

UNIT 3

Magma and Its Properties

Melt, crystals, and gas, and the four controls on how a magma flows

Image: USGS Hawaiian Volcano Observatory

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

- What is a magma made of, and how do its three parts (melt, crystals, gas) change as it rises?
- Why is a rhyolite millions of times more viscous than a basalt at its own eruption temperature, and how much does 100 °C change the viscosity?
- How does dissolved water make a stiff melt runnier, and how does a few percent of water fit into rock at all?
- How much water can a magma hold at 4 km depth, and why does carbon dioxide leave it far deeper than water?
- Why does a lava flow with 50 % crystals behave like a solid, and what gives an ʻaʻā flow its walls?
- How is the viscosity of a flowing lava measured in the field, and what values come out?

A volcano is a machine for moving magma, and everything the machine does depends on what the magma is like: how stiff it is, how heavy it is, how hot it is, and how much gas it holds. Two volcanoes fed by the same amount of melt can behave in opposite ways, one pouring out a river of lava that a person can walk away from, the other blasting the same mass of rock into the stratosphere, and the difference lies in a few properties that can be measured and, to a useful degree, calculated. This unit takes magma apart into its three components, explains what sets its viscosity, density, temperature, and gas content, and shows how each of these is measured, in the laboratory and on a flowing lava.

1 Melt, crystals, and gas

Magma is molten rock below the surface, and it is never a simple liquid, because it is made of three parts, a melt, crystals, and gas. The **melt** is the liquid itself, a hot solution of silicon, oxygen, aluminium, iron, magnesium, calcium, sodium, potassium, and a dozen minor elements. **Crystals** grow in the melt as it cools, so that most magmas arrive at the surface carrying a few percent to half their volume of solid mineral grains, such as olivine and plagioclase feldspar in a basalt or quartz and feldspar in a rhyolite. The **gas** is present in two forms: dissolved in the melt at depth, where the pressure keeps it in solution, and as bubbles once the pressure has fallen enough for it to come out. Water and carbon dioxide make up most of it, and sulfur, chlorine, and fluorine compounds make up most of the rest. The proportions of the three parts change on the way up. A magma that leaves its storage region as a crystal-poor, bubble-free liquid may reach the vent as a froth of bubbles with a quarter of its remaining melt turned to crystals, because cooling makes crystals and decompression makes bubbles, and both happen during ascent.

When magma reaches the surface it is called **lava**, and what is left of the gas escapes into the air. Frozen

lava keeps a record of the three parts. A melt that cools too fast for crystals to grow freezes into **glass**, which is obsidian when it is dense and dark. A melt that freezes while full of bubbles makes **pumice**, a glass foam light enough to float on water, from a silica-rich magma, or **scoria**, a darker and heavier bubbly rock, from a basalt. Crystals that grew before the eruption sit in all of these as visible grains.

KEY RESULT The three parts of a magma

A magma consists of a silicate melt, crystals that have grown in it, and gas that is dissolved in the melt at depth and forms bubbles as the pressure falls. On the way to the surface the share of crystals and bubbles grows at the expense of the melt, and the proportions at the vent decide whether the magma flows or fragments.

1.1 The silicate network

The melt is a liquid, but a strange one. In a basalt some 85 percent of the atoms of the melt are silicon, oxygen, and aluminium, and in a rhyolite 95 percent. Silicon and oxygen bind in a fixed geometry, in which each silicon atom is surrounded by four oxygen atoms at the corners of a tetrahedron, a pyramid with four triangular faces. Two tetrahedra can share one oxygen atom, which then links two silicon atoms and is called a **bridging oxygen**. In pure molten silica every oxygen is a bridge and the tetrahedra form one endless three-dimensional network. To make such a liquid flow, bridges must break and form again, and that costs energy, so molten silica at 1,700 °C is still about a million times more viscous than honey. Figure 1 shows the idea drawn flat.

Aluminium can take the place of silicon in a tetrahedron when a sodium or potassium ion sits beside it to balance the charge, so aluminium counts as part of the network. The other oxides break the network. Calcium, magnesium, iron, and the sodium and potassium that are not paired with aluminium cannot sit inside a tetrahedron, so where they enter the melt a bridge opens and two **non-bridging oxygens** appear, each attached to only one silicon. A basalt has about 50 wt% silica and plenty of calcium, magnesium, and iron, so its network is cut into short pieces that slide past one another easily. A rhyolite has about 75 wt% silica and little of the network breakers, so its tetrahedra are joined in long chains and sheets, and the melt is stiff. This is the reason behind the single most useful rule in volcanology, which Unit 1 states and this unit explains: the more silica a magma has, the more viscous it is. Silica content, to recap, is the mass fraction of silicon dioxide, SiO_2, in the rock, from 45 to 52 wt% in a basalt to above about 69 wt% in a rhyolite.

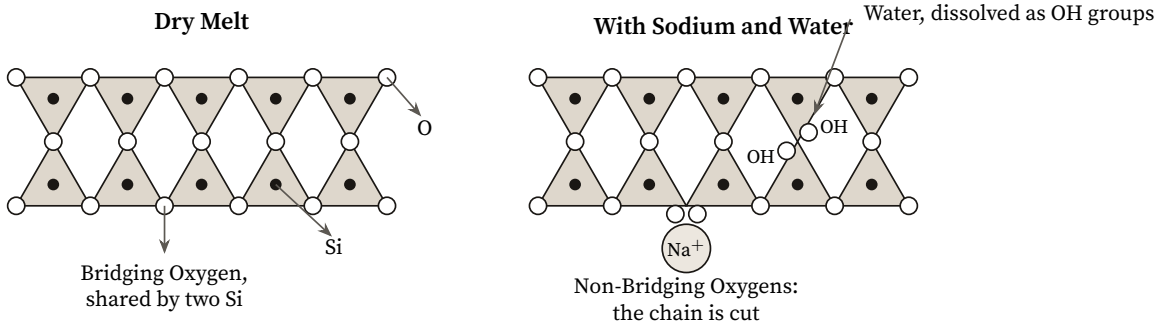


Figure 1. The silicate network of a melt, drawn flat. Each triangle stands for a tetrahedron with a silicon atom (black dot) at its centre and oxygen atoms (circles) at its corners. In a dry, silica-rich melt (left) nearly every oxygen bridges two tetrahedra and the network is continuous. A sodium ion or a dissolved water molecule (right) cuts a bridge into two non-bridging oxygens, which is why both lower the viscosity. A real melt is three-dimensional and disordered; the drawing shows only the principle.

Water does the same job as sodium, and does it better. A water molecule that dissolves in a silicate melt reacts with a bridging oxygen and turns it into two hydroxyl (OH) groups, one on each silicon, and the

bridge is gone. One percent of water by mass is a great many molecules, because a water molecule is light compared with the oxides of the melt (its mass is less than a third of that of a silica unit, SiO_2), so a few percent of dissolved water cuts a rhyolite network into fragments and lowers its viscosity by several orders of magnitude. The reaction has a limit: once the melt holds about 2 wt% of water, most of what is added beyond that stays as whole water molecules between the fragments, as EDWARD STOLPER showed in 1982, which is one reason the first percent does the most work. Section 5 returns to how much water a melt can hold and why it fits into rock at all.

2 Viscosity and its four controls

Viscosity is the resistance of a fluid to flow. Imagine a layer of fluid between two flat plates, and push the upper plate sideways. The force you apply divided by the area of the plate is called the shear stress, and the speed of the upper plate divided by the thickness of the layer is called the rate of shear. The fluid deforms at a rate proportional to the stress, and the viscosity η is the ratio of the two, the shear stress divided by the rate of shear. Its SI unit of measurement is the pascal second (Pa s), because a stress in pascals is divided by a rate in units of one per second. A fluid whose viscosity does not depend on how hard it is pushed is called **Newtonian**, and a crystal-free silicate melt is one, which is what makes the models of this section possible. The values volcanologists meet span more than twelve powers of ten (twelve orders of magnitude, as this is usually said), so they are written as powers of ten, and the model of the next section works with the logarithm of the viscosity rather than with the viscosity itself. Water at 20 °C has about 10^{-3} Pa s, honey about 10 Pa s, a basalt at its eruption temperature between 10 and 1,000 Pa s, which is honey to peanut butter, and a dry rhyolite melt at 800 °C about 10^{11} Pa s, which is stiffer than window putty and closer to a glass that still creeps. Four things set where a magma falls on this ladder.

PRINCIPLES The four controls on viscosity

1. **Temperature.** Viscosity falls steeply as the melt gets hotter. As a rule of thumb it falls by about a factor of ten for every 100 °C, less for a hot basalt and more for a cool rhyolite.
2. **Composition.** The more silica, and the fewer network-breaking oxides, the more viscous the melt. At their own eruption temperatures a basalt has about 10 to 1,000 Pa s, an andesite about 10^3 to 10^5 Pa s, and a rhyolite about 10^6 to 10^9 Pa s.
3. **Dissolved water.** Water breaks the network. A few percent of dissolved water lowers the viscosity of a rhyolite by several orders of magnitude and that of a basalt by less than one.
4. **Crystals.** Suspended crystals raise the viscosity of the mixture, slowly at first and then without limit as their volume fraction approaches about 40 to 60 percent, where the magma locks into a solid. Bubbles can raise or lower the viscosity, depending on how fast the magma is sheared.

2.1 Temperature and composition

The first two controls act together, and the cleanest way to see them is a model that predicts the viscosity of a crystal-free melt from its chemical composition and temperature. The one in general use was published in 2008 by DANIELE GIORDANO, JAMES K. RUSSELL, and DONALD B. DINGWELL, who fitted more than 1,770 laboratory measurements on melts of all compositions, with and without water, to one formula with 18 fitted numbers: one value that every melt shares, the viscosity that all melts approach at very high temperature, and 17 coefficients that carry the composition. Figure 2 shows its curves for four real rocks: a Kīlauea basalt, an andesite and a dacite that serve as reference materials of the United States Geological Survey (rock powders whose compositions have been measured by many laboratories, so that they are known precisely), and the rhyolite of the Bishop Tuff, the deposit of the great eruption of Long Valley in California about 767,000 years ago. Table 1, after the first worked example, gives the

two compositions the examples use.

Three things stand out in Figure 2. First, the curves are ordered by silica at every temperature, with three to four powers of ten between the basalt and the dry rhyolite. Second, every curve falls steeply with temperature, and the slope is steeper for the silica-rich melts, so that the rule of thumb of a factor of ten per 100 °C fits the basalt near 1,150 °C only roughly (the model gives a factor of seven) and understates the rhyolite near 800 °C (a factor of a hundred). Third, the two effects add up in nature, because a rhyolite also erupts 300 to 400 °C cooler than a basalt. A dry rhyolite at its own eruption temperature is therefore a hundred million times stiffer than a basalt at its own eruption temperature, far beyond the thousand to ten thousand times that the silica alone gives at one and the same temperature.

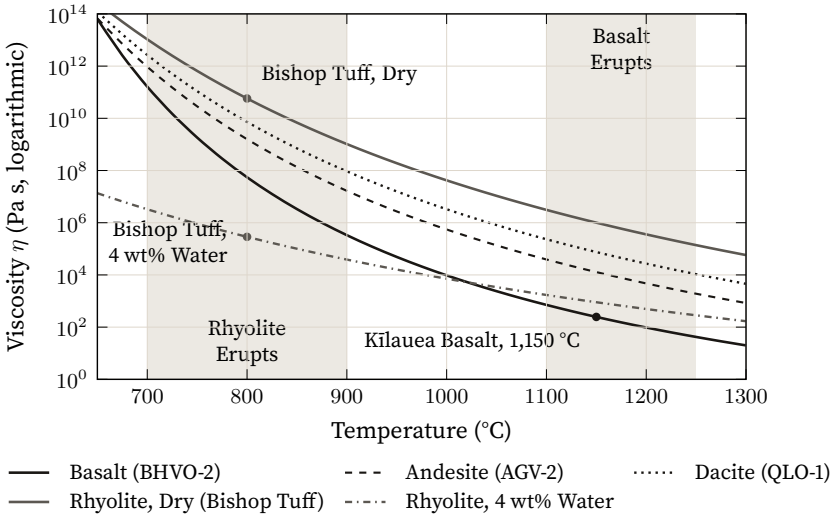


Figure 2. Viscosity of four dry melts against temperature, computed from the model of Giordano, Russell, and Dingwell (2008) for the compositions of Table 1 and for the USGS reference andesite AGV-2 and quartz latite QLO-1, which has the silica content of a dacite. The lowest curve is the same Bishop Tuff rhyolite with 4 wt% of dissolved water. The shaded bands mark the eruption temperatures of rhyolite and basalt, and the dots mark the values of the worked examples. The vertical axis is logarithmic.

WORKED EXAMPLE A Kīlauea basalt and a Bishop Tuff rhyolite

The model of Giordano, Russell, and Dingwell gives the viscosity as $\log_{10}\eta = A + B/(T - C)$, with T the temperature in kelvin, $A = -4.55$ for every melt, and B and C computed from the composition (converted to mole percent, the share of each oxide by number of molecules rather than by mass) with the 17 coefficients of the paper, a recipe this unit does not reproduce; the values below are its results. For the basalt of Table 1 the composition gives $B = 5{,}581$ and $C = 619$, so at 1,150 °C, which is $T = 1{,}423\,\mathrm{K}$,

$$\log_{10}\eta = -4.55 + \frac{5{,}581}{1{,}423 - 619} = 2.39,$$

and the viscosity of the crystal-free basalt melt is about $250\,\mathrm{Pa\,s}$, some 25 times honey. For the dry Bishop Tuff rhyolite the same formula gives $B = 11{,}882$ and $C = 297$, and at 800 °C ($T = 1{,}073\,\mathrm{K}$)

$$\log_{10}\eta = -4.55 + \frac{11{,}882}{1{,}073 - 297} = 10.76,$$

or $6 \times 10^{10}\,\mathrm{Pa\,s}$. The ratio of the two is about 2×10^{8}: the dry rhyolite at its eruption temperature is two hundred million times more viscous than the basalt at its own. The model is good to about a factor of two, so the ratio is an order of magnitude, not a precise value.

Oxide (wt%)	Kīlauea Basalt	Bishop Tuff Rhyolite
SiO_2	49.9	77.4
TiO_2	2.73	0.07
Al_2O_3	13.5	12.4
Iron oxides (as FeO)	11.1	0.78
MnO	0.17	0.04
MgO	7.23	0.07
CaO	11.4	0.51
Na_2O	2.22	3.80
K_2O	0.52	4.89
P_2O_5	0.27	0.01

Table 1. The two compositions used in the worked examples. The basalt is the reference material BHVO-2 of the United States Geological Survey, collected from the pāhoehoe lava that overflowed from Halemaʻumaʻu in 1919 (USGS certificate of analysis). The rhyolite is the early-erupted Bishop Tuff pumice BT-16B analysed by Jolles and Lange (2021), normalized to 100 percent without water.

CHECK YOUR UNDERSTANDING Question 1

A dry rhyolite melt is heated in the laboratory to 1,150 °C, the eruption temperature of a Kīlauea basalt. How does its viscosity then compare with that of the basalt melt at the same temperature?

(A) It is about the same, because temperature is the only control that matters at this level

(B) It is still thousands of times higher, because the silica network remains more connected

(C) It is lower, because a rhyolite contains fewer crystals than a basalt

(D) It is about ten times higher, the usual step of one control on viscosity

The solution is at the end of the unit.

2.2 Dissolved water

The third control is the one that makes silica-rich volcanism possible at all. Figure 2 includes a second rhyolite curve, for the Bishop Tuff melt with 4 wt% of dissolved water, and it lies five powers of ten below the dry curve at 800 °C. Water acts on the viscosity exactly as Section 1 described, by cutting the bridges of the network, and the first percent does most of the work, because it finds the most bridges to cut. The effect on a basalt is far smaller, since its network is already in pieces: 1 wt% of water lowers the viscosity of the Kīlauea basalt at 1,150 °C by a factor of about seven, compared with a factor of 600 for the first percent in the rhyolite.

WORKED EXAMPLE The same rhyolite with its water

Before it erupted, the Bishop Tuff magma held between about 5 and 7 wt% of dissolved water: the composition of the plagioclase crystals that grew in the pumice of Table 1 implies 6.8 wt% (Jolles and Lange 2019, 2021), and droplets of melt trapped inside its quartz crystals, the melt inclusions that Section 5 explains, hold 5 to 6 wt%. The example takes a round 4 wt% rather than the full Bishop Tuff value, because 4 wt% is typical of many rhyolites at the depth where they stall and is the case drawn in Figure 2. Adding 4 wt% water to the composition of Table 1 changes the model's parameters to $B = 10{,}491$ and $C = 25$, so that at 800 °C

$$\log_{10}\eta = -4.55 + \frac{10{,}491}{1{,}073 - 25} = 5.46,$$

which is 3×10^5 Pa s. The water lowers the viscosity by a factor of about 200,000, from 6×10^{10} to 3×10^5 Pa s. The second value is what a rhyolite magma has while it sits in the crust with its water still dissolved. When it rises and the water leaves as bubbles, the melt stiffens by those five orders of magnitude within a few hundred metres of the surface, and Unit 6 follows the consequences.

2.3 Crystals and bubbles

A melt with crystals in it is a suspension, and a suspension is more viscous than its liquid alone because the liquid must flow around the particles. Albert Einstein (1879-1955) showed in 1906 that a small volume fraction ϕ of rigid spheres (the share of the volume that the spheres occupy, so that $\phi = 0.1$ means 10 percent) raises the viscosity by the factor $1 + 2.5\phi$. That formula fails as soon as the spheres are crowded enough to disturb one another's flow, and the extension in use in volcanology, the Einstein-Roscoe relation, which dates from 1952, replaces it by a formula in which the viscosity rises without limit as the fraction approaches the value at which the particles touch.

PRINCIPLE The Einstein-Roscoe relation

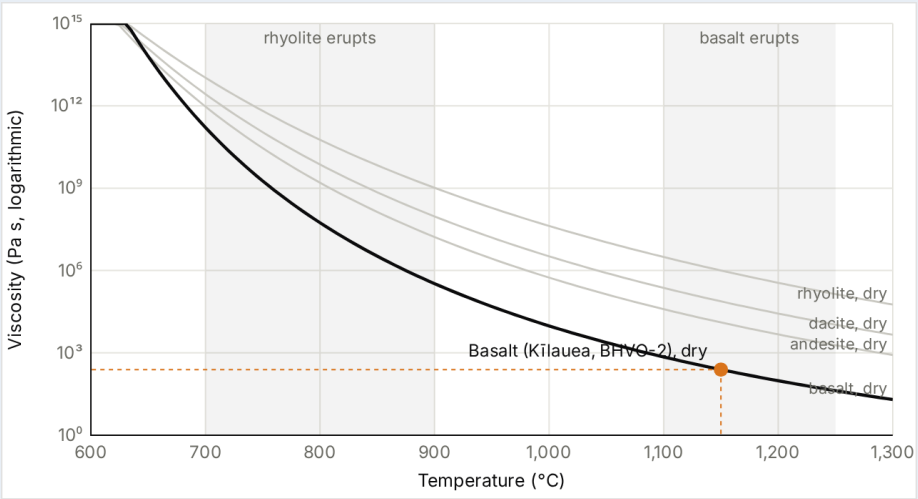
A magma with a volume fraction ϕ of crystals in a melt of viscosity η_0 has the viscosity

$$\eta = \eta_0 \left(1 - \frac{\phi}{\phi_\mathrm{m}}\right)^{-2.5},$$

where ϕ_m is the maximum packing fraction at which the crystals lock together, taken as about 0.6 for volcanic rocks since the work of Bruce Marsh in 1981. The relation is an empirical fit for roughly equant crystals (crystals about as wide as they are long); elongated crystals lock at a lower fraction, and the exponent itself is debated.

Ten percent of crystals raise the viscosity by half, 30 percent by a factor of five or six, and beyond 40 percent the curve turns upward so fast that a magma with 50 to 55 percent crystals behaves as a solid for any practical purpose, which is why lavas with more than about half crystals are almost never seen to flow. Bubbles behave in a more complicated way than crystals, because they can change shape. A bubble that keeps its round shape acts like a rigid particle and raises the viscosity, but a bubble that is stretched by the flow slides along with it and lowers the viscosity instead, and which of the two happens depends on how fast the magma is sheared relative to how fast the bubble can spring back. Field measurements on bubbly Hawaiian lava, in the research spotlight below, show the second case.

INTERACTIVE The viscosity of a magma



Choose one of the four real compositions, then move the temperature, water, and crystal sliders and watch the point travel over the curves of Figure 2. The melt viscosity comes from the model of Giordano, Russell, and Dingwell (2008), the crystals are added with the Einstein-Roscoe relation, and the readout compares the result with a familiar liquid. Notice that the first percent of water does most of the work in the rhyolite and almost nothing in the basalt.

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

WORKED EXAMPLE A basalt with 40 percent crystals

Take the Kīlauea basalt melt of the first example, $\eta_0 = 250\,\mathrm{Pa\,s}$ at 1,150 °C, and let it carry $\phi = 0.4$ of crystals by volume, which is about what a lava has after it has cooled and crystallized for a few kilometres along a channel. With $\phi_\mathrm{m} = 0.6$ the Einstein-Roscoe relation gives

$$\frac{\eta}{\eta_0} = \left(1 - \frac{0.4}{0.6}\right)^{-2.5} = \left(\frac{1}{3}\right)^{-2.5} = 3^{2.5} = 15.6,$$

where $3^{2.5}$ is 3^2 times $\sqrt{3}$, that is 9×1.73. The lava therefore has a viscosity of about $4{,}000\,\mathrm{Pa\,s}$, roughly 16 times its melt. Cooling from 1,150 to 1,100 °C would add another factor of three, from the temperature control, and the two effects together are what turn a fluid pāhoehoe, the Hawaiian lava with a smooth, ropy skin, into a sluggish ‘a‘ā flow, whose surface is a rubble of sharp clinker.

CHECK YOUR UNDERSTANDING Question 2

A basaltic lava cools in a channel until 55 percent of its volume is crystals. With a maximum packing fraction of 0.6, what does the Einstein-Roscoe relation say about it?

(A) Its viscosity is about 2.4 times that of its melt, by Einstein’s rule of 1 plus 2.5 times the fraction

(B) Its viscosity is about 16 times that of its melt, as for the worked example with 40 percent

(C) Its viscosity is several hundred times that of its melt, and the lava is close to locking

(D) Its viscosity cannot be computed, because the relation holds only below 20 percent crystals

The solution is at the end of the unit.

2.4 When lava stops being a liquid

A crystal-free melt is Newtonian, but a lava with crystals and bubbles usually is not. The commonest departure is a **yield strength**, a stress below which the lava does not flow at all, because its crystals have built a loose framework that must be broken before anything moves. A fluid with a yield strength behaves like toothpaste: it holds its shape under a small load and flows under a large one. In a lava flow the stress that drives the lava downhill comes from the weight of the lava above, so it is largest in the deep, fast-moving centre of the channel and smallest at the thin edges. The margins therefore stop while the middle keeps going, and the stopped margins build the walls, called **levées**, that line every ʻaʻā channel. The yield strengths measured on Hawaiian and Etnean lavas are of the order of 100 to 1,000 pascals, tiny by the standards of rock but enough to hold a sheet of lava a few centimetres thick on a gentle slope. Many lavas also become less viscous when they are sheared faster, which is called pseudoplastic, or shear-thinning, behaviour.

RESEARCH SPOTLIGHT

Oryaëlle Chevrel *Laboratoire Magmas et Volcans, Université Clermont Auvergne, IRD*

A century of attempts to measure the viscosity of lava where it flows

Every number in this section was measured in a laboratory, on remelted rock under controlled conditions. Oryaëlle Chevrel, Harry Pinkerton, and Andrew Harris asked what has been measured on real lava while it flowed, and collected every published attempt: eleven, from the iron rod that Trausti Einarsson (1907-1984) pushed by hand into the Hekla lava of 1948 to the motor-driven rotational viscometer that Chevrel refurbished and used on pāhoehoe lobes at Kīlauea in 2016. A penetrometer measures the force needed to push a rod into the lava and suits stiff flows of 10^4 to 10^6 Pa s, but it must reach past the cooled crust, or it measures the crust. A rotational viscometer turns a small paddle, called a vane, inside the lava and records the torque, the twisting force needed to keep it turning. This works below about 10^4 Pa s and shows whether the lava is Newtonian.

The reliable values for Hawaiian basalt at 1,130 to 1,146 °C fall between about 260 and 1,500 Pa s, and the Kīlauea lobes of 2016 gave 380 Pa s at 1,144 °C with 16 percent crystals and 50 percent bubbles. The authors also computed the value for comparison: from the composition of the glass they obtained the melt viscosity with the model of this section, then added the crystals and treated the bubbles as deformable, which gives 220 Pa s, close to what the instrument read. Stiffer lavas scatter from 5×10^4 to 1.5×10^6 Pa s at Hekla, and above 10^6 Pa s no field instrument exists at all.

Read the paper: Chevrel, Pinkerton, and Harris (2019), Measuring the viscosity of lava in the field: A review, Earth-Science Reviews 196, 102852

3 Density and buoyancy

Magma rises because it is lighter than the rock around it, and it stops where it is not. The density of a melt depends on its composition more than on anything else. A basaltic melt, rich in heavy iron and magnesium, has about 2,600 to 2,700 kg/m³ at its eruption temperature, and a rhyolitic melt about 2,300 to 2,400 kg/m³, which is a little more than twice the density of water, 1,000 kg/m³. The rocks they rise through are denser: the mantle rock in which basalt forms has about 3,300 kg/m³, solid basalt about 3,000, and the granite and gneiss of the continental crust about 2,700. The pressure these rocks exert at a depth z is the lithostatic pressure of Unit 1, $P = \rho g z$ with ρ the rock density and $g = 9.81\,\mathrm{m\,s^{-2}}$, which comes to about 27 MPa for every kilometre of crust and 100 MPa at a depth of about 4 km.

The force that drives a melt upward is the difference between the weight of the rock it displaces and its own weight. For each cubic metre of basaltic melt in the mantle it is $(3{,}300 - 2{,}700)\,\mathrm{kg\,m^{-3}} \times 9.81\,\mathrm{m\,s^{-2}} \approx$

5,900 N, a force equal to the weight of 600 kg, which is why melt leaves its source rock as soon as there is a connected path. In the crust the margin shrinks, because the same basalt in granitic crust of 2,700 kg/m^3 is barely buoyant, and where the crust is lighter still, or the melt has grown heavier by losing its light components to crystals, the magma reaches a **level of neutral buoyancy** and spreads sideways instead of rising. This is one reason, though not the only one, why basalt collects in reservoirs a few kilometres below the volcanoes it feeds. A rhyolite melt at 2,300 kg/m^3 is lighter than any crustal rock and would rise freely if its viscosity allowed.

KEY RESULT The density of magma and its buoyancy

A basaltic melt has about 2,600 to 2,700 kg/m^3 and a rhyolitic melt about 2,300 to 2,400 kg/m^3, both lighter than the rocks they rise through (mantle about 3,300, crust about 2,700 kg/m^3). The difference in density times g is the buoyancy force per cubic metre that drives magma upward; where the difference vanishes the magma stops and spreads. Crystals make a magma heavier and bubbles make it much lighter.

The density of a magma with crystals and bubbles is the volume-weighted mean of its parts. Olivine has about 3,300 kg/m^3 and plagioclase about 2,700, so crystals raise the density of a basalt by a few percent at most, but steam at 10 MPa, about 400 m down, has only about 20 kg/m^3, so a magma that is 30 percent bubbles has only 70 percent of the density of its melt. At 75 percent bubbles, the fraction at which a magma breaks into fragments, it has a quarter of the density of its melt, which is less than the density of water. A froth of magma in a conduit therefore gains buoyancy as its bubbles grow, which is why it accelerates upward so violently.

4 The temperature of magma

Basalt erupts at about 1,100 to 1,250 °C, andesite at about 950 to 1,100 °C, dacite at about 800 to 1,000 °C, and rhyolite at about 700 to 900 °C. The order follows the silica content, because silica-rich melts form at lower temperatures, by partial melting of crust or by the crystallization of a basalt, as Unit 2 touches on. The ranges are wide because every volcano has its own history, and a value such as “basalt erupts at 1,200 °C” is a textbook average, not a measurement. The measurements come in four kinds, three made on the hot lava and one on the cooled rock.

A **thermocouple** is a junction of two metals that produces a voltage depending on its temperature. Pushed into flowing lava on a steel rod and left until it stops rising, it gives the most direct value there is, and the Hawaiian Volcano Observatory still takes such readings where a flow can be approached. A **pyrometer** measures the brightness of the glow at one wavelength and converts it to a temperature with the physics of the next section, from a safe distance. A **thermal camera** does the same for every pixel of an image in the infrared, usually at wavelengths of 8 to 14 micrometres, and is the workhorse of monitoring, with one caveat: it sees the surface, and the surface of a lava is always a cooling crust, so a thermal image of a flow shows mostly values far below the interior temperature and only a few hot cracks near it.

The fourth kind reads the temperature from the rock after the event: the composition of the glass between the crystals, and of pairs of minerals that grew together, depends on the temperature at which they formed in ways that laboratory experiments have calibrated, so a quenched sample, one cooled too fast for its glass and minerals to change, carries its own thermometer. The Hawaiian Volcano Observatory uses the magnesium content of the glass in this way for every Kīlauea eruption, and its values for the 2020 to 2023 eruptions were 1,144 to 1,170 °C.

EXCURSION *The first lava thermometer at Kīlauea*

Nobody had put a thermometer into molten lava before 1911; the few earlier figures were optical estimates of the glow of the surface. In that year Frank Perret (1867-1943), an American engineer turned volcanologist, and Ernest S. Shepherd (1879-1949), a chemist of the Geophysical Laboratory in Washington, came to Halemaʻumaʻu, where a lava lake then boiled at the bottom of the pit. They strung a cable 365 m long across the crater, hung their instruments from it, and lowered them into the lake near a persistent fountain that the observers called Old Faithful. The lava destroyed the first sets of equipment, and two electrical resistance thermometers failed, but in July 1911 a platinum thermocouple survived long enough to read 1,010 °C, the first temperature of liquid lava ever recorded. Shepherd's account of the measurement was printed in the first report of the Hawaiian Volcano Observatory, which Thomas Jaggar (1871-1953) founded the following year. Jaggar took up the problem in 1917 with a cruder tool: steel pipes holding Seger cones, the small ceramic pyramids of the porcelain industry that slump at known temperatures, pushed down into the lake. He found that the lake was not uniform, with about 1,000 °C near the surface, lower values a metre or two below it, and about 1,170 °C at the bottom. A century later the glass thermometers of the observatory give 1,140 to 1,190 °C for Kīlauea's lavas, and the thermocouple remains the instrument against which every other method is checked.

4.1 The colour of glowing lava

Every object radiates light whose spectrum, the amount it gives off at each wavelength, depends on its temperature, which is the law that Max Planck (1858-1947) found in 1900 and the astronomers' standard way of taking the temperature of a star. Two consequences of Planck's law are all this unit needs. The wavelength at which the radiation is strongest is inversely proportional to the absolute temperature, by Wien's displacement law, $\lambda_{\max} T = 2.898 \times 10^{-3}\ \mathrm{m\,K}$, where the unit is a metre times a kelvin, and the total power radiated by each square metre of surface grows with the fourth power of the absolute temperature, by the Stefan-Boltzmann law, σT^4 with $\sigma = 5.67 \times 10^{-8}\ \mathrm{W\,m^{-2}\,K^{-4}}$. Both laws hold for an ideal radiator, a black body; a lava surface radiates about 95 percent of that, a fraction called its **emissivity**, and a thermal camera must be told this correction.

For lava at 1,150 °C, which is 1,423 K, the peak lies at 2.0 micrometres, in the near infrared, well beyond what the eye can see. What the eye sees is the short-wavelength tail of the spectrum, and because that tail steepens with temperature, the colour of the glow is a thermometer that works without any instrument. A surface that has just begun to glow, visible in the dark as a faint red, is at about 525 °C; dull red means about 700 °C, cherry red about 900, orange about 1,000 to 1,100, and yellow about 1,100 to 1,200 °C, with the whole glow brightening as the fourth power of the temperature along the way. Figure 3 shows the scale at work on a lava lake, and the interactive element below computes the colour and the radiated power from Planck's law for any temperature a reader chooses.



Figure 3. The lava lake of Erta Ale, Ethiopia, photographed in 2001. The surface is a thin, dark crust of cooled basalt, torn into plates by the convection beneath it; the cracks between the plates and the small fountains glow yellow where lava at about 1,150 °C is exposed, and fade through orange to dull red as a new crust forms in seconds. Credit: Hervé Sthioul, CC BY 2.5.

WORKED EXAMPLE What a square metre of Kīlauea lava radiates

In 2016 a thermocouple in a pāhoehoe lobe of Kīlauea read 1,144 °C, which is $T = 1{,}417\,\mathrm{K}$ (Chevrel and colleagues 2018). Wien's law puts the peak of its radiation at

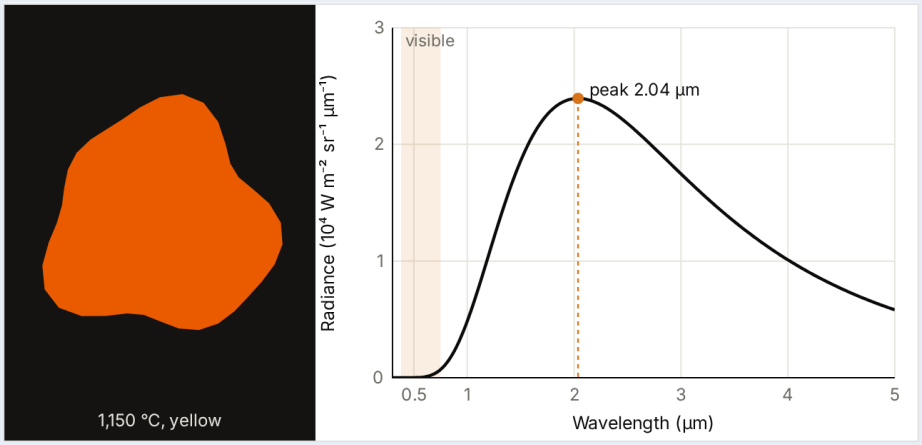
$$\lambda_{\max} = \frac{2.898 \times 10^{-3}\,\mathrm{m\,K}}{1{,}417\,\mathrm{K}} = 2.05 \times 10^{-6}\,\mathrm{m},$$

about 2 micrometres, in the infrared. The Stefan-Boltzmann law gives the power radiated by each square metre of the incandescent surface, with the emissivity of 0.95 of a basalt,

$$\frac{P}{A} = 0.95\,\sigma T^4 = 0.95 \times 5.67 \times 10^{-8}\,\mathrm{W\,m^{-2}\,K^{-4}} \times (1{,}417\,\mathrm{K})^4 = 2.2 \times 10^{5}\,\mathrm{W\,m^{-2}},$$

so a square metre of fresh lava radiates about 220 kW, which is 160 times the roughly 1.4 kW that sunlight delivers to a square metre above the atmosphere. This is why a lava lake is visible from orbit at night and why a flow front can be felt on the face from fifty metres away. It is also why the surface cools so fast: at this rate a crust a centimetre thick loses the heat it holds in a minute or two, and a thermal camera pointed at the same lobe reads a few hundred degrees for most of its surface, with the interior value appearing only in the cracks.

INTERACTIVE **The colour of glowing lava**



Move the temperature slider from 500 to 1,250 °C and watch the glowing patch turn from black through dull red and orange to yellow while the Planck spectrum beside it grows and its peak moves toward the visible band. The readouts give the peak wavelength from Wien's law and the radiated power per square metre from the Stefan-Boltzmann law, and the buttons set the slider to temperatures measured at Kīlauea.

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

CHECK YOUR UNDERSTANDING **Question 3**

On a thermal image of a lava flow, the crust reads 500 °C and a crack into the interior reads 1,150 °C. By roughly what factor does the crack radiate more power per square metre than the crust?

(A) About 2.3 times, the ratio of the two Celsius temperatures

(B) About 5 times, because the radiated power grows with the square of the temperature

(C) About 12 times, because the radiated power grows with the fourth power of the absolute temperature

(D) About 120 times, because the radiated power grows with the fourth power of the Celsius temperature

The solution is at the end of the unit.

5 Water and carbon dioxide in the melt

The gas of a magma is not a separate ingredient at depth. It is dissolved, molecule by molecule, in the melt, as carbon dioxide is dissolved in a bottle of sparkling water before the cap is removed. The amount that a melt can hold, its **solubility**, depends on the pressure above all, and it is the fall of pressure on the way up that turns dissolved gas into bubbles. Section 1 explained how water enters a silicate melt: it reacts with the network and is bound into it chemically as hydroxyl groups, and whole water molecules join them once the melt holds a few percent. That chemical bond is what lets a few percent of water fit into rock at all, and the higher the pressure, the more of both forms the melt takes up. At the pressures of the upper crust, up to a few hundred megapascals, the dissolved water content of a melt grows roughly with the square root of the pressure.

KEY RESULT The solubility of water in a melt

The mass fraction of water C that a melt can hold in solution at the pressure P is approximately

$$C \approx s\sqrt{P}, \qquad s \approx 0.41\ \mathrm{wt\%\,MPa^{-1/2}} \text{ for a rhyolite,}$$

The unit of s looks odd, but it only makes C come out in wt% when P is entered in megapascals: at $P = 100$ MPa, $\sqrt{P} = 10$ and $C = 0.41 \times 10 = 4.1$, so a rhyolite holds about 4 wt% water at 100 MPa, about 4 km down, and about 6 wt% at 200 MPa. A basalt holds less, about 3 wt% at 100 MPa. The square-root law is a simplification of the laboratory measurements, good to a few tenths of a percent below about 200 MPa; the constant and the exponent depend on the melt and, weakly, on its temperature.

Figure 4 draws the law with the pressure increasing downward, so that the surface is at the top of the diagram and the depth can be read on the right. The shape of the curve carries the whole story of an explosive eruption. A melt that holds exactly as much water as it can is **saturated**, and one that holds less is undersaturated. A rhyolite with 4 wt% water is undersaturated at 200 MPa, where the melt could hold 6 wt%, so nothing happens as it rises until it reaches the depth where the curve passes through 4 wt% and the melt becomes saturated. From that depth on, every further metre of ascent forces water out of solution, slowly at first and then faster, because the curve is steepest near the surface: of the 4 wt% a rhyolite holds at 100 MPa, half is still dissolved at 25 MPa, about 1 km down, and nearly all of it has left by the time the pressure has dropped to one megapascal.

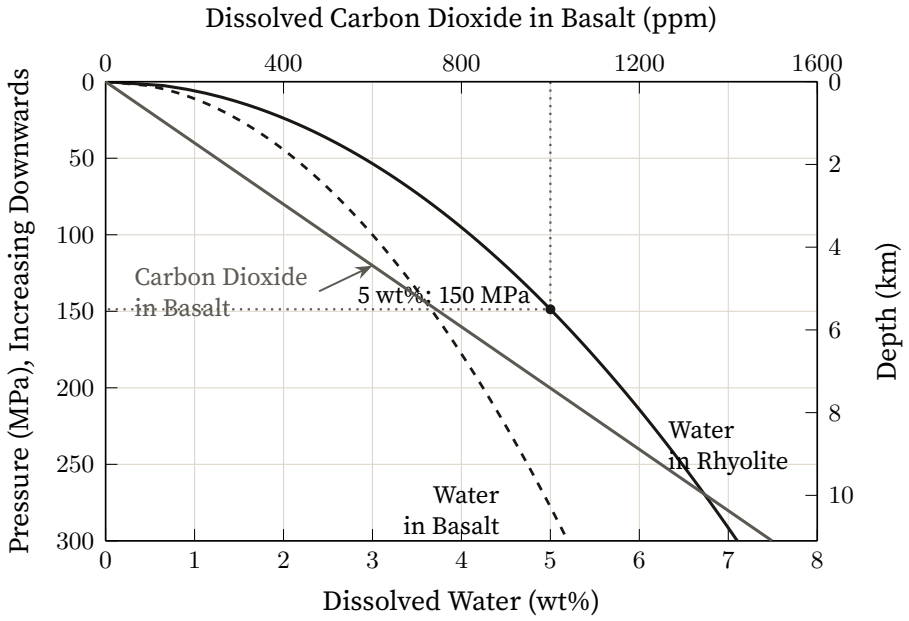


Figure 4. Dissolved water in a rhyolite melt (solid) and a basalt melt (dashed) at saturation, from the square-root law with $s = 0.41$ and $0.30\ \mathrm{wt\%\,MPa^{-1/2}}$, and dissolved carbon dioxide in a basalt (grey, upper scale) from Henry's law at about 5 ppm per MPa after Dixon, Stolper, and Holloway (1995). Pressure increases downward, and the right-hand scale converts it to depth at 27 MPa per kilometre. The dotted lines mark the saturation point of a rhyolite with 5 wt% water, the case of the worked example.

WORKED EXAMPLE The depth of the first bubble

A rhyolite magma holds 5 wt% of dissolved water, within the range of the Bishop Tuff. The

solubility law, solved for the pressure, gives the pressure at which this melt is just saturated:

$$P = \left(\frac{C}{s}\right)^2 = \left(\frac{5\,\text{wt\%}}{0.41\,\text{wt\%}\,\text{MPa}^{-1/2}}\right)^2 = 149\,\text{MPa}.$$

With the lithostatic conversion of 27 MPa per kilometre this is a depth of $149/27 = 5.5\,\text{km}$. Above that depth the magma is oversaturated, so the first bubbles of water vapour appear at about 150 MPa, roughly 5.5 km down. A magma that is stored shallower than this is already bubbly in its reservoir, and the bubbles push it toward the surface, which is one reason why water-rich magmas are so hard to keep underground.

MECHANISM IN DETAIL **Why the first bubble appears at a definite depth**

The pressure in a rising magma falls with the depth as $P = \rho g z$, a straight line, while the water the melt can hold falls as $\sqrt{P}$, a curve. A magma carries a fixed amount of water, C_0, upward. As long as C_0 is smaller than $s\sqrt{P}$ the water stays in solution, and the condition $C_0 = s\sqrt{P}$ holds at exactly one pressure, $P_{\text{sat}} = (C_0/s)^2$, and so at exactly one depth, $z_{\text{sat}} = P_{\text{sat}}/(\rho g)$. Above it the melt can no longer hold what it carries, and the excess, $C_0 - s\sqrt{P}$, must leave as vapour. Because the solubility goes with the square root, doubling the water content quadruples the saturation pressure: a rhyolite with 2 wt% water saturates at 24 MPa, less than 1 km down, and one with 6 wt% at 214 MPa, about 8 km down. The excess grows fastest near the surface, where $\sqrt{P}$ falls most steeply, and that is where the bubbles grow fastest too.

CHECK YOUR UNDERSTANDING **Question 4**

A rhyolite magma holds 6 wt% of dissolved water. Using the solubility law of the unit with $s = 0.41$ wt% per $\text{MPa}^{1/2}$ and 27 MPa per kilometre, where do its first water bubbles form?

(A) At about 15 MPa, less than 1 km below the surface
(B) At about 54 MPa, about 2 km below the surface
(C) At about 100 MPa, about 4 km below the surface
(D) At about 214 MPa, about 8 km below the surface

The solution is at the end of the unit.

5.1 Why carbon dioxide leaves first

Carbon dioxide dissolves in a silicate melt as carbonate groups, and it dissolves poorly. Its solubility grows in proportion to the pressure, which is **Henry's law**, the rule that the amount of a gas a liquid can hold is proportional to the pressure of that gas, and the constant is small: a basalt holds only about 500 parts per million by mass at 100 MPa, about one sixtieth of the water the same basalt holds at that pressure. The grey line in Figure 4 shows it.

PRINCIPLE **Henry's law for carbon dioxide**

The mass fraction of carbon dioxide that a melt can hold is proportional to the pressure,

$$C_{\text{CO}_2} \approx k\,P, \qquad k \approx 5\,\text{ppm}\,\text{MPa}^{-1} \text{ for a basalt,}$$

so that a basalt holds about 500 ppm at 100 MPa and about 1,000 ppm at 200 MPa (Dixon, Stolper, and Holloway 1995). The constant depends on the melt composition, and above a few hundred megapascals the growth is no longer linear.

The consequence is that carbon dioxide leaves a magma far deeper than water does. The basalt that feeds Kīlauea arrives from the mantle with several thousand parts per million of carbon dioxide; the estimate of Gerlach and colleagues in 2002, from the gas output of the summit and the rate of magma supply, is about 0.7 wt%, or 7,000 ppm. A melt with even 2,000 ppm is saturated at 400 MPa (2,000 ppm divided by 5 ppm per MPa), which is about 15 km down, so Kīlauea's magma, with several times that, is making carbon dioxide bubbles before it has left the mantle. By the time it reaches the summit reservoir a few kilometres below the caldera, nine tenths of its carbon dioxide has come out of solution and travels with it as gas, which the summit then breathes out. The water, with its far greater solubility, stays dissolved until the last kilometre. A volcano therefore breathes carbon dioxide from magma that is still deep and water from magma that is nearly at the surface, and the proportion of the two in the gas at the vent tells a volcanologist how deep the magma that is degassing lies. Sulfur, chlorine, and fluorine dissolve in between, and sulfur dioxide is the gas that satellites can measure from orbit; the measurement of these gases and what their fluxes reveal is beyond the Foundation units.

WORKED EXAMPLE The steam in a cubic metre of rhyolite

A cubic metre of rhyolite melt at 2,300 kg/m³ with 4 wt% of dissolved water contains $0.04 \times 2{,}300 = 92\,\mathrm{kg}$ of water, which as a liquid would fill a bucket of 92 litres. Suppose the magma reaches the surface at 850 °C, which is 1,123 K, and all of that water comes out as steam at atmospheric pressure, 0.1 MPa. Water has a molar mass of 18.0 g per mole, so 92 kg is $92{,}000/18.0 = 5{,}110\,\mathrm{mol}$, and the ideal gas law $PV = nRT$ with $R = 8.314\,\mathrm{J\,mol^{-1}\,K^{-1}}$ gives

$$V = \frac{nRT}{P} = \frac{5{,}110 \times 8.314 \times 1{,}123}{1.0 \times 10^{5}\,\mathrm{Pa}}\,\mathrm{m^3} = 477\,\mathrm{m^3}.$$

One cubic metre of magma makes about 480 cubic metres of steam, enough to fill a large house, and about 5 cubic metres even at 10 MPa, some 400 m below the surface. The rock cannot stay in one piece around an expansion of that size; it tears into pumice and ash. This is the reason explosive eruptions exist, and Unit 6 follows the expansion step by step.

5.2 What a crystal remembers

How does anyone know how much water the Bishop Tuff held, or how much carbon dioxide reaches Kīlauea, when the gas has long since escaped? The answer is the **melt inclusion**: a droplet of melt, a few tens of micrometres across, that was trapped inside a growing crystal and sealed off from the rest of the magma (Figure 5). The crystal acts as a pressure vessel, so when the magma erupts and the droplet quenches to glass, it keeps the dissolved water and carbon dioxide it had when the crystal closed around it, and an analysis of the glass with a fine beam of electrons or ions, a microbeam analysis, gives the gas content of the magma at depth. With the solubility laws of this section the volcanologist can then run the first-bubble example backwards and find the pressure, and so the depth, at which the crystal grew. Quartz crystals of the Bishop Tuff hold inclusions that have kept the water of a rhyolite magma for 767,000 years, and the roughly 4 wt% of water that the basaltic magmas of volcanic arcs carry on average, which Unit 2 discusses, was measured in hundreds of inclusions from volcanoes around the world. The method has a well-known trap: as the inclusion cools, the glass shrinks more than its host and a bubble opens in the droplet, and much of the carbon dioxide moves into that bubble. When Penny Wieser and her colleagues measured both the glass and the bubble of inclusions from the 2018 eruption of Kīlauea, they found that the bubbles held most of the carbon dioxide, so that storage depths computed from the glass alone had been too shallow.

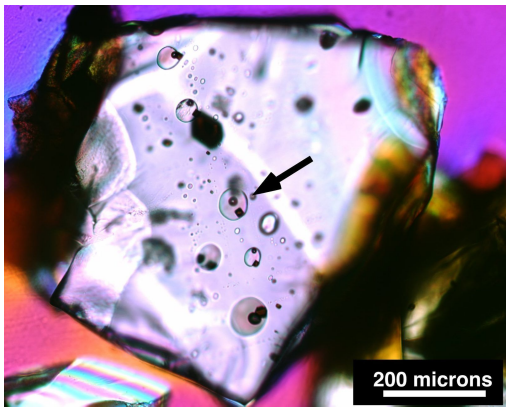


Figure 5. Melt inclusions in an olivine crystal, seen in transmitted light under the microscope. Each inclusion is a droplet of former melt, now glass, with a small round vapour bubble that opened as it cooled; the arrow marks a good example, and the scale bar is 200 micrometres. Credit: Wikimedia Commons user Meltinclusions, public domain.

6 The four magma types

The properties of this unit sort themselves by silica, and Table 2 collects them for the four magma names of Unit 1, with the values the later units use. Every entry is a range, because every magma has its own temperature, water content, and crystal load, and a reader who has followed the unit can reconstruct any entry from the composition, the temperature, and the two laws.

Property	Basalt	Andesite	Dacite	Rhyolite
Silica (wt%)	45 to 52	57 to 63	63 to about 69	above about 69
Eruption temperature (°C)	1,100 to 1,250	950 to 1,100	800 to 1,000	700 to 900
Melt viscosity (Pa s)	10 to 10^3	10^3 to 10^5	10^5 to 10^7	10^6 to 10^9
Melt density (kg/m^3)	2,600 to 2,700	about 2,500	about 2,400	2,300 to 2,400
Typical water (wt%)	0.1 to 1 (arcs about 4)	2 to 5	3 to 6	3 to 7

Table 2. The four magma types and their properties at eruption. The silica boundaries follow the total alkali-silica convention, with basaltic andesite (52 to 57 wt%) left out; the dacite-rhyolite boundary depends on the alkalis too. Viscosities are for melts at their eruption temperatures with the water they still hold at the vent; a rhyolite with all of its 4 to 6 wt% still dissolved is nearer 10^5 to 10^6 Pa s, as the second worked example shows, and dry, cooling, and crystal-rich lavas reach 10^{12} Pa s. The water contents are ranges from melt inclusion studies; ridge basalts hold 0.1 to 0.2 wt%, ocean island basalts 0.5 to 1 wt%, and arc basalts about 4 wt% on average.

The two ends of the table are the two kinds of volcano. A basalt is hot, fluid, and poor in water, so its gas escapes gently through a liquid that flows out of the way. A rhyolite is cool, holds several percent of water, and turns from a flowing liquid into something close to a solid as that water leaves, so the gas cannot escape and tears the magma apart. The andesites and dacites in between can do either, and often do both in one eruption. Where each type comes from, and why basalt is by far the most common, is the subject of Unit 4.

KEY POINTS

- A magma is a silicate melt with crystals and gas, and the share of crystals and bubbles grows as it rises and cools. Silica makes the melt viscous because silicon-oxygen tetrahedra link into a network, which water and the other oxides cut.

- Four controls set the viscosity: temperature (roughly a factor of ten per 100 °C), silica (basalt 10 to 1,000 Pa s, rhyolite 10^6 to 10^9 Pa s at eruption), dissolved water (several orders of magnitude in a rhyolite), and crystals, by the Einstein-Roscoe relation $\eta = \eta_0(1 - \phi/\phi_{\mathrm{m}})^{-2.5}$, which locks a magma near 40 to 60 percent crystals.
- Magma rises because its melt (2,300 to 2,700 kg/m^3) is lighter than rock (2,700 to 3,300 kg/m^3); bubbles lighten it further, and where the density difference vanishes it stops.
- Basalt erupts at 1,100 to 1,250 °C and rhyolite at 700 to 900 °C, measured with thermocouples, thermal cameras, and the composition of quenched glass; the colour of the glow follows Planck's law, from faint red at 525 °C to yellow above 1,100 °C.
- Dissolved water follows $C \approx s\sqrt{P}$, about 4 wt% at 100 MPa in a rhyolite, and leaves in the last few kilometres; carbon dioxide follows Henry's law with only about 500 ppm at 100 MPa in a basalt and leaves tens of kilometres down. One cubic metre of rhyolite with 4 wt% water makes about 480 cubic metres of steam, which is why explosive eruptions exist.

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