

# Magma Composition and Igneous Rocks

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### Introduction

In the following chapters we will find that nearly all of the varied aspects of volcanism are either directly or indirectly controlled by the chemical composition of magma and the rate at which it cools and solidifies. If you are to understand and appreciate subsequent discussions of volcanoes and related volcanic phenomena, it is essential that you have a good grasp of the basic concepts of magma composition and the controls of magma cooling rate on magma crystallization. To provide this important background, this chapter first discusses how the chemical composition of magma (and any resultant igneous rocks) is expressed, followed by a discussion of the compositional characteristics of natural magmas. Following this, the important controls of magma cooling rate on igneous rock texture are discussed. Finally, the dual concepts of magma composition and cooling rate are combined to define a very useful classification scheme for the most common types of igneous rocks.

### Chemical Composition of Magma and Igneous Rocks

#### *Expressing the Chemical Composition of Magma and Igneous Rocks*

Scientists have found it useful to describe the *chemical composition* of magma, as well as any volcanic rock that is derived from magma, in terms of the *weight percentage* of various chemical oxide molecules (called *oxides* for short). The dual concepts of chemical oxide and weight percentage are difficult ones to grasp and it is worth the effort to discuss both of them in some detail before going any further. Let's turn first to the concept of chemical oxide as it is applied to a magma. The following discussion will assume a simple working knowledge of chemical ionization and bonding.

Magma, like water, is a naturally occurring liquid. However, the chemical makeup of magma is much more complex than water. 99% of any given magma is made up of 10 elements, including Silicon (Si), Titanium (Ti), Aluminum (Al), Iron (Fe), Magnesium (Mg), Calcium (Ca), Sodium (Na), Potassium (K), Hydrogen (H) and Oxygen (O). In a magma, the natural state of affairs is for all of these elements to exist as electrically charged ions. The first eight elements listed above exist as positively charged *cations* with charges that range from +1 to +4. The last element of the list, oxygen, occurs as a negatively charged *anion* with a negative charge of -2 (see Table 2.1). Given the presence of both cations and anions, the natural tendency is for the different charged ions to bond with one another to form molecules that are electrically neutral. Silicon, with a +4 charge, bonds with two oxygens to form a molecule with the chemical formula of  $\text{SiO}_2$ . Magnesium, with a +2 charge, bonds with a single oxygen ion to form a molecule with the formula  $\text{MgO}$ . A single aluminum ion, with a +3 charge, cannot bond with a single oxygen ion and create an electrically neutral molecule. Instead, two aluminum ions bond with 3 oxygen ions to yield a slightly more complex molecule with the formula  $\text{Al}_2\text{O}_3$ . The different molecules that are formed in this fashion are called *oxide* molecules because oxygen is the common component of each. It is

assumed that a magma contains just enough oxygen ions to bond with each and every positive ion. In this fashion, a magma is viewed as a mixture of the nine electrically chemical oxides that are listed in Table 2.1.

By convention, the chemical composition of a magma is expressed in terms of the *weight percentage* of the nine chemical oxides listed in Table 2.1. The weight percentage of a given oxide is nothing more than the percentage of the total magma's weight that is made up by that particular oxide. Figure 2.1 presents a useful "thought experiment" that should help illustrate the concept. Imagine that you start off with a bucket that contains exactly 150 grams of magma. Imagine further that you are able to selectively remove the different chemical oxides from the magma. As you do this, you place the different oxides in smaller buckets. When all nine oxides have been separated and placed in their own buckets, you then weigh the contents of each bucket individually. If you have done your separation and weighing carefully, the contents of the nine small buckets should add up to 150 grams. Now, let's determine what the weight percentage of the oxide SiO<sub>2</sub> is. Figure 2.1 shows that you weighed a total of 75 grams of SiO<sub>2</sub>. Thus the weight percentage of SiO<sub>2</sub> in the magma is the 75 grams of SiO<sub>2</sub> divided by the total grams of magma (150 grams) multiplied by 100, or

$$\text{wt.\% SiO}_2 = 75 \text{ g SiO}_2 / 150 \text{ g magma} \times 100 = 50.0 \text{ weight \%}$$

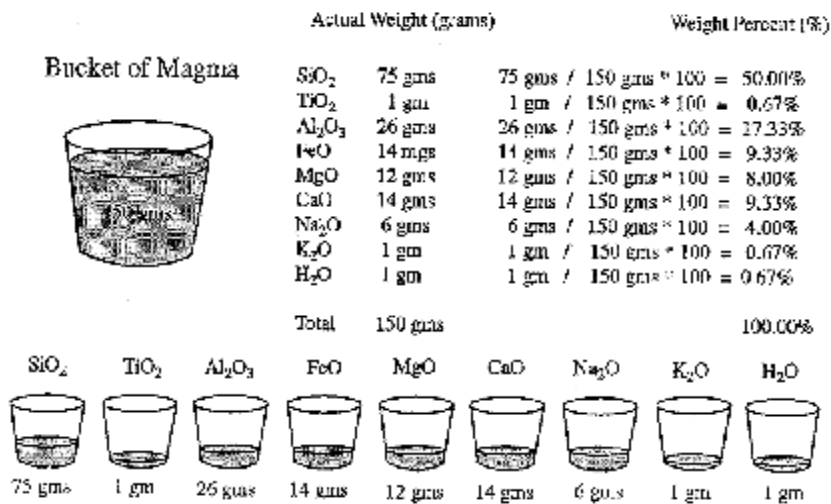


Figure 2.1

A similar calculation for the remaining oxides yields the individual weight percentage values shown in

Figure 2.1. Note that, by definition, the weight percentages of the nine oxides must add up to 100%.

Let's look at the final chemical composition of the magma in the bucket in somewhat more detail,

for it is typical of many (but not all!) natural magmas. The first thing to notice is that the most abundant constituent is SiO<sub>2</sub> which, in this case, makes up one-half of the weight of the magma. The next most abundant constituents are Al<sub>2</sub>O<sub>3</sub>, FeO, MgO and CaO, which make up another 44% of the magma. The remaining 6% consists of Na<sub>2</sub>O, K<sub>2</sub>O, TiO<sub>2</sub>, and H<sub>2</sub>O. These values are typical, but at this point it is important to ask whether or not all magmas, and their resultant volcanic rocks, have the same chemical composition. The answer is a resounding NO! In fact, the chemical composition of different magmas can vary considerably. As will be seen in subsequent chapters, it is this fact that largely controls the wide

range of volcanic rock types, eruption styles, and types of volcanic cones and craters that characterize volcanic activity throughout the world. But for now, let's turn our attention to the characteristics of different magma compositions and how they vary from one to the other.

***Mt. St. Helens - A Case Study of Variable Lava Composition***

Without a doubt, Mt. St. Helens is best known for its 1980 eruption that reawakened our concern about volcanic hazards in the continental United States. Despite its notoriety, the 1980 eruption was only one small episode in a long, drawn-out eruptive history that has lasted over many tens of thousands of years. Over the years, Mt. St. Helens, like nearly all volcanoes of the world, has erupted magmas that cover a very wide range of chemical composition. Thus, Mt. St. Helens also serves our present purpose as a good example of the range of magma (and volcanic rock) compositions that can typically be found among the world's volcanoes. To this end, it is useful to consider in some detail the compositional characteristics of the Mt. St. Helens rocks.

<b>Table 2.1</b>				
Sample #	Lava #1	Lava #2	Lava #3	Lava #4
SiO <sub>2</sub>	49.5	57	66	74.2
TiO <sub>2</sub>	1	1.5	0.5	0.2
Al <sub>2</sub> O <sub>3</sub>	18	16	15	13
FeO	10.5	7.5	4.3	2
MgO	7	4.5	2.5	0.5
CaO	10	7.5	4.5	1.3
Na <sub>2</sub> O	2.5	3.5	4	4.5
K <sub>2</sub> O	1	2	3	4
Gas	0.5	0.5	0.2	0.3
Total	100	100	100	100

Table 2.1, the chemical compositions of four typical volcanic rocks from Mt. St. Helens have been listed. A quick glance at the data points out some very important features. First, it is seen that, once again, SiO<sub>2</sub> makes up at least one-half of the weight of all four rocks. Second, the amount of SiO<sub>2</sub> varies considerably from a low of around 50 wt. % in rock #1 to a high of around 74 wt. % in rock #4. Finally, as the SiO<sub>2</sub> abundance changes from one rock to the next, the abundances of all other oxides change as well. For reasons that are perhaps unclear at this time, it is important to ask *how* these other oxide abundances change with increasing wt. % SiO<sub>2</sub>. Do they change in a random fashion, or are the changes systematic?

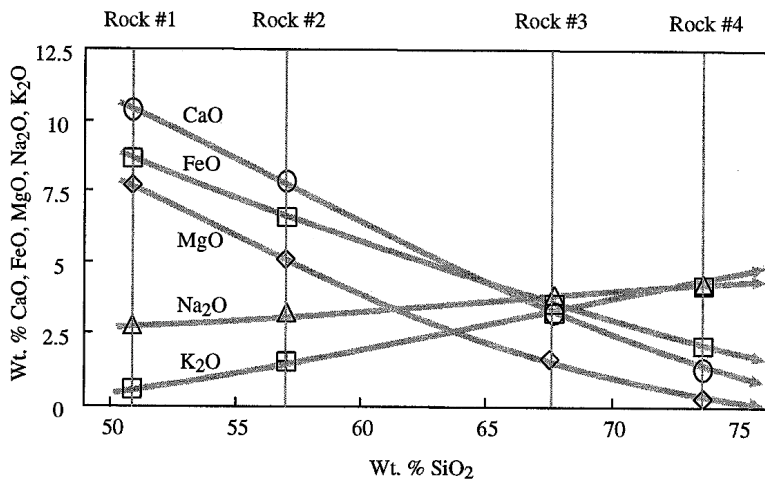


Figure 2.2

2.1. The figure clearly shows that as the SiO<sub>2</sub> content of each successive rock increases, the abundances of FeO, MgO and CaO decrease systematically. At the same time, the abundances of Na<sub>2</sub>O and K<sub>2</sub>O increase systematically. It turns out that these relations hold true not only for the four rocks selected from Mt. St. Helens, but for most magmas and volcanic rocks throughout the world. Why do the oxide abundances vary as they do, and why are the variations so smooth and systematic? The answer to this is an interesting one and is dealt with in great detail two chapters from now. For now, however, it is sufficient to recognize that these systematic compositional variations exist.

### Crystallization of Magma I: Mineral Composition

Magma forms deep within the earth (see Chapter 2). As the magma works its way towards the cooler outer surface of the earth it starts to cool down. When it is erupted onto the earth's surface it cools down very quickly. In much the same fashion as water crystallizes (i.e. solidifies) to ice when it cools, a magma also crystallizes as it approaches the surface of the earth. Unlike water, however, which crystallizes into a single substance, ice, at a single temperature (0<sup>N</sup>C), a magma forms several different substances, called *minerals*, over a very wide range of temperatures (1200<sup>N</sup>C down to around 800<sup>N</sup> or 900<sup>N</sup>C). Thus, while water crystallizing to ice may be a useful analogy for understanding why a magma crystallizes as it cools down, the actual process of magma crystallization is much more complex.

When a magma starts to crystallize, the solid minerals that form as *crystals* have chemical compositions that directly reflect the chemical composition of the magma itself. For example, a magma with the chemical composition of Rock #1 in Table 2.1 is low in SiO<sub>2</sub>, Na<sub>2</sub>O and K<sub>2</sub>O but high in FeO, MgO and CaO. Thus, one might expect that the minerals that crystallize from this magma are themselves rich in FeO, MgO, and CaO and poor in SiO<sub>2</sub>, Na<sub>2</sub>O, and K<sub>2</sub>O. This is precisely what happens. Several different minerals form and they have exotic names like olivine, pyroxene, and calcium plagioclase feldspar. For our purposes, the names of the minerals are not important. What *is* important is that most of these minerals, due to the abundance of iron in them, tend to be dark colored. Thus, magmas that are low in SiO<sub>2</sub> form dark colored minerals upon cooling and the resultant volcanic rock is also dark colored.

It turns out that the changes are very systematic. This conclusion may not be readily apparent to you when looking at a table of chemical data like that shown in Table 2.1. However, it becomes quite obvious when the same data are portrayed graphically on an x-y data plot. Figure 2.2 shows a plot of wt. % SiO<sub>2</sub> on the x-axis versus the Wt. % of FeO, MgO, CaO, Na<sub>2</sub>O and K<sub>2</sub>O on the y-axis for all four rocks listed in Table

In contrast, one might expect a magma with the composition of Rock #4 in Table 2 to crystallize minerals that are rich in  $\text{SiO}_2$ ,  $\text{Na}_2\text{O}$  and  $\text{K}_2\text{O}$  but poor in  $\text{FeO}$ ,  $\text{MgO}$ , and  $\text{CaO}$ . Again, this is exactly what happens. This time the minerals that form have names like potassium feldspar, sodium plagioclase, and quartz. Again, the names are not important. However, due to the relative lack of iron in the minerals, they tend to be light colored. Thus, magmas that are high in  $\text{SiO}_2$  form light-colored minerals that create a light-colored rock. Finally, magmas that have “intermediate” amounts of  $\text{SiO}_2$  tend to crystallize a combination of dark and light minerals and the resulting rock typically has a grayish color to it.

### Crystallization of Magma II: Igneous Rock Texture

The chemical composition of a magma controls what types of minerals (i.e. light versus dark) form upon cooling. However, this is not the only factor that determines what a volcanic rock looks like. Another important factor is the rate (i.e. how fast) at which the magma cools. The cooling rate determines the *texture* of the final rock, which is a measure of how big the individual minerals are that form during cooling. To appreciate how cooling rate controls the size of a crystal, one has to understand a little bit about the process of crystallization. As a magma (or any liquid) cools, the specific cations and anions that ultimately form solid crystals must initially bond with one another to form very small *crystal nuclei*. Based on a great deal of laboratory experimentation, scientists have determined the general relationship between cooling rate and the total number of crystal nuclei that form in most fluids, including magma, and this is schematically shown in Figure 2.3.

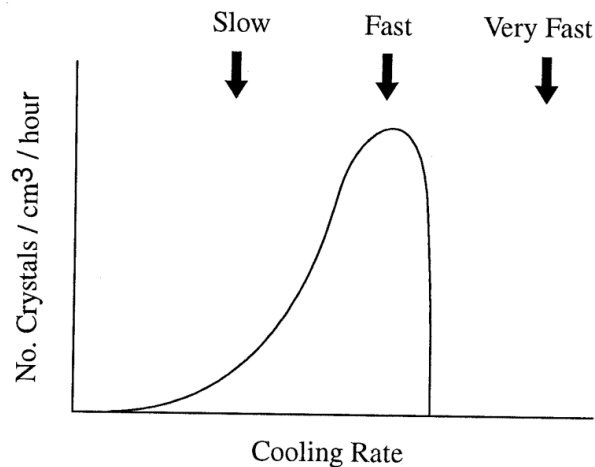


Figure 2.3

Figure 2.3 shows that slow cooling rates generally form only a small number of crystal nuclei. Given this small number of crystals, and abundant time for the crystals to grow, the result is a rock that is dominated by large crystals. Such rocks are referred to as *coarse grained* or *phaneritic*. As the cooling rate increases, so does the number of crystal nuclei that form. Due to the large number of crystals and short amount of time for crystal growth, the result is a *fine grained* or *aphanitic* rock that contains numerous small crystals. If a magma were to go through two stages of cooling, one slow followed by one fast, the

result would be a rock with isolated large crystals enclosed within a predominantly fine-grained rock. This rock texture is referred to as *porphyritic*. The large crystals are referred to as phenocrysts and the fine-grained portion of the rock is called the *matrix* or *groundmass*. Finally, Figure 2.3 shows that *very* fast cooling rates can lead to a situation where no crystals grow at all! Does this mean that the magma remains a liquid? Of course not. What happens in this case is that the magma solidifies to a dark-colored solid that is absent of any crystals. This solid is called *volcanic glass* and the rock is said to have a *glassy* texture. Volcanic glass is absolutely analogous to the man-made glass that is used for so many purposes in our society. The only difference is that man-made glass is formed from compositionally “simple” magma (i.e. nearly pure  $\text{SiO}_2$ ) whereas volcanic glass is derived from a compositionally complex natural magma. Figure 2.4 shows photographs of rocks that display all four of the common textures described

Figure 2.4 - Photographs of Phaneritic, Aphanitic, Porphyritic and Glassy Rocks

above.

Having discussed how different cooling rates can lead to different rock textures, it is important now to consider what types of natural settings can give rise to the different cooling rates. When ascending magma “stalls” within the earth’s crust (i.e. the plutonic environment) the surrounding crustal rocks act as a very efficient insulator and the magma loses its heat very slowly. Thus, the plutonic environment is the one that is characterized by slow cooling rates and coarse-grained phaneritic rocks. In contrast, when magma erupts onto the earth’s surface (i.e. the volcanic environment) the resultant lava lacks any sort of insulation and cools down very quickly. The result is either a fine-grained aphanitic rock or, in some circumstances, a glassy rock. (For reasons beyond the scope of our present discussion, low- $\text{SiO}_2$  and intermediate- $\text{SiO}_2$  magmas typically form aphanitic rocks whereas high- $\text{SiO}_2$  magmas most

commonly form glassy rocks.) An intermediate situation between the two extremes above is where a magma “stalls” within the crust for some period of time and then erupts onto the surface. This process leads to a volcanic rock that has experienced two periods of cooling, the first at a slow rate, the second at a fast rate. The resultant rock has a porphyritic texture. Perhaps surprisingly, most volcanic rocks are porphyritic. Figure 2.5 schematically summarizes the different environments in which phaneritic, aphanitic, glassy, and porphyritic rocks are most commonly formed.

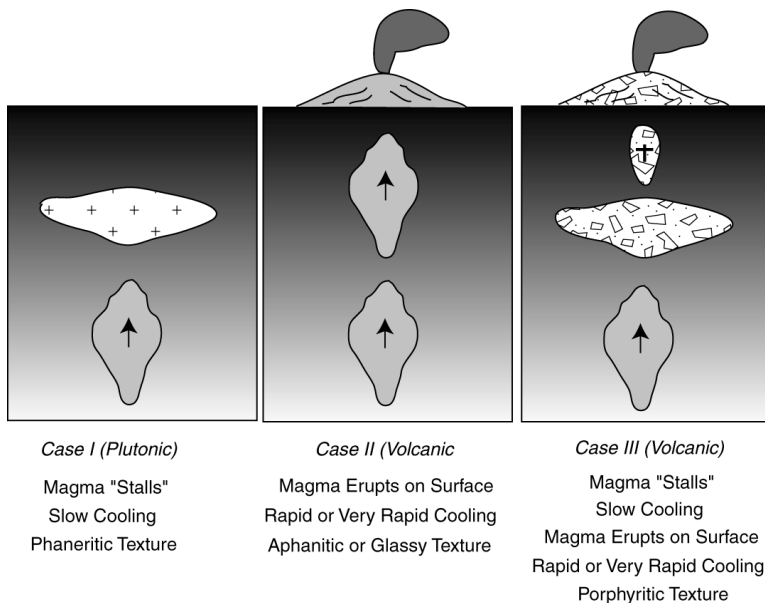


Figure 2.5

## **Igneous Rock Classification: Mineral Composition and Rock Texture Combined**

In the preceding two sections we considered how magma composition controls the *type* of minerals that form (i.e. mineral composition, light versus dark) and how cooling rate controls *what* the minerals look like when they form (i.e. rock texture, phaneritic, aphanitic, porphyritic, glassy). Because the color and size of the minerals that form upon magma cooling are such readily observable features they form a natural basis for classifying igneous rocks into different rock names. The classification scheme that we will use is very straightforward and relies on rock color and texture. The classification scheme itself is shown in Figure 2.6. The first criteria that one looks at is rock texture followed by rock color (dark, intermediate, light). Phaneritic rocks, formed in plutonic environments, are called *gabbro* (dark), *granodiorite* (intermediate), and *granite* (light). Aphanitic rocks, formed in volcanic environments, are called *basalt* (dark), *andesite* (intermediate), and *rhyolite* (light). If a volcanic rock has a porphyritic texture it is called a basalt porphyry, andesite porphyry, and rhyolite porphyry. If, under special circumstances, a volcanic rock has cooled to volcanic glass, the rock is called *obsidian*. There are two additional comments that must be made concerning this classification scheme. First, the rock color is controlled by the color of the minerals which, in turn, are controlled by the chemical composition of the magma from which they form. Thus, magma composition is indirectly represented in the classification scheme. Figure 2.6 shows the approximate wt. % of SiO<sub>2</sub> that corresponds to the various rock names. Second, note that the only difference between basalt and gabbro (or andesite and granodiorite or rhyolite and granite) is the grain size, which is controlled entirely by the cooling rate. Thus, a given magma could give rise to different igneous rocks depending upon where and how it cools and solidifies.

## **PARTIAL MELTING AND THE ORIGIN OF MAGMA**

In the previous chapter it was seen that natural magmas display a wide range in chemical composition, and that the variation in chemical composition is very systematic. Why is the overall range of magma compositions the same regardless of where one looks? Why is the compositional variation so systematic and repetitive from one volcano to the next? Trying to answer these questions has occupied countless man hours during the past century or so. Enough is now known about the origin and evolution of magma that geologists can confidently say that the overall compositional range and systematic compositional variation of natural magmas is primarily a reflection of two important processes at work within the earth. These two processes are called *partial melting* and *fractional crystallization*. Partial melting is what controls the composition of an initial magma when it is formed deep within the earth. Fractional crystallization is a process that changes the composition of magma as it makes its way toward the surface. In this chapter, entitled Partial Melting and the Origin of Magma, we will concentrate on the first of these two very important processes.

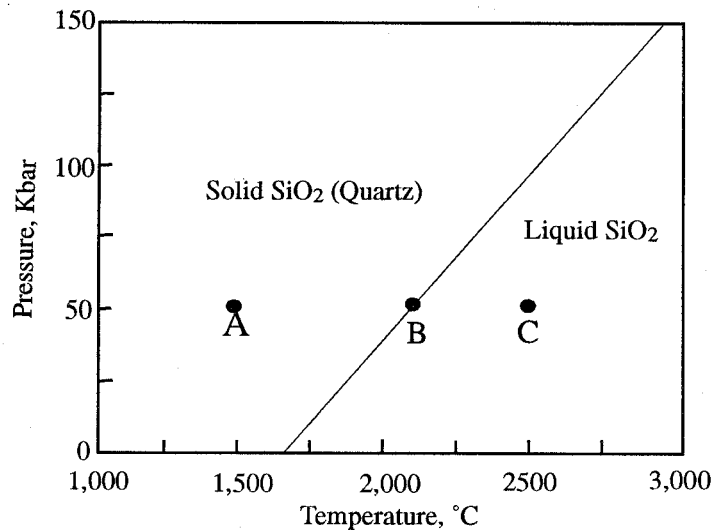
### **Partial Melting Within the Earth's Interior**

Where does magma come from and how does it form? Nearly all magmas originate through partial melting of the upper mantle, though under special circumstances, some magmas result from

melting of rocks in the lower regions of the continental crust. Why do mantle and/or lower crustal rocks melt? Given that the earth's surface is not covered everywhere with volcanoes, there must be various regions of the earth in which something "special" happens to cause melting. It turns out that this "something special" includes three different processes which, either alone or in conjunction with one another, can cause upper mantle (and lower continental crustal) rocks to start melting. These three processes are: (1) an increase in temperature; (2) a decrease in pressure; and (3) the addition of fluids dominated by  $H_2O$ . The first process is obvious. Everyday experience tells us that when a solid is heated it eventually melts. The second and third processes may not be so obvious. However, we now know that these latter two processes are, in fact, the most important when it comes to generating magma within the earth.

### ***T-P Diagrams as Tools for Understanding Temperature and Pressure Controls on Partial Melting***

To emphasize how changes in both temperature and pressure can cause solid rock to melt, it is useful to introduce the concept of the *Temperature-Pressure (T-P) Phase Diagram*, a tool that is used routinely by geologists to understand partial melting processes in both the upper mantle and lower crust. What is a T-P phase diagram? Simply stated, it is a diagram which describes, for a rock of specific (and fixed) chemical composition, the various *phases* (i.e. solid minerals, liquids, gases) that are present at different conditions of temperature and pressure. Figure 3.1 shows a T-P phase diagram for a very simple

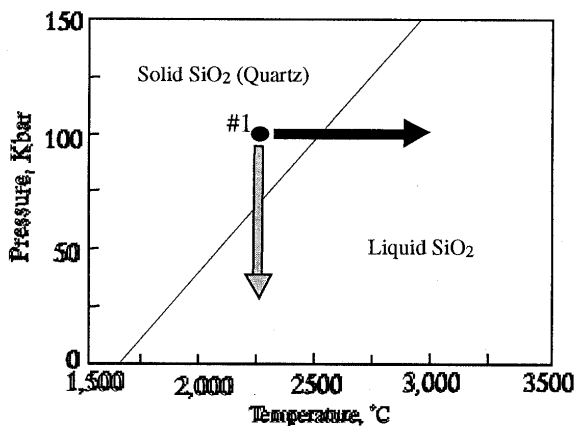


**Figure 3.1. Temperature-Pressure (T-P) Phase Diagram for the  $SiO_2$  system.**

rock that consists entirely of the mineral quartz ( $SiO_2$ ). The diagram shows two different fields, one in which solid  $SiO_2$  (quartz) is the stable phase, the other in which a liquid of pure  $SiO_2$  is the stable phase. The two fields are separated by a diagonal line that is called a *phase boundary*. Note that the position of the phase boundary lies at higher temperatures as the pressure increases. Figure 3.1 tells you what phase (solid or liquid) is stable at a given set of temperature and pressure conditions. For example, if the rock

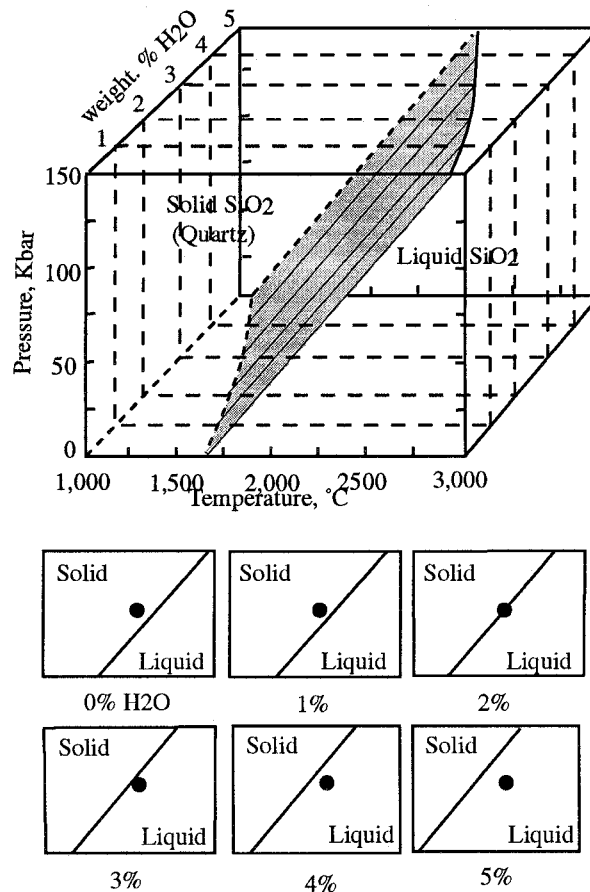
was subjected to a temperature of  $1,500^{\circ}\text{C}$  at a pressure of 50 Kbar (equivalent to approximately 150 km within the earth), indicated by point A, the diagram tells you that the rock will be completely solid. If, on the other hand, the temperature was  $2,500^{\circ}\text{C}$  at a pressure of 50 Kbar (point B), the rock would not be a rock at all, but rather, a liquid. If the rock was at a temperature of  $2,100^{\circ}\text{C}$  at a pressure of 50 Kbar (point C) the rock would lie exactly on the boundary between solid and liquid  $\text{SiO}_2$  and would consist of a mixture of the two phases (i.e. the rock is partially molten).

T-P diagrams such as Figure 3.1 are useful for predicting what the stable phase (or phases) will be for a given set of temperature and pressure conditions for a rock of fixed chemical composition. They can also be used to predict what happens to a rock when one or both of the variables of temperature and pressure are changed. Figure 3.2 shows the same T-P phase diagram as before. Imagine a chunk of rock made entirely of the mineral quartz that is approximately 300 km deep within the mantle ( $P = 100$  Kbar) at a temperature of  $2,250^{\circ}\text{C}$  (labeled point 1). At this set of T-P conditions, Figure 3.2 shows that the rock will be completely solid. Now imagine that, for some reason, the temperature starts to increase. If the rock has not changed its position in the earth, the rock would follow a “trajectory” on the T-P phase diagram shown by the solid black arrow. As the rock started to heat up it would remain completely solid until the temperature reached a value of  $2250^{\circ}\text{C}$ . At this point, the rock would start to melt. When the temperature passed beyond  $2250^{\circ}\text{C}$  the rock will be completely molten. This scenario fully describes how an increase in temperature (process #1 above) can cause a rock to melt. Now, imagine that instead of remaining stationary within the earth, the chunk of rock started to move towards the earth’s surface. (Given that the interior of the earth is constantly in motion, this idea is not as far fetched as might seem.) Imagine further that the temperature of the rock does not change as it moves towards the surface. This time, the rock would be following a trajectory on the T-P phase diagram shown by the solid gray arrow. As the rock started its trip towards the surface it would remain solid. However, note that when the rock reaches a pressure of 70 Kbar (corresponding to a depth of around 210 km), it encounters the “phase boundary” separating the solid from liquid field and the rock starts to melt. At it moves to even shallower depths, the rock is now completely molten. This scenario fully describes how a decrease in pressure (process #2 above) can also cause a rock to melt. This process of melting is known as *decompression melting*.



The third way in which a rock can be caused to melt (process #3 above) is through the addition of fluids that are rich in  $\text{H}_2\text{O}$ . This process cannot be demonstrated with a T-P phase diagram because it involves a change in the composition of the rock, which violates one of the “rules” of a T-P diagram. Instead, let’s consider a useful analogy that should, at the very least, convince you that such a process can indeed cause a rock to melt. Unless you have lived your entire life in an area where the temperature never drops

below freezing, you have experienced a time when ice covers the roads and sidewalks in your hometown. To combat the dangers of the ice, local road crews usually adopt one of two tactics. The first is to simply spread sand on the ice, thus increasing the traction for both people and automobiles. The second approach is to throw salt (NaCl) onto the ice, which causes the ice to melt. Pure ice melts at  $0^{\circ}\text{C}$ . But a mixture of salt and ice melts at  $-24^{\circ}\text{C}$ ! As long as the temperature of the ice is between  $0^{\circ}\text{C}$  and  $-24^{\circ}\text{C}$  before salt is put on it, the mixture of salt and ice will melt to a very salty liquid. Why this happens is beyond the scope of this text. Suffice it to say that this example points out a fundamental principle of chemistry. When two substances are put in contact with one another, the temperature at which the two substances melt together is much lower than the melting temperature of each substance alone. This same principle applies to rocks and  $\text{H}_2\text{O}$ . As  $\text{H}_2\text{O}$  is added to a rock, the melting temperature of the rock (+  $\text{H}_2\text{O}$ ) goes down. Even without a change in temperature, this could cause a rock to start melting.



**Figure 3.3**

This is demonstrated in Figure 3.3 which is a schematic T-P-%  $\text{H}_2\text{O}$  volume that shows what happens to the phase boundary of Figure 3.1 as  $\text{H}_2\text{O}$  is added to the pure quartz rock. Note that as more

and more H<sub>2</sub>O is added to the rock, the position of the solid-liquid phase boundary is progressively shifted to lower and lower temperatures. If one was to take “slices” through this three-dimensional T-P-% H<sub>2</sub>O volume each slice would represent a unique T-P phase diagram for a rock + H<sub>2</sub>O of constant (but different) compositions. Five such diagrams are shown in Figure 3.3 and are used to emphasize how addition of H<sub>2</sub>O can cause a rock to melt. Assume that a pure quartz rock is present within the earth at constant T-P conditions indicated by the solid dot. As more and more H<sub>2</sub>O is added to the rock the solid-liquid phase boundary is shifted to lower temperatures. Somewhere between the addition of 1% and 3% H<sub>2</sub>O, the solid-liquid phase boundary has moved past the prevailing T-P conditions and the rock (+ H<sub>2</sub>O) melts. Thus, Figure 3.3 shows how the addition of H<sub>2</sub>O to a rock can cause it to melt.

### *Partial Melting Within the Upper Mantle: A Plate Tectonic Control*

In the first chapter on Volcanism it was shown that the vast majority of the world’s volcanoes are associated with either divergent plate margins or convergent plate margins, though a volumetrically significant amount of volcanism also occurs in intra-plate environments far removed from any plate boundaries. Most divergent and convergent margins, as well as sites of intra-plate volcanism (hot spot volcanism), are located in regions that are far removed from continental crust. In such cases, the only available source for generating magma is the upper mantle. In those few cases where significant magmatism is found in regions where continental crust is present, an additional lower crustal source is also available. Because the bulk of the world’s magmas are generated through partial melting of the upper mantle, this is the appropriate place to start a detailed discussion of magma generation.

The upper mantle is composed of a rock called *peridotite*, which includes many of the “dark” minerals discussed in the previous chapter. The specific minerals present are olivine and pyroxene, and two minerals not mentioned in Chapter 2, garnet and/or spinel. Once again, the names of the minerals are not important. What is important is that these minerals are all very low in SiO<sub>2</sub>, Na<sub>2</sub>O and K<sub>2</sub>O, and relatively rich in Al<sub>2</sub>O<sub>3</sub>, FeO, MgO and CaO. When upper mantle peridotite starts to melt it produces a magma that reflects the compositional characteristics of the original solid rock. Thus, the magma is also poor in SiO<sub>2</sub>, Na<sub>2</sub>O, and K<sub>2</sub>O, and high in Al<sub>2</sub>O<sub>3</sub>, FeO, MgO and CaO. You will recall that this type of magma is one that produces either basalt or gabbro upon solidification, and is most commonly called *basaltic* magma. Does partial melting of the mantle produce more SiO<sub>2</sub>-rich magmas that create rocks like andesite, diorite, dacite, granodiorite, rhyolite, or granitite? The answer is no. The process that produces these more SiO<sub>2</sub>-rich magmas is fractional crystallization (and also lower crustal melting) which will be discussed in greater detail later on.

Given that partial melting of the upper mantle generates a basaltic magma, the obvious question of importance is why basaltic magma is not generated and erupted over the entire surface of the earth? Why are such magmas restricted to divergent and convergent margins, as well as intra-plate hot-spot settings? The answer is that these regions are the only places where upper mantle peridotite is either being: (1) heated up (i.e. process #1 above); (2) brought towards the surface without being cooled down significantly (process #2); or (3) having large amounts of H<sub>2</sub>O added to it (process #3). In fact, it is the unique physical setting of divergent and convergent plate margins and “hot spots” that permit one or more of these three processes to occur. *That* is why nearly all of the earth’s volcanic activity is restricted to

these three environments. But what is it about these various environments that is conducive to magma generation? We now turn to what is known about divergent and convergent plate margins and intra-plate “hot spot” settings in order to understand how magma is generated in the three different environments.

Figure 3.4 shows a highly idealized cross-section through the earth and points out, in a highly schematic fashion, the relationship between mantle convection and the location of both convergent and divergent plate margins. The mantle convects because the very hot outer core heats up the rocks in the lower mantle. When a rock heats up, it expands very slightly. As it expands, the density of rock goes down. This leads to a situation where hot, low density rock in the lower mantle lies beneath cooler, higher density rock in the upper mantle. Just as a low-density hot air balloon will rise through higher-density cold air, the lower density rock rises towards the earth’s surface. Because the volume of the earth does not increase, any rock that rises towards the earth’s surface must be replaced by rock somewhere else that descends towards the core-mantle boundary. The result

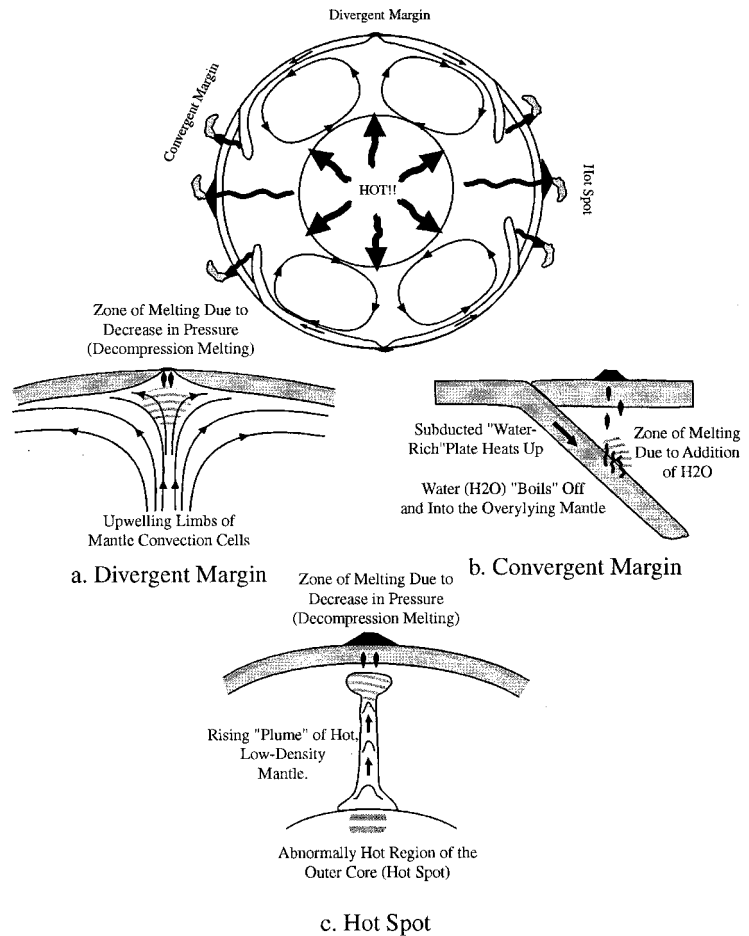


Figure 3.4

is a *convection cell*. Typically, convection cells in the mantle occur in pairs where up-welling material approaches the surface of the earth and then spreads laterally in opposite directions before once again descending towards the core-mantle boundary. The relationship between divergent and convergent plate margins is a direct one. Divergent plate margins occur at those sites where the up-welling limbs of a convection cell pair spreads laterally in opposite directions. Convergent plate margins occur where a convection cell starts its downward descent towards the core-mantle boundary. What goes on in both divergent and plate margin settings that leads to significant mantle melting and the generation of basalt magma? We now turn to this question.

Figure 3.4a shows an enlarged view of a typical divergent plate margin. Up-welling limbs of a pair of mantle convection cells lie beneath the point where two lithospheric plates move away from one another. Numerous theoretical studies have demonstrated that the rock moves upwards much faster than

it can cool down. This leads to a situation where the decrease in pressure causes the rock to start melting. Once melting begins, the small volumes of molten rock collect into discrete bodies of basaltic magma. Because magma is less dense than the surrounding upper mantle peridotite, these bodies of magma rise towards the surface where they eventually erupt as a volcano. This classic example of decompression melting is what generates the large volumes of basaltic magma that is eventually erupted on the sea floor and is continuously creating new oceanic crust. Figure 3.4b shows an enlarged view of a typical convergent plate margin. The subducted lithospheric plate follows the downward descending limb of a single convection cell. The upper portion of the subducted lithospheric plate is rich in water as it has been sitting at the bottom of the ocean for millions of years. As the lithospheric plate is subducted, it moves to greater depths and greater temperatures. Eventually a point is reached where the water “boils” off the plate and escapes into the overlying mantle. (This typically occurs at a depth of around 120 km). The addition of water lowers the melting temperature of the mantle peridotite and causes the rock to melt. Once again, the molten rock collects into discrete bodies of basaltic magma that rise to the surface by virtue of their low density. Thus, subduction zones represent a classic example where the addition of water has caused upper mantle peridotite to melt.

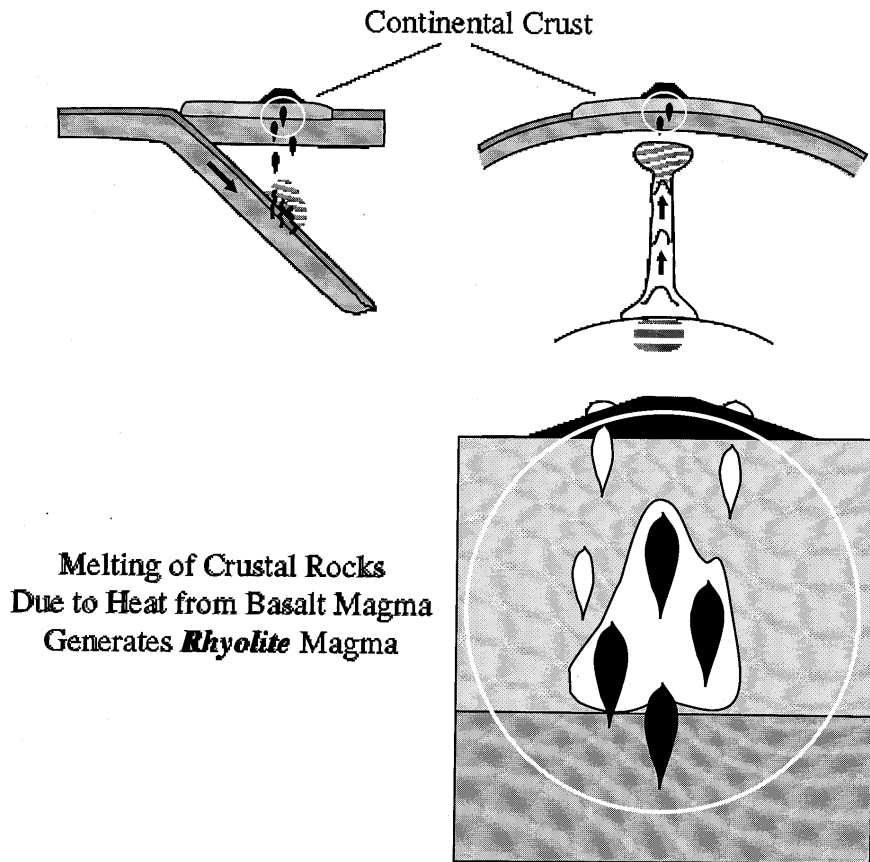
One final question is what causes basaltic magma to be generated beneath a hot spot. The answer to this question is less clear for geologists still do not have a clear idea of what exactly a hot spot is. Most geologists believe that a hot spot represents an abnormally hot region in the outermost core. The extra heat causes a small region of the lower mantle to become hotter. The extra heat is not sufficient to cause this region of the lower mantle to melt, but it *is* sufficient to cause a decrease in rock density which permits a low-density columnar “plume” of lower mantle material to rise towards the earth’s surface. Once again, numerous theoretical studies have shown that the rising material moves faster than it can cool down. This sets the stage for decompression melting to occur as the rising plume nears the earth’s surface. As in the previous two examples, the molten portion of the rock collects as discrete bodies of basalt magma that rise to the earth’s surface. Thus, the generation of basaltic magma within a hot spot environment is due to combination of the addition of heat (at the core-mantle boundary) and a decrease in pressure within a rising mantle plume.

### ***Partial Melting in the Lower Crust***

The previous section dealt with partial melting in the upper mantle and the generation of basaltic magma. At that time it was pointed out that magma can also be generated through the partial melting of lower crustal rocks. Such a process is restricted to regions where volcanism occurs on continental crust. Though much less significant than partial melting of the mantle, continental volcanism is sufficiently abundant that it is important to consider where and how lower crustal melting can be involved.

Unlike the mantle, crustal rocks are rigid and do not undergo convection. Thus, decompression melting is not involved in the melting of the lower crust. There is not available source of water that can be added to the lower crust. Thus, addition of H<sub>2</sub>O is also not a viable melting process. The only available process for causing rocks in the lower crust to melt is through an increase in temperature. Under normal conditions, the temperature at the base of the crust is not sufficient to cause crustal rocks to melt. Thus, crustal melting can only occur in regions where additional heat is provided from some other

source. The only known source of heat is from hot, basaltic magma that is generated in the upper mantle beneath the continental crust. The only regions where mantle-generated basalt magma can rise into overlying continental crust is where a subduction zone occurs along the edge of a continent (e.g. the



Andean volcanic arc along the western margin of South America) or where a hot spot lies beneath continental crust (e.g. Yellowstone National Park). What type of magma results from the partial melting of the continental crust? Recall that the compositional characteristics of a magma reflects the composition of the rock from which it is generated. In contrast to upper mantle peridotite, rocks of the continental crust contain, on average, considerably more  $\text{SiO}_2$ ,  $\text{Na}_2\text{O}$ , and  $\text{K}_2\text{O}$ , and lesser amounts of  $\text{FeO}$ ,  $\text{MgO}$ ,  $\text{Al}_2\text{O}_3$  and  $\text{CaO}$ . Any magma that is generated from such rocks will reflect this difference. In fact, when continental crustal rocks begin to melt, they

Figure 3.5

normally generate a very  $\text{SiO}_2$ -rich magma which, upon crystallization, would produce a rock known as rhyolite or granite. This magma type is normally referred to as *rhyolitic magma*. Figure 3.5 schematically outlines where and how continental crustal melting can be expected to occur.

### Fractional Crystallization and Magma Evolution

In the previous chapter it was demonstrated how partial melting of the earth's interior generates large volumes of low- $\text{SiO}_2$  basaltic magma and, in continental environments, lesser but still significant volumes of high- $\text{SiO}_2$  rhyolitic magma. It is important to recognize, however, that partial melting of mantle or crustal rocks cannot explain the overall continuous range of magma compositions that characterize nearly every volcanic region. The process that is primarily responsible for this feature is fractional crystallization. What is fractional crystallization? It is a process that occurs when most liquids

cool and start to crystallize. Unless the liquid is compositionally very simple, like pure water (H<sub>2</sub>O), when the liquid cools to the point where it starts to solidify, it forms a solid that has a *different chemical composition* than that of the liquid from which it is forming. As more and more of the solid forms, the *chemical composition of the remaining liquid necessarily changes*. Thus, fractional crystallization is a process that causes the chemical composition of a liquid to change. To illustrate this concept, consider the following account that was given to me by one of my colleagues at Indiana, Dr. Robert Wintch. I have tried (unsuccessfully) to verify its truthfulness. True or not, however, it is an amusing story that describes the process of fractional crystallization in understandable, down-to-earth terms.

### ***The Making of Apple Jack: Fractional Crystallization at Its Best!***

The time is that of pioneer America, perhaps in the late 1600s or early 1700s. The winters during those years were much more severe than they are now and the early pioneers did not have all of the modern conveniences that we have to ward off the winter cold. They could keep themselves warm from the outside, with a good fire and heavy wool clothes, or from the inside with a good stiff drink of alcohol. Because distilleries (and distilled spirits) were few and far between, the early pioneers had to make their own alcoholic beverages. The beverage of choice was Apple Jack, which could be made from apple cider, a common drink of the times. How did the pioneers transform apple cider, which was non-alcoholic, into Apple Jack, which was extremely alcoholic? They used their common sense and the process of fractional crystallization!

Late fall was the time in which the pioneers would harvest their apples and make large quantities of apple cider. Normally, the cider would be stored in large wooden barrels and could keep for some time. Eventually, however, the unused cider would begin to ferment in the barrels and turn into “hard” cider that had a bitter taste due to the presence of a small amount of alcohol. Either by accident or design, the pioneers found that if they left the barrels of hard cider outside in the cold for a long time, the contents of the barrel would freeze into a giant “ice cube” in the center of which was a small pocket containing a dark, thick, alcohol-rich liquid that was the Apple Jack. How did the simple process of freezing transform hard cider, with only a small amount of alcohol, into Apple Jack, with a great deal of alcohol? The answer is outlined in Figure 1, below.

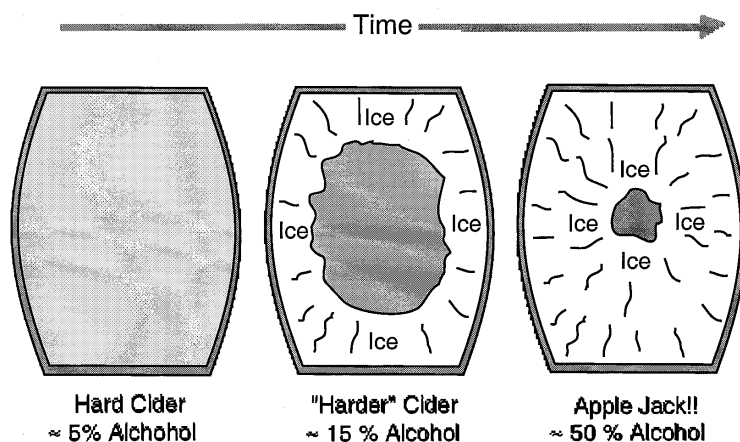


Figure 1

Hard apple cider consists of a complex mixture of water, alcohol, and apple flavoring. Water freezes to solid ice at 0<sup>N</sup>C (or 32<sup>N</sup>F). Alcohol, on the other hand, does not freeze until much lower temperatures, on the order of -80<sup>N</sup>C. As the temperature of the barrel drops below 0<sup>N</sup>C, the water in the cider will freeze but the alcohol will not (i.e. the first solid to form has a different chemical composition than the liquid from which it is forming). As the temperature continues to drop and more and more of the water freezes to ice, the volume of remaining cider gets smaller and the concentration of alcohol in the cider increases (i.e. the chemical composition of the remaining liquid necessarily changes.) Eventually, when all the water has frozen to ice, the remaining liquid consists of essentially nothing but alcohol and apple flavoring. This is the secret to making Apple Jack...and is also a classic example of fractional crystallization at work.

What does all this have to do with generating a continuous range of magma compositions? The general process of fractional crystallization outlined in the Apple Jack example has everything to do with it! Consider a basaltic magma that nears the surface of the earth and starts to cool and crystallize. You already know that a low-SiO<sub>2</sub> magma like basalt will crystallize “dark” minerals that are poor in SiO<sub>2</sub>, Na<sub>2</sub>O, and K<sub>2</sub>O and rich in FeO, MgO, CaO and Al<sub>2</sub>O<sub>3</sub>. The chemical composition of the “dark” minerals broadly reflects the chemical composition of the basalt magma, but the minerals are not exactly the same as the magma. Thus, when the “dark” minerals start to crystallize the chemical composition of the remaining liquid portion of the magma *must* change. This is exactly analogous to the hard cider (i.e. the original basalt magma), the ice (i.e. the “dark” minerals) and the Apple Jack (i.e. a new composition liquid magma). How does the liquid portion of the magma change as more and more minerals crystallize? To answer this, it is necessary to know the chemical composition of the crystallizing minerals relative to that of the original basalt magma. Table 1 lists a typical basalt magma composition along with a reasonable composition for the “dark” minerals that can be expected to crystallize in a basalt.

<b>Table 1</b>		
	<b>Basalt Magma</b>	<b>“Dark” Minerals</b>
<b>SiO<sub>2</sub></b>	<b>50.0</b>	<b>40.1</b>
<b>TiO<sub>2</sub></b>	<b>0.7</b>	<b>1.4</b>
<b>Al<sub>2</sub>O<sub>3</sub></b>	<b>17.3</b>	<b>20.1</b>
<b>FeO</b>	<b>9.3</b>	<b>11.3</b>
<b>MgO</b>	<b>8.0</b>	<b>13.0</b>
<b>CaO</b>	<b>9.3</b>	<b>15.0</b>
<b>Na<sub>2</sub>O</b>	<b>4.0</b>	<b>1.0</b>
<b>K<sub>2</sub>O</b>	<b>0.7</b>	<b>0.1</b>
<b>H<sub>2</sub>O</b>	<b>0.3</b>	<b>0.0</b>
<b>Total</b>	<b>100.0</b>	<b>100.0</b>

It is important to note that the dark minerals contain less SiO<sub>2</sub>, Na<sub>2</sub>O and K<sub>2</sub>O than the basalt magma, but more TiO<sub>2</sub>, Al<sub>2</sub>O<sub>3</sub>, FeO, MgO, and CaO. This means that when the dark minerals crystallize from basalt, they remove from the liquid magma *more* TiO<sub>2</sub>, Al<sub>2</sub>O<sub>3</sub>, FeO, MgO and CaO, and *less* SiO<sub>2</sub>, Na<sub>2</sub>O, and K<sub>2</sub>O than is present in the liquid to begin with. Because of these relationships, the original basalt will change in composition towards a liquid that is *richer* in SiO<sub>2</sub>, Na<sub>2</sub>O and K<sub>2</sub>O and *poorer* in TiO<sub>2</sub>, Al<sub>2</sub>O<sub>3</sub>, FeO, MgO and CaO. This result may not be so obvious, so Figure 2 attempts to make it clear through use of the same “bucket analogy” that was used in Chapter 2 to describe the concept of magma composition and wt. % oxides.

Compare the composition of the original magma with that of the remaining magma after the removal of 25 gm of “dark” minerals. You will see that, as predicted, the percentages of SiO<sub>2</sub>, Na<sub>2</sub>O and K<sub>2</sub>O (and H<sub>2</sub>O) have increased while those of FeO, MgO, CaO (and Al<sub>2</sub>O<sub>3</sub> and TiO<sub>2</sub>) have all decreased.

### Fractional Crystallization in Natural Magmas

You have now seen how fractional crystallization can cause a magma to change in chemical composition. How, then, does this process explain the formation of magmas that cover a wide and continuous range of chemical compositions? The answer lies in the fact that the *chemical* process of fractional crystallization can be combined with the *physical* process of crystal-liquid segregation to generate a continuous spectrum of natural magma compositions. Figure 3 summarizes how most geologists believe this takes place.

Consider a basaltic magma that is generated in the upper mantle that then rises towards the earth’s surface. The magma rises upwards because it is less dense than the surrounding rock of the upper mantle.

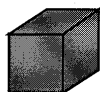
### Fractional Crystallization and Magma Evolution

Bucket of Magma  
(150 gms)



SiO <sub>2</sub>	50.00% of 150 gms = 75 gms
TiO <sub>2</sub>	0.67% of 150 gms = 1 gm
Al <sub>2</sub> O <sub>3</sub>	17.33% of 150 gms = 26 gms
FeO	9.33% of 150 gms = 14 gms
MgO	8.00% of 150 gms = 12 gms
CaO	9.33% of 150 gms = 14 gms
Na <sub>2</sub> O	4.00% of 150 gms = 6 gms
K <sub>2</sub> O	0.67% of 150 gms = 1 gm
H <sub>2</sub> O	0.67% of 150 gms = 1 gm

"Dark" Minerals  
(25 gms)



SiO <sub>2</sub>	40.10% of 25 gms = 10.0 gms
TiO <sub>2</sub>	1.40% of 25 gms = 0.3 gms
Al <sub>2</sub> O <sub>3</sub>	20.10% of 25 gms = 5.0 gms
FeO	11.30% of 25 gms = 2.7 gms
MgO	13.00% of 25 gms = 3.2 gms
CaO	15.00% of 25 gms = 3.6 gms
Na <sub>2</sub> O	1.00% of 25 gms = 0.2 gms
K <sub>2</sub> O	0.10% of 25 gms = 0.0 gms
H <sub>2</sub> O	0.10% of 25 gms = 0.0 gms

Bucket of Magma  
(150 gms)



Minus

"Dark" Minerals  
(25 gms)



SiO <sub>2</sub>	75 gms	minus 10.0 gms = 65.0 gms
TiO <sub>2</sub>	1 gm	minus 0.3 gms = 0.7 gms
Al <sub>2</sub> O <sub>3</sub>	26 gms	minus 5.0 gms = 21.0 gms
FeO	14 gms	minus 2.7 gms = 11.3 gms
MgO	12 gms	minus 3.2 gms = 8.8 gms
CaO	14 gms	minus 3.6 gms = 10.4 gms
Na <sub>2</sub> O	6 gms	minus 0.2 gms = 5.8 gms
K <sub>2</sub> O	1 gm	minus 0.0 gms = 1.0 gm
H <sub>2</sub> O	1 gm	minus 0.0 gms = 1.0 gms

### Final Magma Compositions

SiO <sub>2</sub>	65.0 gms / 125 gms * 100 = 52.00%
TiO <sub>2</sub>	0.7 gm / 125 gms * 100 = 0.56%
Al <sub>2</sub> O <sub>3</sub>	21.0 gms / 125 gms * 100 = 16.80%
FeO	11.3 gms / 125 gms * 100 = 9.00%
MgO	8.8 gms / 125 gms * 100 = 7.04%
CaO	10.4 gms / 125 gms * 100 = 8.32%
Na <sub>2</sub> O	5.8 gms / 125 gms * 100 = 4.64%
K <sub>2</sub> O	1.0 gm / 125 gms * 100 = 0.80%
H <sub>2</sub> O	1.0 gm / 125 gms * 100 = 0.80%
Total	125 gms 100.00%

If, however, the rising magma was to suddenly encounter rocks that are less dense than the magma itself, the magma could no longer rise towards the surface and it would “stall”. This commonly happens when a rising basalt magma encounters low density crustal rocks. When the magma stops rising, it forms a stationary *magma chamber* in which the magma slowly cools down. As cooling continues, the basalt magma eventually starts to crystallize “dark minerals” and, due to the process of fractional crystallization, the liquid becomes enriched in  $\text{SiO}_2$ ,  $\text{Na}_2\text{O}$ , and  $\text{K}_2\text{O}$ , and depleted in  $\text{FeO}$ ,  $\text{MgO}$  and  $\text{CaO}$ . (If this liquid were to erupt onto the surface it would form a rock called andesite.) A very important aspect of this entire process is that the “dark” minerals are more dense than the basalt liquid in which they formed and, given sufficient time, they will sink to the bottom of the magma chamber. This process, known as gravitation crystal settling, is believed by many to be the most important factor in bringing about a separation of crystals and liquid. (Other physical processes of crystal-liquid separation also take place in

a magma chamber but they are beyond the scope of this discussion). Note that, after the “dark minerals” have sunk to the bottom of the magma chamber, the magma consists of a crystal-free liquid of andesite composition. If the andesite liquid were to cool and crystallize, it would form mostly dark minerals and a few light minerals. Crystallization of these minerals would force the liquid to change in composition towards higher  $\text{SiO}_2$ ,  $\text{Na}_2\text{O}$ , and  $\text{K}_2\text{O}$ , and less  $\text{FeO}$ ,  $\text{MgO}$  and  $\text{CaO}$ . (This liquid would produce a dacite lava.) Once again, the higher density minerals would settle to the magma chamber bottom and the magma would now consist of a crystal-free liquid of dacite composition. It does not take much imagination to recognize that this process could continue until the last bit of liquid magma is one that is very high in  $\text{SiO}_2$ ,  $\text{Na}_2\text{O}$  and

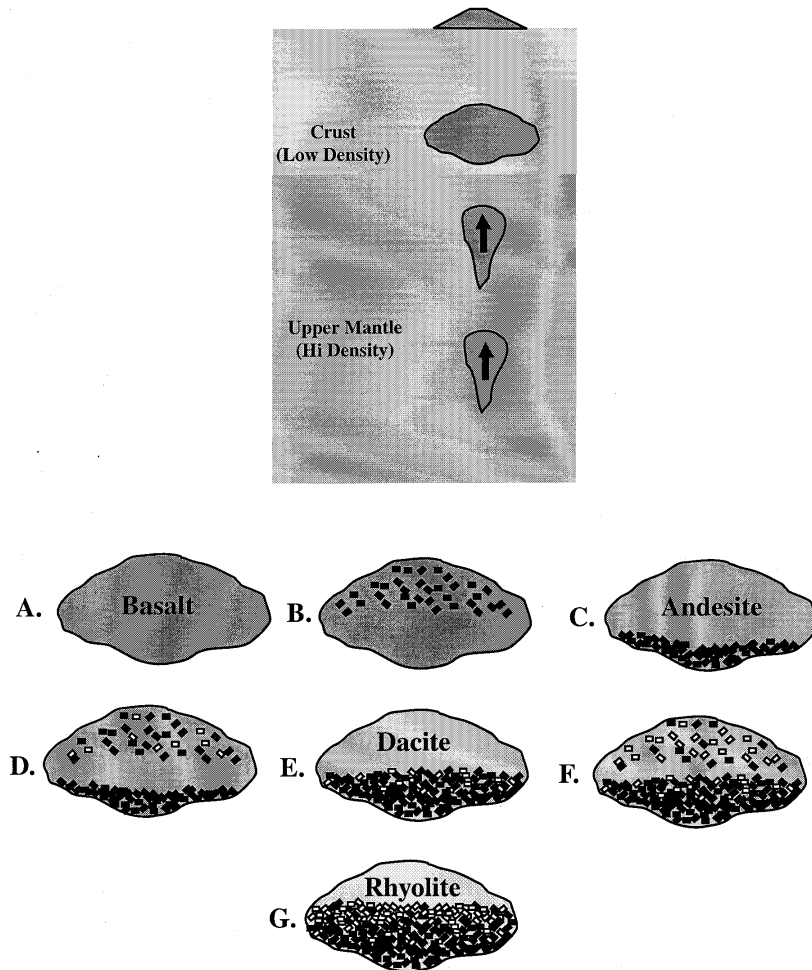


Figure 3

$K_2O$  and very depleted in  $FeO$ ,  $MgO$  and  $CaO$ . In fact, this magma would produce a rhyolite lava upon eruption. In nature, the process outlined above is a continuous one. The compositionally evolving liquid that always resides at the top of the magma chamber could periodically (and arbitrarily) erupt onto the surface as lava. If the eruptions were closely spaced together, the result would be a series of lavas that cover nearly a complete compositional spectrum from basalt to rhyolite.