Unit 3

Magma and Igneous Rocks

Why rock melts, how melt cools, and what the crystals record

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By the end of this unit, I should be able to answer these questions:

- Why does the mantle melt under a mid-ocean ridge although it gets cooler on the way up?
- What decides whether a melt becomes granite or basalt, and why is "granite lava" so rare while basalt lava is everywhere?
- Why does a slowly cooled magma make a coarse rock while the same magma erupted makes fine-grained basalt or glass?
- What does Bowen's reaction series say, and how does it let one magma produce several different rocks?
- How do I tell a dyke from a sill and a pluton from a lava flow in the field and on a map?
- What are the hexagonal columns of the Giant's Causeway, and why are they hexagonal?

Every rock on Earth began as a melt. The oldest continental crust, the floor of every ocean, and the ground under every volcano all passed through a liquid state before they became the solid rock a geologist can hold. This unit follows that history from the first drop of melt in the hot mantle to a crystal under the microscope. It explains why rock melts at all, what decides whether the melt ends up as granite or basalt, how the size of the crystals records the speed of cooling, and how to read the bodies that magma leaves behind in the field.

1 Why rock melts

Magma is molten rock below the surface. It is never a pure liquid. The melt itself is a hot liquid of silicon, oxygen, aluminium, iron, magnesium, calcium, sodium, and potassium, and it carries crystals that have already formed, fragments of the rocks it passes through, and dissolved gases, above all water vapour and carbon dioxide. When magma reaches the surface and the gases escape, the liquid that flows out is called **lava**. The crystals in a magma are minerals, that is, naturally formed solids with a definite chemical composition and an ordered arrangement of atoms. Almost all the minerals in this unit are silicates, which are built from the silicon-oxygen tetrahedron, a silicon atom surrounded by four oxygen atoms. Basaltic lava, the dark lava of Hawaii and Iceland, erupts at about 1,100 to 1,200 °C, which is hot enough to glow orange in daylight. Lavas that are rich in **silica**, that is, in silicon dioxide, SiO_2, which is what the tetrahedra of a silicate add up to, are the kind that build steep volcanoes, and they are cooler, at about 700 to 900 °C.

The surprising fact about melting is how rarely it happens. The temperature in the crust rises by about 25 °C for every kilometre of depth, with a range of 15 to 30 °C per kilometre in stable continental crust, and the deep mantle has a temperature of several thousand degrees. Yet the mantle is solid rock, and we know

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this because earthquake waves of a kind that cannot pass through liquids travel through it. The reason is that the melting point of rock rises with pressure, and in the mantle it rises faster than the temperature does. Two curves decide everything in this section, and Figure 1 draws them together with the three ways in which a rock can be made to cross from one to the other. The **geotherm** is the temperature of the rock at each depth. The **solidus** is the temperature at which a rock of a given composition begins to melt at each depth, and the **liquidus** is the higher temperature at which the last crystal disappears; between the two the rock is partly molten. The solidus in the figure is that of **peridotite**, the olivine-rich rock that makes up the upper mantle. Under a stable continent the geotherm stays below the solidus at every depth, and that is why continents do not sit on a sea of magma.

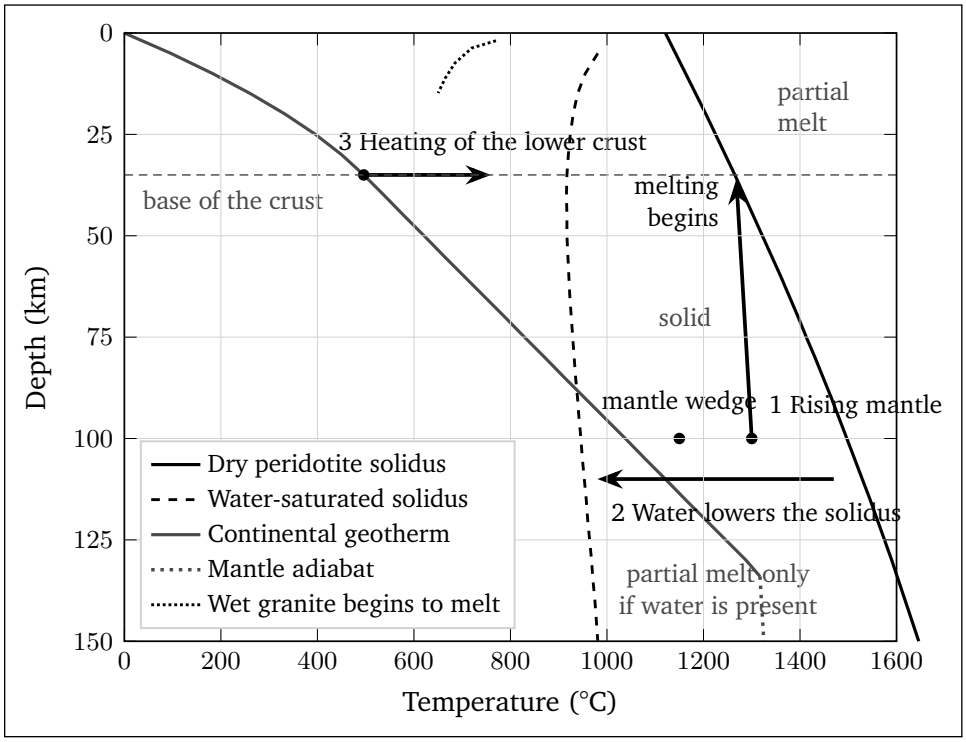


Figure 1. *Temperature against depth in the upper mantle and crust, with the three ways to melt rock. The dry peridotite solidus is the fit of Hirschmann (2000) to melting experiments, the water-saturated solidus follows the formula of Katz, Spiegelman, and Langmuir (2003), and the continental geotherm is a steady conductive model with a surface heat flow of 60 mW per square metre, a conductivity of 3 W per metre and kelvin, and 1 μW per cubic metre of radioactive heat production in a 35 km crust. Below the crust the geotherm joins the mantle adiabat, the temperature curve of mantle whose temperature changes by compression and expansion alone, which would be at 1,250 °C at the surface and warms by 0.5 °C per kilometre of depth. Water-rich granitic crust begins to melt along the short dotted curve (the experiments of Tuttle and Bowen, 1958). Depth is converted to pressure with a density of 3,300 kg per cubic metre.*

There are only three ways to bring a rock across its solidus, and each has its address on the plate map. The first is to lower the pressure. Hot mantle that rises, for example under a mid-ocean ridge where two plates pull apart, or in a hot column of mantle under Hawaii or Iceland, loses pressure far faster than it loses heat. A rising parcel cools by only about 0.5 °C per kilometre, because expansion is the only thing that cools it (this rate of cooling by expansion alone is called the adiabatic gradient, and estimates of it run from 0.3 to 0.5 °C per kilometre), while its solidus drops by about 4 °C per kilometre. The parcel therefore meets its solidus on the way up and begins to melt, and this is called **decompression melting**. It makes most of the magma on Earth, including the basalt of the entire ocean floor. The second way is to add water. Water dissolved in a silicate melt breaks the bonds between the silicon-oxygen tetrahedra, and a wet peridotite begins to melt several hundred degrees below the dry solidus, as the dashed curve in Figure 1 shows. Above a subducting plate, the sinking plate, which geologists call the slab, heats up and releases the water it carries in its minerals, the water rises into the mantle above it, and that mantle melts although its temperature has not changed. This **flux melting** feeds the volcanoes of the Andes, Japan, and every other volcanic arc. The third way is to heat the rock, and it is the rarest of the three. Basaltic magma from the mantle that

pools at the base of thick continental crust can raise the temperature there by several hundred degrees, and water-rich crust begins to melt at only about 650 to 700 °C. Melting of this kind produces some of the granites of mountain belts.

Key Result **The three ways to melt rock**

A rock melts when its temperature rises above its solidus, and this happens in three ways. Decompression melting occurs where hot mantle rises and its solidus falls faster than it cools, as under mid-ocean ridges and hot spots. Flux melting occurs where water released from a subducting plate lowers the solidus of the mantle above it, as under volcanic arcs. Melting by heating occurs where hot basaltic magma warms the base of thick crust. Decompression makes most of the Earth's magma.

Worked Example **Where a rising parcel of mantle begins to melt**

Consider a parcel of mantle at 1,300 °C at a depth of 100 km that starts to rise under a mid-ocean ridge. The pressure at that depth is the weight of the rock above, $P = \rho g z = 3{,}300\,\mathrm{kg\,m^{-3}} \times 9.81\,\mathrm{m\,s^{-2}} \times 100{,}000\,\mathrm{m} = 3.24\,\mathrm{GPa}$. A gigapascal is a billion pascals, so this is about 32,000 times the pressure of the atmosphere at sea level. The solidus of dry peridotite from the compilation of melting experiments by Hirschmann (2000) is

$$T_\mathrm{s} = 1{,}120.7 + 132.9\,P - 5.10\,P^2$$

in degrees Celsius with P in gigapascals, which gives $T_\mathrm{s} = 1{,}498\,^\circ\mathrm{C}$ at 100 km. The parcel is almost 200 °C too cold to melt there. As it rises it cools by about 0.5 °C per kilometre, while its solidus falls by about 120 °C per gigapascal, which is the slope of the formula above at these pressures. Every kilometre of depth adds $\rho g \times 1{,}000\,\mathrm{m} = 32\,\mathrm{MPa}$, or 0.032 GPa, so the solidus falls by about 3.9 °C for every kilometre the parcel rises. The gap of 198 °C therefore closes at about 4.4 °C per kilometre, and a first estimate puts the start of melting after about 45 km of rise, near 55 km depth. The formula is less steep at high pressure than the 120 °C per gigapascal used here, so the gap closes a little more slowly at first and the exact crossing lies a little shallower than 55 km. At 50 km the solidus is 1,322 °C and the parcel, which has cooled by 0.5 °C for each of the 50 km it has risen, is at 1,275 °C; at 40 km they are 1,284 and 1,270 °C; at 36 km both are 1,268 °C. The parcel begins to melt at a depth of about 36 km, at a pressure of 1.2 GPa, and every further kilometre of rise melts a little more of it. The mantle under ridges is probably somewhat hotter than the parcel of this example, and estimates vary; a parcel starting at 1,350 °C would begin to melt at about 51 km and one at 1,400 °C at about 66 km.

Check Your Understanding **Question 1**

A parcel of mantle rises under a mid-ocean ridge and cools as it goes. Why does it nevertheless begin to melt?

(A) Its solidus temperature falls with the falling pressure faster than the parcel itself cools

(B) Friction against the surrounding mantle heats the parcel as it rises

(C) Seawater seeps down through the ridge and lowers the melting point of the parcel

(D) The parcel's temperature rises as the pressure on it drops, and that starts the melting

The solution is at the end of the unit.

A rock that crosses its solidus does not turn into liquid all at once, because a rock is a mixture of several

minerals and the mixture melts over a range of temperatures. Under a mid-ocean ridge, for instance, the rising mantle melts by roughly 5 to 20 per cent before the melt escapes upward, and the rest stays solid. This is called **partial melting**, and it has a consequence that shapes the whole planet: the melt does not have the composition of the rock it came from. The mechanism is explained in the box below; its outcome can be stated in two sentences. Partial melting of peridotite, which contains about 45 per cent silica by mass, yields a basaltic melt with about 50 per cent silica. Partial melting of basaltic or of older continental crust yields melts with 60 to 75 per cent silica, which solidify as the rocks called diorite and granite, which Section 2 introduces, and as their volcanic equivalents. Each round of melting pushes the melt toward more silica, and this is the first of two processes (the second is in Section 3) that have sorted the Earth into a silica-poor mantle and a silica-rich continental crust.

Mechanism in Detail **Why the first melt is richer in silica than the rock it leaves**

A mixture of two substances melts at a lower temperature than either substance alone, which is why salt spread on an icy road makes the ice melt below 0 °C. The same holds for the minerals of a rock. Pure diopside, a pyroxene, melts at about 1,390 °C and pure anorthite, the calcium plagioclase, at about 1,550 °C, yet a mixture of the two begins to melt at about 1,270 °C, as Bowen measured in 1915, and a rock of quartz and feldspar begins to melt at only about 650 to 700 °C when water is present. The first liquid to appear always has the composition of the lowest-melting mixture of the minerals present, and in silicate rocks that mixture is rich in silicon, aluminium, sodium, and potassium, the elements that bind most weakly into the crystal structures, and poor in magnesium and iron, which bind tightly into olivine and pyroxene. As melting continues the liquid dissolves more of the refractory minerals, those with the highest melting points, and its composition moves back toward that of the whole rock, but the melt escapes long before that point. A 10 per cent melt of peridotite is therefore a basalt, and what stays behind is a peridotite even richer in olivine than before.

Key Result **Partial melting makes the melt richer in silica**

A rock melts over a range of temperatures, and the first melt has the composition of its lowest-melting mixture of minerals, which is richer in silica, aluminium, and the alkalis (sodium and potassium) than the rock itself. Partial melting of mantle peridotite with about 45 per cent silica gives basalt with about 50 per cent, and partial melting of crust gives melts with 60 to 75 per cent silica. The residue is poorer in silica than before.

2 The four groups by silica

Geologists sort magmas and the rocks they form by the mass fraction of silica, SiO_2, in a chemical analysis of the whole rock. Four groups are in use, and their boundaries are conventions agreed for convenience rather than natural breaks. A rock is called **ultramafic** below 45 per cent silica, **mafic** between 45 and 52 per cent, **intermediate** between 52 and 66 per cent, and **felsic** above 66 per cent. The words come from the chemistry: mafic rocks are rich in magnesium and iron (ferrum in Latin), felsic rocks in feldspar and silica. Table 1 lists the four groups with the properties that follow from the silica content, and the rest of this section explains why they follow. The minerals of each group are the ones that Bowen's reaction series in Section 3 assigns to it: olivine and pyroxene in the ultramafic rocks, plagioclase and pyroxene in the mafic ones, plagioclase and amphibole in the intermediate ones, and quartz and the feldspars in the felsic ones.

Property	Ultramafic	Mafic	Intermediate	Felsic
Silica (per cent by mass)	below 45	45 to 52	52 to 66	above 66
Volcanic rock	komatiite	basalt	andesite	rhyolite
Plutonic rock	peridotite	gabbro	diorite	granite
Colour	dark green	black, dark grey	grey	pale, pink
Eruption temperature (°C)	above 1,500	1,100 to 1,200	950 to 1,050	700 to 900
Viscosity (Pa s)	below 10	10 to 100	10^3 to 10^5	10^6 and above
Density (kg m^{-3})	about 3,300	about 3,000	about 2,800	about 2,700

Table 1. *The four groups of igneous rocks by silica content and the properties that go with them. The boundaries are conventions. Temperatures and viscosities are typical values for melts with little water and few crystals; dissolved water lowers the viscosity by orders of magnitude, and the densities are those of the solid rocks.*

The most important of these properties is the **viscosity**, the resistance of a liquid to flow, measured in pascal seconds. Water has a viscosity of about 0.001 Pa s and honey about 10 Pa s. In a silicate melt the silicon-oxygen tetrahedra link into chains and networks, exactly as they do in the crystals of Unit 2, and the more silica a melt contains, the more complete the network and the harder it is for the liquid to flow. A basaltic melt at 1,200 °C has a viscosity of 10 to 100 Pa s, which is thicker than honey, and basalt still flows for tens of kilometres down a volcano. A rhyolitic melt at 800 °C has a viscosity of a million pascal seconds or more, a hundred thousand times stiffer than honey, and such a melt piles up over its vent in a dome rather than flowing away. Two other things change the viscosity. A higher temperature makes every melt thinner, and dissolved water makes it thinner still, because the water breaks up the network of tetrahedra. Crystals suspended in the melt make it thicker, and a magma that is more than about half crystals no longer flows at all, which Section 5 comes back to.

The silica content also explains why “granite lava” hardly exists while basalt lava covers the ocean floors. A felsic magma is cool, stiff, and rich in dissolved water. It rises slowly through the crust, loses heat on the way, and usually stops and solidifies at depth as granite. When it does reach the surface, the water comes out of solution as the pressure falls, and a melt too stiff to let the bubbles escape is torn apart into ash and pumice. Felsic volcanoes therefore erupt mostly fragments, and rhyolite lava forms only short, thick flows and domes. A basaltic magma is hot, fluid, and poor in water, so its gas escapes quietly and the lava runs. The same difference in viscosity decides what an eruption looks like, which is a subject beyond the Foundation units.

Check Your Understanding **Question 2**

Which of these melts flows most easily?

(A) A rhyolitic melt with 74 per cent silica at 850 °C

(B) A basaltic melt with 49 per cent silica at 1,150 °C

(C) An andesitic melt with 60 per cent silica at 1,000 °C

(D) A rhyolitic melt with 74 per cent silica at 1,000 °C

The solution is at the end of the unit.

Each group has its place on the map of the plates, which Unit 6 describes in full. Basalt forms wherever the mantle melts by decompression: at the mid-ocean ridges, which produce about 20 cubic kilometres of new basaltic crust every year, a cube about 2.7 km on each side, over hot spots such as Hawaii, and in the flood basalts that have covered regions the size of countries, such as the Antrim plateau of Northern Ireland in Section 5. Andesite is the typical lava of volcanic arcs above subduction zones, where the melts that added water produces in the mantle change their composition on the long way up through the crust. Granite forms in the thick crust of continental arcs and collision zones, and the great granite bodies of the Andes, the Sierra Nevada of California, and the Himalaya are all of this kind. Peridotite is the mantle itself,

and it reaches the surface only in slices of ocean floor pushed onto land and in fragments carried up by basalt. Komatiite, the ultramafic lava, is almost entirely restricted to rocks older than 2,500 million years, when the mantle was hotter than today.

Key Result **The four silica groups**

Igneous rocks are grouped by their silica content as ultramafic (below 45 per cent), mafic (45 to 52), intermediate (52 to 66), and felsic (above 66 per cent); the boundaries are conventions. With rising silica the melt is cooler, far more viscous, richer in dissolved water, less dense, and paler, because the silicon-oxygen tetrahedra link into a stiffer network. Mafic magma flows out as lava; felsic magma mostly solidifies at depth or erupts as ash.

3 Bowen's reaction series

As a basaltic magma cools, its minerals appear one after another in a definite order, which was worked out between 1912 and 1928 by Norman L. Bowen (1887-1956), as the excursion below tells. Figure 2 shows the result, which every geologist knows as Bowen's reaction series.

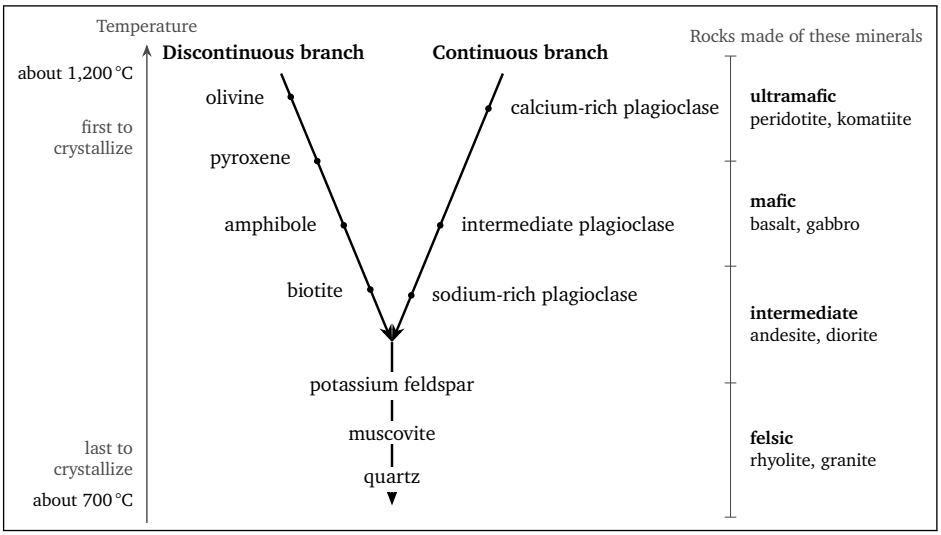


Figure 2. *Bowen's reaction series. On the discontinuous branch each mineral reacts with the remaining melt to form the next one as the temperature falls; on the continuous branch plagioclase keeps its structure and changes its composition from calcium-rich to sodium-rich. The two branches meet, and potassium feldspar, muscovite, and quartz crystallize last. The temperatures are approximate and depend on pressure and water content. Drawn after Bowen (1922, 1928).*

The series has two branches that run at the same time. On the **discontinuous branch**, the first mineral to crystallize from a basaltic melt is olivine. As the temperature falls, the olivine crystals react with the melt around them and are converted into pyroxene, the pyroxene later reacts to form amphibole, and the amphibole reacts to form biotite. Each of these is a different mineral with a different structure, which is why the branch is called discontinuous. On the **continuous branch**, plagioclase feldspar crystallizes from beginning to end, but its composition changes. The first plagioclase is rich in calcium, and as the melt cools the crystals react with it and take up more and more sodium, so that a single crystal that grew during cooling has a calcium-rich centre and a sodium-rich rim. Plagioclase is a solid solution between a calcium end member and a sodium end member, as Unit 2 explained for olivine. The two branches meet when the melt has cooled to roughly 800 °C, and the last minerals to crystallize, from a melt that is now rich in silica, potassium, and water, are potassium feldspar, muscovite, and quartz.

Principle **Bowen's reaction series**

As a basaltic magma cools, its minerals crystallize in a fixed order. On the discontinuous branch olivine forms first and reacts with the melt to give pyroxene, then amphibole, then biotite. On the continuous branch plagioclase forms throughout, changing its composition from calcium-rich to sodium-rich. Potassium feldspar, muscovite, and quartz crystallize last, from the silica-rich melt that remains. Minerals near the top of the series are therefore found together in mafic rocks and minerals near the bottom in felsic rocks, and minerals from the two ends rarely occur together.

The series explains at once why a basalt contains olivine, pyroxene, and calcium-rich plagioclase and hardly ever quartz, while a granite contains quartz, potassium feldspar, sodium-rich plagioclase, and a little biotite and hardly ever olivine (the exceptions, such as the little quartz in some sills of dolerite, the medium-grained basalt of sills and dykes, come from the last drops of melt or from unusual compositions). Minerals that crystallize at similar temperatures occur together, and minerals from opposite ends of the series do not. It also explains a second thing, which Bowen regarded as his main result. If the early crystals are removed from the melt, the melt that remains has a new composition, poorer in magnesium, iron, and calcium and richer in silica, sodium, and potassium than the magma it started as. Crystals are removed when they sink to the floor of a magma body because they are denser than the melt, when they are left behind on the walls, or when the melt is squeezed out of a mass of crystals. This process is called **fractional crystallization**, and the sorting of a magma into parts of different composition is called **differentiation**. A single parent basalt can in this way produce a gabbro from its first crystals, then a diorite, and finally a small amount of granite from its last melt, and the layered intrusions of the world, where the sunken crystals lie in beds like sediments, are the record of it. Two other processes change a magma on its way up: it can melt and dissolve the rocks it passes through, which is called assimilation, and two magmas can mix.

Worked Example **The magnesium left in a melt after olivine has settled out**

Suppose a melt freshly extracted from the mantle contains 18 per cent magnesium oxide, MgO, by mass, as the most magnesium-rich basaltic melts do. The olivine that crystallizes from it has the composition $(\mathrm{Mg_{0.9}Fe_{0.1}})_2\mathrm{SiO_4}$. With the atomic masses Mg 24.3, Fe 55.8, Si 28.1, and O 16.0, the mass of one formula unit is $1.8 \times 24.3 + 0.2 \times 55.8 + 28.1 + 4 \times 16.0 = 147.0$ in atomic mass units. The 1.8 magnesium atoms count as 1.8 units of MgO at $24.3 + 16.0 = 40.3$ each, which is $1.8 \times 40.3 = 72.5$, so MgO makes up $72.5/147.0 = 49.4$ per cent of the olivine. Let 20 per cent of the mass of the melt crystallize as this olivine and settle to the floor. The magnesium oxide must be conserved, so with m the mass of the original melt,

$$0.18\,m = 0.20\,m \times 0.494 + 0.80\,m \times x,$$

where x is the mass fraction of MgO in the melt that remains. Solving gives $x = (0.18 - 0.0988)/0.80 = 0.102$. The remaining melt contains about 10 per cent MgO, close to the most magnesium-rich basalts of the ocean floor, and its silica content has risen from 48 to about 50 per cent, because olivine contains only 41 per cent silica (the same conservation sum with 0.48 for the melt and 0.41 for the olivine gives 0.498). Removing a quarter of the original melt in all, rather than a fifth, would bring the MgO down to about 7.5 per cent, which is a common ocean-floor basalt; the MgO falls fast because every gram of olivine carries away about five times as much MgO as a gram of the melt holds. Olivine settling alone can thus turn one melt into a family of basalts.

Check Your Understanding **Question 3**

A geology student reports an igneous rock in which olivine and quartz grew together from the same melt. Why would a petrologist ask to see the specimen?

(A) Because olivine forms only in volcanic rocks and quartz only in plutonic ones, so they never share a melt

(B) Because olivine crystallizes at the start of Bowen's series and quartz at the end, from melts of different composition

(C) Because quartz cannot crystallize from any melt and must have been deposited by groundwater afterwards

(D) Because olivine reacts with water and would have turned into quartz long ago

The solution is at the end of the unit.

Excursion ***Bowen at the Geophysical Laboratory***

Bowen was born in Kingston, Ontario, and as a student at Queen's University there he spent his summers mapping rocks by canoe for the Ontario Bureau of Mines, before he went on to the Massachusetts Institute of Technology. In 1912, at the age of 25, he joined the Geophysical Laboratory of the Carnegie Institution of Washington, which had been founded in 1905 to study rocks with the methods of physics and chemistry. He ground pure chemicals to the composition of a mineral mixture, held the mixture at a chosen temperature in an electric furnace until crystals and liquid had stopped changing, that is, until the mixture had reached equilibrium, dropped it into water to freeze whatever liquid was present into glass, and then identified the crystals and the glass under the microscope. Dozens of such runs, one temperature at a time, gave the melting diagram of the mixture. His 1913 diagram of the plagioclase feldspars showed that the crystals and the melt in equilibrium with them always differ in composition, with the crystals richer in calcium, and that is the continuous branch of the series. In 1915 he published "The later stages of the evolution of the igneous rocks", which argued that fractional crystallization of basalt could produce the whole range of igneous rocks, and his 1928 book *The Evolution of the Igneous Rocks* became the handbook of a generation of petrologists. Field geologists resisted for years, and modern work has added other processes to his, but the reaction series itself has stood for a century.

4 Textures and the names of igneous rocks

The size and arrangement of the crystals in an igneous rock are called its **texture**, and texture is a record of how fast the melt cooled. The box below explains why, and the rule it gives is worth stating first: the slower the cooling, the larger the crystals. Geologists describe igneous textures with a handful of words.

- A **glassy** rock has no crystals at all, because the melt cooled too quickly for any to form. Obsidian is a felsic glass, and the dark crust of a basalt flow is glass a few millimetres thick.
- An **aphanitic** rock is made of crystals too small to see without a lens, smaller than about a millimetre. This is the texture of lava flows and of thin dykes and sills, which cool in hours to years.
- A **phaneritic** rock is made of crystals large enough to see with the unaided eye, from about a millimetre to a few centimetres, all of roughly similar size and interlocking. This is the texture of rocks that cooled slowly at depth over thousands of years.
- A **porphyritic** rock has large crystals, called **phenocrysts**, set in a finer groundmass. It records two

steps of cooling: a slow one at depth, during which the phenocrysts grew, and a fast one when the magma rose or erupted and the groundmass froze around them.

- A **pegmatitic** rock has crystals larger than a few centimetres, sometimes metres long. Pegmatites form from the last, water-rich melt of a granite, in which atoms move fast enough to build giant crystals.
- A **vesicular** rock is full of small cavities left by gas bubbles that were frozen into the lava. Pumice is so full of them that it floats.
- A **pyroclastic** rock is made of fragments thrown out by an explosive eruption, from ash to blocks, that settled and were welded or cemented together. Tuff is the common name.

Mechanism in Detail — Why slow cooling makes large crystals

A crystal grows in two steps. First a few atoms must come together in the right arrangement to form a seed, which is called nucleation, and then atoms from the melt attach to the seed, which is growth. Nucleation is hard, because a tiny seed has a large surface for its volume and the surface costs energy, so seeds form only when the melt has been cooled some way below its freezing temperature. Growth, on the other hand, needs only that atoms diffuse through the melt to the seed, and diffusion is fast in a hot melt and slow in a cool one. When a melt cools slowly, it stays just below the freezing temperature for a long time. Few seeds form there, but atoms move freely, so the few seeds grow large, and the result is a phaneritic rock. When a melt cools fast, it is driven far below the freezing temperature before much has happened. Seeds now form in great numbers, but the melt has become stiff and the atoms slow, so each seed grows little, and the result is an aphanitic rock. Cooled fast enough, the melt becomes too stiff for seeds to form at all and freezes as glass. The same competition between nucleation and growth makes large crystals in a sugar syrup left for a week and fine ones in syrup cooled on ice.

Key Result — Grain size records cooling rate

The size of the crystals in an igneous rock records how fast the melt cooled, because slow cooling gives few seeds that grow large and fast cooling gives many seeds that stay small. Glass means cooling within hours, crystals below a millimetre (aphanitic) mean cooling within days to years in a lava flow or thin intrusion, and crystals above a millimetre (phaneritic) mean cooling over thousands of years at depth. Phenocrysts in a fine groundmass record slow cooling followed by fast cooling.

How long a body of magma takes to solidify can be estimated from a single rule of heat conduction. Heat travels through rock by conduction, and the time t it takes to pass through a thickness d is

$$t \approx \frac{d^2}{\kappa}, \tag{1}$$

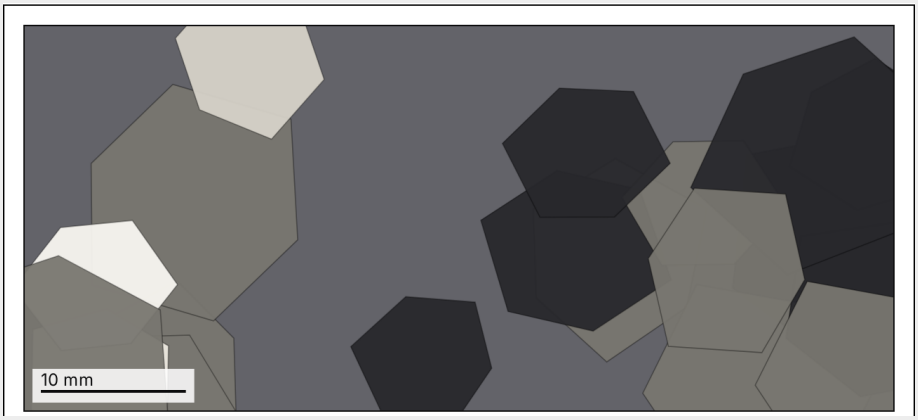
where κ (kappa) is the **thermal diffusivity** of rock, a measure of how fast heat spreads through it, about $10^{-6}\,\mathrm{m^2\,s^{-1}}$. A thickness in metres, squared and divided by this number, gives a time in seconds, and the value of κ means that heat needs about 10^6 seconds, roughly twelve days, to pass through one metre of rock. The time grows with the square of the size, so a body ten times thicker takes a hundred times longer to cool, and this is why the cooling times of igneous bodies span from hours to a million years.

Worked Example — The cooling time of a dyke and of a pluton

A basalt dyke one metre thick loses its heat through both walls, so the heat has to travel about half a metre. Equation (1) gives $t \approx (0.5\,\mathrm{m})^2/10^{-6}\,\mathrm{m^2\,s^{-1}} = 2.5 \times 10^5\,\mathrm{s}$, which is about three days; for the whole body to reach the temperature of the surrounding rock takes a few times longer, say a week or two. A granite pluton one kilometre across has a thousand times the distance, and the square makes it a million times the time: $t \approx (500\,\mathrm{m})^2/10^{-6}\,\mathrm{m^2\,s^{-1}} = 2.5 \times 10^{11}\,\mathrm{s}$, which is about 8,000 years for the centre to begin to cool and several tens of thousands of years for the body to solidify throughout. The pluton therefore takes about a million times longer than the dyke, tens of thousands of years

against days, and that difference in time is the difference between granite and basalt in the hand. The estimate ignores the heat released when the melt crystallizes, which lengthens the time, and the circulation of groundwater through the hot rock, which shortens it, but neither changes the factor of a million by much. The largest batholiths, hundreds of kilometres long, were not single melts at all; they grew from many separate pulses over a million years or more.

Interactive **Cooling rate, texture, and the name of the rock**



Move the time slider from a few hours to a million years and watch the texture change from glass to a mass of invisible crystals to a coarse mosaic; the readouts give the typical crystal size, the thickness of a body that takes that long to solidify by the conduction rule of the unit, and the name of the rock once the silica slider has set its group. Notice that the time, not the composition, decides the texture, and that the composition, not the time, decides whether the rock is basalt or rhyolite, gabbro or granite.

Click the picture or this link to try it in the online version of this unit.

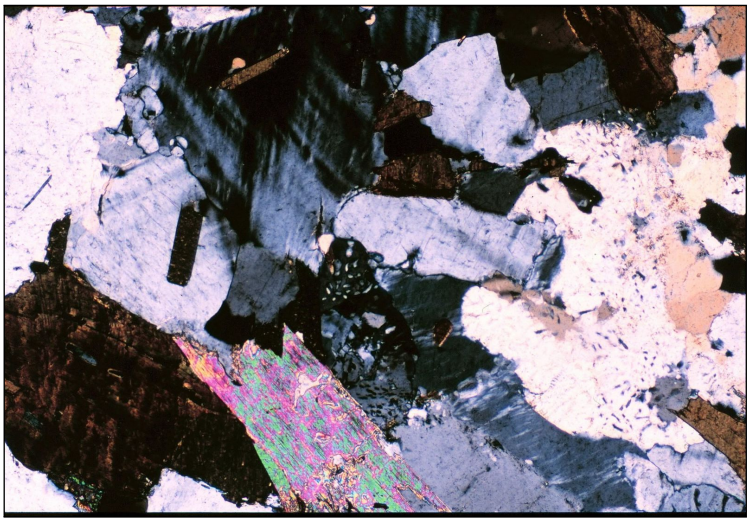


Figure 3. *A thin section of the Woodbury Granite from near Barre, Vermont, in cross-polarized light. The grey and white crystals with no stripes are quartz, the striped grey crystals are plagioclase, whose stripes are twinning, two orientations of the same crystal grown together, the brown crystals are biotite, and the bright pink and green crystal is muscovite. The photographer gives no scale; Dale (1909) measured the feldspars of this granite at about 2.7 mm across on average, which gives the scale of the picture. Credit: Darla Sondrol, Geoscience Digital Image Library, CC0.*

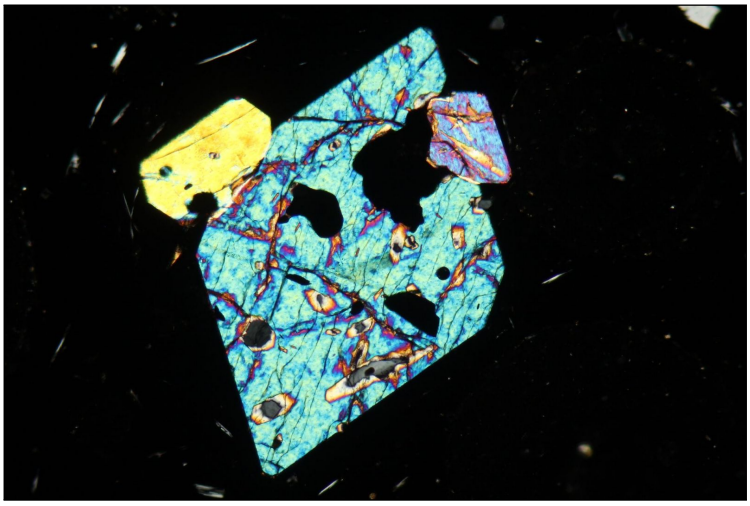


Figure 4. *A thin section of an olivine basalt in cross-polarized light, 2 mm across. The large bright crystal is an olivine phenocryst, its shape complete because it grew freely in the melt at depth, and the groundmass around it is a mass of plagioclase laths and pyroxene too small to see without a microscope, which froze when the lava erupted. Credit: Strekeisen, Wikimedia Commons, CC BY-SA 4.0.*

Figures 3 and 4 show the two commonest igneous rocks under the polarizing microscope of Unit 2, in slices ground to a thickness of 0.03 mm so that light passes through the crystals. In the granite every crystal is a millimetre or more across and the crystals interlock without any gaps, which is the phaneritic texture of a melt that cooled over thousands of years underground. In the basalt a single large olivine crystal sits in a groundmass of crystals a hundredth of its size, which is the porphyritic texture of a lava that carried its early crystals to the surface and then froze within days.

Texture and composition together give an igneous rock its name, and Table 2 is the grid that geologists use. The columns are the four silica groups of Section 2, and the rows are the two textures, so that each column holds a pair of rocks of the same composition: a fine-grained volcanic rock that cooled at the surface and a coarse-grained plutonic rock that cooled at depth. Basalt and gabbro are chemically the same rock, and so are andesite and diorite, and rhyolite and granite. The names in the table are the ones in everyday use; the full scheme of the International Union of Geological Sciences names plutonic rocks by the proportions of quartz, alkali feldspar (the group to which the potassium feldspar of Bowen's series belongs), and plagioclase in a triangular diagram, and granite is one of its fields, as the last worked example of this section shows.

Texture	Ultramafic	Mafic	Intermediate	Felsic
Aphanitic (volcanic)	komatiite	basalt	andesite	rhyolite
Phaneritic (plutonic)	peridotite	gabbro	diorite	granite
Glassy	(none)	basaltic glass	(rare)	obsidian
Pyroclastic	(none)	basaltic tuff	andesitic tuff	rhyolitic tuff
Silica (per cent)	below 45	45 to 52	52 to 66	above 66
Dark minerals (per cent)	above 70	about 40 to 70	about 20 to 40	below 20

Table 2. *The classification grid of the common igneous rocks. A rock is named from its silica group (columns) and its texture (rows). Dolerite, also called diabase, is the medium-grained basalt of dykes and sills, and the proportions of dark minerals are typical values rather than definitions.*

Worked Example **Naming the rock of Figure 4**

The thin section in Figure 4 shows a groundmass of crystals far smaller than a millimetre, so the texture is aphanitic, with one crystal a millimetre across set in it, so the rock is also porphyritic. The phenocryst is olivine, and the groundmass is plagioclase and pyroxene; there is no quartz. Olivine, pyroxene, and plagioclase are the minerals at the top of Bowen's series, and a rock made of them with no quartz is mafic, in the column of Table 2 with 45 to 52 per cent silica. Mafic and aphanitic together give the name basalt, and the full description is a porphyritic olivine basalt. Had the same melt cooled slowly at depth, the same minerals would have grown to several millimetres and the rock would be called gabbro. For comparison, a rock in which the crystals are all several millimetres across and the minerals are quartz, potassium feldspar, plagioclase, and a little biotite, with an analysis of about 70 per cent silica, is felsic and phaneritic, and its name is granite.

Worked Example **The mode of the Woodbury Granite**

The proportions of the minerals in a rock by volume are called its mode, and they are measured on a thin section by counting what lies under each point of a regular grid. In 1909 the survey geologist T. Nelson Dale measured the mode of the Woodbury Granite of Vermont, the rock of Figure 3, with the older Rosiwal method, which adds up the lengths of the crystals of each mineral along straight lines. Along 38.1 inches of lines, just under a metre in all, across a specimen from the lower quarry of the Woodbury Granite Company he found feldspar 65.6 per cent, quartz 27.1 per cent, and mica 7.3 per cent. To name a plutonic rock, the scheme of the International Union of Geological Sciences uses only quartz (Q), alkali feldspar (A), and plagioclase (P), recalculated to 100 per cent. The quartz fraction is

$$\frac{Q}{Q+A+P} = \frac{27.1}{27.1+65.6} = 0.29,$$

or 29 per cent, which lies in the band from 20 to 60 per cent that holds granite, granodiorite, and tonalite. The band is divided by the share of plagioclase among the feldspars: below 10 per cent the rock is an alkali-feldspar granite, from 10 to 65 per cent a granite, from 65 to 90 per cent a granodiorite, and above 90 per cent a tonalite. Dale lists the potassium feldspar as the most abundant mineral of this rock and the milk-white plagioclase as its third, after quartz, so the plagioclase makes up less than half of the feldspar and clearly more than a tenth. The Woodbury stone is therefore a granite in the strict sense of the scheme, and not only in the quarryman's sense, in which every hard crystalline building stone is a granite.

Check Your Understanding **Question 4**

A dark rock from a lava flow contains feldspar crystals 1 cm long in a groundmass whose crystals cannot be seen with a lens. What cooling history does the texture record?

(A) Fast cooling at the surface followed by slow cooling after burial

(B) Steady slow cooling at great depth over thousands of years

(C) Slow cooling at depth, during which the large crystals grew, followed by fast cooling at eruption

(D) Cooling in a single step over about a year, the time a 1 cm crystal needs to grow

The solution is at the end of the unit.

5 Intrusions and their field relations

Magma that solidifies below the surface forms an **intrusion**, and the rocks it invades are called the **country rock**. Intrusions come in a few shapes, which Figure 5 draws together, and the shapes are named by their relation to the layers of the country rock. A **dyke** is a sheet of igneous rock that cuts across the layering, usually steep, from centimetres to tens of metres thick and sometimes hundreds of kilometres long; dykes are the cracks through which magma rose. A **sill** is a sheet that was injected between the layers and lies parallel to them. A **laccolith** is a sill that grew thick enough in the middle to dome the layers above it into a lens. A **pluton** is any large body of coarse-grained igneous rock, and a **batholith** is a pluton, or a cluster of plutons, that is exposed over more than 100 square kilometres; the Sierra Nevada batholith of California is 600 km long, and a pluton exposed over a smaller area is called a stock. Plutons are the frozen magma reservoirs of former volcanic regions, exposed after the kilometres of rock above them were eroded away.

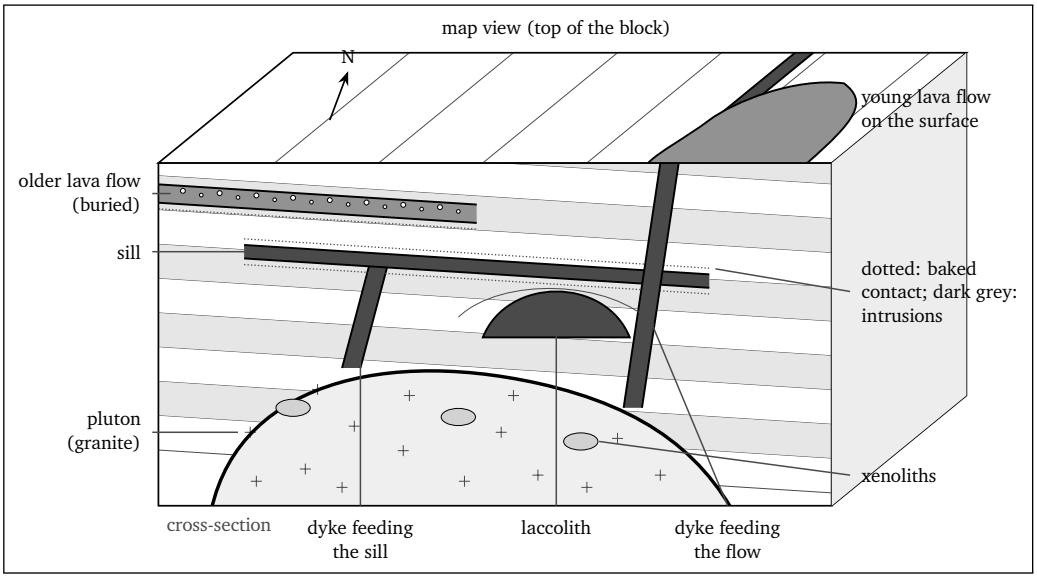


Figure 5. *Intrusive bodies and lava flows in a block diagram, schematic and not to scale. The front face is a cross-section, the top face the map of the same ground. A pluton of granite with xenoliths of the country rock feeds two dykes. One ends in a sill that follows the bedding, the other reaches the surface and feeds a young lava flow; a laccolith domes the beds above it. An older lava flow, now buried, has a baked zone only beneath it and gas bubbles at its top. On the map the dyke appears as a straight band crossing the beds, while the sill is hidden.*

Three features of an intrusion can be read in an outcrop. The first is the **chilled margin**, a rim of fine-grained or glassy rock a few centimetres thick along the edge of a dyke or sill, where the magma froze fast against cold country rock while the interior cooled slowly and grew larger crystals. The second is the **baked contact**, a zone of country rock next to the intrusion that was hardened and recrystallized by its heat, the beginning of the contact metamorphism of Unit 5. The third feature consists of **xenoliths**, fragments of country rock torn off the walls and frozen into the magma, from a few centimetres to the size of a house. All three say the same thing about age: the intrusion is younger than the rock it chills, bakes, or carries as fragments. This is the principle of cross-cutting relationships of Unit 1, which states that a body of rock that cuts across another is the younger of the two, and its companion, the law of included fragments, which states that a rock that contains fragments of another is younger than the rock the fragments came from. In Figure 5 the dykes cut the beds and the sill, so they are younger than both, and the pluton holds fragments of the beds, so it is younger than the beds.

The same features separate a sill from a buried lava flow, which can look alike in a cliff, since both are sheets of fine-grained basalt lying between sedimentary beds. A sill was injected into beds that already existed on both sides, so it has chilled margins and baked contacts above and below, and the beds above it are baked. A lava flow poured out over the ground, so it bakes only the beds beneath it, and the beds above were laid down later on its cold top. The top of a flow is full of frozen gas bubbles and often rubbly or weathered, and

the sediment laid on it may contain fragments broken from that top, whereas a sill is dense from top to bottom and may contain xenoliths of its roof. On a geological map a dyke appears as a straight band that cuts across the outcrop pattern of the beds, because it is a steep sheet that the ground surface cuts through, while a sill appears as a band that follows the beds exactly and can be mistaken for one of them.

Key Result **Reading an intrusion in the field**

A dyke cuts across the layering of the country rock, a sill lies between the layers, and a pluton is a large body of coarse-grained rock. An intrusion is younger than the rock it cuts, bakes, chills against, or carries as xenoliths. A sill bakes and chills against the rocks above and below it; a lava flow bakes only the rocks beneath it, has gas bubbles and a rubbly top, and is covered by sediment laid down on its cold surface. On a map a dyke crosses the beds in a straight band and a sill follows them.

Excursion ***Neptunists against Plutonists***

In the late eighteenth century the most influential geological teacher in Europe was Abraham Gottlob Werner (1749-1817) of the Mining Academy at Freiberg in Saxony, and he taught that all rocks, granite and basalt included, had been precipitated, that is, had settled out as solids, from a primeval ocean that had once covered the Earth. His followers were called Neptunists after the Roman god of the sea. In Edinburgh James Hutton (1726-1797) argued instead that granite and basalt had been injected as melts from the hot interior of the Earth, and his side was called the Plutonists after the god of the underworld. Hutton had his proof on his doorstep. In the cliff of Salisbury Crags, below Arthur's Seat in the middle of Edinburgh, a sill of dolerite lies in sandstones of the Carboniferous period, and at the spot now called Hutton's Section the sandstone at the contact is bent, broken, and baked, and tongues of the igneous rock reach into it. A rock precipitated from water could not have done that, and the sill had to have been forced in as a hot liquid after the sandstone had formed. The Neptunists had one good objection: molten rock cooled in a furnace turns into glass, and basalt is not glass. Hutton's friend James Hall (1761-1832) answered it with experiments that he published in 1805. He melted basalt in the furnace of an iron foundry, let it cool quickly, and got glass; then he let the same melt cool over many hours, and it crystallized into a stony rock like the original. Slow cooling made crystals, as Section 4 explains, and the last argument of the Neptunists was gone.

Shallow intrusions and thick lava flows share one more feature that any visitor to Northern Ireland can see. When a sheet of basalt cools from its two surfaces inward, it contracts, and the contraction sets up a tension that the rock relieves by cracking. The cracks begin at the cooling surfaces and advance into the sheet, step by step, as the cooling front moves inward, and they organize themselves into a network of polygons, because a network of three cracks meeting at angles of 120 degrees relieves the tension with the least total length of crack. The result is **columnar jointing**, a mass of polygonal columns standing at right angles to the cooling surfaces, most of them with five or six sides. Figure 6 shows the most famous example. The columns are not crystals, and the hexagons have nothing to do with the atomic structure of the rock; they are the same pattern as the cracks in drying mud, scaled up. Experiments on basalt from Iceland by Lamur and colleagues, published in 2018, showed that the cracks form at 840 to 890 °C, well below the temperature at which the lava finished crystallizing, so a column is already solid rock when it breaks from its neighbours. Slower cooling makes wider columns, because the tension builds up more gradually and fewer, more widely spaced cracks are enough to relieve it, which is why the columns of a thick flow are half a metre across while those in a thin dyke may be a few centimetres.



Figure 6. *The Giant's Causeway on the coast of County Antrim, Northern Ireland, where about 40,000 basalt columns stand in a lava flow of the Antrim plateau that erupted in the Paleocene epoch, about 60 to 55 million years ago. The columns are about half a metre across and mostly six-sided; the visitors at the left give the scale. Credit: Ajay Suresh, Wikimedia Commons, CC BY 2.0.*

Worked Example **The columns of the Giant's Causeway**

The Giant's Causeway is a flow of the Antrim lava plateau, which erupted about 60 to 55 million years ago (geologists write 60 to 55 Ma, where Ma means millions of years ago) when the North Atlantic began to open. The flow in the cliffs is nearly 30 m thick in places, and about 40,000 columns are exposed. In Figure 6 the visitors are about 1.7 m tall and the columns beside them are about 0.45 m across. The area of a regular hexagon is 0.87 times the square of its width across the flats, so one column covers $0.87 \times (0.45\,\mathrm{m})^2 = 0.18\,\mathrm{m}^2$, so 40,000 columns cover about $7{,}000\,\mathrm{m}^2$, a patch roughly 85 m on each side, which is the size of the headland on which the columns stand. How long did the flow take to cool? Equation (1) with half the thickness, 15 m, gives $t \approx (15\,\mathrm{m})^2/10^{-6}\,\mathrm{m}^2\,\mathrm{s}^{-1} = 2.3 \times 10^8\,\mathrm{s}$, which is about seven years for the middle of the flow to cool through the temperature at which the columns crack, and a few decades for the whole flow to cool to the temperature of the ground. The slow advance of the cooling front over those years is what allowed the cracks to space themselves half a metre apart; a dyke a metre thick, which cools in days, makes columns a few centimetres wide.

The textbook picture of the reservoir under a volcano has long been a cavern of liquid, the magma chamber, which empties in an eruption and finally freezes into a pluton, but that picture has changed. Geophysical surveys under active volcanoes rarely find large bodies of liquid; they find zones of partly molten rock, a few per cent to perhaps a quarter melt, that extend through much of the crust. Petrologists find that the crystals in a single lava have different origins and ages, and that some of them grew tens of thousands of years before the eruption that carried them up. The picture that fits these findings is a crystal **mush**, a framework of crystals with melt in the spaces between them, which cannot erupt as it is because it is too stiff, until melt is squeezed out of it and gathers into a lens that can. The research spotlight presents the review that drew this picture together; a pluton, in this view, is a frozen mush column.

Research Spotlight

Katharine V. Cashman *University of Bristol*

Magma lives in crystal mush that reaches through the whole crust

Katharine Cashman, with Stephen Sparks and Jonathan Blundy, reviewed in the journal Science in 2017 what three decades of observations had done to the classical magma chamber, the large pool of liquid under a volcano that introductory books still draw. They gathered three kinds of evidence. Seismic and electrical surveys of volcanic regions, such as the ones under the Yellowstone caldera and the central Andes, find bodies thousands of cubic kilometres in size through which earthquake waves travel slowly (bodies of low seismic velocity), but with estimated melt fractions of only 2 to 25 per cent, so that the melt sits in a framework of crystals rather than in a chamber. The crystals carried by erupted lavas have compositions that change from centre to rim, which is called zoning, and isotope ratios that differ from grain to grain, which shows that they grew in different parts of the system and were gathered together shortly before the eruption; at Mount St Helens zircon crystals dated the magmatic activity to tens of thousands of years before 1980, while the final assembly of the erupted magma took months to a few years. The dissolved gases trapped in melt inclusions, tiny droplets of melt sealed inside growing crystals, record storage over a range of depths down to the middle crust. From this evidence the authors describe a trans-crustal magmatic system, a column of crystal mush from the lower crust to the volcano in which melt segregates slowly, over thousands to hundreds of thousands of years, and is extracted in episodes into shallow, short-lived chambers that feed eruptions. The pluton of Section 5 is in this view the frozen body of such a mush, and the magma chamber only its uppermost part.

Read the paper: Cashman, Sparks, and Blundy (2017), Vertically extensive and unstable magmatic systems: A unified view of igneous processes, Science 355, eaag3055

Key Points

- Rock melts in three ways: by decompression where hot mantle rises, by the addition of water above a subducting plate, and by heating at the base of thick crust. Decompression makes most magma, and the first melt is always richer in silica than its source.
- Igneous rocks are grouped by silica as ultramafic, mafic, intermediate, and felsic, with boundaries at 45, 52, and 66 per cent; higher silica means a cooler, stiffer, paler, and more water-rich magma, which is why basalt flows and felsic magma mostly solidifies at depth or erupts as ash.
- Bowen's reaction series gives the order in which minerals crystallize from a cooling basaltic magma, olivine and calcium-rich plagioclase first and quartz last, and the removal of early crystals, fractional crystallization, lets one magma produce several rocks.
- Grain size records cooling rate, from glass and aphanitic lava that cooled in hours to years to phaneritic plutons that cooled over thousands of years; cooling time grows with the square of the size, $t \approx d^2/\kappa$.
- A rock is named from its silica group and its texture: basalt and gabbro, andesite and diorite, rhyolite and granite are the same compositions cooled fast and slowly.
- Dykes cut the layering and sills follow it; an intrusion is younger than the rocks it cuts, chills against, bakes, or carries as xenoliths; and columnar jointing records the slow cooling of a thick sheet of basalt.

Solutions to the questions in the text: 1-A, 2-B, 3-B, 4-C. The online version of this unit explains each answer.