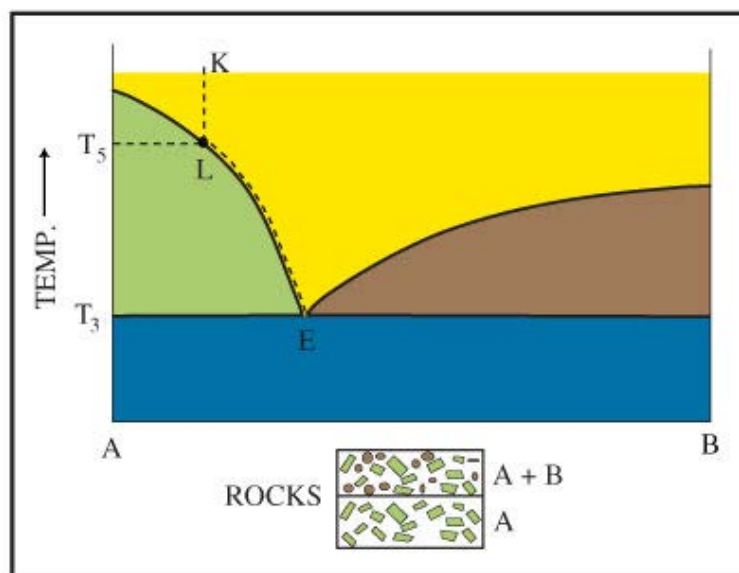


Fig. 5.19: Crystallization relationships in a eutectic system.

Mineral A crystallizes from L to E where it is joined by mineral B. These two minerals then crystallize together in constant (eutectic) proportions at constant temperature until all the melt has disappeared. To start with (from L to E) a rock consisting of mineral A will be formed. At E a rock consisting of minerals A and B (in their eutectic proportions) will be formed.

Eutectic crystallization of basaltic magma in a magma chamber should produce a gabbro consisting of the crystallizing minerals in their eutectic proportions. If we consider a simplified system this would be plagioclase and clinopyroxene in the proportions ~60:40. Many gabbroic bodies, however, consist largely of plagioclase-rich layers alternating with clinopyroxene-rich layers in so-called layered gabbros. Layers are commonly graded with clinopyroxene-rich bases grading into plagioclase-rich tops. The bulk composition of all the layers gives the eutectic composition which means that some physical process has been superimposed on the eutectic system to form the layers. Plagioclase and clinopyroxene have different densities so that one possibility is that they nucleated in bursts and settled onto the magma chamber floor according to their density.

5.6.5.2 Fractional crystallization

The removal of crystallizing minerals from magma means that they can no longer react with the melt. Different minerals crystallize successively at progressively lower temperatures. The process whereby minerals are removed in sequence and the magma changes in composition in consequence is called fractional crystallization. Fractional crystallization can result in, for example, the evolution of basaltic magma to andesitic magma (via basaltic andesite, intermediate between basalt and andesite; Fig. 5.20) which in turn can evolve to rhyolitic magma (via dacite, a magma composition intermediate between andesite and rhyolite; see Table 5.4). The process of fractional crystallization can be illustrated in a TAS diagram (used for the classification of volcanic rocks in Fig. 5.4).

We will consider a typical *basaltic magma* with 50wt.% SiO_2 and 2.5wt.% $\text{Na}_2\text{O} + \text{K}_2\text{O}$ which is located at A in Fig. 5.20A. This basalt resides in a magma chamber. The first mineral to crystallize from basaltic magma is magnesium-rich *olivine*. The magnesian end-member of the olivine group (forsterite: $\text{Mg}_2[\text{SiO}_4]$) contains 42wt.% SiO_2 and the iron-rich end-member (fayalite: $\text{Fe}_2[\text{SiO}_4]$) has 30wt.% SiO_2 . Olivine contains no sodium or potassium so that its composition plots as shown in Fig. 5.20B. As olivine with a composition of about Fo_{75} crystallizes the magma composition moves away from that of olivine towards B in Fig. 5.20B. The olivine will become gradually more iron-rich (recall the “cigar” diagram for the crystallization of olivine) and so the path from A to B will in fact be slightly curved. Olivine crystals will sink to the floor of the magma chamber where they will accumulate to form a rock consisting entirely of olivine – a *dunite* (see the classification scheme for ultramafic rocks in Fig. 5.2).

The next mineral to start to crystallize will be *clinopyroxene*, typically augite with composition $\text{Ca}(\text{Mg,Fe})[\text{Si}_2\text{O}_6]$. This contains 54–47wt.% SiO_2 and, like olivine, augite contains no sodium or potassium; it plots to the right of olivine in Fig. 5.20C. As olivine and augite crystallize the magma will change in composition with a vector like BC in Fig. 5.20C. Augite and olivine will sink to the floor of the chamber where a *wehrlite* (ultramafic rock consisting of olivine + clinopyroxene) will form above the dunite. Olivine and augite will crystallize in their eutectic proportions (Fig. 5.19).

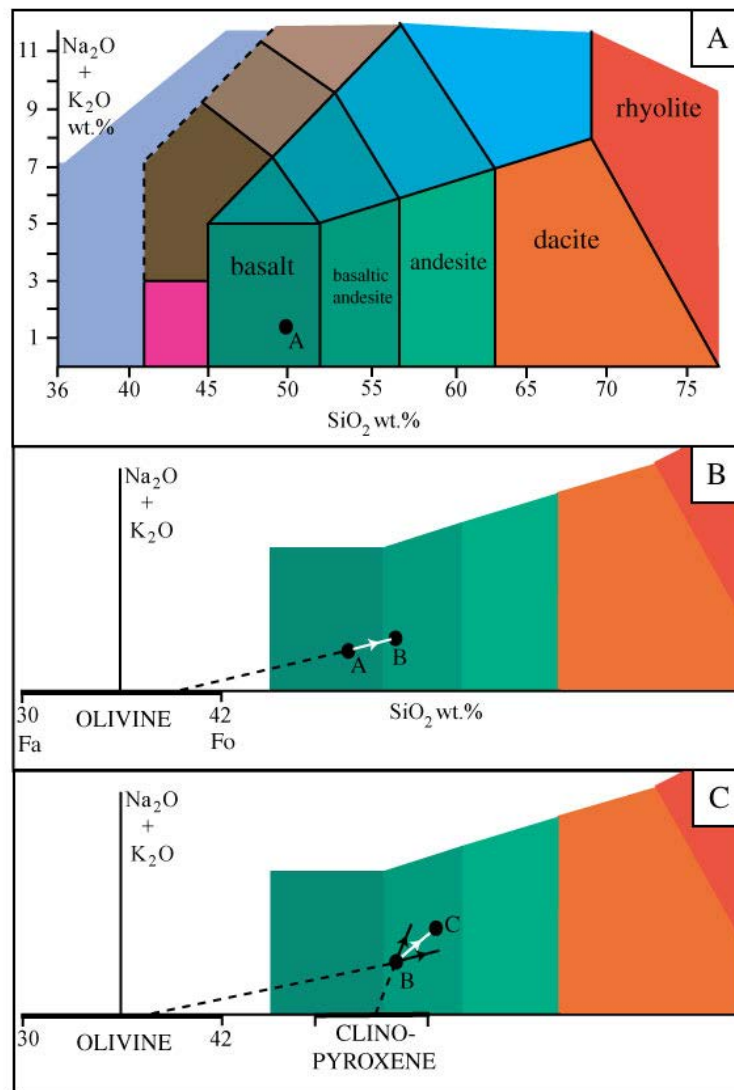



Fig. 5.20A-C: Compositional evolution of magma as a result of fractional crystallization illustrated in a TAS diagram. A. Starting composition of basaltic melt. B. Magma A changes composition towards B as magnesium-rich olivine crystallizes. C. Magma B changes composition towards C as olivine and clinopyroxene crystallize together.

After olivine and augite have crystallized for some time they will be joined by *plagioclase feldspar*. Plagioclase, like olivine, forms a solid solution series. The calcium end-member anorthite ($\text{Ca}[\text{Al}_2\text{Si}_2\text{O}_8]$) contains 44wt.% SiO_2 whereas albite ($\text{Na}[\text{AlSi}_3\text{O}_8]$) contains 68wt.% SiO_2 and 11wt.% Na_2O . Plagioclase will therefore plot along a long line with a positive slope in Fig. 5.20D. The first plagioclase to crystallize will have a composition of about An_{75} . The effect of plagioclase removal will be very similar to that of olivine removal in Fig. 5.20D. As plagioclase, olivine and augite are removed from melt C it will change in composition in the direction of D in Fig. 5.20D. The minerals will accumulate on the floor of the magma chamber and form an *olivine gabbro* (a plutonic rock composed of olivine, plagioclase and clinopyroxene) overlying the earlier-formed dunite and wehrlite.

The next mineral to start to crystallize will probably be *magnetite* (Fe_3O_4) and olivine may cease to crystallize. Magnetite is not a silicate mineral and contains no SiO_2 , K_2O or Na_2O . Crystallization of magnetite will therefore have a large effect on the composition of the evolving melt, driving it to the right in Fig. 5.20E. Augite (more iron-rich than previously) and plagioclase (more sodium-rich than previously) will continue to crystallize together with magnetite and the melt composition will move up and to the right in Fig. 5.20E. At this stage of crystallization other phases, such as amphibole, may start to form.

The compiled compositional evolution of the melt is shown in Fig. 5.20F. The initial melt (A) which was in the *basalt* field ends (E) in the *rhyolite* field, having passed through the *basaltic andesite*, *andesite* and *dacite* fields on the way. The melt has been progressively depleted in mafic components and enriched in felsic components. When the melt approaches the rhyolite field, K-feldspar and quartz will begin to crystallize.

The compositional evolution of the melt changes in direction when new phases begin to crystallize, such as at B in Fig. 5.20. The $\text{Na}_2\text{O} + \text{K}_2\text{O}$ vs. SiO_2 diagram has been chosen here to illustrate fractional crystallization because it is used to define volcanic rock names (as the TAS diagram; Fig. 5.4). Other diagrams could have been chosen to show the evolution of melt composition during fractional crystallization e.g. MgO vs. CaO .



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How much rhyolite can be formed from basaltic magma by fractional crystallization? A rough estimate can be made from inspection of the melt compositions in Table 5.4. K_2O does not enter any of the crystallizing minerals until K-feldspar starts to form in rhyolite. The basaltic magma in Table 5.4 contains 0.51% K_2O , increasing by a factor of ca. 9 to 4.50% in the rhyolite. This means, as a first approximation, that *about 10% rhyolite can be formed by the perfect fractional crystallization of basalt.*

Fractional crystallization, which is a very important process in igneous petrology, takes place in magma chambers.

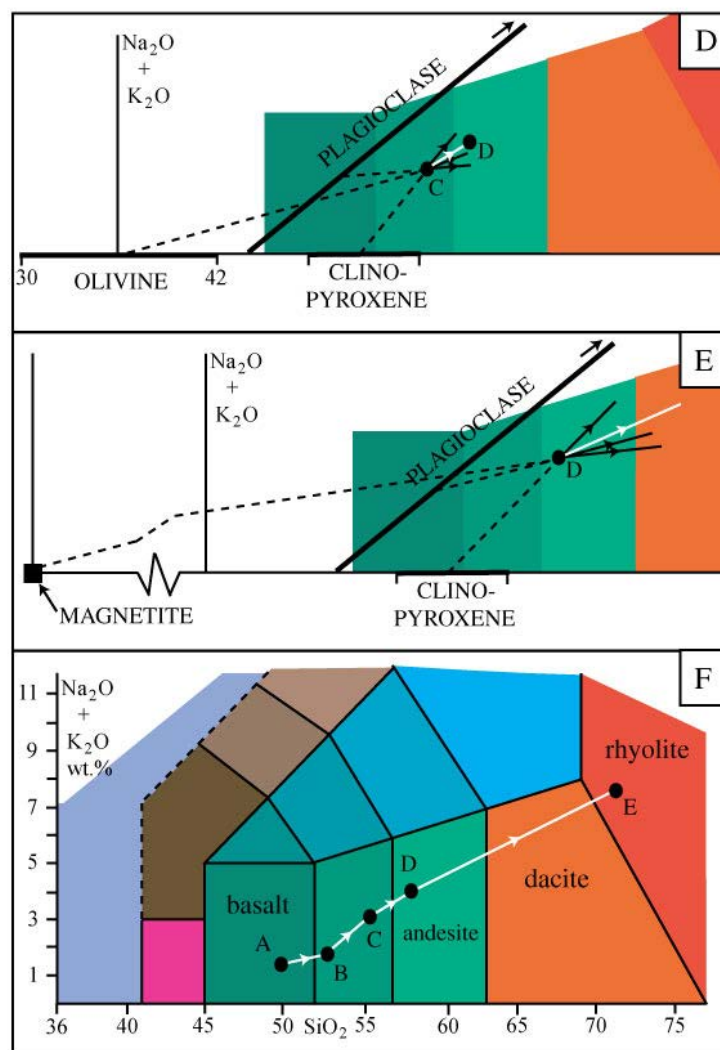


Fig. 5.20D-F: Compositional evolution of magma as a result of fractional crystallization illustrated in a TAS diagram.

D. Magma C changes composition towards D as olivine, clinopyroxene and plagioclase crystallize together. E. Magma D changes composition towards E as magnetite crystallizes together with plagioclase and clinopyroxene. F. Summary of the fractional crystallization process. Efficient fractional crystallization can result in basaltic magma evolving to rhyolitic magma.

5.6.5.3 Bowen's reaction series

As we have seen when discussing olivine in the mineralogy notes, this is a solid solution mineral which changes in composition as the melt crystallizes because of continuous reaction between crystals and melt. Early, high temperature crystals are magnesian; later, lower temperature crystals are enriched in iron. If early-formed crystals are removed from the melt it will be able to become extremely iron-rich. *Plagioclase* also forms a *solid solution series*; early-formed plagioclases are calcium-rich (i.e. rich in the anorthite component) whereas late-formed ones are sodium-rich (albite-rich). The same relationship is valid for, for example, pyroxenes; early-formed pyroxenes are Mg-rich and they become increasingly Fe-rich as crystallization proceeds. All these variations in composition are a result of the continuous reaction between crystals and melt. The most important of these solid solution minerals in igneous petrology is plagioclase because this mineral crystallizes over an extremely wide range of magma compositions, from basaltic to rhyolitic. In Fig. 5.21 plagioclase represents a *continuous reaction series*.

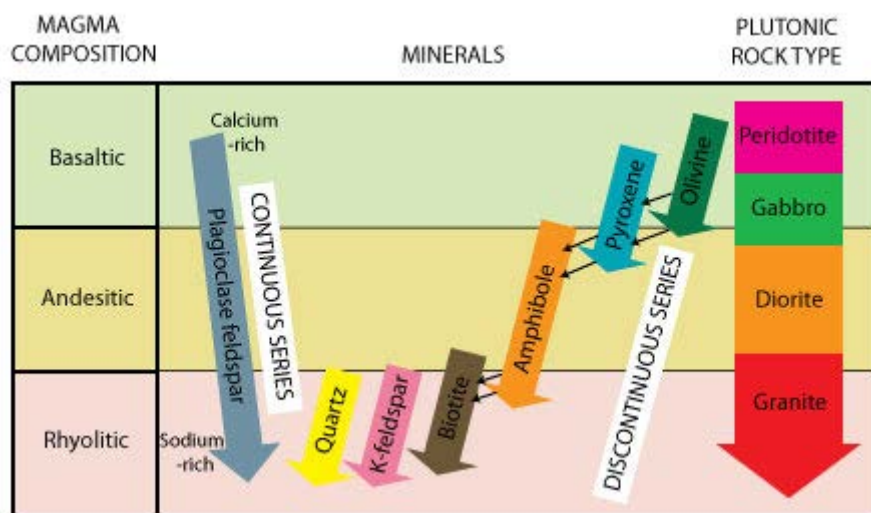


Fig. 5.21: Bowen's reaction series.

This comprises a "discontinuous" series as one mineral is followed by another and a "continuous" series as a mineral gradually changes composition. Plagioclase gradually becomes less calcium- and more sodium-rich. Olivine and pyroxenes likewise gradually become less magnesium- and more iron-rich (not illustrated).

The first mineral to crystallize from basaltic magma is usually Mg-rich olivine which gradually becomes more Fe-rich as crystallization proceeds (because of continuous reaction). Olivine forms a continuous reaction series, but much shorter than plagioclase in Fig. 5.21. Olivine contains less SiO_2 than the magma which will become enriched in this component (as we have seen in connection with Fig. 5.20). When the amount of SiO_2 in the magma reaches a critical level, *olivine reacts with the melt to form pyroxene* which contains more SiO_2 than olivine (Fig. 5.21). The minerals that have crystallized so far are all anhydrous so that the amount of H_2O dissolved in the remaining melt will increase. When this reaches a critical level *pyroxene reacts with the melt to form amphibole* which is a hydrous mineral. Amphibole can in turn react with the evolving melt to form *biotite*. By this time the melt will have become so enriched in silica and potassium that *quartz and K-feldspar* begin to crystallize from rhyolitic magma to form granite. This sequence of minerals develops as a result of the discontinuous reaction between crystals and melt and is called the *discontinuous reaction series*. It is important to realise that these reaction processes take place in magma chambers.

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These two types of reaction series form the basis for fractional crystallization and were first recognised by Norman L. Bowen after whom they are named. Bowen thought that basalt was the parental magma for all more evolved compositions i.e. all igneous rocks were ultimately formed as a result of the fractional crystallization of basaltic magma. We now recognise fractional crystallization as an important process, but it is not responsible for the formation of e.g. most andesitic volcanism and huge granitic batholiths. As we have seen, the perfect fractional crystallization of basaltic magma produces only 10% of rhyolitic magma which can crystallize to form granite. If granitic batholiths were formed by fractional crystallization of basalt, this implies that batholiths are underlain by magma chambers containing the early-crystallized products of the fractional crystallization of basaltic magma and these have volumes some ten times greater than the batholiths themselves. This is clearly not the case.

5.7 Igneous mineral deposits

Magmatic processes are responsible for the formation of a variety of economic mineral deposits. Some pegmatites contain economically important minerals. As fractional crystallization nears its conclusion in rhyolitic magma, some elements that were present in the original magma in small quantities become greatly concentrated in the highly evolved residual melt. This is because they do not enter the structures of the crystallizing minerals, either because of their large (e.g. uranium) or small (e.g. beryllium, lithium, boron) size. Granitic pegmatites may therefore contain exotic minerals that allow small or large elements to enter their structure e.g. beryl and tourmaline (contain beryllium and boron respectively).

Some layered gabbros contain layers that are enriched in economically important minerals. Layered intrusions are commonly formed by multiple influxes of magma. The new magma can mix with the old, resident magma to produce a variety of features, including deposits of economic minerals, such as chromite. Two very important examples are the Great Dyke in Zimbabwe and the Bushveld Intrusion (the largest intrusion in the world) in South Africa. The latter also contains the world's largest economic deposit of platinum-group minerals.

Basaltic magma contains a small quantity of sulphur in solution. Since early-formed minerals (olivine, pyroxene, plagioclase) do not contain any sulphur in their structure it will become concentrated in the melt. When the amount of dissolved sulphur reaches a critical level a sulphide liquid separates from the silicate liquid. The magma separates into two *immiscible liquids* – like oil and water. Some of the elements in the magma are rapidly transferred to the sulphide liquid – like iron, nickel and copper. The sulphide droplets have a much higher density than the silicate liquid and quickly sink to the floor of the magma chamber. Most of the world's economic deposits of nickel were formed in this way.

6 Sedimentary rocks

6.1 Introduction

All rocks on the surface of the Earth are exposed to the effects of erosion. The products of erosion accumulate to form sediments. Young sediments are unconsolidated (loosely connected) but with time they become consolidated and form sedimentary rocks. Sediments form on the surface of the Earth (most often under water) and occur in layers – with the youngest layers at the top.

The thickness of sediments on the surface of the Earth varies widely, from totally absent (for example at mid-ocean ridges) up to a maximum of ca. 20 km. Sedimentary rocks cover ca. 80% of the surface of the Earth, but contribute less than 1% to its mass. Sediments are, however, economically very important because they contain the major energy sources – coal, oil, and gas.

6.2 Weathering

There are two types of process that cause the alteration of rocks exposed on the surface of the Earth – *physical* and *chemical*. Physical or mechanical alteration results in rocks breaking into smaller fragments; these rock fragments are called *detritus*. Different grain sizes of detrital fragments are named as shown in Table 6.1.

Term	Size range (mm)
boulder	>256
cobble	64–256
pebble	2–64
sand	$\frac{1}{16}$ –2
silt	$\frac{1}{256}$ – $\frac{1}{16}$
mud	< $\frac{1}{256}$

Table 6.1. Grain size of unconsolidated sediments

Boulders, cobbles and pebbles are coarse-grained, sand is medium-grained, and silt and clay are fine-grained.

There are a variety of processes that result in *physical weathering*. Rocks buried in the crust are under pressure because of the weight of the overlying deposits and hotter than at the surface (the increase in temperature with depth is usually in the range 20–40°C/km). *Joints* form when pressure is reduced as a result of uplift and erosion. The type of joints depends on the kind of rock involved (Fig.6.1).

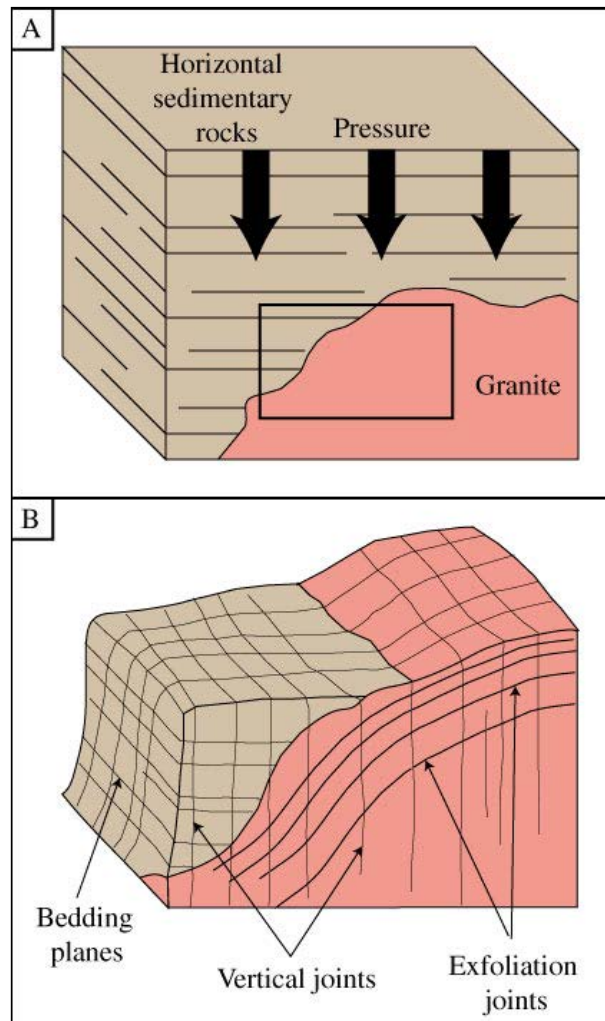


Fig.6.1: Development of joints in granites and sediments as a result of uplift and erosion. Granites (homogeneous, massive rocks) commonly form onion-like sheets, whereas sediments develop vertical fractures.

Loose blocks of rock resulting from jointing can accumulate at the base of rock exposures to form *talus*. There are several agents that can cause joints to open and rocks to form fragments. These include frost-wedging (water enters cracks and expands on freezing; most effective where freezing and thawing alternate rapidly) (Fig.6.2), root-wedging, salt-wedging, thermal expansion and animal influence (including humans).

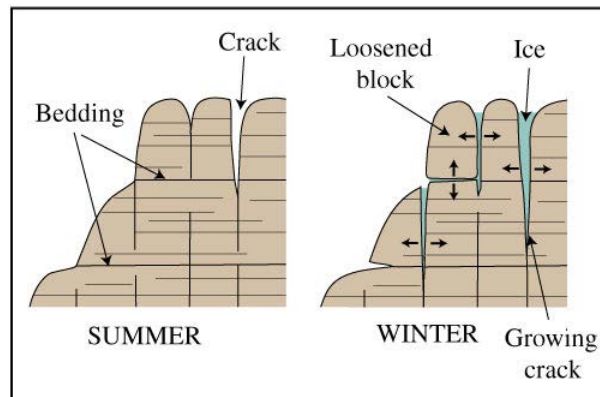
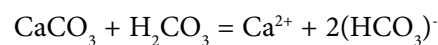


Fig.6.2: Cracks opening as a result of frost wedging.
This is a very efficient way of fracturing rocks.


Rocks are also affected by *chemical weathering*. Water plays a key role here because of chemical reactions with rock-forming minerals. Chemical weathering is most effective in warm climates. Halite (rock salt, NaCl) dissolves in pure rainwater, whereas calcite (CaCO_3) is only soluble in acidic water. Rainwater dissolves CO_2 from the air and forms carbonic acid H_2CO_3 . This can react with calcite:



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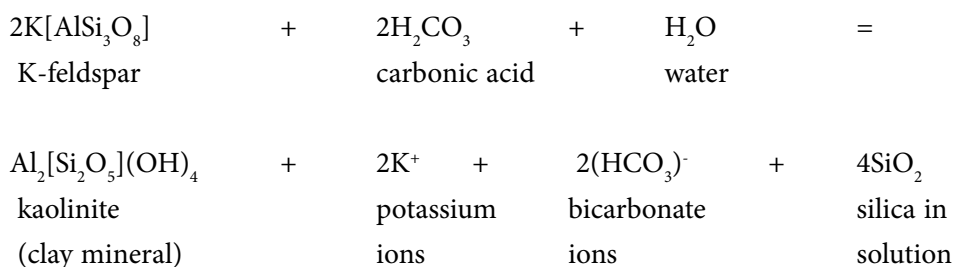
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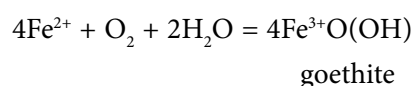
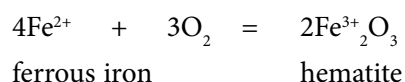
Because of the solubility of calcite, the rocks limestone and marble (metamorphosed limestone) dissolve in wet climates. Joints become wider and underground rivers and caves can form.

Water can react with other minerals. K-feldspar (important in e.g. granite) reacts with slightly acidic water to form kaolinite (a clay mineral).



Many other silicate minerals also react slowly with (acidic) water to form clay minerals (a process called hydrolysis). These include olivine, pyroxenes, amphiboles and micas (all dark minerals). Quartz, however, is chemically resistant and is often the only mineral to survive extensive chemical weathering.

Oxidation can play an important role in the chemical alteration of minerals. Iron is the critical element in this context because it exists in two oxidation states, Fe^{2+} and Fe^{3+} . The oxidation (rusting) of iron is a well-known phenomenon, not least to car owners. Two reactions are involved:



Both the minerals hematite and goethite are components of rust.

Physical and chemical weathering operate together. Mechanical processes give more cracks (fractures) and thereby a greater surface area. Chemical processes, which operate on surfaces, are therefore enhanced by extensive fracturing (Fig.6.3). Minerals become altered at different rates. Quartz is a very resistant mineral and is often the only one to survive chemical weathering; all the others alter to clay (Fig.6.4). Sandy beaches are usually dominated by quartz.

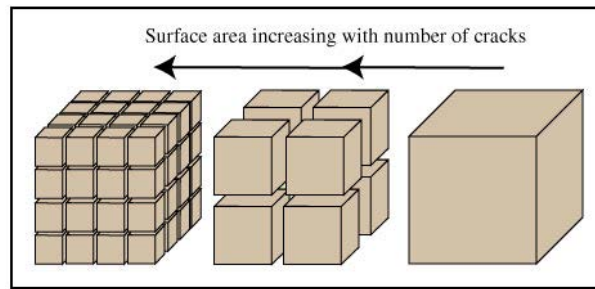


Fig.6.3: The efficiency of chemical weathering increases as the surface area increases because of mechanical weathering. The two weathering processes work together.

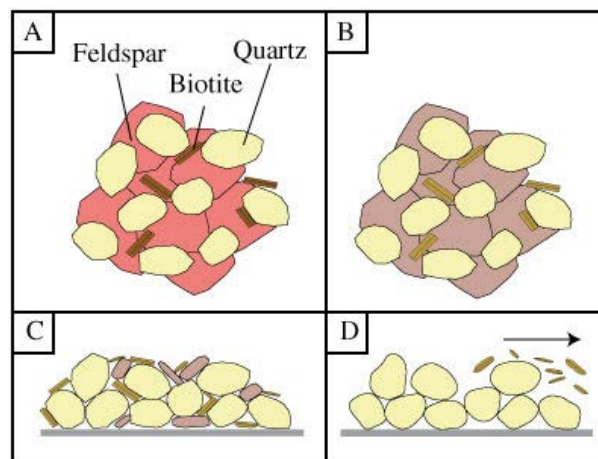


Fig.6.4: Minerals alter at different rates. Feldspar and dark minerals in granite alter relatively rapidly to clay minerals whereas quartz is very resistant. The clay fraction is easily removed, leaving quartz sand as the final product of weathering.

Both mechanical and chemical alteration proceed faster on edges – and faster still on corners. This commonly results in rectangular blocks becoming rounded – so-called spherical weathering (Fig.6.5).

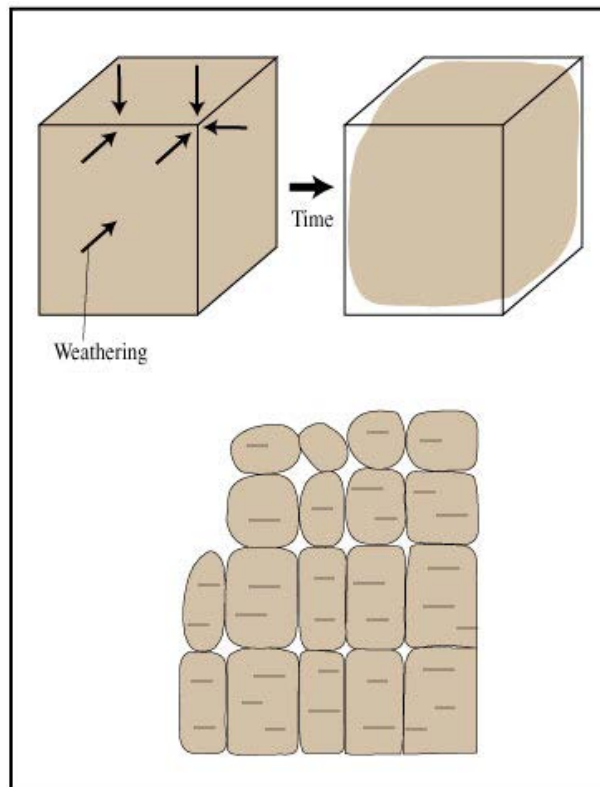


Fig.6.5: Extensive weathering results in the development of rounded boulders. This is a consequence of weathering being a surface effect.

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6.3 Classification of sedimentary rocks

Weathering processes form rock fragments, grains of resistant minerals, new minerals (e.g. kaolinite, a clay mineral) and ions in solution. Particles can be transported by wind, water or ice until they are deposited. Dissolved ions can enter the groundwater and/or be transported by streams and rivers to the sea. These ions form new minerals in the sea or in pore spaces underground. Organisms can play an important role.

There are four different types of sedimentary rocks:

- *clastic* – (or detrital) – consist of consolidated fragments
- *biochemical* – consist of the shells of organisms
- *organic* – consist of carbon-rich plant remains
- *chemical* – consist of minerals deposited directly from hydrous solutions

6.3.1 Clastic sedimentary rocks

Loose detritus becomes a clastic sedimentary rock via a series of events (Fig.6.6).

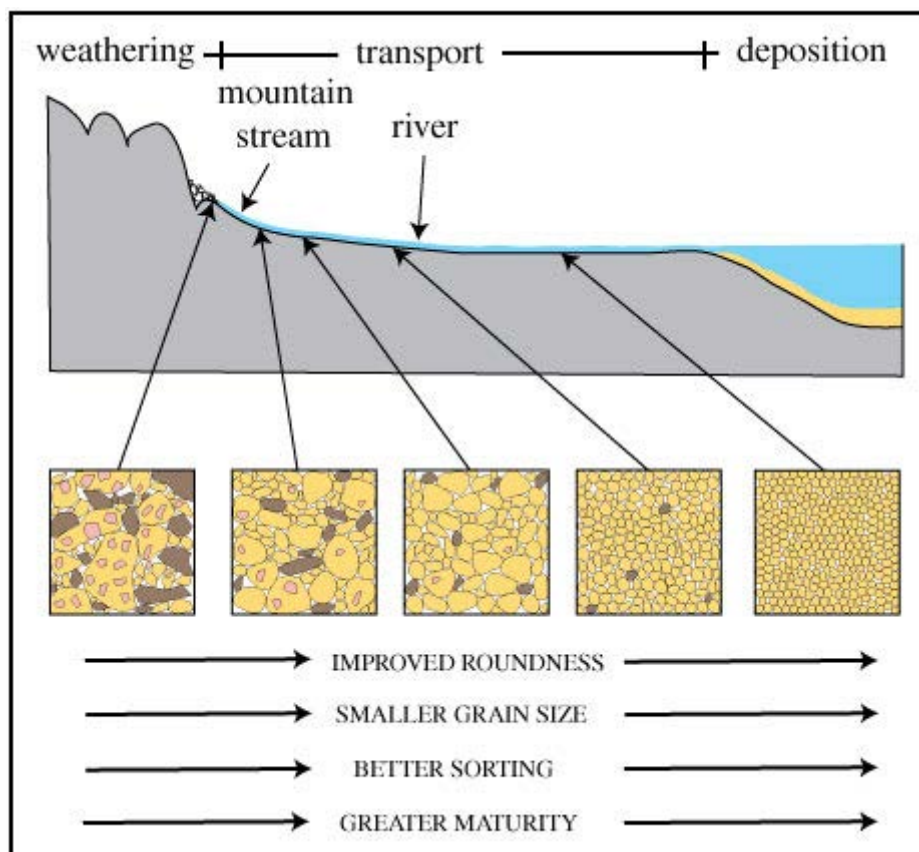


Fig.6.6: Formation of clastic sedimentary rocks involving weathering, transport and deposition.

An immature sediment (breccia) consists of angular fragments with a variety of grain size (poorly sorted). As a sediment becomes more mature the grains become increasingly rounded, smaller, and more uniform in size. The final product illustrated here is a mature sandstone.

- a) *Weathering* (dominantly physical) forms fragments (detritus).
- b) *Transport*. Rock fragments fall off the outcrop and/or are moved by wind, water or ice. Since water and wind move small particles further than large ones they become sorted by size during transport. The combined processes of weathering and transport are referred to as *erosion*.
- c) *Deposition*. Particles settle out of the medium of transport. For example when a glacier melts it deposits its load of rock particles. When flowing water slows down the coarsest particles settle out, or when the wind dies down sand particles are deposited.
- d) *Lithification* (consolidation) is the process whereby loose sediment is transformed into solid rock. The pressure resulting from burial under younger sediments presses water and air out of pore spaces and results in *compaction*. Sand can compact by 10–20%, whereas mud (a mixture of clay and water) can compact by as much as 50–80%. Compacted sediment may then be *cemented* by the deposition of new minerals (usually quartz or calcite) that are derived from groundwater. Lithification is the combined result of compaction and cementation (Fig.6.7).

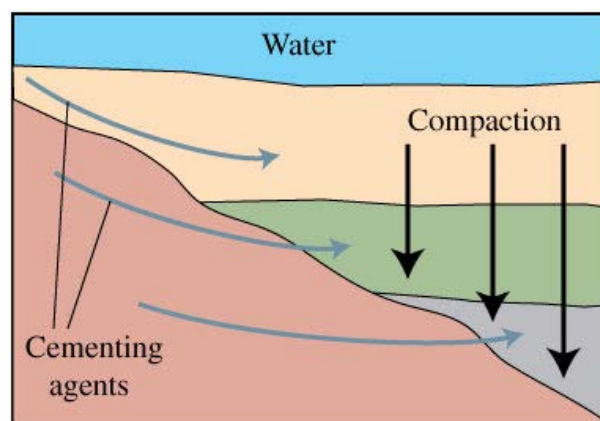


Fig.6.7: The sedimentary rock-forming process (lithification).

This involves a combination of compaction and cementation. The cementing agents are components carried by groundwater.

As rock (and mineral) fragments are transported away from their source they become smaller and change their shape. Sediments deposited near the source will be very different from those deposited far away after considerable transport.

A clastic sedimentary rock that forms very close to its source will contain large, angular fragments. These large fragments (clasts) will occur in a matrix (or groundmass) consisting of smaller particles. This type of rock is a *breccia*. The fragments can consist of igneous, metamorphic or sedimentary rocks (or a combination of these), depending on the nature of the rock type(s) that have been eroded.

During transport the angular fragments become rounded. A clastic sedimentary rock with large, rounded fragments in a finer grained matrix is called a *conglomerate*. Like breccias, the nature of the large particles in a conglomerate reflect the nature of the rock types that have been eroded.

With greater transport the fragments become smaller because of collision with each other, and they become exposed to further alteration. Dark minerals and feldspar become hydrolysed and turn into clay minerals that are transported away. The clasts are commonly dominated by quartz grains, and the rock which forms is a *sandstone* (i.e. particles in the size range $\frac{1}{16}$ –2 mm). In a river, sand-sized particles may be deposited in bars. If sand-sized particles reach the estuary of a river they will be deposited in the coastal region. A special type of coarse-grained sandstone containing feldspar as well as quartz grains is called *arkose*.

Smaller particles can be transported further but will be deposited at some stage to form *siltstone* after lithification (particles in the size range $\frac{1}{256}$ – $\frac{1}{16}$ mm). The smallest, clay-sized particles are transported even further and are deposited as mud. After lithification mud becomes *mudstone*. Clay particles, however, commonly form minute flakes that are orientated parallel with the bedding plane. After compaction (and cementation) these very fine-grained clastic rocks have a well-developed parting and the rock is called a *shale*.



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There is one other rock type that forms in a marine environment as a result of the redistribution of unconsolidated sediments (by submarine avalanches) and their accumulation “down-slope” on the deep sea floor. These contain a range of grain sizes in a fine-grained matrix and have the unusual name “*greywacke*”. These commonly show a graded structure in which the grain size decreases upwards.

During transport we observe that particles become smaller. As we get further from the source, the sedimentary rocks that form vary from breccia (angular fragments) and conglomerate (rounded clasts), through sandstone (and arkose) to siltstone and finally shale.

In addition to variations in grain size, sediments show varying degrees of *sorting* (Fig.6.6) i.e. the range of particle sizes present. A breccia, consisting of large fragments in a finer grained matrix, is very poorly sorted. Sandstone consisting of rounded quartz grains of equal size is very well sorted.

When describing the nature of a sedimentary rock, a widely used term is *sediment maturity* (Fig.6.6). The degree of maturity of a sediment expresses how much it has evolved from being just a crushed-up version of its source rock to a sediment that has lost its easily weathered components and has become well sorted and rounded. A breccia (coarse angular fragments in a finer grained matrix) is an immature sediment whereas a sandstone consisting entirely of well rounded quartz grains of uniform size is a very mature sediment.

6.3.2 Biochemical sedimentary rocks

Organisms can play a major role in the formation of sedimentary rocks. Many organisms have shells of CaCO_3 (calcite or its polymorph aragonite). Others have shells of silica (SiO_2). When these organisms die their shells can accumulate to form a biochemical sediment. The soft part of the organism rots away – or becomes transformed into oil. Plants can also contribute an organic component to sediments, as we will see in section 6.3.3.

The environment around a coral reef is extremely rich in organisms such as corals, algae, oysters, clams and snails. Plankton float in the water. All these have shells of CaCO_3 (calcite or aragonite). When the organism dies, its skeleton remains where it is (e.g. coral reef) or is transported away. During transport the skeletal material can break into small fragments. When this material is deposited it forms a calcium carbonate-rich sediment – *limestone*. Since this is largely composed of the remains of organisms it is a biochemical sediment. A different type of limestone can be precipitated directly from aqueous solutions without the influence of organisms (e.g. stalagmites and stalactites in caves). This type of limestone is of chemical origin and will be dealt with later. Limestones are sometimes formed in an environment where clay is deposited simultaneously. This gives rise to a rock type known as *marl* – a mixture of limestone and clay.

There are many varieties of limestone of biochemical origin. Some are dominated by coral reefs in their original position (*reef limestone*). Others consist largely of shells and shell fragments (*fossiliferous limestone*). Some contain small spheres of calcite (oolites; *oolitic limestone*) formed by coatings of calcite around small particles (usually shell fragments or quartz grains). These form in agitated, shallow water. Some consist of microscopic shells of plankton called foraminifera (*chalk*). Lime mud can consolidate to give very fine-grained limestone called *micrite*.

The shells of some organisms, most notably those of a type of plankton called radiolaria, are composed of silica (SiO_2). These tiny shells accumulate on the deep sea floor as a silica-rich ooze. After burial beneath younger deposits these solidify to form a type of cryptocrystalline quartz called *chert*. These chert deposits only develop in the deep ocean and commonly form bands (banded chert).

6.3.3 Organic sedimentary rocks

Coal is an organic sedimentary rock. It is formed from the remains of plants that grew in swamps or forests. The plant remains became buried and, after being subjected to elevated temperatures and pressures, were converted to the black, combustible rock known as coal which consists of >50% carbon.

The soft parts of plankton can mix with mud on the sea floor and be incorporated into shale. This organic material (which is gradually converted into oil) colours the shale black. Such black shales are called *oil shales*.

6.3.4 Chemical sedimentary rocks

Chemical sediments form by the precipitation of minerals from aqueous solutions. There are three main types of chemical sediments:

- *evaporites* – formed by the evaporation of salt water
- *travertine* – carbonate rocks precipitated from water
- *dolomite* and *chert* – formed by the replacement of other rocks

Evaporites. The evaporation of salt water leaves a residue of salt. For salt water to evaporate it requires a closed system (e.g. a salt lake with no outlet) and a warm climate. Salt water contains many other ions in solution than Na^+ and Cl^- , and a variety of minerals form in a regular sequence during the evaporation of salt water. The first mineral to form is *gypsum* ($\text{CaSO}_4 \cdot 2\text{H}_2\text{O}$) when about 80% of the water has evaporated, followed by *halite* (NaCl) when about 90% has evaporated. After this, a sequence of relatively rare evaporite minerals may form (including the potassium equivalent of salt, sylvite KCl).

As we have seen, limestone can form by the accumulation of biochemical material. It can also, however, form by direct precipitation from water without organisms being involved. This chemical variety of limestone is called *travertine*. Water, especially acidic water, can dissolve calcite in limestone. The carbonate material is, however, commonly precipitated again, often in limestone caves (as stalagmites and stalactites etc.) or around hot springs. Travertine, which is usually banded and beige in colour, is widely used as a facing stone.

Some rock types are formed by the replacement of pre-existing sediments. The question then arises as to whether it is reasonable to call them sedimentary rocks? The processes of burial (and the resulting compaction) and cementation are included in the “sedimentary” realm. Processes that take place after deposition, but not involving particularly elevated temperatures and/or pressure (which result in *metamorphism* which is dealt with in Chapter 7), are referred to as *diagenesis* and play a very important role in the formation of solid rocks from loose sediments.

Dolomite ($\text{CaMg}(\text{CO}_3)_2$) is a carbonate mineral in which half the calcium in calcite is replaced by magnesium. Dolomite forms as a result of reaction between calcite and Mg-bearing groundwater. Calcite can become partially *replaced* by dolomite. This replacement can take place soon or long after formation of the limestone. The term dolomite is used both for the mineral $\text{CaMg}(\text{CO}_3)_2$ and the rock that is formed.

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Chert is an extremely fine-grained (cryptocrystalline) variety of quartz. Black chert is called *flint*. Chert/flint are very fine-grained and almost glassy with a conchoidal fracture. Most plankton have shells composed of carbonate, but some have shells composed of SiO_2 . This silica, which is distributed throughout biochemical limestone deposited on the sea floor, becomes dissolved by percolating water and may be deposited elsewhere. The deposition of chert/flint usually starts around/on “impurities” present in the limestone – such as larger shell fragments (for example sea urchins). Deposition may continue along bedding planes, form nodules or take place in an irregular fashion. Reddish chert is called jasper. Fossilised wood has usually been replaced by chert, and detailed structures (such as tree rings) may be superbly preserved during this process. Agate is banded chert that has been precipitated in a cavity (usually in lava) and has been deposited inwards from the walls. Many commercial agates have been artificially coloured.

6.4 Sedimentary structures

Most sedimentary rocks contain structures that bear witness to process that took place during their formation. The most obvious structure is layering.

6.4.1 Layering (bedding)

A layered sequence consists of many *layers* (also called *beds*). The layers are separated from each other by *bedding planes*. Several layers are jointly referred to as *strata*. An alternative term for “layering” is “stratification”. *Stratigraphy* is the study of strata. A *stratigrapher* is a geologist who studies strata.

Different layers reflect changes in the type of material deposited and/or in the conditions during deposition (Fig.6.8). Depositional conditions change rapidly in a stream, resulting in many, thin, laterally discontinuous layers of different types of clastic sediment. Marine conditions can be stable over long periods of time, giving thick sequences with constant composition – like thick layers of chalk (e.g. the White Cliffs of Dover).

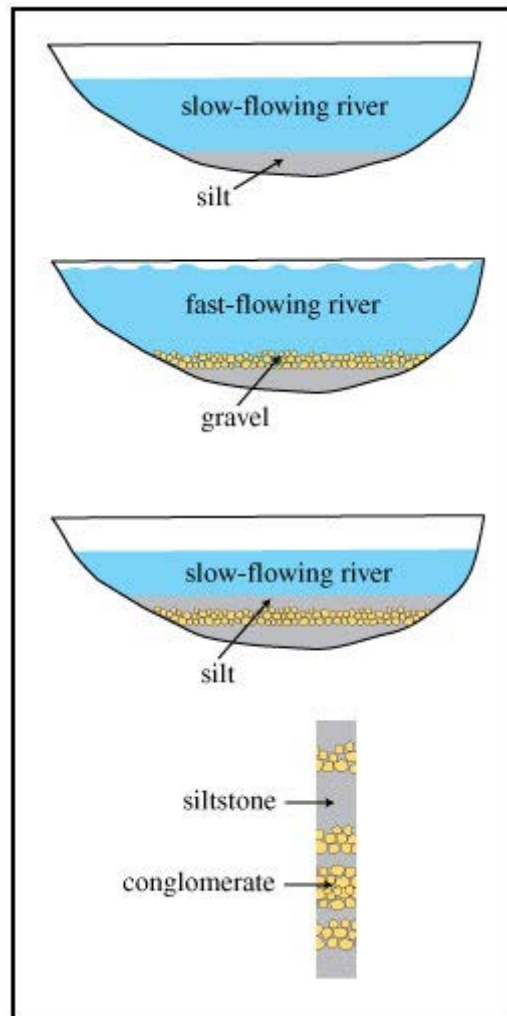


Fig.6.8. Bedding forms as a result of changes in the conditions of deposition. An alternating sequence of layers of siltstone and conglomerate may reflect changes in the strength of current in a stream or river.

A layer (or sequence of layers) that is so characteristic that it can be followed through the landscape for a large distance, comprises a stratigraphic formation. A *formation* is usually named after a place (the “type locality”) and the rock type. A group comprises several formations.

6.4.2 Surface markings

The bedding planes of sediments sometimes contain structures that reflect processes that took place during their formation. The tidal zone of beaches commonly show *ripple marks*. These form when a current (water or wind) moves sedimentary particles. A series of elongate ripple marks are formed perpendicular to the current direction. Such ripple marks on sandstone surfaces can be preserved during extensive metamorphism and deformation in quartzites and yet reflect gentle current activity on an ancient shallow sea floor.

6.4.3 Graded bedding

Loose sediment lying on a slope can become unstable and move down slope as a “mud flow” – consisting of different sized particles mixed with water. This process commonly takes place in a marine environment and is referred to as a *turbidity current* (Fig.6.9). When the rate of flow decreases, sediment particles are deposited in a sequence according to their size. The new sediment that forms is *size-graded* with sand (or gravel) at the base and mud/clay at the top.

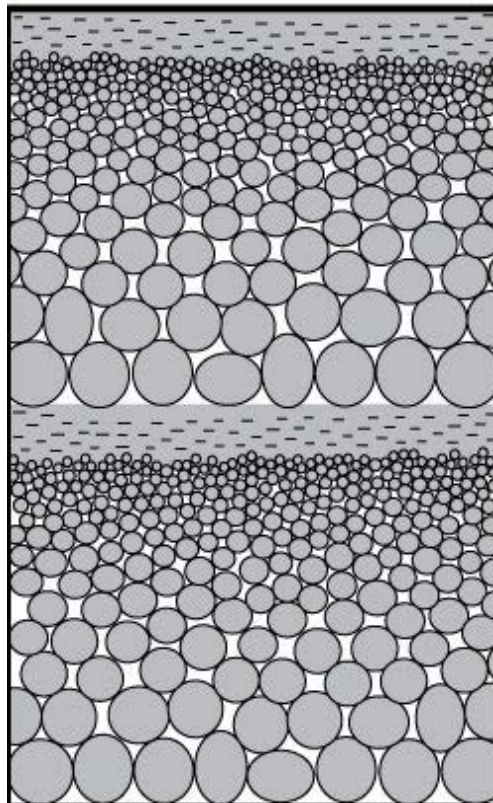


Fig.6.9. Size-graded bedding reflects deposition from a so-called turbidity current. Turbidity currents represent the down-slope flow of a slurry of loose sedimentary material on the sea floor.

Wet sediments that are exposed to the air can dry out and form a characteristic series of polygonal cracks – *mud cracks*. Organisms can leave traces of their existence on sediments. The most obvious is their remains after death when they become fossils. Some organisms that live on the sea floor leave marks on the sediment surface (or just below the surface) in connection with burrowing activity or tracks. These traces of organic activity without the creature itself being preserved are known as *trace fossils*. Different kinds of fossil material can also reflect water depth (e.g. crabs) or a terrestrial environment (e.g. plants remains).

6.5 Where do sediments form?

Sediments can form, for example, on a beach, in a river, in the sea, in connection with an ice age etc. There are two main environments: on land or *terrestrial* and under the sea or *marine*.

6.5.1 Terrestrial environments

These include *glacial*, *mountain stream*, *mountain front*, *desert (or sand dune)*, *lake* and *river* environments.

6.5.1.1 Glacial environment

A glacier moves material of all sizes, fragments that fall onto its surface or are “plucked” from its underlay. The material that is deposited when the glacier melts is called *till*. A till comprises unsorted and un-layered material with a wide range of grain sizes – from large angular blocks to clay. The fragments show a very limited degree of alteration.

6.5.1.2 Mountain stream environment

Streams on steep slopes and with (periodically) large volumes of water can move large blocks, while small particles are transported away. The resulting sediment is a *breccia* (angular fragments) or *conglomerate* (rounded fragments).

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6.5.1.3 Mountain front environment

When a mountain stream reaches the edge of a mountainous area, the material deposited forms a characteristic shape resembling that of a fan with the handle pointing up the stream. The deposits forming an *alluvial fan* consist dominantly of coarse clasts; the resulting sediment is a conglomerate. Alluvial fans can be formed in climates where Fe^{2+} is oxidized to Fe^{3+} so that the resulting sediment has a reddish colour because iron is present as the red-coloured mineral hematite (Fe_2O_3). Red, oxidized clastic sediments are known as *red beds*. Red beds can, however, also be formed in other types of environment, such as deserts.

6.5.1.4 Desert environment

Deserts are characterized by a lack of vegetation and by winds. Sand, silt and dust particles can be transported, depending on the wind strength. Sand dunes are commonly formed. Deposits are well-sorted and sand grains are well-rounded. The deposits commonly show *cross bedding* – beds that are inclined relative to a thicker stratum in which they occur (Fig.6.10). Cross bedding forms when sand/silt particles are deposited on a slope. This happens when sand dunes are formed. Particles are blown to the top of a dune and tumble down the slope at the front of the dune to be deposited on a slope. The slope dips (by up to about 35°) in the direction in which the dune is moving i.e. in the direction in which the wind blows. Changes in the wind direction can result in beds dipping in different direction overlying each other. Sand grains transported by the wind can polish larger rock fragments lying on the surface and give very smooth, polished surfaces known as *desert varnish*. Rock fragments on the surface sometimes develop characteristic shapes reflecting different wind directions; these polished rocks are called *dreikanter*.

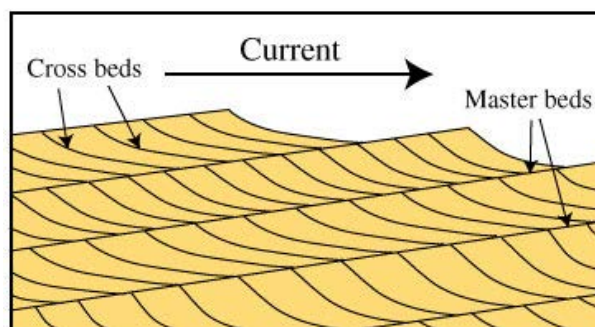


Fig.6.10. Cross beds form on the leeward side of sand dunes or on the downstream side of flowing water.
The sedimentary sequence consists of successive layers (master beds) of cross-bedded strata.

Cross bedding is a common feature in desert environments (in connection with the formation of sand dunes) but may form in response to other types of current activity (stream currents or ocean waves).

6.5.1.5 Lacustrine environment

Sediments deposited in lakes (lacustrine deposits) are relatively fine grained and laminated (i.e. have many thin layers). Any fossils present will be of fresh water varieties. Evidence of plant life may be preserved in material washed into the lake. Lakes in areas that are cold in the winter may freeze over. Deposition of relatively coarse clastic material takes place when the lake thaws, but in the winter, when no new material enters the lake, fine grained material (clay) may settle out. These types of layers are called *varves*. One varve (a thin, graded layer) reflects deposition in one year.

6.5.1.6 Fluvial environment

Many sediments are deposited from rivers. These are called *fluvial deposits*. As the rate of flow of a river decreases with distance from its source the size of the particles that can be transported becomes smaller. Small particles can be deposited on a *flood plain* when a river overflows its banks. These fine grained flood plain deposits may develop mud cracks when they dry out. Material deposited in the course of the river itself may show ripple marks and small cross bedding features. As is the case with other terrestrial sediments, fluvial deposits may be oxidized and develop a reddish colour.

During their transport, the majority of minerals (with the important exception of quartz) in the fragments become altered to clay minerals. Fluvial sediments therefore consist dominantly of *sandstone*, *siltstone* and *shale*.

6.5.2 Marine environments

Marine environments include *delta*, *coastal areas*, *shallow sea* and *deep sea* environments.

6.5.2.1 Delta environment

Deltas form where rivers meet the sea. Large proportions of the huge volumes of clastic material transported by rivers are deposited in deltas. The internal structure of a delta is divided into three portions. Going from the river towards the sea these are *topsets*, *foresets* and *bottomsets* (Fig.6.11).

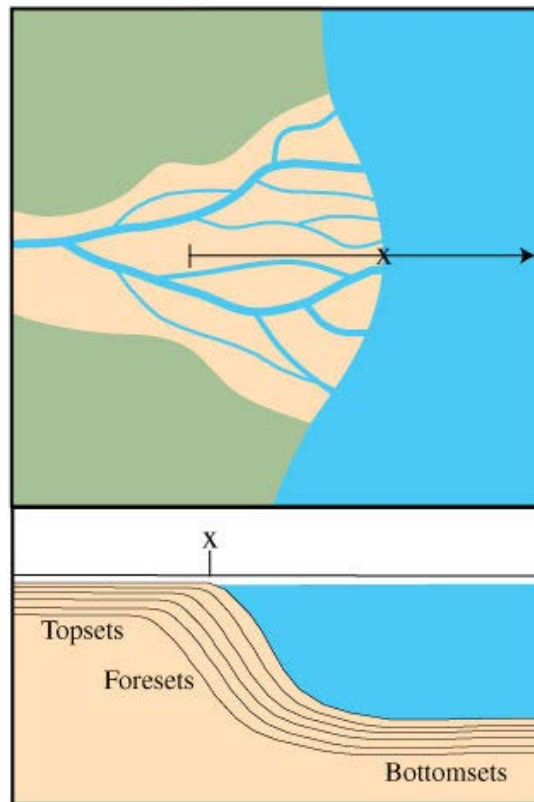


Fig.6.11. Simplified map view of a delta and a cross section showing its structure. Delta deposits are divided into *topsets*, *foresets* and *bottomsets*.

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Topsets comprise more or less horizontal layers of mud and clay. The remains of swamps that develop on the topset leave plant remains that may be altered to coal. As they advance, topsets bury foresets. *Foreset* deposits accumulate on a slope and consist of silt and mud. During development of the delta they become overlain by the topsets and themselves overlie the bottomsets. *Bottomsets* comprise horizontal layers of mud and clay deposited in deeper water; they become buried by the foreset deposits. Changes in sea level over time can result in delta deposits being extremely complex.

6.5.2.2 Coastal environment

Currents transport sand and silt along the coastline. If the sea level rises, these can be deposited. We observe these deposits on modern beaches with well-rounded, well-sorted sand/silt with ripple marks.

6.5.2.3 Shallow marine environment

Finer grained material can become washed out to sea where it is deposited on the sea floor in water depths up to about 100m. These *shallow marine clastic deposits* consist of well-sorted, well-rounded silt. Marine organisms are active (molluscs, worms etc.) and leave their shells as well as a variety of tracks (trace fossils). These submarine deposits do not contain mud cracks, ripple marks or the fossil remains of terrestrial organisms.

In warm coastal regions where there is a limited supply of clastic material, deposition is dominated by shell fragments to form *shallow marine carbonate deposits*. These occur today on tropical islands, far from the mainland which could supply clastic material. Much of the carbonate material may be supplied by *coral reefs* (Fig.6.12). The spherical carbonate structures called *oolites* are a common feature of shallow marine carbonate deposits.

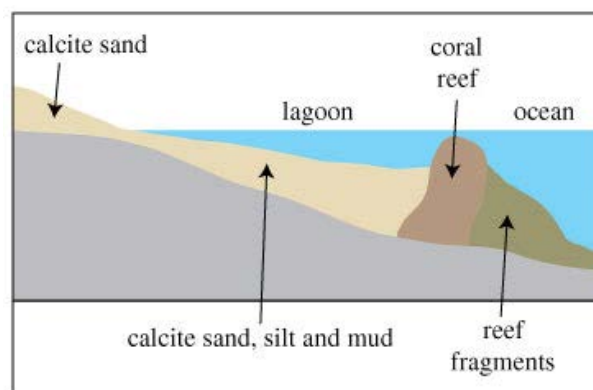


Fig.6.12. Various carbonate environments associated with a coral reef. Deposits include carbonate sand, silt and mud and material that develops into coraliferous limestone.

6.5.2.4 Deep ocean environment

In the region between the coast and the deep sea, periodic *turbidity currents* deposit graded, fine grained sediments. Submarine fans may develop at the “deep” ends of submarine valleys. The graded sediments that form as a result of deposition from turbidity currents are referred to as *greywacke*. In the deep ocean, the only sources of sedimentary material are fine-grained clay and mud and the shells of plankton. Accumulation takes place very slowly to form thin layers of calcareous shales and/or chert (from silica-rich plankton shells). The deep sea floor is locally covered by numerous “manganese nodules” which are small metal-rich spherical bodies that represent potential ore-reserves.

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7 Metamorphic rocks

7.1 Introduction

All kinds of rocks can be subjected to changes in temperature and/or pressure. They can be heated and/or deformed so that their appearance changes drastically. New rocks are formed from old by the process of *metamorphism* (*meta* = *change*; *morph* = *form*). Whereas sedimentary and volcanic processes take place at the surface of the Earth, metamorphic processes take place outside our range of visibility, below the surface. It is therefore often challenging to reconstruct metamorphic processes in detail.

Metamorphic processes take place in the *solid state*. There is no melt involved. If temperatures are so high as to melt a rock, the products on cooling are igneous. Melting marks the upper limit of metamorphism. Metamorphic rocks form a vital part of the *rock cycle* (Fig. 1.1) on the way, for example, from sedimentary rocks to igneous rocks.

Metamorphic rocks are commonly very different from the original rock from which they formed (the *protolith*). For example, a clay-rich sedimentary rock (a shale) can be metamorphosed to a coarse grained, banded rock containing shiny micas, quartz and grains of garnet (a garnet mica schist). Black, vesicular, fine-grained basaltic lava can be metamorphosed to a heavy green and red rock consisting of red garnet and green pyroxene (an eclogite).

The Earth's mantle consists largely of ultramafic rock (peridotite) which is at a high temperature and is subjected to flow (large scale convection currents). The peridotite is in a plastic state (but below its melting temperature) and is deformed in the solid state – it is a metamorphic rock. We will, however, concentrate on crustal rocks. Metamorphism usually involves the formation of new minerals (a *metamorphic mineral assemblage*) and the development of a new structure (commonly a banding or *metamorphic foliation*). The term foliation comes from the Latin word folium, which means leaf. A simple comparison between igneous, sedimentary and metamorphic rocks is shown in Table 7.1.

	Igneous rocks	Sedimentary rocks	Metamorphic rocks
Starting material	Melt	Rock Aqueous solutions Biological material	Rock
Formation processes	Solidification Crystallization from melt	Weathering Transport Deposition Lithification	Alteration in solid state
Temperature and pressure conditions	650–1250°C Surface to ca. 10 kb	<150°C Surface conditions Very low pressure	ca.150–800°C Usually up to ca.10 kb

Table 7.1. Comparisons between the three main rock types

7.2 Metamorphism – causes and effects

Rocks become metamorphosed because of the influence of heat, hot water, pressure, or differential stress.

7.2.1 Heat

Pure *limestone* consists of calcite (CaCO_3). If limestone is heated, calcite grain boundaries migrate and the grain size increases. Any fossils that were present become indistinct and finally all trace of them disappears. The limestone has *recrystallized* in the solid state. The rock has not melted. It has not changed its chemical composition. A textural change has taken place and the rock has become a *marble* – the product formed by the metamorphism of limestone. Most rocks, however, form *new minerals* during metamorphism. Clay minerals, for example, are not stable at elevated temperatures (partly because they contain so much water in their structure) and break down to form, for example, mica. Newly formed large mineral grains in metamorphic rocks are called *porphyroblasts*; these are the metamorphic equivalents of phenocrysts in igneous rocks.

7.2.2 Pressure

Pressure increases with depth below the surface of the Earth simply because of the weight of the overlying rocks. This is called lithostatic pressure. The density of the rocks involved plays a role, but on average (in the continental crust):

lithostatic pressure increases by 1 kilobar per 3.3 kilometers depth

Pressure units used in textbooks vary somewhat. Here we use kilobars which can be converted to other units as shown in Table 7.2.

1 atmosphere	ca. 1 bar
1 kilobar (Kb)	1000 bar
1 kilobar (Kb)	100 megapascal (Mpa)
1 kilobar (Kb)	0.1 gigapascal (Gpa)

Table 7.2. Conversion of pressure units

The effect of pressure alone can produce some metamorphic reactions, but both pressure and temperature increase with depth. The geothermal gradient is the increase in temperature with depth. This varies considerably, depending on the plate tectonic setting. *Average values for the geothermal gradient are in the range 15°C–50°C per km.*

7.2.3 Water

Hot water that passes through rocks at depth below the surface (*hydrothermal solutions*) dissolves some materials and deposits others. The composition of the hydrothermal solution depends on the type of rocks involved. In limestones and marbles they will be carbonate-rich; in granites they will be silica-rich. Hydrothermal solutions that pass through relatively cold rocks that can fracture commonly deposit minerals in joints to form *veins*. These veins are commonly filled with quartz (or calcite in carbonate-rich rocks). Quartz is such a common vein-filling mineral since SiO_2 is a product of many metamorphic reactions and readily enters hydrothermal solutions. Water is also a product of many metamorphic reactions. This water contributes to the hydrothermal solutions.

7.2.4 Differential stress

Material subjected to different pressures in different directions becomes deformed. There are two types of differential stress: *normal stress* and *shear stress*.

Normal stress operates perpendicular to a surface. The most usual situation involves compression (where rocks are squashed), but tension can be involved (where rocks are stretched). Shear stress moves one part of a material sideways relative to another.



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Rocks near the surface of the Earth react in a brittle way to deformation – they break (to form faults or fractures). Rocks at depth that are undergoing metamorphism, however, are not brittle: they are ductile and react plastically. Deformation under elevated pressure and temperature conditions commonly results in the development of a preferred mineral orientation or *foliation*.

Metamorphic foliation forms by several processes. Grains can deform plastically and become “squashed”. Part of a grain can dissolve where the pressure is greatest, and be re-deposited where the pressure is least by a process called pressure solution. Grains may rotate, and new metamorphic minerals can form. These can all contribute to the development of a metamorphic foliation.

7.3 Types of metamorphic rocks

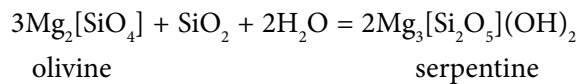
There are two main groups of metamorphic rocks, those that are *foliated* and those that are not. Most *non-foliated* rocks have been subjected to the effects of temperature (under constant pressure), whereas foliated rocks have been deformed during heating.

7.3.1 Non-foliated metamorphic rocks

Non-foliated metamorphic rocks form when metamorphism takes place under constant pressure. An intrusion of granitic magma will heat its so-called “country rock envelope”. The temperature of the country rocks depends on the distance from the intrusion. The size of the intrusion and the composition of the magma are also very important (e.g. granitic (rhyolitic) magma at 700°C will not heat the country rocks as much as gabbroic (basaltic) magma at 1200°C). The country rocks furthest away from the intrusion will simply recrystallize whereas new metamorphic minerals will form as the heat source is approached. Non-foliated metamorphic rocks that have been “baked” because of their proximity to an intrusion are called *hornfels*. Hornfels is a relatively fine-grained, massive rock. Any kind of protolith can be subjected to contact metamorphism to produce, for example, metapelitic (clay-rich sedimentary rock prior to metamorphism) hornfels, metabasaltic hornfels etc. Metamorphism resulting from the heating of rocks by an intrusion is called *contact metamorphism*.

Two important non-foliated rock types are *marble* (metamorphosed limestone) and *quartzite* (metamorphosed sandstone). Both may, of course, occur as foliated varieties if they are deformed during metamorphism.

The metamorphism of igneous rocks shows some important differences from that of sediments. This is because many sediments contain H₂O-bearing minerals (e.g. clay), and water is present in the pore spaces. Metamorphic reactions in sedimentary rocks involve dehydration. Gabbro, however, consists dominantly of plagioclase and pyroxene, neither of which contains water. Gabbros crystallize at 1000–1200°C and the minerals are stable at all temperatures below this, as long as the system is dry. The metamorphism of a gabbro requires the presence of water. The same applies for most igneous rocks. Here we will consider the metamorphism of *dunite* (an ultramafic rock consisting of >90% olivine; a kind of peridotite). Most dunites consist of Mg-rich olivine with composition close to forsterite Mg₂[SiO₄]. In the presence of water, forsterite breaks down at >200°C to form serpentine.



A rock consisting largely of serpentine is a *serpentinite*. Serpentinites can also occur as foliated rocks. Another hydrous, Mg-rich silicate mineral that can be formed from the metamorphism of olivine-rich rocks is talc ($\text{Mg}_3[\text{Si}_4\text{O}_{10}](\text{OH})_2$). Talc-rich rocks are known as soapstone.

7.3.2 Foliated metamorphic rocks

The progressive metamorphism and deformation of clay-rich sedimentary rocks (shales) produces a characteristic series of rock types. In metamorphic petrology, clay-rich sedimentary rocks are called *pelites* and the metamorphic products are *metapelites*. The key feature of pelites in terms of their chemical composition is that they are *aluminium-rich*. During the metamorphism of pelitic rocks, a series of minerals with layered structures (phyllosilicates) are formed. When these phyllosilicates become orientated in a parallel fashion they impart a foliated structure to the rock.

In shale the clay minerals (phyllosilicates) are orientated in the bedding plane – they are responsible for the characteristic cleavage of shales. When a shale is pressed from two sides it becomes folded and the (new) phyllosilicates become orientated perpendicular to the direction of pressure to form a distinct parting called slaty cleavage (Fig.7.1).

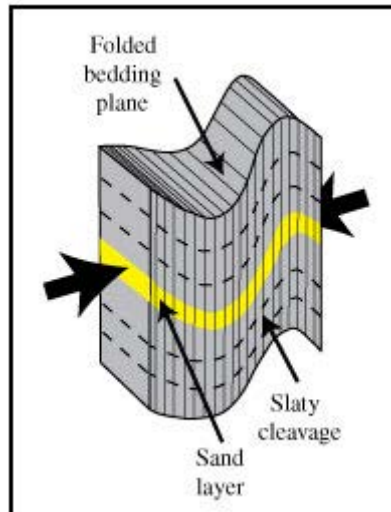


Fig.7.1. Development of slate.

Clay minerals in shale are orientated in the bedding plane. After deformation (folding) and metamorphism new platy minerals define a slaty cleavage perpendicular to the direction of pressure. An original sandy layer will recrystallize and become quartzite.

The first foliated metamorphic rock to form during metamorphism of pelitic rocks is a *slate*. Some of the clay minerals become unstable under the elevated temperature and break down to form new phyllosilicates. These develop with their layered-structure perpendicular to the direction of pressure (Fig.7.1). Slates used to be widely utilized as roofing material. It is sometimes possible to see the trace of the original bedding plane in a slate where, for example, sand-rich layers can be preserved (Fig.7.1). Slaty cleavage forms at the same time as folds during the deformation and metamorphism of pelitic rocks.

With higher temperature the white mica muscovite begins to form from the breakdown of clay minerals. The rock becomes slightly coarser-grained and the surface develops a shiny appearance (a silky lustre). The slate has become a *phyllite*. The surface of a phyllite is commonly crenulated. As the temperature increases the grain size increases further, the muscovite flakes become larger, and dark mica (biotite) may form. The phyllite has become a mica *schist*. In addition to quartz, mica schists commonly contain other metamorphic minerals (section 7.4.1). The term “schistosity” is sometimes used for the metamorphic foliation in schists.

As the conditions of metamorphism increase the rock becomes even coarser-grained and a banding (layering) develops. The banding commonly consists of light and dark layers in which the light layers are dominated by quartz and feldspar whereas the dark layers contain minerals such as biotite, hornblende, garnet, pyroxene etc. The schist has become a *gneiss*.



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With increasing grade of metamorphism a metapelitic gneiss will begin to melt. The first melt to be formed will have the composition of a granite. This melt may migrate out of the rock and accumulate to form an intrusion. In other cases it will remain more or less where it formed and on cooling will crystallize to form patches of granite. The rock will now consist of two components – a granitic part and a “restite” part (the part that did not melt). Such a mixed rock is called a *migmatite*. We have now reached the upper limit of metamorphism since the granitic part of the migmatite is an igneous rock.

All these foliated rock types form as a result of the metamorphism and accompanying deformation of pelitic protoliths (shale). Other protoliths give different metamorphic rock types. For example, *granite* can be metamorphosed to *granitic gneiss*. A characteristic type of gneiss is formed by the metamorphism and deformation of porphyritic granites (with large alkali feldspar phenocrysts). While the matrix develops a foliated structure the phenocrysts react as resistant blocks and are preserved as “eye-shaped” particles. This special kind of granitic gneiss is called *augen gneiss*.

The metamorphism and deformation of mafic rocks (*basalt*, andesite, gabbro, diorite) gives four important rock types. The first is a *greenschist*. The greenish colour of this foliated metamorphic rock is due to the presence of chlorite and epidote (\pm actinolite). White albite is also present. As the metamorphic grade increases these minerals break down as hornblende (black) and plagioclase (white) are formed. A metamorphic rock consisting largely of these two minerals is an *amphibolite*. Other minerals that may be present in smaller quantities include garnet.

Under conditions of relatively high pressure and low temperature (e.g. 8 kb and 200°C) basaltic rocks are metamorphosed to *blueschists*. The bluish colour comes from the mineral glaucophane (an amphibole). Under very high pressure conditions plagioclase feldspar is unstable. The mineralogy of the mafic rock changes from plagioclase and augite (a clinopyroxene containing Ca, Mg and Fe) to garnet and omphacite (a clinopyroxene containing Na and Al in addition to Ca, Mg and Fe). Since garnet has a reddish colour and omphacite is green, the resulting high pressure metamorphic rock (*eclogite*) is red and green. Both these minerals have high densities (omphacite $D = 3.3$; garnet $D = >3.6$) so that eclogite has a density of ca. 3.5. Since eclogite contains >90% mafic minerals it is an ultramafic rock. But its chemical composition will be the same as its protolith (basalt, a basic rock with 45-52% SiO_2). So eclogite is unusual in being an ultramafic basic rock (most ultramafic rocks are ultrabasic i.e. with <45% SiO_2).

The metamorphism of limestone produces *marble*. The metamorphism of pure limestone (consisting of close to 100% calcite) does not produce any new minerals; the rock simply recrystallizes. Many limestones, however, are impure. The presence of quartz grains means that silicate minerals can form during metamorphism. And if dolomite ($\text{CaMg}(\text{CO}_3)_2$) is also present it is possible to form a range of metamorphic minerals (such as tremolite (an amphibole), diopside (pyroxene) and forsterite (olivine)). Foliated marbles are often very attractive rocks that have widely been used for sculptures etc.

7.3.3 Types of protolith

All kinds of sedimentary or igneous rock can become metamorphosed. Rocks can also be metamorphosed more than once. This gives an extremely wide range of potential protoliths and products. In order to simplify matters we consider a very restricted variety of important types of protolith (Table 7.3).

COMPOSITIONAL TYPE	PROTOLITH
Pelite	Shale
Basic	Mafic igneous rock (<i>basalt</i> , andesite, gabbro, diorite)
Ultrabasic	Ultramafic (peridotite etc.)
Calc-silicate	Limestone, dolomite (often quartz-bearing). Marl – a mixture of clay and limestone
Quartzo-feldspathic	Granite, rhyolite

Table 7.3. The main types of metamorphic protoliths

7.4 Grades of metamorphism

Metamorphism takes place in response to changes in temperature and pressure (Fig.7.2). Sedimentary processes (lithification) cease and metamorphism begins at about 150°C (depending on the pressure). The upper limit of metamorphism is when rocks begin to melt. This depends on the composition of the rock involved (pure quartz sandstone will not begin to melt until ca. 1700°C; a metapelite (metamorphosed shale) will begin to melt at ca. 650°C if the pressure is high enough), the presence of H₂O, and the pressure. The PT space between these two limits define the conditions under which metamorphism normally takes place.

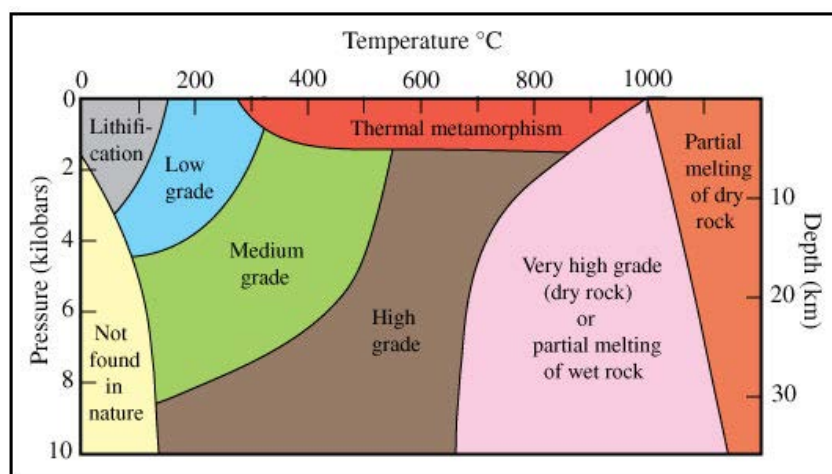


Fig.7.2. Pressure (P) – Temperature (T) conditions of metamorphism.

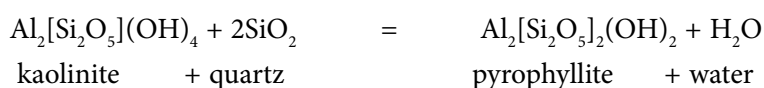
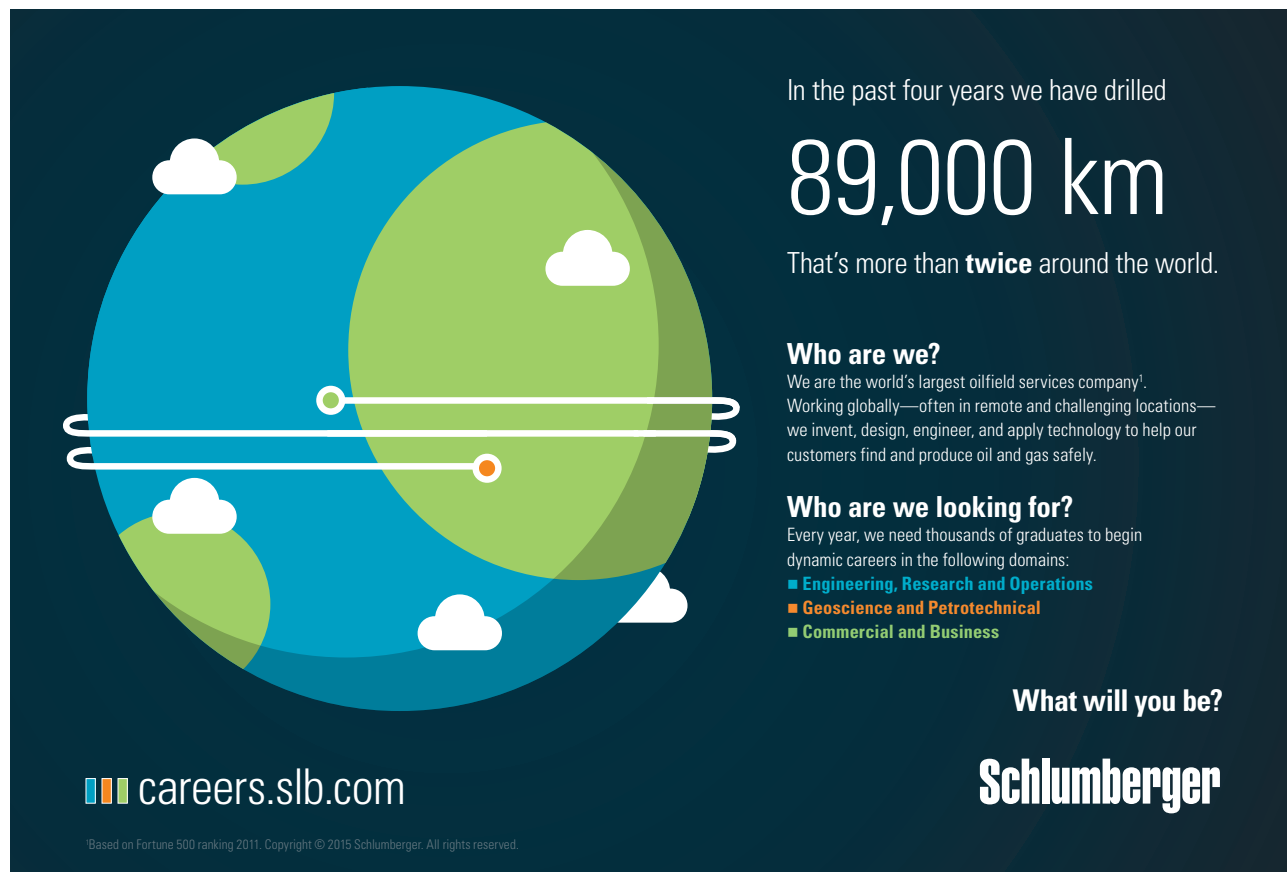
The lower limit of metamorphism is where sedimentary processes (lithification) end. The upper limit is where rocks begin to melt. The field of metamorphism is divided into low-, medium- and high grade. A field of very high grade occurs where rocks of appropriate composition begin to melt. A field of thermal metamorphism is recognized at low pressure.

The pressure – temperature space where metamorphism takes place is divided into regions of *low*-, *medium*- and *high grade*. A field of *very high grade* occurs where rocks of appropriate composition begin to melt. Metamorphism at low pressures is referred to as *thermal metamorphism* and is due to the heating of rocks by magma.

As metamorphism progresses from low to high grade, rocks generally become coarser grained and “drier” – water-bearing minerals break down with increasing metamorphic grade.

7.4.1 The progressive metamorphism of shale

Shale consists dominantly of clay minerals and minute quartz grains, together with water in the pore spaces and in the structure of the clay minerals. Clay minerals become unstable with increasing temperature. For example, the mineral kaolinite (a common component in shale) reacts with quartz at 200°C–250°C (depending on the pressure) and forms a new mineral (pyrophyllite which is also a phyllosilicate) with the liberation of water (Fig.4.3).

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This reaction takes place during the transition of shale into slate. Another new mineral to form in slates is chlorite (Fig.7.3). This is also a phyllosilicate. Flakes of both pyrophyllite and chlorite will lie in the cleavage plane of the slate. The formation of both pyrophyllite and chlorite takes place under low grade metamorphic conditions.

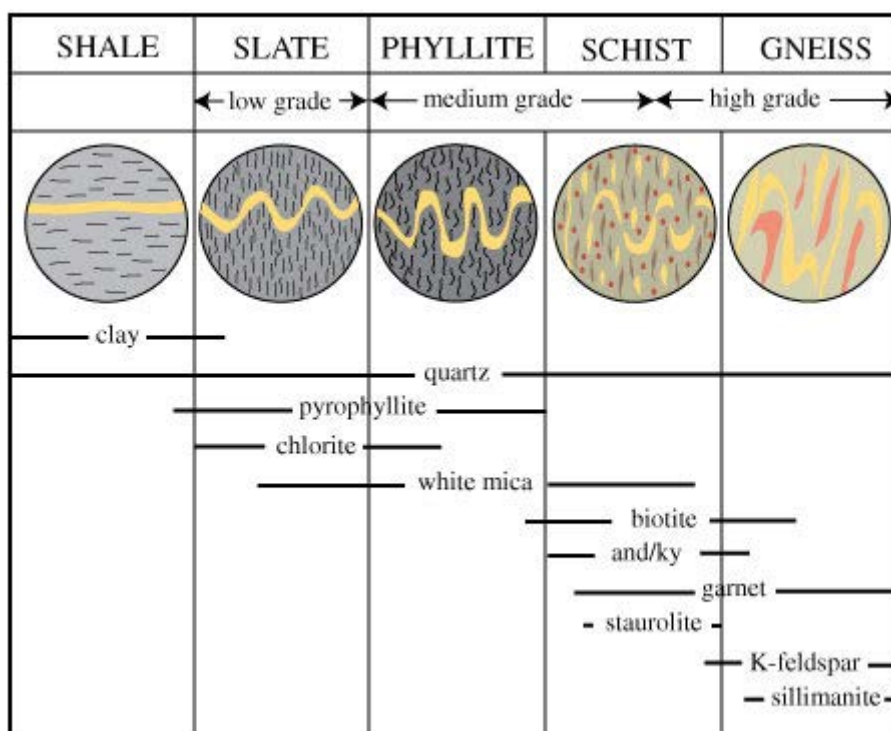
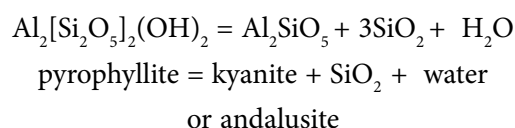


Fig.7.3. Progressive regional metamorphism of clay-rich sediments (shales).

The metamorphism of shale leads first to slate, then phyllite, schist and gneiss. This structural evolution is accompanied by a series of mineralogical changes. New minerals form in the sequence pyrophyllite, chlorite, white mica, biotite, andalusite/kyanite, garnet, staurolite, K-feldspar, sillimanite. The shale is illustrated with a thin layer of sandstone. This becomes quartzite in phyllite and unrecognizable in schist and gneiss.

The next new mineral to form is white mica (Fig.7.3). Small flakes of muscovite will be orientated parallel with the other phyllosilicates. When there is sufficient white mica to provide a silky lustre, the rock becomes a phyllite. As the metamorphic grade increases, white mica flakes become larger and pyrophyllite becomes unstable (at 300°C–420°C depending on the pressure) and breaks down to form an Al_2SiO_5 polymorph, SiO_2 and water:



If the pressure is below ca. 2Kb (equivalent to a depth of about 7km) the Al_2SiO_5 polymorph will be andalusite. At >2Kb it will be kyanite. The polymorphic transition between andalusite and kyanite is more dependent on pressure than temperature (Fig.4.3). Note that both the reactions above release water. This water contributes to the hydrothermal solution in the rock. The SiO_2 released by the breakdown of pyrophyllite can enter the hydrothermal solution and be deposited elsewhere as quartz veins.

Dark mica (biotite) is the next new mineral to form. The grain size of the minerals continues to increase and the rock becomes a *schist*. A variety of aluminium-bearing minerals can occur in schists. Which one(s) develop(s) depends on the precise composition of the original shale and the pressure – temperature conditions. Possible minerals include garnet, staurolite and cordierite.

As the metamorphic grade continues to increase, K-feldspar forms from the breakdown of muscovite. Sillimanite is also formed by this reaction, and water is released.



Under these high grade conditions the rock develops an irregular banded structure which is a characteristic feature of *gneisses*. Note that quartz was a constituent of the shale protolith and takes part in some reactions and is a product of others. It is the only important mineral that survives the entire metamorphic process.

As the metamorphic grade increases even further the metapelitic gneiss will begin to melt. The first melt to form will have a granitic composition. The rock will consist of two parts – a “restite” (the part that did not melt) and a granite. Mixed rocks of this type are called *migmatites*.

The progression from shale through slate, phyllite, schist and gneiss reflects metamorphism under increasing temperature and pressure conditions – so-called *prograde* metamorphism. A typical PT trend is illustrated in Fig.7.4. Many other PT paths are possible. In nature the PT path will usually be curved (Fig.7.5). The prograde metamorphism may not reach the maximum conditions but stop at any stage. Slates are a product of low grade metamorphism. Phyllites reflect medium grade metamorphic conditions. Schists overlap the medium – high grade boundary. Gneisses are high grade metamorphic rocks. Migmatites have been subjected to partial melting and are very high grade metamorphic rocks.

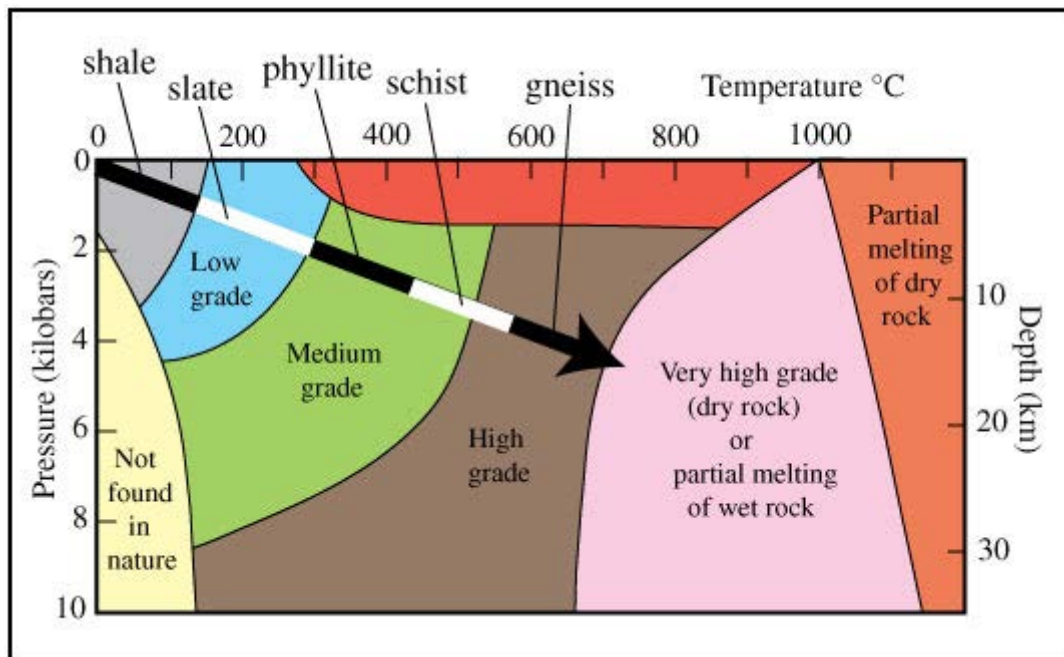


Fig.7.4. The progressive metamorphism of shale in a P-T diagram.

The geothermal gradient (the rate at which temperature increases with depth) here is about 45°C/km. Many other gradients are possible.

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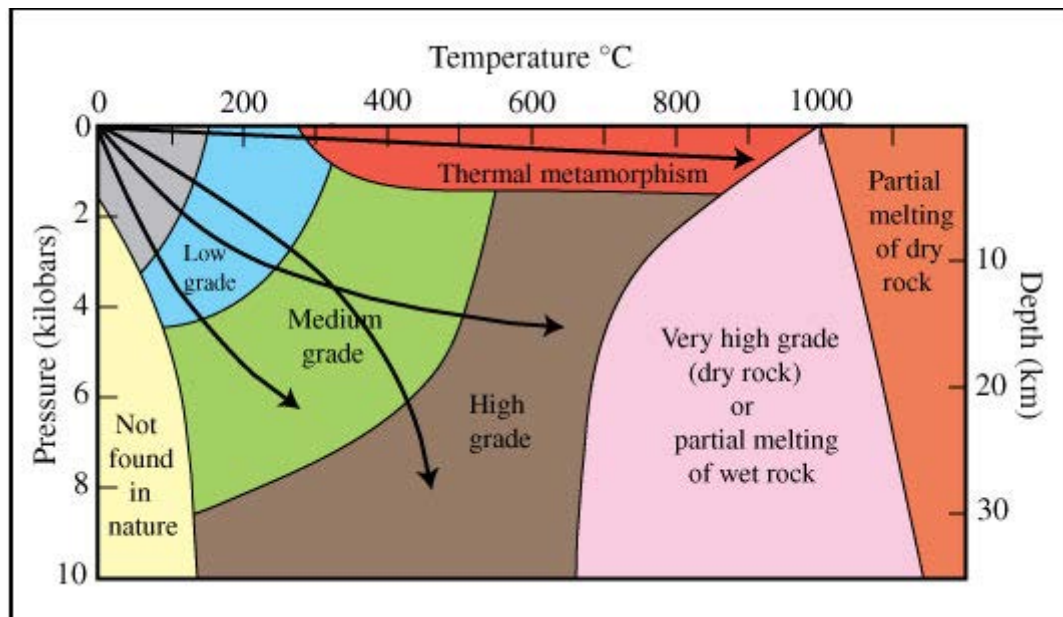


Fig.7.5. Possible paths of metamorphism in a P-T diagram.

Prograde metamorphism usually follows a curved PT path and may stop at any stage.

7.4.2 Index minerals and metamorphic zones

Some of the minerals that form in metamorphic rocks (e.g. garnet, Al_2SiO_5 polymorphs) do not occur (or are relatively rare) in sedimentary and/or igneous rocks. Certain minerals are a good indicator of metamorphic grade since they only appear when specific PT conditions are reached. These are called *index minerals*. When studying an area of metamorphic rocks the presence of index minerals is noted and a line is drawn on the map indicating the first occurrence of a particular mineral. A line separating two index minerals is called an *isograd* (Fig.7.6). An isograd indicates constant metamorphic conditions. For index minerals like the Al_2SiO_5 polymorphs (andalusite, kyanite or sillimanite) to occur requires that the PT conditions are appropriate (Fig.4.3) but it also requires that we are dealing with metapelitic rocks. Al_2SiO_5 polymorphs cannot form in non-pelitic rocks (such as metamorphosed limestones or sandstones which often occur together with metapelites).

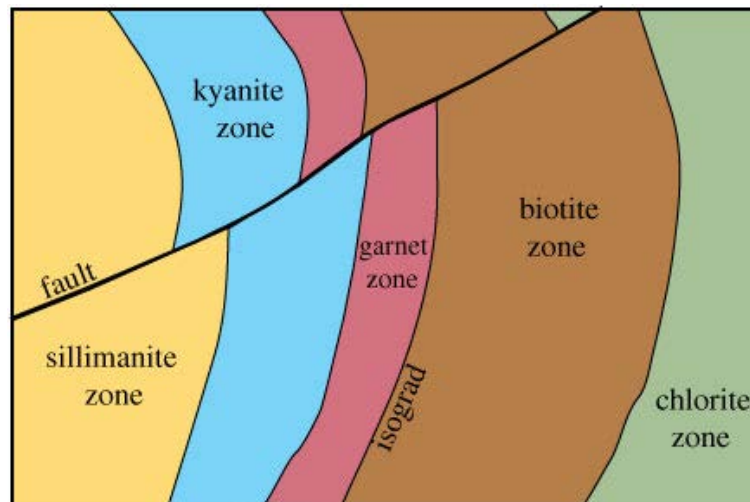


Fig.7.6. Metamorphic zones, index minerals and isograds.

A metamorphic zone is characterized by the presence of an index mineral (e.g. kyanite). Lines on a map separating metamorphic zones are isograds. The metamorphic zones here are based on metapelitic rocks. Note that the metamorphic zones are offset across a fault that is younger than the metamorphism.

As mentioned above, metamorphism that takes place under increasing temperature and pressure conditions is called prograde metamorphism. Water leaves the rock during the prograde metamorphism of shale through slate, phyllite and schist to gneiss. A gneiss may be formed at, for example 650°C and 5 Kb. But when we find it in, for example, Norway or Scotland it is at $10 \pm 15^\circ\text{C}$ and 1 bar. Yet the minerals that formed at the maximum metamorphic conditions (*peak metamorphism*) are still present in the rock. The reverse reactions did not take place, even though the rock gradually cooled through high-, medium- and low-grade metamorphic conditions. This is because *retrograde metamorphism* requires water (Fig.7.7). Under certain conditions hot water is added to metamorphic rocks while they cool so that retrograde metamorphism takes place. But this is far less common than prograde metamorphism. It is usual for the peak metamorphic conditions to be recorded by the minerals present in the rock.

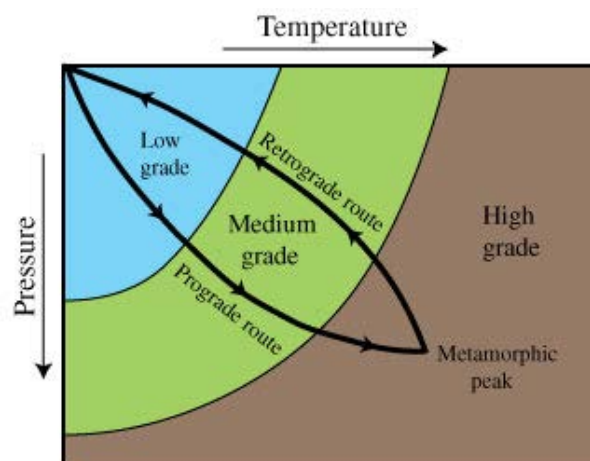


Fig.7.7. Prograde metamorphism takes place during increasing temperature and pressure. The maximum pressure and temperature conditions reached are called “peak metamorphism”. Retrograde metamorphism can take place if water is supplied to the rock. This is relatively uncommon.

7.4.3 Metamorphic facies

Which minerals occur in a metamorphic rock depend on the grade of metamorphism and the nature of the protolith. A pelite metamorphosed at 5Kb and 550°C may contain quartz, muscovite, biotite, almandine garnet and sillimanite (it will be a garnet sillimanite mica schist). A quartz-bearing dolomitic limestone metamorphosed under the same T and P conditions will consist of calcite, diopside (a clinopyroxene) and grossular garnet (it will be a foliated marble). A (wet) basalt metamorphosed at 5Kb and 550°C will consist largely of plagioclase feldspar and hornblende (it will be a foliated amphibolite). The mineralogy of these three metamorphic rocks is very different – but they were all formed under the same temperature and pressure conditions. These three rocks all belong to the same *metamorphic facies* (the *amphibolite facies*, as we will see later). A metamorphic facies includes rocks that are metamorphosed under the same conditions. There are seven main metamorphic facies (Fig.7.8).

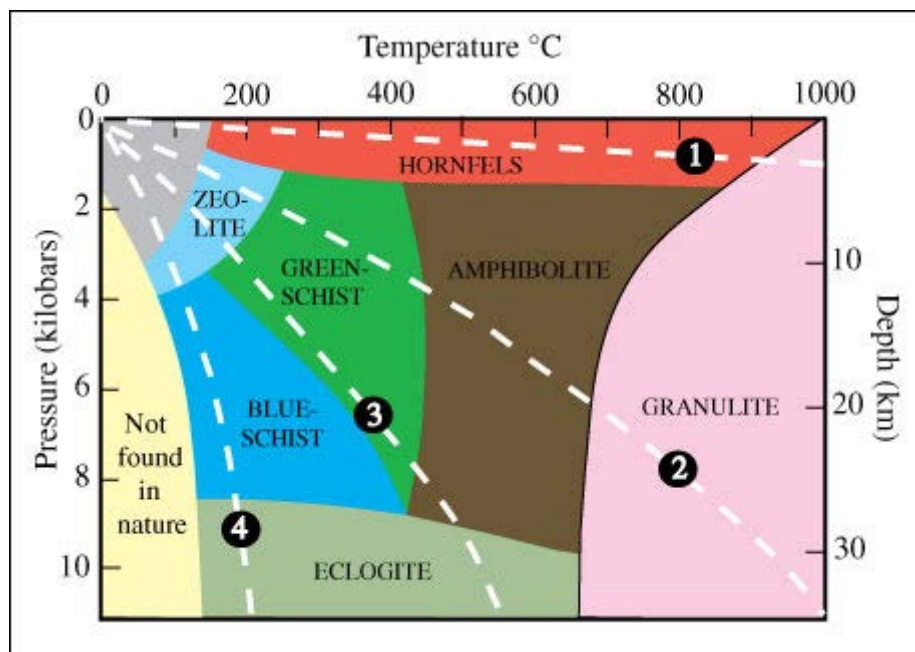


Fig.7.8. Location in P-T space of the seven main metamorphic facies.

Note that the different facies (coloured fields) develop under different geothermal gradients. The dashed lines are different geothermal gradients (numbered 1 to 4).

The ZEOLITE facies is named after the zeolite minerals that form in rocks of appropriate composition at a very low grade of metamorphism.

The HORNFELS facies forms as a result of thermal metamorphism (i.e. heating at low pressure). The rocks are not foliated and are usually fine grained and brittle. They are commonly “spotted” because of the growth of new porphyroblasts (commonly andalusite and/or cordierite in metapelites).

The GREENSCHIST facies develops under low to medium grades of metamorphism. The name comes from the characteristic appearance of metamorphosed basaltic rocks that become “green schists” under low- to medium-grades of metamorphism. The green colour is mainly due to the presence of chlorite and epidote.

The AMPHIBOLITE facies develops under medium- to high-grades of metamorphism. It is also named after metamorphosed basaltic rocks that develop plagioclase feldspar and hornblende (an amphibole) under amphibolite facies conditions.

The GRANULITE facies forms under high grades of metamorphism; the rocks are commonly granular in appearance. Many rock types will begin to melt (in the presence of water) under granulite facies conditions. Many granulite facies rocks were formed under dry conditions so that melting could not take place. A characteristic mineral of many rocks in granulite facies is orthopyroxene.

The BLUESCHIST facies forms under relatively high pressure and low temperature conditions. It is also named after the characteristic appearance of metamorphosed basaltic rocks which become blue(ish) schists because of the presence of glaucophane, a blue amphibole.

The ECLOGITE facies develops under very high pressure conditions. Eclogites consist largely of reddish garnet (pyrope-rich) and greenish clinopyroxene (omphacite) and are formed when plagioclase in metabasaltic rocks becomes unstable because of the high pressure.

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7.4.4 Geothermal gradients

It is evident from Fig.7.8 that different metamorphic facies form under different PT conditions. A key factor is the rate at which the temperature increases relative to the increase in pressure – the geothermal gradient. The hornfels facies forms under very high geothermal gradients, usually more than $100^{\circ}\text{C}/\text{km}$. The blueschist facies, on the other hand, develops under low geothermal gradients, usually below $15^{\circ}\text{C}/\text{km}$. Conditions under which the greenschist, amphibolite and granulite facies develop cover a wide range of geothermal gradients between these two extremes, and usually in the range $50\text{--}20^{\circ}\text{C}/\text{km}$.

7.5 Environments of metamorphism

Different metamorphic facies series clearly develop under different geothermal gradients. These occur in specific geological settings (or geological environments). Several of these environments are related to subduction zones (Fig.7.9).

7.5.1 Burial metamorphism

This type of metamorphism occurs simply as a result of the increase of temperature with depth. In sedimentary basins, the sediments become lithified at a certain depth. With greater depth the conditions may approach those of low-grade metamorphism. Zeolite facies conditions are reached in the lower parts of deep sedimentary basins and the lower part of the greenschist facies can even be reached in very deep basins. The only deformation of the rocks associated with this type of metamorphism is due to the weight of the overburden (normal stress). Burial metamorphism only effects sedimentary rocks (and any minor intrusions (dykes, sills) that may have been emplaced in the basin).

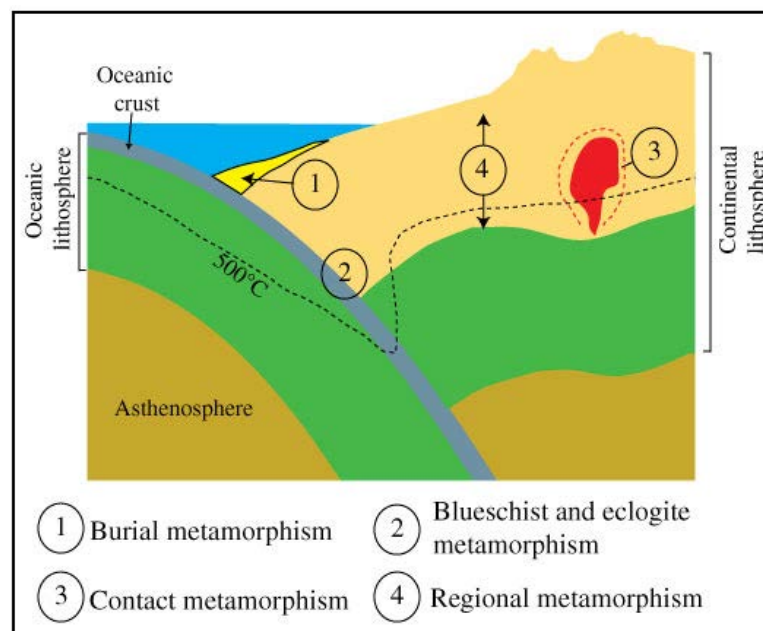


Fig.7.9. Several environments of metamorphism are related to subduction zones.

These include burial metamorphism in the lower parts of deep sedimentary basins; blueschist facies and eclogite facies metamorphism associated with the subducting plate; regional metamorphism of continental crustal rocks; and contact metamorphism associated with intrusions of magma in the continental crust.

7.5.2 Blueschist facies and eclogite facies metamorphism

The low geothermal gradient necessary for blueschist facies conditions requires relatively rapid pressure increases in relatively cold rocks. This situation occurs in subduction zones where the subducting oceanic crust (and any overlying sediments) moves downwards. The adjustment to the increase in pressure with depth is instantaneous, but the oceanic crust is relatively cold compared with the mantle. This is evident from the curvature of the 500°C isotherm in Fig.7.9. Under these PT conditions, basaltic rocks of the oceanic crust (which will largely have been metamorphosed to greenschists as a result of ocean floor metamorphism to be dealt with in section 7.5.6) will become blueschists as the mineral glaucophane becomes stable. Since it is rocks of the oceanic crust (dominantly basalts) that are subducted, most blueschist facies rocks are metabasalts. As oceanic crustal material is subducted to greater depths both the pressure and temperature rise so that conditions for the formation of eclogites are reached (Fig.7.8). Eclogites are very dense rocks (ca. 3.5g/cm³), denser than mantle peridotites, and will therefore tend to sink. This is also the major reason why they are such rare rocks at the surface of the earth; exceptional circumstances are necessary to bring eclogites to the surface.

7.5.3 Regional metamorphism

Regional metamorphism takes place as a result of the combined effects of deformation and heating and is also known as *dynamothermal* metamorphism. We have seen the effects of deformation of pelitic rocks in section 7.3.2 (the textural progression from shale to slate to phyllite to schist to gneiss); these are dynamic effects. The change in mineralogy of pelites was dealt with in section 7.4.1 where minerals appear (and some disappear) in a particular sequence as temperature rises. The effects of temperature and deformation cannot, of course, be separated. For example, garnet cannot form in slate and schist cannot contain clay minerals. We have concentrated on the metamorphism of pelitic rocks because these develop a good foliation (or schistosity) since they contain a variety of phyllosilicates. All rocks types can, of course, be subjected to regional metamorphism.

Regional metamorphism can take place over a wide range of geothermal gradients and the greenschist facies, amphibolite facies and granulite facies occupy large PT fields in Fig.7.8. The most widespread regional metamorphism takes place as a result of continental collision that is the main mountain-building process. The Caledonian mountain belt (which includes much of the Scottish Highlands and the “backbone” of Norway) was formed by the collision between two continents some 400 million years ago, consists largely of regionally metamorphosed rocks. There are, of course, some intrusions since these also develop during mountain-building processes.

7.5.4 Thermal metamorphism

This is also known as *contact metamorphism* because rocks are heated in the proximity of an intrusion of magma. A large granite pluton will initially have a temperature of ca. 800°C. Depending on their depth below the surface, the country rocks will have a temperature of, for example, 100°C. The rocks nearest to the pluton will be heated most and the thermal effect will decrease with distance from the heat source. A *contact metamorphic aureole* will develop around the intrusive body (Fig.7.10).

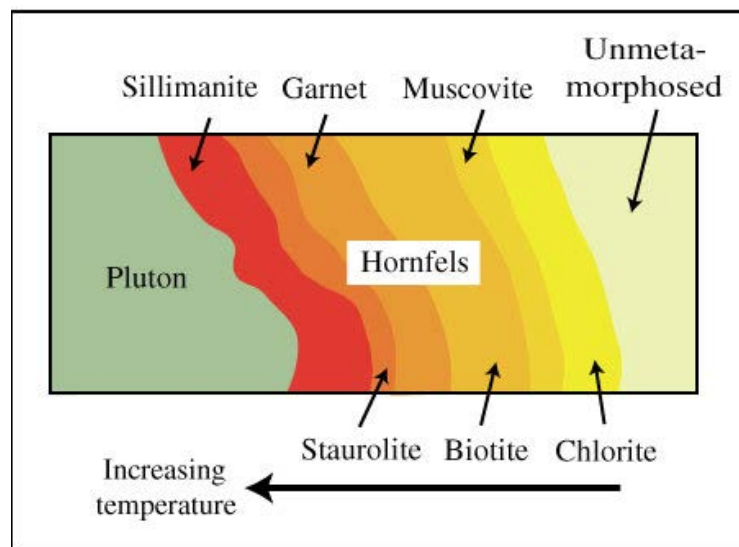


Fig.7.10. Contact metamorphism around a pluton.

The country rock envelope develops a contact metamorphic aureole. The rocks are not deformed during heating and hornfels facies rocks develop. High temperature minerals will develop close to the heat source and progressively lower temperature minerals with distance from the pluton. The minerals named here are appropriate for the contact metamorphism of pelitic rocks. As in Fig.7.6, the lines separating the different metamorphic zones are isograds.

The size of the aureole and the minerals that develop will depend on a variety of factors. These include the temperature contrast between the magma and the country rock envelope. Basaltic magma (which will form gabbro on slow cooling) may have a temperature of 1200°C, whereas rhyolitic magma (which will form a granite) may be ca. 800°C. If the magma is emplaced at depth, the country rock envelope may be as hot as 400°C; near the surface the envelope will be <100°C. Another factor is the size of the intrusion; a narrow dyke will not have the same thermal effect as a large pluton. The nature of the country rocks will also be important. The sequence of minerals that can develop in shales will be very different from those that can develop in, for example, calc-silicate rocks.

7.5.5 Dynamic metamorphism

The geological environment in which dynamic metamorphism occurs is in the vicinity of faults. Any heat involved is either due to the depth below the surface or to heat supplied by friction. Rocks near the surface react in a brittle way to faulting and can become crushed or brecciated (broken into angular fragments) to form a fault breccia. At depth, where rocks behave plastically, recrystallization can take place to form fine-grained, strongly-foliated rocks called mylonite (Fig.7.11).

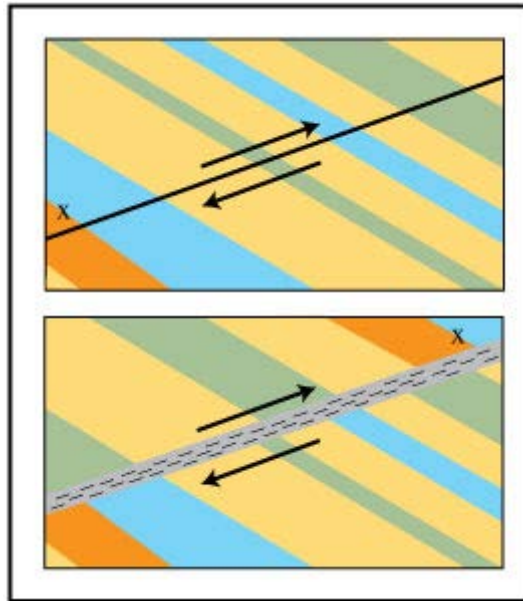


Fig.7.11. Rocks on either side of a deep, major fault plane can recrystallize to a fine-grained “mixed rock” called a mylonite.

The composition of a mylonite depends on the nature of the rocks on either side of the fault. Here the mylonite (grey) will be a mixture of four rock types illustrated with different colours.

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7.5.6 Metamorphism at mid-ocean ridges

The oceanic crust forms at mid-ocean ridges and consists dominantly of basaltic rocks. There is a thin covering of deep sea sediments, increasing in thickness away from the ridge. Seawater sinks into the oceanic crust through cracks and faults. This water seeps down into hot rocks and a *convection system* is set up (Fig.7.12). The hot water passing through the basalts reacts with them (at 300–400°C) and the basaltic mineralogy changes into that equivalent to a greenschist (chlorite, epidote, albite). There is no deformation during the metamorphism, so no schistosity develops. This type of metamorphism is also known as “ocean floor metamorphism”.

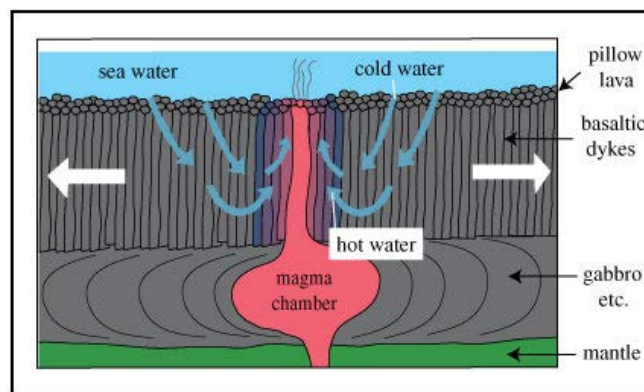


Fig.7.12. Hot water reacts with basaltic rocks near mid-ocean ridges.

A convection system is set up as seawater sinks into hot oceanic crustal rocks. The hot water reacts with the basaltic rocks and gives rise to metamorphism under conditions equivalent to the greenschist facies (at low pressure). The basaltic mineralogy becomes transformed into chlorite + epidote + albite. The resulting rocks are called “greenstones” since they are mineralogically similar to greenschists but are not foliated.

This means that most of the oceanic crust comprises metabasalts, and the oceanic crust that is subducted is “wet”. As the wet oceanic crust is subducted it becomes metamorphosed in blueschist facies (section 7.5.2.). But some of the water is released into the overlying wedge of mantle peridotite. The presence of water in the peridotite means that it can begin to melt at a lower temperature than dry peridotite. This is a major reason why subduction zones are regions of volcanic activity.

7.6 Where do metamorphic rocks occur?

Most metamorphic rocks at the surface of the earth occur in the Precambrian shield areas (ancient continents) and mountain belts. The Precambrian shields are dominated by gneissic rocks. Mountain belts are formed as a result of continental collision. The “young” mountain belt that stretches roughly east-west from the Alps to the Himalayas was formed (and is still being formed) as a result of the collision of the African and Indian continental plates with the European and Asian continental plates. Gneissic areas represent the “root zones” of ancient mountain belts.