CHAPTER 4: IGNEOUS ROCKS

Most rocks in the Fells are igneous rocks. This chapter will introduce you to specific types of igneous rocks and how they form. For more detail on the formation of igneous rocks I suggest Winter (2010) and Jerram (2021).
   **(Note: Terms in red and italics appear as entries in the companion glossary.)**

Igneous rocks have several important characteristics that give us clues about where, when, and how they have formed. Knowing the chemical constituents of the rock, i.e., what minerals are in it, and whether it is intrusive or extrusive tells us about how and where magma was generated, the magma's flow as a liquid, and the conditions (temperature, pressure, water content, depth, etc.) under which the magma crystallized. Another important consideration about igneous rocks in the Fells is how the different units might be related to each other. For example, were they from the same magma chamber or formed at about the same time? We will have to make observations about mineral types, composition, and *textures* (grain size and grain arrangement) to get meaningful information about igneous rocks.

4.1 IS IT IGNEOUS?

One thing that distinguishes most igneous rocks from other rock types is that they are made of silicate and aluminosilicate minerals (see section 3.3.1) that form *interlocking crystals*. It is unusual for other types of minerals to be abundant, except for some secondary oxides and sulfides. The *interlocking texture* means that the mineral grains grow against each other and at least partly surround each other, fitting together like the pieces of a jigsaw puzzle (Fig. 4.1). Many times, we can see flat boundaries on some of the mineral grains where they are in contact with each other. These are crystal faces, and they represent the shape of the mineral grains as they grew in the magma (see Fig. 2.5). These relationships may be difficult to see in an outcrop or in hand specimens unless the mineral grains are coarse, but it becomes much clearer in sawed rock slabs or in microscope views (see Figs. 2.2-2.4). A hand lens is very helpful in this regard. Although igneous rock recognition is not difficult, it takes a little practice, especially in the Fells, where igneous rocks can be hard to identify in the field. To get useful information, we will have to go beyond simply identifying a rock as igneous and learn some specific igneous rock types.

Figure 4.1 – Thin section view with crossed polarizers showing interlocking crystals of pyroxene (colorful grains), biotite (dark brown) and plagioclase feldspar (white to black striped grains) in a coarse-grained igneous rock. See also Figure 2.5.
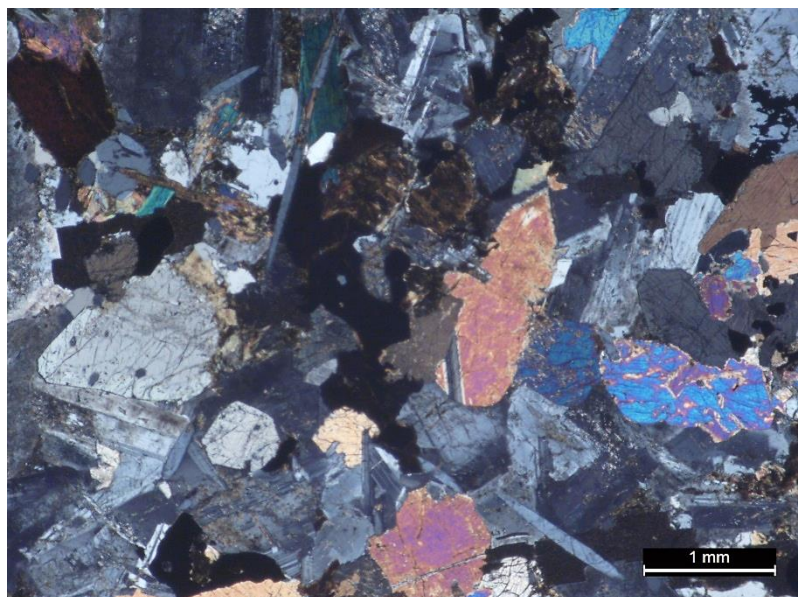
In Chapter 2, Figure 2.5 represents a highly idealized version of the crystallization of an igneous rock. This theoretical rock contained three minerals, each of which began to crystallize at a different temperature, resulting in an interlocking texture as the magma cooled and its composition changed. In real magmas it is not quite so simple. The first mineral to form may become unstable later as temperature falls. At lower temperatures, or perhaps when surrounded by magma that has a changing composition as ions are extracted to form new minerals, some of the earlier crystals may react with the magma and get *resorbed*. This *resorption* process will result in crystals with rounded edges or rounded holes resembling Swiss cheese (Fig. 4.2A). Alternatively, a crystal could react with *hydrothermal* (hot water) fluids at lower temperatures, which can alter a crystal's composition as some ions get replaced and are used to make new minerals, or as water reacts with the mineral. These reactions may preserve the original crystal shape but replace it with a mineral of a different composition or crystal structure, giving the new crystal a dirty or impure appearance. This is called *corrosion*, and it is common in igneous rocks of the Fells where rocks experienced later periods of heating and metamorphism. In the Fells, plagioclase feldspar is commonly corroded (Fig. 4.2B).
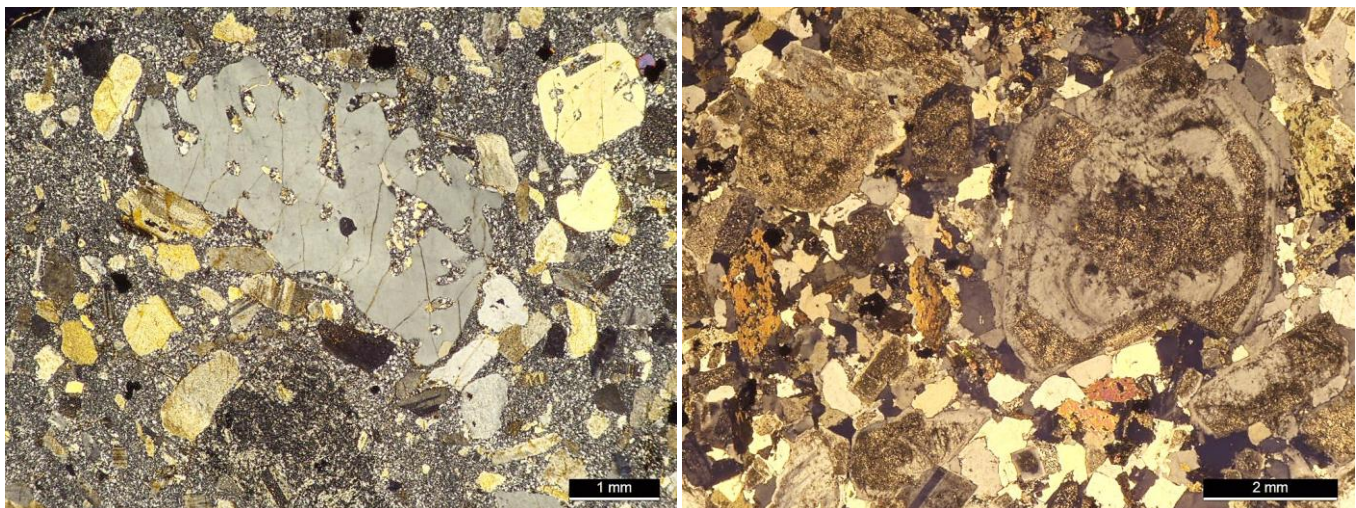

Figure 4.2 – *Thin section views with crossed polarizers of resorbed and corroded crystals. (left) Volcanic rock with a rounded and deeply embayed, resorbed quartz crystal. This crystal was resorbed into a magma by melting or chemical reactions. (right) Corroded plagioclase feldspar crystals with fuzzy interiors made of fine-grained muscovite mica (also called sericite). After the crystals formed, they reacted with the surrounding magma or fluids. The reaction of the feldspar is only detectable in a microscopic view. Note the concentric growth bands in the large plagioclase crystal on the right side of the image as are depicted in Figure 2.5. The orange grains are an altered amphibole called hornblende. The clear grains are quartz.*

4.2 GRAIN SIZE DISTINCTIONS

 A simple classification of igneous rocks in the field (Fig. 4.3) is based on grain size (part of a rock's texture) and its overall color (Fig. 3.1), i.e., whether it is dark (*mafic*) or light (*felsic*). Sometimes, this is all you can do in the field because applying more specific rock names requires a microscopic view of the rock in thin section and/or a detailed chemical analysis and precise mineral identification. *Grain size* is the size of the mineral grains that make up the rock (Fig. 4.4). If you can see individual mineral grains with your naked eye, the rock is considered *coarse-grained* or *phaneritic*. You may be able to see different mineral colors or the reflection of light off cleavage planes when you turn the rock in direct light. Occasionally, we find extremely coarse-grained igneous rocks in which mineral grains are greater than 2.0 cm in size, sometimes reaching several meters across or in length. A rock with this very coarse texture is called a *pegmatite*. If a rock appears to

have a solid color, maybe with the hint of mineral grains, but the grains are so small that you cannot distinguish them without a hand lens, we call the rock *fine-grained* or *aphanitic*. It is also possible for an igneous rock to be non-crystalline (*natural glass*) or so fine-grained that it is not possible to identify crystals, even in a microscope.

Grain size: / Color:	COARSE-GRAINED	Fine-grained	Extremely fine-grained or non-crystalline
Dark (mafic)	GABBRO	basalt	glass
Light (felsic)	GRANITE	felsite	

Figure 4.3 – *Simple field classification of igneous rocks based on just grain size and color.*

Figure 4.4 – *Examples of igneous rocks with different grain sizes and color from pegmatite to glass. Scales in cm. Starting on the left: granite pegmatite (igneous rock with crystals larger than 2 cm), gabbro and granite (coarse-grained igneous rocks), basalt and felsite (fine-grained igneous rocks), and natural glass (obsidian).*

4.3 GRAIN SIZE VS. CRYSTALLIZATION AND COOLING

Grain size tells us something about how igneous rocks formed. For crystals to grow in magma, it takes time for individual ions to migrate to a place where they can attach or assemble themselves onto a growing crystal. If given enough time, crystals can grow to enormous sizes; up to several meters across, in extreme cases. If there is too little time for ion migration, large crystals will not be able to form, and the ions will be forced to locally form a crystal of their own in the magma. This generates many tiny or fine crystals (see Fig. 2.5). So, what controls whether there is time for ion migration and coarse crystal growth? It's mostly about how fast the rock loses heat and cools. If magma loses heat quickly and cools fast, it will have to crystallize fast. Fast crystallization occurs when magma reaches the land surface, or is close to it, and is exposed to cooler rocks, air and water that absorb heat quickly. These are *extrusive igneous rocks,* and they tend to be fine-grained. In extreme cases, where magma cools super-fast, i.e., too fast for significant ion migration, cooling will force a non-crystalline igneous rock or *natural glass* to form. Conversely, when we see a coarse–

grained igneous rock, it represents slower heat loss and cooling. This happens when magma intrudes other rocks deep in the crust, where the intruded rocks are not very different in temperature from the magma. The magma slowly loses heat to the surrounding *host rocks*. This is what generally happens down 10's of kilometers beneath Earth's surface with what are called *intrusive igneous rock* bodies. Rocks that crystallize at great depth tend to be coarser than intrusions that crystallize closer to the land surface from the same magma because of faster rates of heat loss in cooler near-surface environments. There can be a depth gradation between coarse and fine-grained igneous rocks, but there is also a third possibility. Some igneous rocks are composed of coarse crystals surrounded by much finer crystals or glass. This is a *porphyritic texture* (or *porphyry*) and results from an initial stage of slow crystal formation (slow heat loss) at depth followed by a later stage of rapid crystallization (rapid heat loss) in the remaining magma. The second stage of rapid cooling occurs when the partly crystallized magma moves toward the surface or when it is extruded. You will hear more about this below. Thus, the important point is that there can be variations in grain size between intrusions at different depths and variations in a single magma triggered by its crystallization at different depths while it moves toward the surface.

The shapes of intrusions also influence their cooling rate. Intrusive igneous rock bodies formed in large magma chambers, sometimes covering many kilometers, are called *plutons*, and their resulting rocks are called *plutonic igneous rocks* (Fig. 4.5). Intrusions that squeeze into fractures or cracks in the subsurface are called *dikes*, and they end up having planar or tabular shapes (Fig. 4.6). Magmas forming dikes are exposed to relatively large surface area of host rock that can rapidly absorb heat away from a small magma volume, leading to a much higher rate of heat loss for dikes than in large magma chambers. Thus, thin dikes are not as coarse as rocks formed in thicker dikes or plutons from the same magma. This relationship is shown on a crude graph (Fig. 4.7). You can see evidence of more rapid cooling along the contacts of intrusive igneous rocks, especially dikes, where areas near the contact are finer than areas in the interior (Fig. 4.6). Heat was lost quickly along the magma's contact with a cool host rocks and an area of finer grain size called a "*chill zone*" develops.
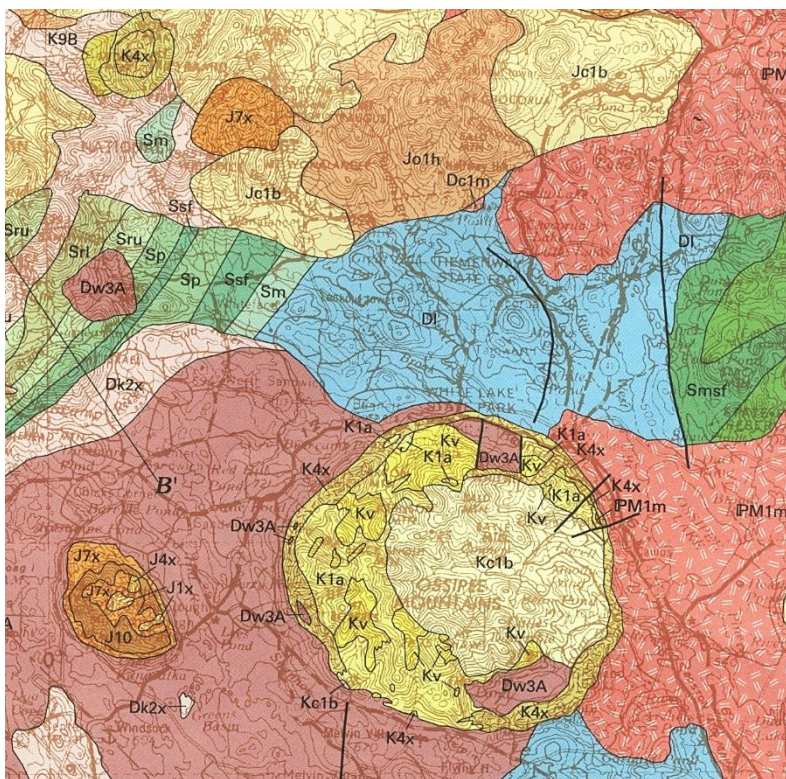

Figure 4.5 – *Plutonic igneous rocks on the Geologic Map of New Hampshire (Lyons and others, 1997). In order from oldest to youngest: the area marked Dk2x is the Kinsman Granodiorite, Dw3A is a quartz and plagioclase intrusion (Winnipesaukee Tonalite), PM1m are bodies of granite that contain both muscovite and biotite (Effington Pluton), Jc1b is the pink Conway Granite, Jo1h is the Mount Osceola Granite, and the units marked K are mostly pink granite and rhyolite of the Ossipee Mountains. The yellow units of the Ossipee Mountains are about 13 km across. Other units (green and blue) are highly metamorphosed sedimentary rock units that served as the host rocks for the intrusions. The capital letters in the abbreviations for the rock units stand for time periods: Silurian (S), Devonian (D), Mississippian (M), Pennsylvanian (P), Jurassic (J), and Cretaceous (K). Heavy black lines are faults.*

Figure 4.6 – (left) A dike of basalt in gabbro at Nahant, Massachusetts. The dike has well defined chill zones on its edges (widths marked with yellow bars), which are very fine-grained basalt. The dike's lumpy interior texture is due to vesicles (bubbles) that had minerals precipitate in them. Rock hammer for scale. (right) A very thin basalt dike cutting across felsic volcanic rock in the Fells. Scale is in cm.
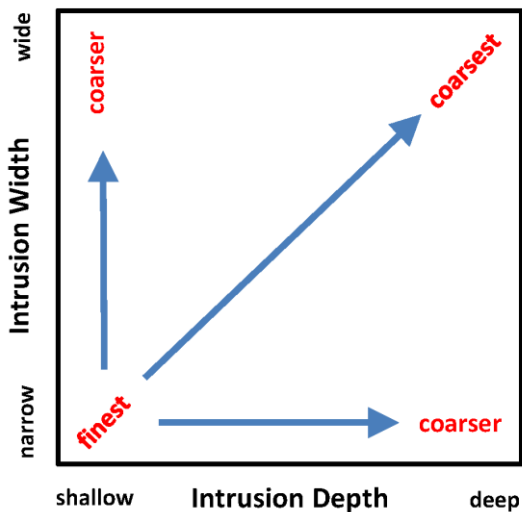

Figure 4.7 – Simple factors controlling the grain size of a body of intrusive igneous rock. These factors control how fast heat is lost from magma and how fast the magma crystallizes. This is a very general depiction used as a starting point. There are other factors that make crystallization and crystal growth rates more complex.

    You may ask: How fast is fast cooling and how slow is slow cooling of magma? In the case of natural glass, extremely fast cooling may take a few minutes. For example, when hot lava is exposed to air, water, or ice it almost instantly causes a crust of natural glass to form. At the other extreme, slow cooling that produces coarse-grained igneous rocks deep in the crust may take half a million years when the host rock has a temperature that is not that much lower than the temperature of the cooling magma. Fast cooling, as occurs in fine-grained igneous rocks, may be on the order of days to 10's of years where the temperatures of surrounding host rock, air, or water bodies are significantly lower than the cooling magma. Overall, the chemical and temperature changes that allow magmas to

crystallize are exceedingly slow chemical reactions as compared to those you do in a high school chemistry lab. In the chemistry lab, the reactions must be faster. Can you imagine completing an experiment that lasted several years let alone thousands, or millions of years? No one would graduate from high school! We should not forget that though heat loss, or the cooling rate, is perhaps the most easily identifiable (and understandable) component in controlling an igneous rock's grain size, there are other factors that make crystallization more complicated. Changes in pressure, differences or changes in the composition of magma, the shape of the magma chamber or spaces that are intruded, and the amount of water dissolved in the magma all play roles in the rate of crystallization.

4.4 COLOR VS. CHEMISTRY

Identifying the color of igneous rocks as either dark or light can be tricky, but, with a little practice, color can be categorized in the field with consistency (Figs. 4.3 and 4.4). Any rock that is black to very dark greenish-gray on a fresh surface is defined as "dark". They can also be referred to as *mafic* rocks, a name that indicates an abundance of mafic or dark-colored minerals. Any other color - including tan, dark red, medium to light gray, or pale to medium green - on a fresh surface is referred to as "light" or *felsic*. The color distinctions are crude, but they have some meaning in terms of rock chemistry and mineralogy. Mafic rocks have higher abundances of iron, magnesium, and calcium and lower concentrations of aluminum and silicon than light or felsic rocks. On the weathered surfaces of mafic rocks, you will often see a rusty color, indicative of limonite, that has formed from the oxidation of iron. The light or felsic rocks have much less iron and are generally richer in silicon, aluminum, sodium, and potassium. We sometimes call these rocks *sialic* rocks after their higher silicon (Si) and aluminum (Al) abundance. The light rocks are also formed from magmas that tend to be much stickier, or exhibit a higher *viscosity*, due to their high silica content. Magmas of high viscosity do not flow as easily and therefore do not as readily form lava flows. You can now guess which magma type dominates in Hawaii and Iceland where you see lava fountaining out of the ground or flowing downhill at high speeds. Light and dark igneous rock types and their chemistry are crudely at two ends of the spectrum in the igneous rock world, but there can be even more extreme cases, such as *ultramafic rocks* composed of more than 90% mafic or dark minerals. There are also lots of minerals, as well as many rock types, in the gray area between mafic and felsic.

4.5 MORE ON PORPHYRIES

So far, we have learned some simple igneous rock names that can be used in the field (Figs. 4.3 and 4.4). However, there are other igneous rock types that we should know something about that can also be identified in the field. We may encounter a rock with coarse crystals surrounded by finer crystals or glass called a *porphyry* (Fig. 4.8). The individual large crystals are called *phenocrysts*, while the surrounding finer material is the *groundmass*. The most common phenocryst mineral is plagioclase feldspar, which usually occurs as white or light gray crystals that often have well defined crystal faces. Crystals that show flat crystal faces are known as *euhedral* grains (Fig. 4.9). Quartz, amphibole, pyroxene, and olivine also commonly occur as phenocrysts. We classify porphyries by the type of groundmass that the rock has, so, for example, the name porphyritic basalt indicates that a groundmass of basalt has phenocrysts in it. Porphyries often tells us that there were two stages of cooling involved in the rock's history, an initial phase of slow cooling where individual coarse crystals slowly grew in the magma, followed by rapid cooling of the remaining magma, thereby encasing the phenocrysts. This may happen to magma that starts cooling slowly in the subsurface, forming crystals that are suspended in the magma ("crystal slush"), and then the entire

magma is forced to an area where it can cool more quickly, as might occur when the magma rises, approaching the land surface, or is extruded.
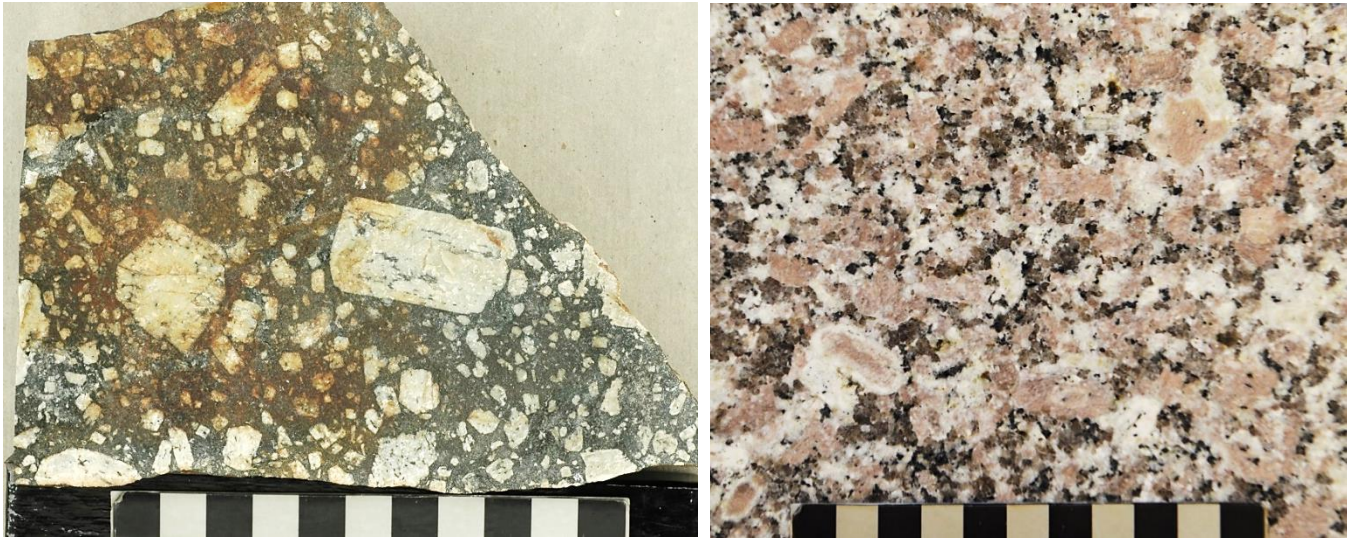

Figure 4.8 – *Porphyries are rocks composed of a ground mass of finer crystals surrounding isolated larger crystals (phenocrysts). Porphyries can be fine or very fine-grained rocks (left) with large crystals in them. They can also have a coarse ground mass (right) with very coarse-grained phenocrysts. On the right is granite with very coarse alkali feldspar phenocrysts (pink) in finer plagioclase feldspar (white) and quartz (gray). Some of the alkali feldspar crystals have rims of white plagioclase, like a Tootsie Pop (right image, lower left). Scales in cm.*

Figure 4.9 – *Euhedral grains in an igneous rock are crystals that have flat crystal faces and outlines preserved. Shown here are euhedral plagioclase phenocrysts in a porphyritic basalt dike. The opaque (black) mineral grain is pyrite, which is not euhedral. Thin section view in plane polarized light. Note also the euhedral phenocrysts in the porphyry images above (Fig. 4.8A).*
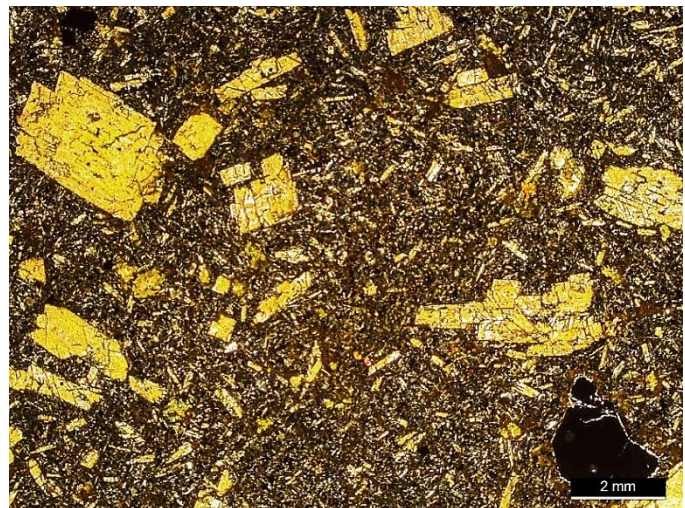
4.6 VARIETIES OF VOLCANIC ROCKS

When magma reaches the surface, it leads to volcanic eruptions that produce fine-grained igneous rocks, which may have either a light or dark color. Although generally fine-grained or porphyritic, these rocks can form in several different ways. One way is for the liquid to spill out on to the land surface to produce *lava flows*. Basaltic lava is more common than felsic ones because basaltic magma is less *viscous* and flows more easily and felsic magmas tend to have explosive eruptions that produce fewer flows. Volcanic rocks can also be deposits composed of particles of magma that have been blown apart and thrown upwards into the air during volcanic eruptions. This volcanic material accumulates on the land surface to produce *pyroclastic volcanic rocks*. For more detail on volcanic rocks and eruptions, I suggest Jerram (2021). There is also a catalog of felsic volcanic rock types, components, and textures in the Fells on the Fells Geology web site: VOLCANIC ROCKS.

4.6.1 Volatiles and Gases in Magmas

In addition to the molten minerals carried in a magma there are also dissolved water and gases of various types, such as sulfur gases and carbon dioxide. Dissolved water and gas constituents are very reactive and together are known as *volatiles*. Magma under high pressure can hold large volumes of dissolved volatiles. As magma rises and pressure on the magma decreases, the volatiles are released from solution as vapors that will form bubbles, much like what happens when you release pressure by opening a soda bottle. Lava with volatiles can behave like a soda bottle that is shaken and loses its dissolved CO_2 gas very rapidly when opened. Evidence of the release of volatiles are the gas bubbles or *vesicles*, that get trapped in solidified magma (Fig. 4.10). Vesicles are common features of lava flows but can also occur in shallow dikes. In extreme cases, where *obsidian* is full of bubbles, a highly vesicular (frothy) glass is formed called *pumice* that resembles a sponge or Styrofoam. This material, even though made from rock, will float because of trapped air bubbles.

Figure 4.10 – (left) Vesicles, or open cavities, that result from crystallization or solidification of bubbly magma. (right) Vesicles in basalt that had minerals grow in them to form *amygdaloidal* basalt. The *amygdules* (brown spots) are now weathering and are heavily oxidized. Basalt dike at East Nahant, Massachusetts.

4.6.2 Ejecta Formation

Rapid depressurization of magma that contains dissolved volatiles can trigger a rapid gas bubble expansion in the magma that leads to an explosive eruption. This causes the most violent, and dangerous, volcanic eruptions. These are the eruptions that make the evening news when they occur in populated areas and force large areas to evacuate, or they interfere with air traffic. Expanding gas blows the magma apart into tiny pieces of glass that are shot skyward. Any crystals that have already formed in the magma also become airborne. In addition, shattered *fragments* of pre-existing intruded rock (*country or host rock*) also become incorporated into the magma. The fragments may be pieces of volcanic rock from earlier eruptions that are blasted apart and sent skyward or pieces of older non-volcanic rock formations in the subsurface that the magma passed

through to get to the surface. All the airborne material, or *ejecta*, is thrown into the atmosphere by the force of the blast and rising heat of the eruption, with the smallest particles reaching highest into the atmosphere and traveling the longest distances. Ejecta are classified by their different sizes as *ash* (less than 2 mm), *lapilli* (2-64 mm), and *bombs* or *blocks* (greater than 64 mm). *Blocks* were solid when they were ejected, while *bombs* were molten material. Near the eruption, material pushed vertically into the atmosphere may later fall back to the land surface and descend the flank of a volcano, creating *ejecta deposits*. This material may simply settle, or it may be deposited by violent currents of ejecta in a hot dense descending cloud, or a jet created by a volcanic blast.

4.6.3 Pyroclastic rocks

When ejecta is deposited, it forms *pyroclastic rocks*. Pyroclastic rocks are a mixture of four components: 1) *natural glass* in the form of broken fragments and *shards* (the shattered, curved walls of glassy magma bubbles); 2) *crystals* that were phenocrysts already developed in the magma at the time of eruption; 3) solid *lithic fragments* (rock fragments); and 4) *molten lava and pumice fragments* (Fig. 4.11). Rocks made of a mixture of ash plus any of the other components is called *tuff* (Fig. 4.12). Glass particles and pumice are sometimes still molten, soft, and sticky when they are deposited. They will adhere to each other and other particles while deforming under the accumulating weight of more ejecta to produce a hard deposit when they cool that is known as *welded tuff*. Pieces of soft pumice can be deposited before they harden, and they may become buried by more ejecta. The load of overlying material on these soft pumice pieces will produce *flattened pumice fragments* or *fiammé*. It's like when you step on a Styrofoam packing peanut or a piece of popcorn. Near an eruption, tuff deposits have a mixture of lots of materials and can be rich in lava lapilli, bombs, and large pumice pieces, but far from an eruption, only very fine volcanic particles will blanket the land surface as a layer of dust-like *volcanic ash* that is usually *non-welded tuff*.

Figure 4.11 *– Types of ejecta. A) Blocks (large ejecta fragments) made of dark banded felsite in pyroclastic rock on Marblehead Neck, Massachusetts. B) Sialic lava bombs in pyroclastic rock with abundant lapilli fragments at Castle Rock on Marblehead Neck. Dimes for scale (in yellow circles). (cont.)*
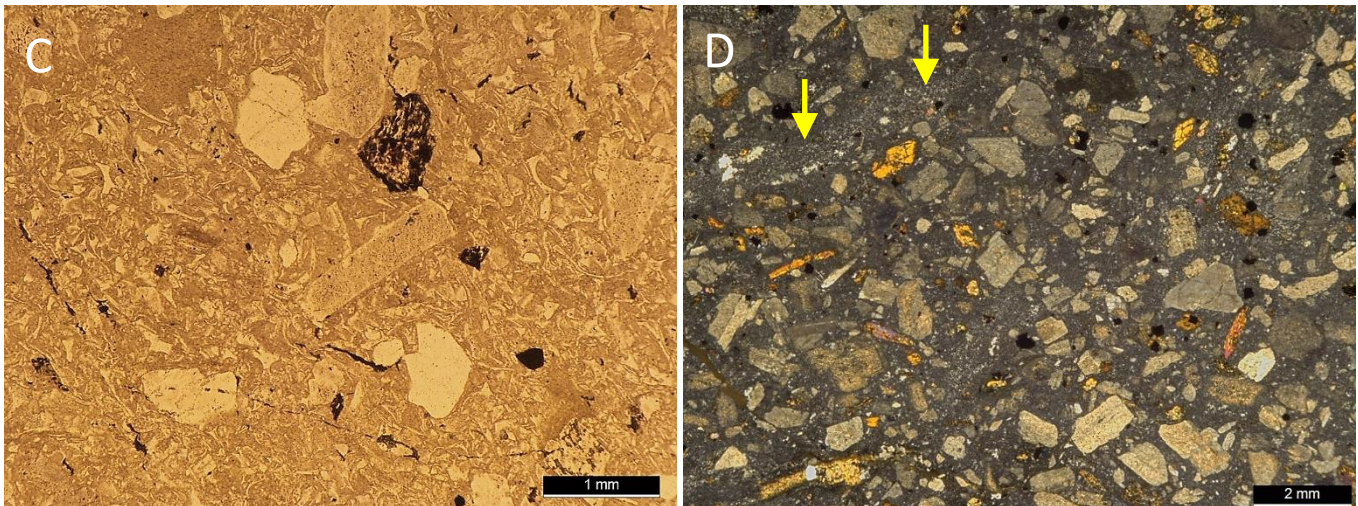
Figure 4.11 (cont.) - Types of ejecta. Both images are views of thin sections. C) Pyroclastic rock from the northern Fells with abundant glass shards showing curved bubble wall structures surrounding broken mineral grains (crystals) and a few lithic fragments. The dark grain is a volcanic lithic fragment. View in plain polarized light. D) Pyroclastic rock (tuff) from the southern Fells with abundant broken crystals of plagioclase and hornblende (orange grains). In the upper left corner (yellow arrows on upper side) is a flattened pumice fragment (area with no crystals). View with crossed polarizers.
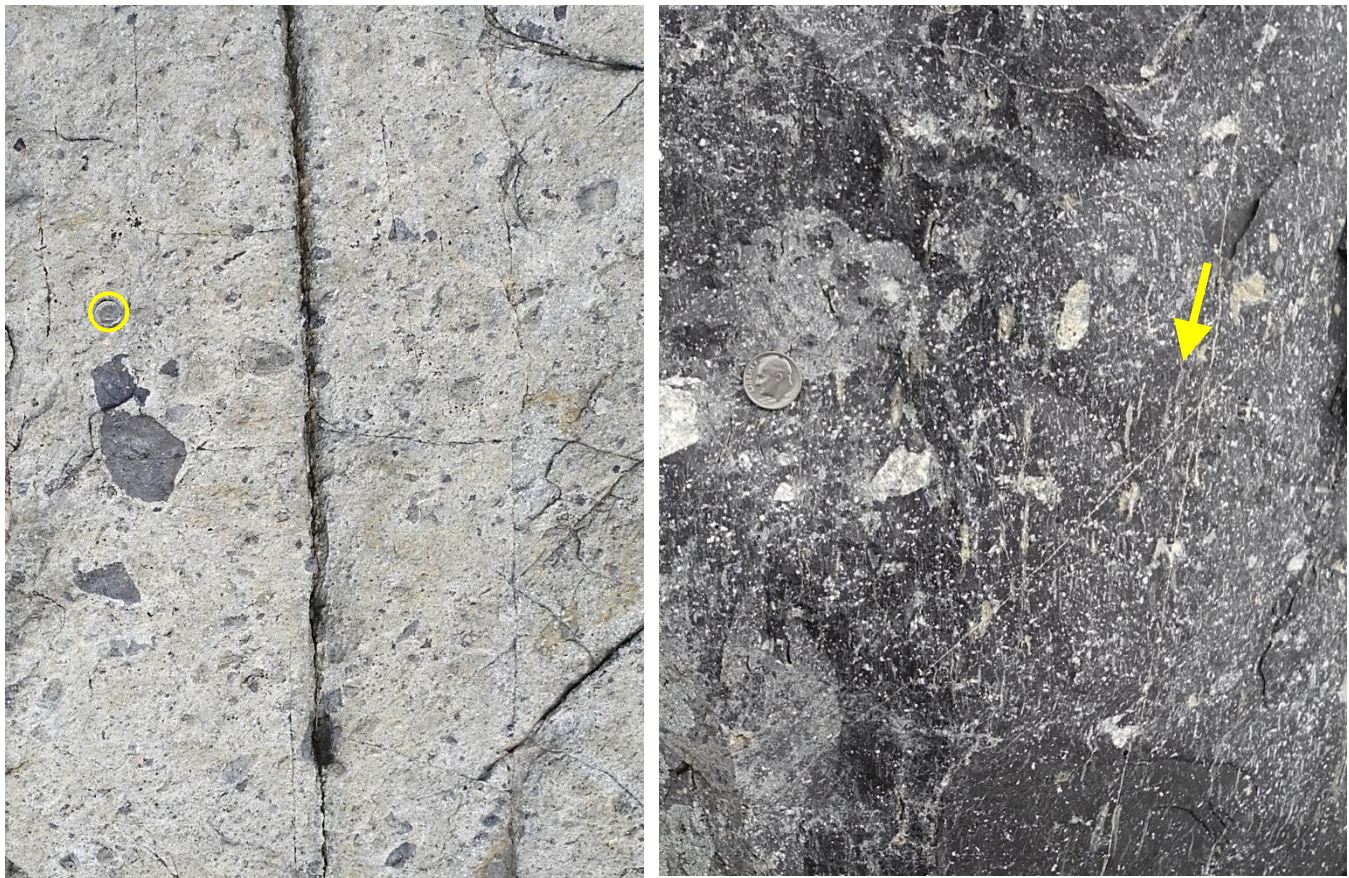

Figure 4.12 – Tuff at Castle Rock on Marblehead Neck. (left) Tuff with volcanic lapilli fragments surrounded by volcanic ash (light gray). Dime for scale (yellow circle). (right) Welded tuff with lapilli fragments, plagioclase crystals (white specks), and flattened pumice fragments that show up as the elongate and pinched gray blobs oriented vertically on the image (arrow). Dime for scale.

As if things were not complex enough already, after the deposition of ejecta, these materials may be eroded and re-deposited by surface processes, possibly along with some other, non-volcanic materials, producing *volcaniclastic sediment* and later *volcaniclastic rock*. This is sedimentary rock dominantly composed of volcanic materials that are eroded and redeposited by surface processes. The particles in volcaniclastic rocks may be more rounded than in pyroclastic rocks, but it is often very difficult when looking at ancient rocks to tell them apart because they can contain many of the same components. Pyroclastic materials that are eroded and moved down the flanks of a volcano by erosional processes to form landslides, debris flows, or mud flows become volcaniclastic deposits at the base of the volcano or in adjacent valleys. This can be a tremendous hazard in volcanic areas as these materials are unstable on steep slopes.

4.6.4 Obsidian

Another rock type we must consider in volcanic settings is the natural glass produced by very rapid chilling of magma. It is not uncommon for light-colored (*felsic*) magmas to cool too quickly for crystals to form, even at a microscopic level. This produces a type of natural glass called *obsidian* (Fig. 4.13). Obsidian and extremely fine crystalline rocks may form from the rapid cooling of sialic lava flows, but they can also form by the accumulation of layers of molten ash particles that weld together to form a thick molten mass. Both these rock types commonly have layering called *banding*. In lava flows, banding results from the flow and mixture of areas in the lava of slightly different composition or oxidation. Banding may also be from the original layering in a deposit of molten ash particles. Welded, sticky, hot ash can start to flow if it is soft and thick enough. Both sialic lavas and layered welded ash can produce marble cake-like banding in the form of highly contorted folds (Fig. 4.13).

Figure 4.13 – (left) Natural glass, in this case folded obsidian, formed by the quenching (extremely rapid cooling) of lava. Obsidian is a common product of the rapid cooling of sialic lava flows or welded ash deposits. Scale in cm. (right) Banded rhyolite on Marblehead Neck in recrystallized (devitrified – see below) obsidian. Dime for scale (yellow circle).

4.6.5 Devitrification

Some very fine-grained volcanic rocks in the Fells started out as obsidian but their texture changed not too long after they formed. Obsidian is an unstable substance over millions of years, especially when exposed to elevated temperatures for an extended time. Even though it appears to already be a solid, the transition from glass to very fine interlocking crystals of quartz and feldspar that are visible in a microscope occurs while the glass is still hot but below the crystallization temperature of these minerals. We call this process *devitrification*, or the crystallization of glass, as opposed to *vitrification*, which is the formation of glass. Deep burial and temperatures and pressures associated with metamorphism can accelerate devitrification. Devitrification makes the rock much harder than if it were still glass. None of the rock units in the Fells are young enough to still have obsidian or any remaining glass; they have all devitrified. However, we can still sometimes see the outlines of the original glass shards, pumice fragments, and banding in these rocks preserved as mineral grains of different sizes and shapes (Figs. 4.11C, 12, and 13).

4.7 INCLUSIONS IN IGNEOUS ROCKS

All igneous rocks may enclose pieces of other rock formations. These enclosed pieces are known as *inclusions* or *xenoliths*. Inclusions are just what the name suggests: they are pieces of host rocks that were surrounded by magma or included in it (Fig. 4.14). Inclusions are common near the edges of an intrusion where pieces of the surrounding host rock broke off into the magma. Inclusions can also occur in extrusive igneous rocks, where they represent pieces of subsurface rock transported to the surface in the rising magma. Inclusions in igneous rocks are always pieces of older rock units than the surrounding rock and help us establish relative ages. Inclusions are investigated further in Chapter 8.
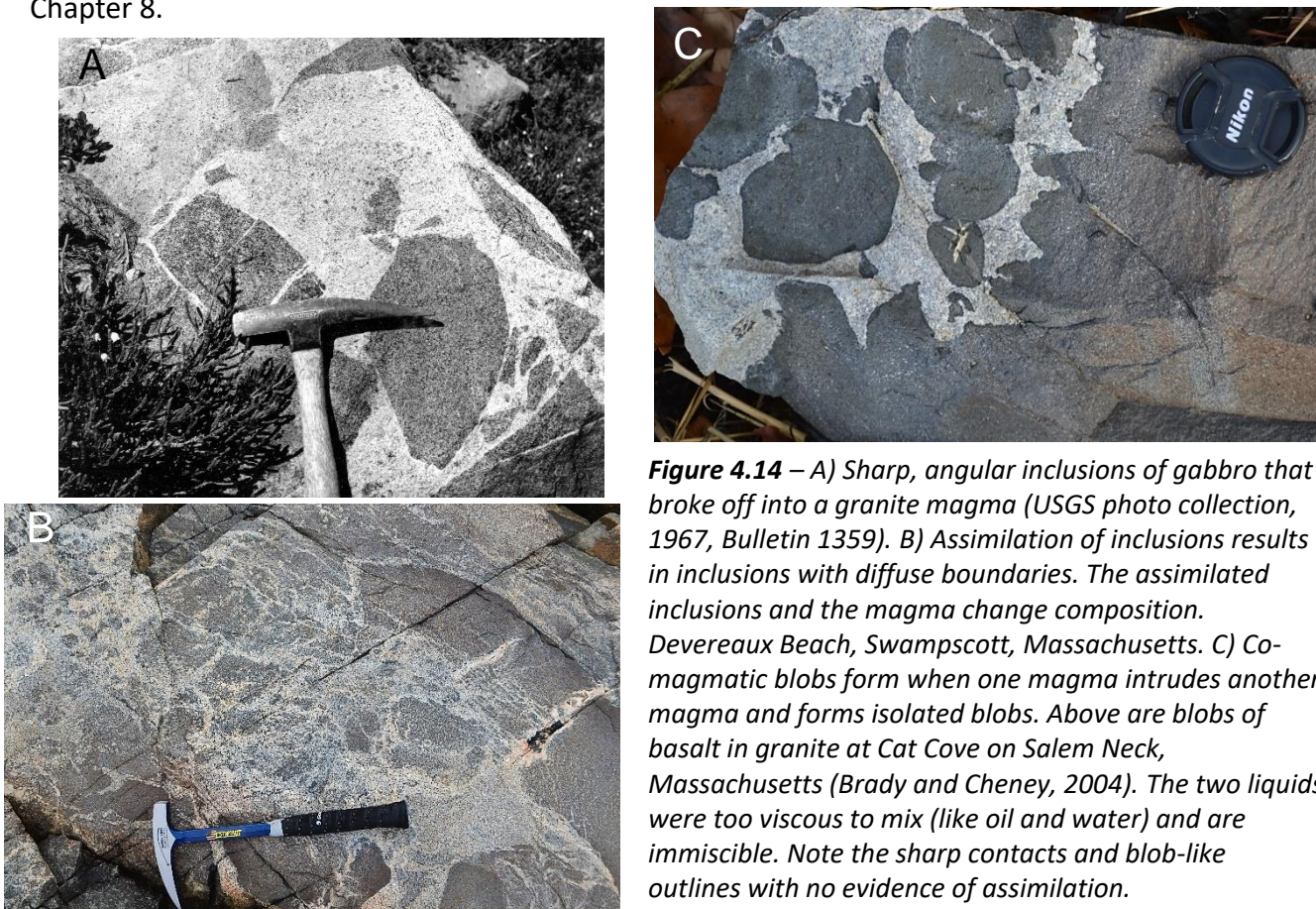
Figure 4.14 – A) Sharp, angular inclusions of gabbro that broke off into a granite magma (USGS photo collection, 1967, Bulletin 1359). B) Assimilation of inclusions results in inclusions with diffuse boundaries. The assimilated inclusions and the magma change composition. Devereaux Beach, Swampscott, Massachusetts. C) Co-magmatic blobs form when one magma intrudes another magma and forms isolated blobs. Above are blobs of basalt in granite at Cat Cove on Salem Neck, Massachusetts (Brady and Cheney, 2004). The two liquids were too viscous to mix (like oil and water) and are immiscible. Note the sharp contacts and blob-like outlines with no evidence of assimilation.

Inclusions can have several forms. They can break off the sides of a magma chamber and drift into the liquid with the inclusion being minimally affected by the very hot surrounding magma (Fig. 4.14A). This is often the case in dikes where the dike does not stay hot long enough to significantly melt or react with inclusions of the host rock. The inclusion will be chemically unchanged and angular with its sides being the flat surfaces of fractures that allowed the inclusion to break off into the magma.

Alternatively, heat from the magma may alter the inclusion chemically, perhaps with some of the chemical constituents of the inclusion undergoing *assimilation* into the melt (Fig. 4.14B). This is when inclusions release ions or melted constituents into the magma. If this happens in a major way, it can alter the composition of the magma. The inclusion will also have an altered composition, and its shape may be rounded as outer layers of the inclusion are removed to or react with the magma. It may deform like taffy as it gets soft. The magma may show evidence of assimilation with streaks of melted material from an inclusion in it after the magma crystallizes. Water and other fluids in fractures in the country rock that mix with the magma can aid this process by speeding up chemical reactions. Water may also absorb or contribute chemical constituents during the assimilation.

A third possibility is that inclusions can be formed as a result of two magmas contacting each other while both are still hot liquids. Two magmas existing simultaneously are said to be *co-magmatic* (Fig. 4.14C). Where two magmas come into contact, the liquids might not mix, and blobs of both magmas may appear in each other with sharp contacts. The liquids have trouble mixing because of how long this may take, and one may have a high viscosity that prevents it from flowing easily. You can see this pattern in a lava lamp where there are blobs of one liquid in another. Mixing of magmas may result in some confusing field relationships where two rocks appear to simultaneously form blob-like inclusions in each other.

4.8 DETAILED CLASSIFICATION OF IGNEOUS ROCKS

Thus far, we have learned a simplified classification of igneous rocks that can be used in the field, and we have learned about some of the structures that can tell us whether igneous rocks are intrusive or volcanic (extrusive). However, professional geologists that study igneous rocks need to classify them in more detail according to their mineral and chemical compositions. Igneous *petrologists*, the scientists that study igneous rocks, may sometimes have enough experience to get suspicious about the exact chemical or mineral composition of an igneous rock in the field. However, the exact composition of an igneous rock is something that is only precisely studied by looking at rocks in a microscope or by obtaining detailed chemical or mineralogical analyses. This is because the color of igneous rocks can vary and may be deceiving when viewed in hand specimen. It is also very difficult to determine the percentages of minerals in igneous rocks when the crystals are very fine and not clearly visible, even in a microscope. When detailed mineralogical or chemical data are available, it is possible to distinguish rocks according to the most widely used classification scheme created by the International Union of Geological Scientists (IUGS).

4.8.1 IUGS Classification of Coarse-grained Igneous Rocks

The most widely used classification of coarse-grained igneous rocks was first formulated in 1979 and is continuously revised by the IUGS (Le Bas and Streckheisen, 1991; Le Maitre and others, 2002). Before this classification scheme, and the development of quantifiable rock names, rock names where sometimes used differently by different petrologists, resulting in confusion. The IUGS classification is based on quantified mineral contents of igneous rocks, or an idealized mineral

content inferred from a detailed chemical analysis. The main minerals used for the classification of igneous rocks are the light-colored or *felsic minerals*, including (with their abbreviations): Q = *quartz*, A = *alkali feldspar* (orthoclase, microcline, and plagioclase that has greater than 95% cations of Na and K), P = *plagioclase* (with cations of Na and Ca in varying percentages), and F = *feldspathoids*. Feldspathoids are the minerals leucite and nepheline, which only occur in rocks that have a low abundance of silicon as compared to most felsic igneous rocks. These rocks have little or no quartz, and they also do not have enough silicon in them to allow the whole magma to crystalize to feldspar. Feldspathoids are not common, and as far as I know, they do not occur in the rocks of the Fells, and we will no longer worry about them in the classification scheme.

To classify a coarse-grained igneous rock using the IUGS scheme, we must know the percentages of the felsic minerals in the rock. Mineral percentages can be estimated using a *point count*. To perform a point count, we need to identify the minerals that appear at the intersection points of a grid over a large enough area to get a statistically meaningful count of different minerals. To do this with a coarse-grained igneous rock, we can place a clear plastic grid over a polished slab of the rock and record the mineral that occurs at each point on the grid, which can then be tabulated into the percentages of different minerals in the rock. In a thin section it is possible to count different minerals at the crosshairs of a microscope view as the thin section is mechanically moved a uniform increment across the field of view (Fig. 4.15).
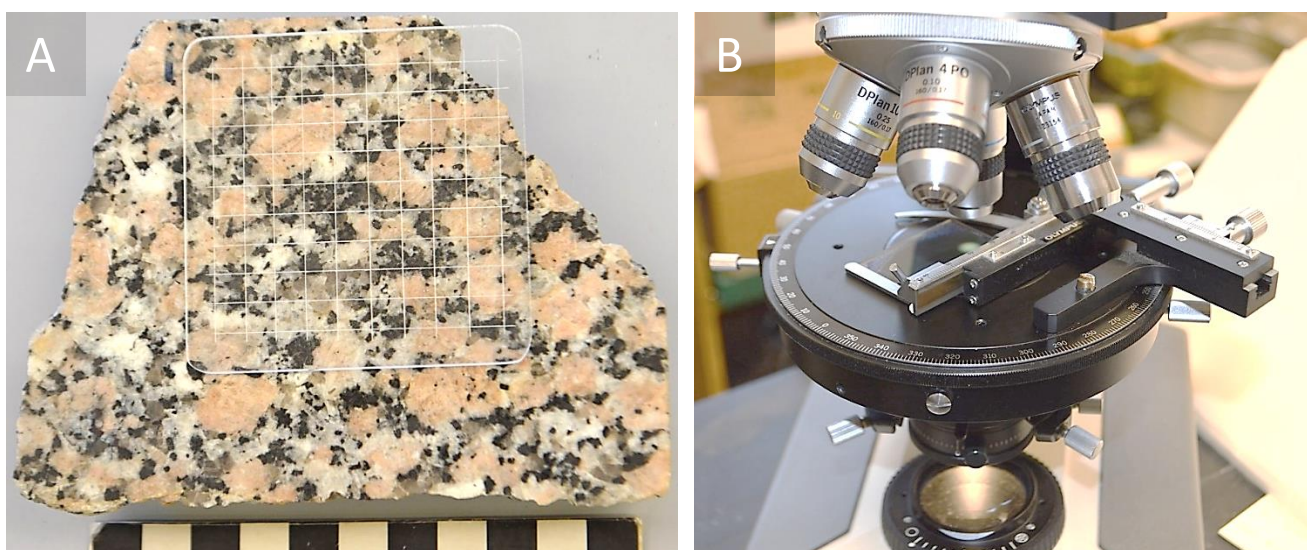

Figure 4.15 *– Point counts are used to record the mineral percentages of rocks by tabulating the mineral types that occur at the intersection points of a grid. Point counting can be done by (A) counting points on a plastic grid overlay of a polished coarse-grained rock surface or (B) using a point counting device mounted on the stage of a petrographic microscope. This device advances a thin section in the field of view of a microscope at a regular increment. If enough points are counted and the rock has a uniform distribution of minerals, both techniques provide accurate estimates of mineral percentages in a rock.*

If there are no feldspathoid minerals in a rock, and mafic minerals are less than 90%, a ternary QAP plot (triangular graph of Q, A, and P percentages only) is used to plot a rock's IUGS composition. A ternary plot is a way of simultaneously plotting the relative percentages of three components that total to 100% (Fig. 4.16). On the IUGS QAP plot of rock compositions (Fig. 4.17), areas with a specific range of composition are given standard igneous rock names (Le Bas and Streckeisen, 1991; Le Maitre and others, 2002). If a rock has more than 90% mafic minerals, a different graph is used to classify the rock (see Le Maitre and others, 2002 or Winter, 2010), which is beyond the scope of information we need for the study of rocks in the Fells. One of the weaknesses

of the IUGS classification scheme is that as mafic mineral percentages become high, the classification is based on mineral percentages (Q+A+P) that could make up less than half of the rock. It would be like ranking baseball teams based on only the players that throw or bat right-handed. This is the reason that, in the lower right-hand corner of the IUGS plot (Fig. 4.17), there are sometimes two or three different rock names for one field on the graph in order to distinguish some mafic-rich rock types.

Figure 4.16 – *(right) A ternary plot of the light-colored or felsic minerals in an igneous rock: quartz (Q), alkali feldspar (A), and plagioclase (P). The composition of a coarse-grained igneous rock is plotted on the graph. The percentages for each sialic mineral are: 50% quartz, 30% plagioclase, and 20% alkali feldspar. (bottom) To plot the composition of the rock, the relative percentage of each mineral component (quartz, alkali feldspar, plagioclase) is read for each mineral between its origin, or corner on the plot (100%), and the axis opposite the origin (0%).*
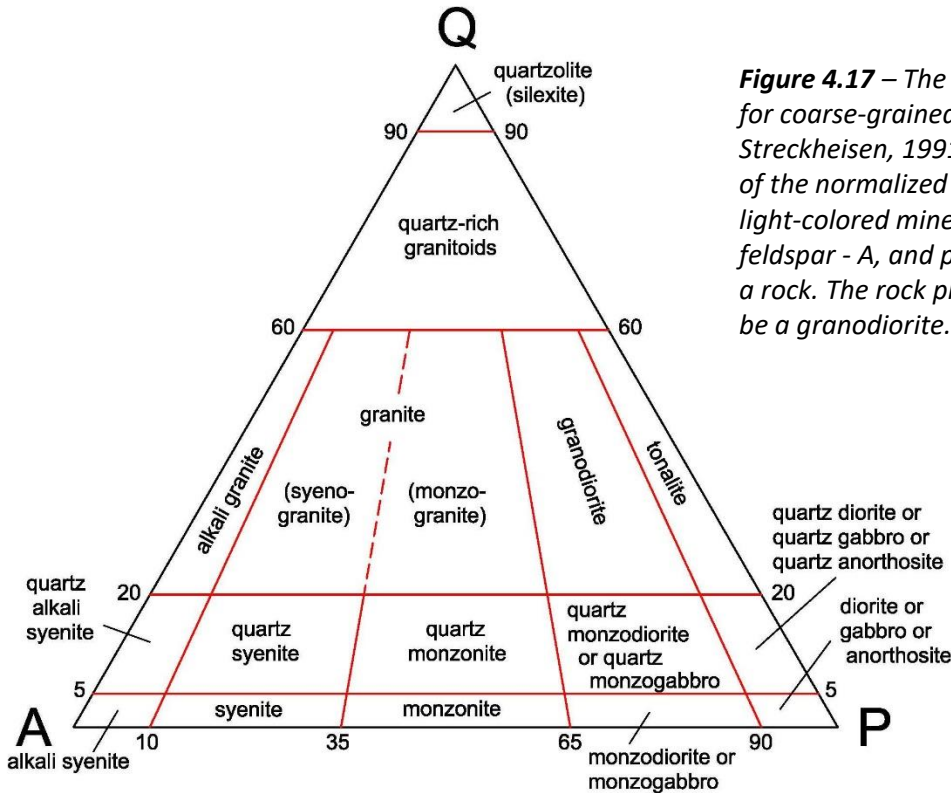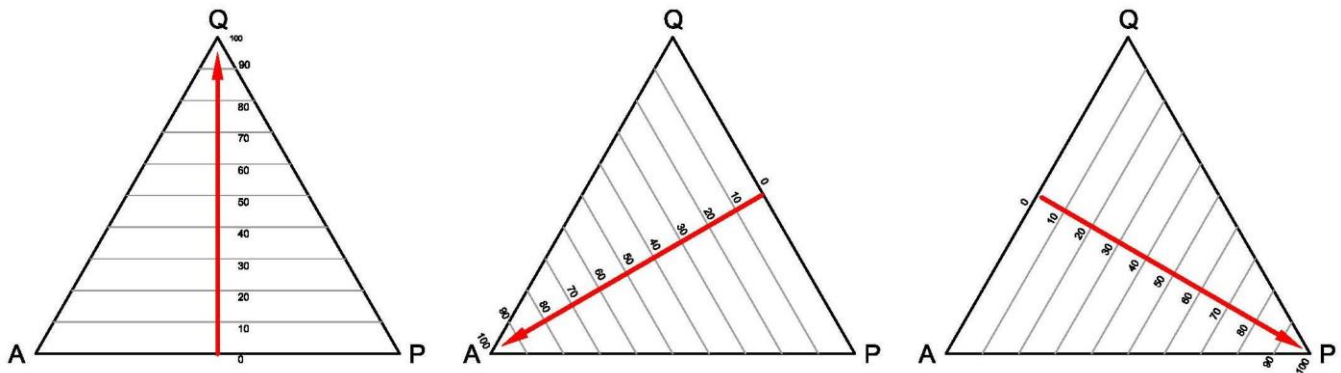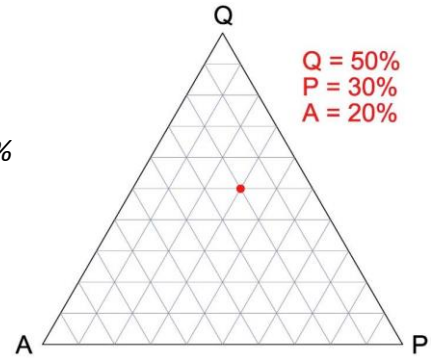

Figure 4.17 – *The IUGS classification scheme for coarse-grained igneous rocks (La Bas and Streckheisen, 1991) is based on a ternary plot of the normalized percentages of felsic or light-colored minerals (quartz - Q, alkali feldspar - A, and plagioclase - P) that occur in a rock. The rock plotted in Figure 4.16 would be a granodiorite.*

To plot a rock sample on the ternary QAP plot, the Q, A, and P percentages for the whole rock must be re-calculated relative to each other. This is called *normalization* of the felsic mineral percentages. Normalized percentages are calculated against the total of all three components (bold letters indicate normalized percentages):

$\mathbf{Q} = [Q/(A+P+Q)] \times 100\%$	An example point count:	Normalization calculations:
$\mathbf{A} = [A/(A+P+Q)] \times 100\%$	44.0% quartz (Q)	Total (T) = Q + P + A = 88.0%
$\mathbf{P} = [P/(A+P+Q)] \times 100\%$	26.4% plagioclase (P)	$\mathbf{Q}$ = Q/T = 44.0% / 88.0% = 50%
	17.6% alkali feldspar (A)	$\mathbf{P}$ = P/T = 26.4% / 88.0% = 30%
	7.9% biotite	$\mathbf{A}$ = A/T = 17.6% / 88.0% = 20%
	4.1% hornblende	(These are the %'s plotted on Fig. 4.16)

Once the QAP percentages are normalized, there are two methods for plotting the position of the rock sample on the ternary QAP graph. On a QAP graph, each corner of the plot represents 100% of the component at that corner and the axis opposite the corner represents 0% of that component. As a starting point, both methods require plotting of the normalized Q% (**Q**). This is done by counting away from the bottom axis vertically toward the Q corner and drawing a horizontal line at this position (Initial Step on Fig. 4.18).

To complete the plot using Method #1 (Fig. 4.18), do the same as above for either the normalized A% (**A**) or normalized P% (**P**), again counting away from the axis opposite the corner for this component. Draw a straight line parallel to the axis opposite this component's corner. The point where the A or P line intersects the quartz line is the rock sample's position on the QAP graph. You can test your answer by determining whether the third component's normalized % is also at the intersection.
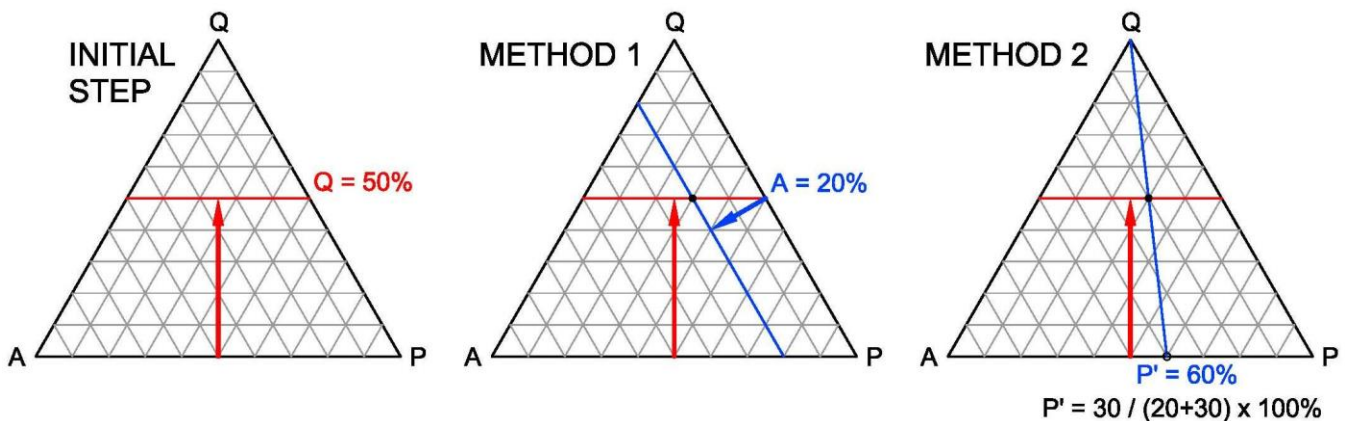

Figure 4.18 – To use the IUGS classification scheme for coarse-grained igneous rocks, normalized percentages of quartz, alkali feldspar, and plagioclase relative to their total must be calculated and plotted by one of two methods. Both methods start with the initial step of plotting the normalized percentage of quartz (*Q* in red). Method 1 is completed by plotting either the normalized percentage of plagioclase (*P*) or, as is shown here, alkali feldspar (*A* in blue). The composition of the rock plots at the intersection of the quartz and feldspar percentage lines. Method 2 is completed by calculating the percentage of plagioclase vs. total feldspar (*P'* = P/(P+A) x 100%). This percentage (*P'*) is plotted along the base of the triangle (*A-P* axis) from 0% at *A* to 100% at *P* and then connected by a line to the *Q* origin (blue). The rock composition is plotted at the intersection of this line (blue) and the quartz percentage line (red). The example on this figure is a plot of *Q* = 50%, *A* = 20%, and *P* = 30% from Figure 4.16.

To complete the plot using Method #2 (Fig. 4.18), calculate the percentage of **P** against just **A + P**:

$$P' = [P/(A+P)] \times 100\%$$

On the bottom axis of the graph (from **A** to **P**), plot this number (**P'**) as a point by counting away from **A** (lower left) corner toward the **P** corner. Connect this point and the **Q** corner with a straight line. Where this line intersects the original **Q** line is where the rock plots on the QAP graph.

If the rock plots in the lower right of the QAP diagram (Fig. 4.17), it will be in a field with multiple possible names. If the rock has greater than 90% plagioclase and less than 10% mafic minerals, it is an *anorthosite*. If it has plagioclase with Ca>Na and more than 10% mafic minerals, it is *gabbro*, and if it has plagioclase with Na>Ca and more than 10% mafic minerals, it is *diorite*. This distinction also applies to rocks that have high alkali feldspar, the monzogabbro and monzodiorite box, and the quartz diorite, quartz gabbro, and quartz anorthosite fields on the QAP plot. Some of the rocks on the IUGS plot are exceedingly rare. There are also separate IUGS ternary diagrams for different types of gabbro and for ultramafic rocks (greater than 90% mafic minerals) but this goes beyond the scope of what we need to know about the classification of rocks in the Fells.

4.8.2 Classification of Fine-grained Igneous Rocks

It is not practical to apply the IUGS classification of fine-grained igneous rocks to hand samples or even many thin sections because mineral grains may be too small to accurately identify them. However, if mineral percentages can be determined, their percentages can be applied to a simplified IUGS ternary plot for rock identification (Le Maitre and others, 2002 as simplified by Winter, 2010; Fig. 4.19). The beauty of this scheme is that it is simple. However, for naming fine-grained igneous rocks, we generally rely on chemical analyses. It also means that giving a fine-grained igneous rock a precise rock name in the field is highly unlikely, and all our field identification will have to be based on color and the types of phenocrysts in the rock.
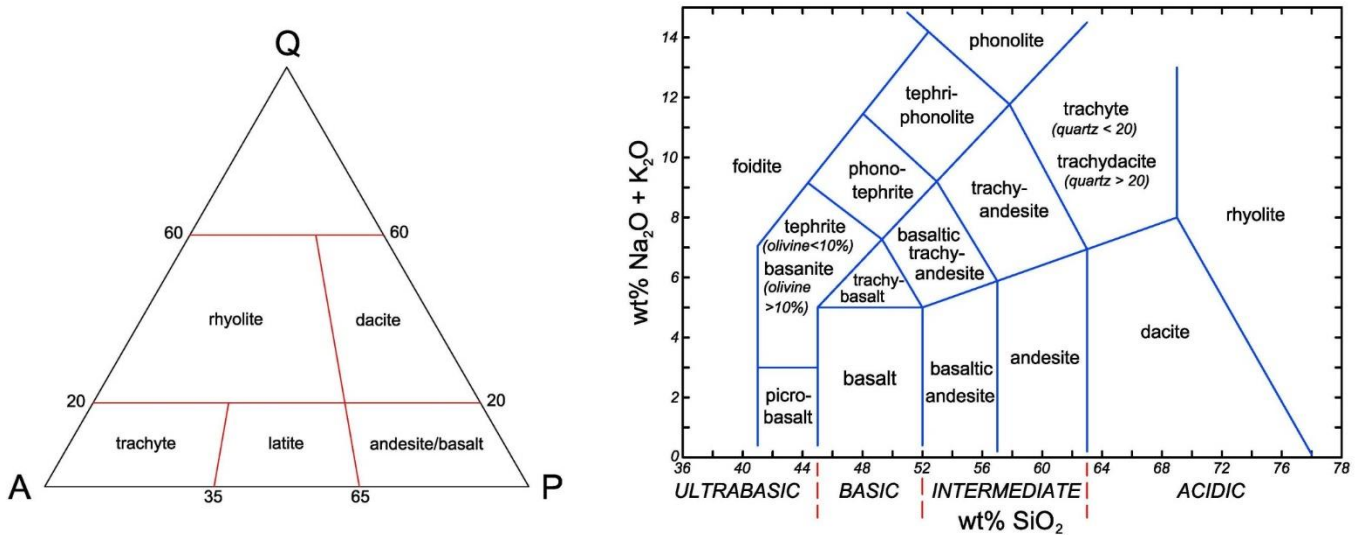

Figure 4.19 *– Classification schemes for fine-grained igneous rocks. Left: A simple mineralogical classification modified from Le Maitre and others (2002) and Winter (2010). This classification parallels the IUGS classification of coarse-grained rocks and uses only normalized QAP percentages. Because of variations in mafic mineral percentages, rocks in the lower right field near P can have different names. Right: A chemical classification scheme of La Bas and others (1986) and Le Maitre and others (2002) based on a plot of wt% ($Na_2O + K_2O$) vs. SiO_2. Because of all its subdivisions, using this classification requires an accurate chemical analysis. This scheme tries to address the distinction of rocks with different mafic contents as well as different normalized quartz and feldspar percentages (represented by $SiO_2\%$).*

A different naming scheme for fine-grained igneous rocks, other than the IUGS classification, is now widely used that depends on rock chemistry (LeBas et al., 1986; Le Maitre and others, 2002; (Fig. 4.19). Fine-grained igneous rocks have a relatively limited range of chemical compositions with some key elemental percentages that define major differences between rock types. Most important is the percentage of silica (SiO_2) in the rock. A second consideration is whether a rock is low or high in its combined percentage of sodium and potassium ($Na_2O + K_2O$). The rock may be *subalkaline* (low $Na_2O + K_2O$) or *alkaline* (high $Na_2O + K_2O$). They may also be further refined as *sodic* if sodium dominates the $Na_2O + K_2O$ percentage or *potassic* if potassium dominates the $Na_2O + K_2O$ percentage. Like the IUGS classification the *LeBas classification* is shown on a graph, though this time the axes are weight % $Na_2O + K_2O$ vs. weight % SiO_2, with different names applied to different fields on the graph. Note that the IUGS and LeBas nomenclature share some names, but they are used slightly differently. At this point, the differences are too small for us to worry about. Our original simple field classification named all the dark-colored, fine-grained rocks (LeBas's ultrabasic and basic) as basalt and all the light-colored, fine-grained rocks (LeBas's intermediate and acid) as felsite.

4.8.3 Types of Feldspar

Another important step in classification of both coarse- and fine-grained igneous rocks is recognizing varieties of feldspar beyond simply alkali feldspar or plagioclase. This is important because the occurrence of different chemical varieties of feldspar can help distinguish rock types. These distinctions are sometimes easily made when looking at a thin section in a petrographic microscope, which allows the separation of different rock units based on types of feldspar.

First, let's look at the subdivision of *plagioclase*. Previously, we gave the chemical formula for plagioclase as $NaAlSi_3O_8$ to $CaAl_2Si_2O_8$. These compositions indicate two end members, or extremes, for the composition of plagioclase. The pure sodium (Na) end member is *albite,* and the pure calcium (Ca) end member is *anorthite*. However, plagioclase can have any composition between these two extremes with varying percentages of sodium and calcium. How we describe the composition of plagioclase is to indicate the percentages of the two end members that produce the resulting composition. To do this we abbreviate the two end members as 'Ab' for albite and 'An' for anorthite. Six names for varieties of plagioclase are recognized as listed below:

Anorthite	An 90-100%	Ab 0-10%
Bytownite	An 70-90%	Ab 10-30%
Labradorite	An 50-70%	Ab 30-50%
Andesine	An 30-50%	Ab 50-70%
Oligoclase	An 10-30%	Ab 70-90%
Albite	An 0-10%	Ab 90-100%

The high calcium plagioclase types at the top of the list above tend to occur in mafic rocks such as basalt and gabbro, while the high sodium plagioclases at the bottom of the list tend to occur in sialic rocks such as granite, granodiorite, dacite, and rhyolite.

There are also several types of *potassium* or *alkali feldspar*, which are *sanidine*, *orthoclase*, and *microcline*. Alkali feldspar types are more a function of the temperature at which the mineral formed than compositional differences. Different alkali feldspars have different crystal forms. Also, keep in

mind that although the chemical formula for the potassium feldspars is written as $KAlSi_3O_8$, there can be some sodium in this mineral that substitutes for potassium. *Sanidine* is a high temperature alkali feldspar found almost exclusively in volcanic and fine-grained sialic rocks as phenocrysts. *Orthoclase* is common in plutonic rocks but can also occur in volcanic rocks and dikes. Orthoclase forms at a higher temperature than *microcline*, which is found exclusively in plutonic rocks. *Microcline* is often formed as a solid-state transformation of orthoclase, with microcline becoming more stable at lower temperatures during cooling.

During the cooling of solid orthoclase and microcline that contain some sodium, a special transformation can occur. Sodium can be excluded from the solid alkali feldspar structure to form separate areas of sodium plagioclase (albite). The albite separates into thin white platy streaks (*lamellae*) that are often in distinct contrast to the usually pink, tan, orange, or salmon color of orthoclase or microcline. This structure of albite lamellae is known as a *perthitic texture* or *perthite*. While clearly visible in microscope slides, perthites can also be seen in some hand specimens (Fig. 4.20). It is also possible for very sodium-rich plagioclase (albite) to crystallize with some potassium. In this case, the separation of the alkali feldspar and albite results in lamellae of alkali feldspar in albite. This forms an *antiperthitic texture* or *antiperthite*.
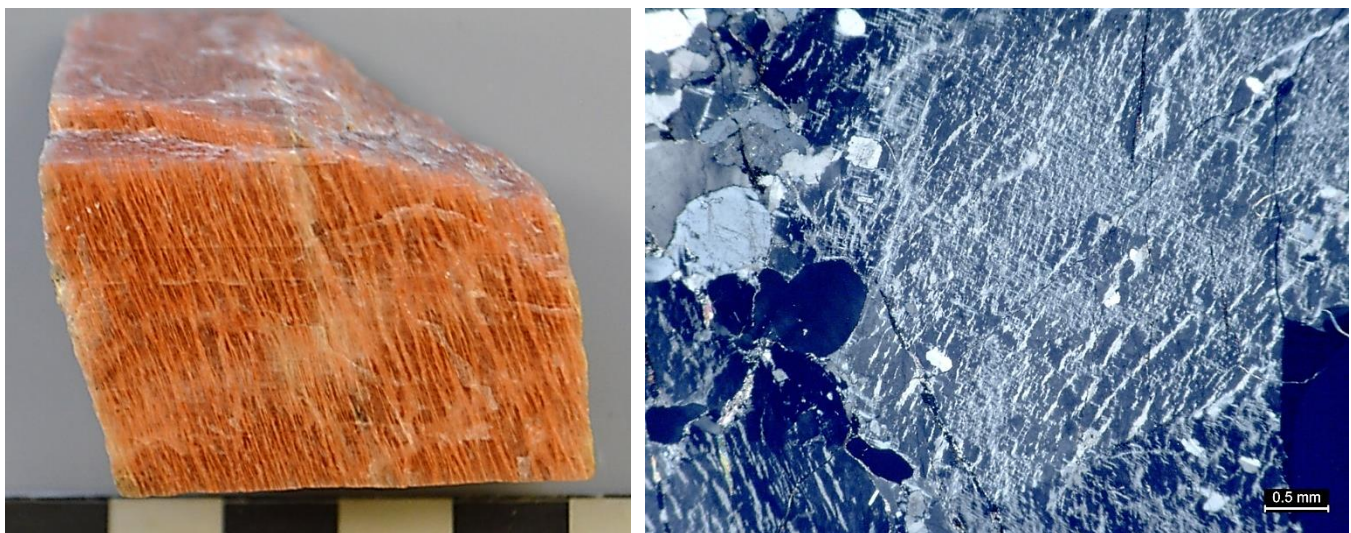

Figure 4.20 – (left) Perthite viewed in hand specimen. Potassium or alkali feldspar, which is orange, has light-colored lamellae (wavy lines) of sodium plagioclase or albite. Scale in cm. (right) Microperthite in a thin section of granite viewed with crossed polarizers. The white streaks in this alkali feldspar crystal are the albite lamellae. Scale in lower right corner.

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