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The Geology of Magma Systems Background and Review

Alan R. Peterfreund

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THE GEOLOGY OF MAGMA SYSTEMS
BACKGROUND AND REVIEW

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ABSTRACT

A review of basic concepts and current models of igneous geology is presented. This review is intended for the nongeologists involved in magma energy studies. Emphasis is centered on studies of magma generation, ascent, emplacement, evolution, and surface or near-surface activity. An indexed reference list is also provided to facilitate future investigations.

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THE GEOLOGY OF MAGMA SYSTEMS BACKGROUND AND REVIEW

Introduction

The objective of this report is to summarize the major geologic models pertinent to the development of energy utilization from magma bodies in the western United States. The geologic subjects relevant to the development of magma energy encompass the fields of volcanology, geochemistry, igneous petrology, tectonophysics, geophysics, and regional geology. In each of these fields, volumes have been dedicated to discussing subjects presented only briefly here. My intent is to review the major findings of some of the recent research in these fields as it relates to magma energy. Complementary to this review, an indexed bibliography is presented that hopefully will be useful in facilitating future research in individual topics.

The potential for extracting energy from active magma bodies in the upper crust has been discussed by Colp (1974, 1980), Fedotov et al. (1975), and Heffington (1977) and others. One of the major needs to achieve this potential is an appreciation of the complex nature of magma systems. However, as an active magma body has never been directly observed, except by its surface manifestations, geologists must rely on models to understand the associated processes and properties of the system. These models are based on observable features of magma bodies, which include: associated volcanic activity and products, solidified chambers (plutonic bodies), and geophysical properties observed remotely. These models are also supported by experimental studies of the physical and chemical properties of

silicate melts and magmas. Recent advances in our understanding of the Earth's crustal activity, plate tectonics, has given a new framework into which igneous activity can be related. Regional studies, such as those of the western United States, can therefore consider volcanic and plutonic activity in the broader context of crustal and mantle activity. This has increased and will continue to increase our understanding on the generation, ascent, emplacement and evolution of magma bodies.

This report is divided into two main sections. The first section is an abbreviated primer in igneous geology. A review of rock classifications, plate tectonics, magma associations, and volcanism is provided for the non-geologist. The second section discusses models of magma processes. Emphasis is placed on those models pertinent to igneous activity in the western United States.

Due to the enormity of the published research relevant to magma systems, some selectivity has been necessary. Topics that are only briefly mentioned in this report, or not covered, are included in the indexed reference lists. The reference lists provided are by no means exhaustive, but rather include the most recent works, as well as review papers, in these important magma-related subjects. It is my hope that this report and the indexed bibliography give future investigators a helpful starting point to embark on future research.

A Primer in Igneous Geology

Rock Classifications

Igneous rocks are those that have solidified from a molten state. Classifications of igneous rocks are based on chemistry, mineralogical composition, and texture. Chemically, the most important distinction is based on the silica content of the rock. Silica comprises 40 to 75 percent of most igneous rocks. Silica-rich rocks, such as rhyolites, have in the past been referred to as "acidic" and rocks with less silica, basalts, as "basic," with intermediate rocks, such as dacite, being called "intermediate." These terms, which unfortunately persist, are technically incorrect, although the silica content of the rock remains an important classifying criterion.

Mineralogically, a broad classification of igneous rocks considers the presence of light and dark minerals. Light minerals are typically quartz and feldspar, both silica-rich, and collectively referred to as felsic. Dark minerals, which include pyroxenes, amphiboles, and olivines, are rich in magnesium and iron, and are referred to as mafic. The minerals associated with this scheme are shown in Figure 1. A continuous series exists between the felsic and mafic rocks with end-members similar to those based on silica content. Table 1 illustrates one such classification scheme. Table 2 gives the average chemical composition for some of the major igneous rocks.

OLLIER, 1969

Figure 1. Mineral Content of Common Igneous Rocks

Table 1. Classification of common igneous rocks (Names of aphanitic rocks are given in capitals, those of phaneritic rocks in lower-case letters)

Felsic Rocks Light-colored Minerals more than 60% of Rock				Mafic Rocks Dark-colored minerals 40% or more of rock			
	Alkali feldspar ¹ predominant	Alkali and soda-lime feldspar approximately equal	Soda-lime feldspar predominant	Alkali feldspar 10% or more of the total feldspar	Soda-lime feldspar more than 90% of the total feldspar		Feldspar less than 10% of the rock
					Feldspar more than 30% of the rock	Feldspar 30-10% of the rock	
More than 10% quartz	RHYOLITE granite	RHYODACITE quartz monzonite ⁴	DACITE ³ quartz diorite ⁵ granodiorite ⁶	QUARTZ BASALT quartz gabbro granogabbro			
Less than 10% quartz or feldspathoid ²	TRACHYTE syenite	TRACHYANDESITE ⁷ monzonite	ANDESITE diorite anorthosite ¹⁰	TRACHYBASALT ⁸ SPILITE ⁸ shonkinite	BASALT ⁹ gabbro diabase ¹¹	PICRITE-BASALT picrite	dunite peridotite ¹² perknite ¹² carbonatite ¹³
More than 10% feldspathoid ²	PHONOLITE ¹⁴ nepheline syenite ¹⁶	TEPHRITIC PHONOLITE nepheline monzonite	PHONOLITIC TEPHRITE	TEPHRITE ¹⁵ BASANITE ¹⁵ essexite ¹⁷ theralite ¹⁷			NEPHELINE LEUCITITE ijolite

- 1 Generally orthoclase, microcline, or albite in coarse-grained rocks; sanidine or anorthoclase in fine-grained rocks.
- 2 Silica-deficient equivalents of the feldspars, including nepheline, leucite, sodalite, kalsilite, analcite, melilite, etc.
- 3 Most dacites correspond in chemical composition to granodiorites rather than to quartz diorites.
- 4 Adamellite is approximately equivalent.
- 5 Tonalite is approximately equivalent.
- 6 In granodiorite alkali feldspar is approximately half as abundant as soda-lime feldspar.
- 7 Latite is approximately equivalent.
- 8 In spilite the feldspar is predominantly alkalic, usually albite.
- 9 When the rock contains more than 5% olivine it is called olivine basalt. Tholeiitic basalts are relatively low in alkalis and high in silica, whereas alkalic basalts are relatively high in alkalis and low in silica. Tholeiitic basalts usually lack alkali feldspar; alkalic basalts contain alkali feldspar, but in smaller amounts than the trachybasalts.
- 10 In anorthosite soda-lime feldspar makes up more than 90% of the rock.
- 11 Diabase tends to be coarser grained than basalt and is characterized by diabasic texture, in which well-formed (euhedral) lath-shaped grains of feldspar penetrate grains of augite.
- 12 Dunite contains more than 90% olivine, peridotite 90% to 10% olivine, and perknite less than 10% olivine. These ultramafic rocks, particularly peridotite and dunite, are very commonly altered to serpentine.
- 13 Carbonatites consist primarily of carbonates of calcium, magnesium, and iron, with variable amounts of other minerals such as nepheline, melilite, apatite, and magnetite.
- 14 Phonolites containing many large phenocrysts of leucite are leucitophyres.
- 15 Tephrite contains 10% or less olivine, basanite more than 10% olivine.
- 16 When alkali feldspar and feldspathoid are about equally abundant the rock may be called foyaite.
- 17 Theralite contains more feldspathoid and less soda-lime feldspar than essexite.

After Macdonald, 1972

Table 2. Average chemical compositions of volcanic rocks (weight percent)¹

	Rhyo- lite	Trachyte	Phono- lite	Rhyoda- cite	Dacite	An- desite	Tholeiitic basalt	Alkalic basalt	Nephe- line	Pyroxene- nite	Peri- dotite	Dunite	Ser- pentine
SiO ₂	73.6	58.3	56.9	66.3	63.6	54.2	50.8	45.8	40.3	50.5	43.5	40.2	40.4
TiO ₂	0.2	0.7	0.6	0.7	0.6	1.3	2.0	2.6	2.9	0.5	0.8	0.2	tr.
Al ₂ O ₃	13.4	18.0	20.2	15.4	16.7	17.2	14.1	14.6	11.3	4.1	4.0	0.8	1.9
Fe ₂ O ₃	1.2	2.5	2.3	2.1	2.2	3.5	2.9	3.2	4.9	2.4	2.5	1.9	2.8
FeO	0.8	2.0	1.8	2.2	3.0	5.5	9.1	8.7	7.7	7.4	9.8	11.9	4.3
MgO	0.3	2.1	0.6	1.6	2.1	4.4	6.3	9.4	13.3	21.7	34.0	43.2	36.0
MnO	0.3	0.1	0.2	0.1	0.1	0.1	0.2	0.2	0.2	0.1	0.2	0.2	--
CaO	1.1	4.2	1.9	3.7	5.5	7.9	10.4	10.7	13.0	12.0	3.5	0.8	0.7
Na ₂ O	3.0	3.8	8.7	4.1	4.0	3.7	2.2	2.6	3.1	0.4	0.6	0.3	
K ₂ O	5.4	7.4	5.4	3.0	1.4	1.1	0.8	1.0	1.4	0.2	0.2	0.1	0.2
H ₂ O+	0.8	0.5	1.0	0.7	0.6	0.9	0.9	0.8	1.1	0.5	0.8	0.4	10.5
P ₂ O ₅	0.1	0.2	0.2	0.2	0.2	0.3	0.2	0.4	0.8	0.1	0.1	0.1	tr.

¹ After Nockolds, 1954

The textural characteristics of igneous rocks help determine whether they had an intrusive (plutonic) or extrusive (volcanic) origin. The principal differentiations are based on grain size and crystallinity. The more rapidly a rock is cooled, the finer the grain size and the poorer the crystallinity. Coarse-grained rocks, called phaneritic, are typically intrusives and fine-grained rocks, aphanitic, are extrusives. The same distinctions can be made based on the degree of glassiness or, conversely, on the degree of crystallinity, with glassiness being a characteristic of rapidly extruded rocks. The terms aphanitic and phaneritic are synonymous with extrusive and intrusive in Table 1.

Plate Tectonics

The plate tectonic model provides a conceptual framework that accounts for where magmatic processes occur and their chemical and physical framework. Plate tectonics describes the Earth's crust and underlying lithosphere as a series of rigid plates (Figure 2) in motion with respect to each other, driven by activity of the underlying mantle. The processes occurring at a given plate boundary are the result of two dominating factors; the nature of the crust of the adjoining plates and their relative motion.

The Earth's crust can be divided into two major types: oceanic and continental. Oceanic crust is primarily basaltic, thin (5 to 15 km), and relatively dense ($\sim 2.9 \text{ g/cm}^3$). Continental crust is more silicic, thicker (30-50 km), and less dense ($\sim 2.7 \text{ g/cm}^3$). Oceanic crust is considerably younger than continental as it is

Figure 2. The principal active volcanoes of the world are not distributed randomly on the Earth's surface; they tend to be associated with plate boundaries.

being "continuously" created at mid-oceanic rift zones and destroyed at subduction zones. The plates, mentioned above, can be either continental or oceanic or both. It is the nature of the crust at the plate boundary, however, that influences the type of processes that occur.

Three types of plate-boundary systems exist; divergent, convergent, and transform (Figure 3). Divergent motion of plates produces extensional features resulting in the formation of rifts. The majority of the world's magmatic activity occurs within the rifts formed by diverging oceanic plates. A buildup of volcanic products produces topographic ridges referred to as oceanic rises. Continental rifts are also the result of extensional forces and have associated magmatic activity. The complete development of a continental rift, in fact, is thought to lead to the production of new oceanic crust.

Convergent plate-boundary systems are most dependent on the crustal nature of the two interacting plates. Processes associated with the subducting plates result in magmatic activity producing island-arc volcanism. In the western United States, the Cascade volcanoes typify this process. The convergence of two oceanic plates also produces magmatic activity but with considerably different characteristics. The convergence of two continental plates causes crustal compression resulting in mountain-building episodes and relatively minor volcanism.

Figure 3. Diagrammatic cross section showing the relationship of magma generation to plate tectonics (modified after Ringwood, 1969b).

Transform boundaries exist where two plates slide past each other horizontally. An example of this interaction is the San Andreas fault in California. Transform boundaries occur with all crustal variations, a characteristic significant to the nature of the associated seismic activity. Magmatism is a minor process in this tectonic region.

The driving mechanism for plate tectonics is poorly understood, but is generally agreed to be the result of convection within the underlying mantle. The cause and effect relationship between plate boundaries and mantle convection is a subject of considerable debate and well beyond the scope of this report.

The spatial relationship between plate boundaries and volcanism is quite evident in Figures 2 and 3. Volcanism occurs at spreading centers within the oceans and on the continents, above subduction zones, particularly well defined along the edges of the Pacific plate in what is called "The Ring of Fire," and at anomalous intra-plate locations. This latter class is referred to as "hot spots" and is widely thought to be the product of localized mantle plumes. These plumes apparently persist for considerable periods of time, as is evident from the series of volcanic islands in the Pacific that form as the Pacific plate passes over the "stationary" plume. The Hawaiian Islands-Emperor Seamounts chain, has progressively younger volcanic rocks toward the southeast demonstrating the validity of this concept. The Yellowstone volcanic complex has been hypothesized to be an example of an intracontinental hot spot.

Magma Associations

Three distinct magma series can be identified that relate to the tectonic environment in which they occur (Table 3). A magma series is a sequence of igneous rocks of differing composition that are implied to be genetically related. The three principal magma sources are: the tholeiite, calc-alkaline, and alkali series. The tholeiite series are dominated by basalt that is poor in alkalis (sodium and potassium). Tholeiites are the principal products of oceanic rift zones and subsequently are the dominant component of the oceanic crust. The calc-alkaline magma series are associated with subduction zones. Andesites are the dominant rock type, although the series include basalts, dacites and rhyolites. The alkali series, which are least abundant, are also characterized by basalts, but are distinguished from the tholeiite series by the relative enrichment in alkali elements. The alkali series, as seen in Table 2, appear in a variety of tectonic environments, often in association with tholeiites.

Volcanism

Volcanology is the study of magma processes at or near the Earth's surface. As this is the most visible of magma processes, much of our knowledge of magma systems must be extrapolated from observations of volcanic processes and products. Volcanic processes can be observed while a system is active or inferred from surface morphology and volcanic products. The opportunity to observe active systems is limited. The information currently being collected at Mount St. Helens in Washington will be a very significant

TABLE 3. PLATE TECTONIC CLASSIFICATION OF MAGMAS

PLATE SETTING	PLATE MARGIN		INTRAPLATE			
	CONVERGING (SUBDUCTION ZONE)	DIVERGING (OCEANIC RISE)	OCEANIC		CONTINENTAL	
			MARGINAL- SEA BASIN	LARGE OCEAN BASIN	RIFT SYSTEM	CRATONIC AREA
MAGMA SERIES	<p>CALC-ALKALINE</p> <pre> graph TD A[CALC-ALKALINE] --- B[ARC-THOLEIITE] A --- C[SHOSHONITE] D[CALC-ALKALINE] --- E[LOW-K] D --- F[HIGH-K] </pre>	THOLEIITE (LOW-K)	THOLEIITE (LOW-K)	ALKALI THOLEIITE	THOLEIITE ALKALI (SOMETIMES BIMODAL)	ALKALI

CONDIE, 1976

contribution to our knowledge of volcanism. We have already introduced the subject of volcanic products and will not extend that discussion here. Surface morphology of volcanic constructs is the combined product of a number of factors. The physical and chemical properties of the lava (magma at the surface) strongly influence the type of constructs formed. The proportion of the liquids, gases and solids in a system controls the profile of a construct. The more liquid, the lower the profile. Complementary to this is the influence of viscosity which in turn is controlled by the temperature of eruption, composition, amount and state of the volatiles, and degree of crystallization. The more viscous the fluid phase is, the less distance it would be expected to travel from the vent. Basalts have typically low viscosity, i.e. very fluid, and rhyolites high viscosities. Physical factors of eruption style, vent characteristics, rate, quantity, and duration of effusion, and pre-flow topography, also influence the final morphology. Table 4 shows a classification of volcanoes based on the factors discussed here. Figure 4 is a diagram illustrating the morphology of a variety of plutonic and volcanic structures.

Complementary to this classification, types of eruptions can be delineated based on the nature of the magma and the characteristics of the explosive and effusive activity. Table 5 illustrates one such scheme that is widely accepted. Much of the current research in volcanology is in modeling volcanic processes. This will be discussed later in this report.

TABLE 4.

CLASSIFICATION OF VOLCANOES (MODIFIED FROM RITTMANN, 1962)

QUALITY OF MAGMA	SMALL ← QUANTITY OF MAGMA → GREAT				TYPE OF ACTIVITY	
FLUID VERY HOT, BASIC	LAVA FLOWS		SHIELD VOLCANOES	BASALTIC PLAINS	LAVA FLOODS	EFFUSIVE
INCREASING GAS CONTENT AND SILICA ↓ VISCOUS, "COOL," ACIDIC	TEPHRA-AND-SPATTER CONES	COMPOSITE CONES	STRATOVOLCANOES	STRATO-VOLCANIC CHAIN		MIXED
	TEPHRA CONES					
	ENDOGENOUS DOMES, PLUG DOMES	DOMES WITH THICK FLOWS	NONE KNOWN		EXPLOSIVE	
	MAARS	TUFF CONES				
VERY VISCOUS, ABUNDANT CRYSTALS	DIATREMES	EXPLOSION CALDERA	VOLCANO-TECTONIC SINKS	IGNIMBRITE SHEETS		
	← SINGLE VENT →			FISSURE VENT ←

Figure 4. Block diagram of plutonic and volcanic structures. (After Q R.G. Schmidt and H.R. Shaw, U.S. Geological Survey.

Table 5 Classification of volcanic eruptions (modified after Macdonald 1972)

Eruption type	Physical Nature of the magma	Character of explosive activity	Nature of effusive activity	Nature of dominant ejects	Structures built around vent
Basaltic flood	Fluid	Very weak ejection of very fluid blebs; little lava fountaining	Voluminous wide-spreading flows of very fluid lava	Cow-dung bombs and spatter; very little ash	Small spatter cones and ramparts of limited extent; broad lava plain
"Plains" or Icelandic (sensu stricto)	Fluid	Weak ejection of very fluid blebs, little lava fountaining	Moderately wide-spreading flows of thin sheets, often through lava tubes and channels	Cow-dung bombs and spatter; very little ash	Very broad flat lava cone, (shields) which frequently coalesce, some spatter cones
Hawaiian	Fluid	Weak ejection of very fluid blebs; lava fountains	Thin, often extensive flows of fluid lava, frequently involving lava tubes and/or channels.	Cow-dung bombs and spatter; very little ash	Spatter cones and ramparts; very broad flat lava cones
Strombolian	Moderately fluid	Weak to violent ejection of pasty fluid blebs	Thicker, less extensive flows of moderately fluid lava; flows may be absent	Spherical to fusiform bombs; cinder; small to large amounts of glassy ash	Cinder cones
Vulcanian	Viscous	Moderate to violent ejection of solid or very viscous hot fragments of new lava	Flows commonly absent; when present they are thick and stubby; ash flows are rare	Essential, glassy to lithic, blocks and ash; pumice	Ash cones, block cones, block-and-ash cones
Peleean	Viscous	Like Vulcanian, commonly with glowing avalanches	Domes and/or short very thick flows; may be absent	Like Vulcanian	Ash and pumice cones; domes
Plinian (exceptionally strong Vulcanian)	Viscous	Paroxysmal ejection of large volumes of ash, with accompanying caldera collapse	Ash flows, small to very voluminous; may be absent	Glassy ash and pumice	Widespread pumice lapilli and ash beds; generally no cone building
Rhyolitic flood	Viscous	Relatively small amounts of ash projected upward into the atmosphere	Voluminous wide-spreading ash flows; single flows may have volume of tens of cubic miles	Glassy ash and pumice	Flat plain, or broad flat shield, often with caldera
Ultravulcanian	No magma	Weak to violent ejection of solid fragments of old rock	None	Accessory and accidental blocks and ash	Block cones; block-and-ash cones
Gas eruption	No magma	Continuous or rhythmic gas release at vent	None	None; or very minor amounts of ash	None
Fumerolic	No magma	Essentially nonexplosive weak to moderately strong long-continued gas discharge	None	None; or rarely very minor amounts of ash	Generally none; rarely very small ash cones

Greeley and King, 1976

Magma Processes

The varieties of igneous rocks sampled at the Earth's surface are the final products of a complex series of processes. The generation of magma, its ascent through and interaction with the upper mantle and crust, evolution in magma chambers at various depths, and its final emplacement and/or eruption, involves complex physical and chemical processes, most of which are unobservable to the geologist. In this section we will consider these processes, particularly as they pertain to the occurrence of magma in the western United States.

Magma generation

Petrologists like to think of magmas as either primary or derivative. Primary magmas are those most like their source rocks, while derivative magmas have undergone some sort of differentiation or contamination. Yoder and Tilley (1962), have shown from experimental work that tholeiitic and alkali basalts are the primary products of partial melting of a garnet peridotite mantle between 50 and 100 km, with tholeiites forming closer to the surface than alkali basalts. Green and Ringwood (1967) have also suggested that tholeiitic and alkali basalts are primary but differ due to the degrees of partial melting of a pyrolite mantle, with alkali basalts being a higher percent melt. The variations on these schemes are many. The nature of the source rock, depth of origin, degree of partial melting, pressure-temperature environment, and physical process of magma separation, all combine as factors considered in any primary magma generation model. The overwhelming occurrence

in space and time of basalts and their relative consistency in composition, however, has led to most models accepting basalts as being closest to primary magmas, and as such indicative of a fairly homogeneous global layer in the mantle.

Magma generation is initiated by melting due to: 1) a rise in temperature, 2) a release of pressure, 3) an influx of volatiles, or a combination of these mechanisms. A consideration of the generation of magma in a subduction zone will serve to demonstrate the wide variety of mechanisms that have been proposed as well as focus our attention on the tectonic environment appropriate for the western United States.

Thermal energy is created by frictional heating above the downgoing slab in the Benioff zone, which is the foci for most earthquakes in subduction zones. Due to the cool nature of most downgoing slabs, thermal models suggest that this alone would not produce a melt. However, as the slab descends, volatiles are released and rise through the overlying mantle decreasing the liquidus temperature (the melting temperature). This process may be aided by the generation of convection cells in the overlying mantle. The depth of melting would be a function of chemical composition, volatile abundance, strength of the convection, and thermal energy production, the latter being due to rate of subduction and properties of the mantle and lithosphere.

The petrogenic models for explaining the observed rocks above a subduction zone must account for: 1) the predominance of basaltic-andesite, 2) the position of the volcanic front, which is usually 100-200 km inward of the oceanic trench and marks the first appearance of volcanism, 3) regular chemical trends of increasing K_2O/Na_2O and decreasing Ca away from the line of initial subduction, 4) the calcalkaline magma series, and 5) isotopic and rare earth element patterns. All petrogenic models consider the magmas to be a product of mantle and crustal processes. Within the mantle two main models prevail, the direct melting of the subducted slab (see Marsh, 1979) and partial melting of the overlying mantle (see Nichols and Ringwood, 1973). The direct melting is thought to occur due to frictional heating, and hydrated state of the minerals. Thermal models (see Anderson et al., 1980) have shown, however, that frictional heating will be insufficient to melt the slab. Volatiles contained within the downgoing slab, however, are considered to play a key role in melting the overlying mantle (Wyllie, 1978). Fractional crystallization, a process by which a liquid becomes increasingly silicic due to settling out of mafic crystals, has been suggested as a method of producing the calc-alkaline series (Green and Ringwood, 1967). This process is unlikely to be major, however, as many fractionation trends are not observed.

The generation of magma at the base of the overlying crust has also been suggested, but thermally is difficult to defend. However, the interaction of magma generated within the mantle with the crust is considered, by some models, the means of producing

the calc-alkaline series. Figure 5 schematically shows possible origins of the calc-alkaline series. A summary table of models of andesite magma generation is presented (Table 6).

Magmatic processes change with time as the rate of subduction and dehydration changes. Figure 6 describes one scenario of magma evolution with time above a subduction zone. Also, convection seems to be more important at "mature" subduction zones and facilitates the development of "back-arc basins."

Ascent of Magma

The ascent of magma through the mantle and crust is controlled by properties of the magma as well as characteristics of the material through which the magma ascends. Two distinct phases of magma rise are: 1) the segregation of the initial melt and ascent through the plastic mantle and 2) the continued ascent through a brittle crust or lithosphere.

Within the mantle three scenarios of magma collection and ascent are considered (Williams and McBirney, 1979). 1) Magma could form and migrate through a dendritic system eventually coalescing into a single channel. 2) As the magma forms, an entire blob may become gravitationally unstable, because it is less dense than the surroundings. This would lead to a diapiric rise of the liquid with or without the solid phase. 3) A mechanism of zone melting could occur whereby a liquid rises by melting the overlying rock while crystallization occurs at the base of the mass.

Figure 5. Summary of possible origins of igneous rocks in continental margins and "mature" island arcs.

Green, 1980

Table 6. Some models for andesitic volcanism and their consequences for crustal evolution

	Model	Primary basis	Crustal evolution if dominant process	Comments	References (Partial List)
Mantle process	Andesitic liquid from partial melting of ultramafic mantle or subducted basaltic crust or deep fractionation of basaltic melt from mantle. Melt source in or near Benioff zone.	Voluminous nature of andesitic volcanism in subduction zones. Feasibility indicated by experimental data.	Addition of andesitic material to crust. Subduction of water-rich crust required for development of continental crust.	Does not by itself account for genesis of andesite outside subduction zones, more voluminous character and radiogenic composition of continental silicic magmas, or granitic character of upper continental crust.	Yoder (1969), Green and Ringwood (1968), Kushiro (1972, 1973, 1974), Mysen and Boettcher (1975).
	Melting of subducted sediments.	Transport of lowmelting material to mantle depth	No new continental material generated.	As above, but limited by isotopic data.	Coats (1962), Donnelly et al. (1971).
Crustal process	Andesitic liquid from large degree of melting of lower crust.	Voluminous nature of andesitic volcanism and intermediate batholiths in areas of thickened crust.	Mild fractionation of crust but no growth.	Does not account for thickening of crust accompanying volcanism. Requires excessively high temperature for lower crust.	Forbes et al. (1969), Bateman and Wahrhaftig (1966).
	Andesitic mush from lower crust consisting of rhyolitic liquid and refractory residuum.	As above.	As above.	As above, but avoids problem of high temperature.	Piwnskii (1973), Pressnall and Bateman (1973).
Mantle and crustal processes	Assimilation of granitic rocks by basaltic magma from mantle.	Xenocrysts in andesite.	Gradual oceanization of continents by dilution with basalt.	Does not account for andesites in island arcs. Limited by isotopic data.	Turner and Verhoogen (1960).
	Low pressure fractionation of basaltic magma from mantle to produce andesite. Partial melting of lower crust to produce dacitic liquid after thickening of crust.	Association of basalt with andesite. Development of batholiths in mature island arcs and continental margins.	Growth and fractionation of crust with time.	Yields basaltic bulk composition of crust unless refractory material returns to mantle. Requires very high temperature for lower crust. Does not account for higher $^{87}\text{Sr}/^{86}\text{Sr}$ of continental andesite.	Kuno (1968).
	Mixing of basaltic magma from mantle with rhyolitic magma from lower crust.	Phase assemblages and textures in andesite.	As above.	As above, but avoids problem of high temperature in lower crust while accounting for correlation of isotopic composition of andesite with crustal setting.	Piwnskii and Wyllie (1968), Younker and Vogel (1976). this paper.
	Andesitic melt from subducted slab and overlying mantle plus rhyolitic melt from lower crust.	Voluminous nature of andesitic volcanism in subduction zones. Silicic character of continental volcanism. Sr isotopic data.	As above.	Similar to above except that mafic parent magma is andesitic. Yields continental crust of andesitic composition. Does not account for higher $^{87}\text{Sr}/^{86}\text{Sr}$ of continental andesite or genesis of andesite outside subduction zones.	Brown (1977), Younker and Vogel (1976).

Eichelberger, 1978a

Fig. 6. Model for crustal evolution based on evidence of mixing in andesite. (1) Young island arc: thickening and heating of oceanic crust with emplacement of basaltic sills. Melting of amphibolite begins at base of crust and andesitic magma forms by mixing of mantle and crustal melts. Surface lavas are basaltic to andesitic, e.g. Mariana. (2) Mature island arc: Zone of melting expands as crust approaches continental thickness. Material undergoing melting is still generally amphibolite of basaltic composition. Surface lavas are basalt to rhyolite, andesitic on average. e.g. Lesser Antilles. (3) Continental margin or old island arc: Large volume of lower crust undergoes partial melting, including earlier emplaced basaltic material and silicic material from earlier episodes of crustal melting. Some refractory material may sink back into mantle. Surface lavas are basalt to rhyolite, dacitic on average. Granite batholiths accumulate in upper crust and average more silicic than volcanic piles because mixing favors eruption, e.g. Andes.

The mechanisms of each of these processes are dependent on physical properties of the magma, which change during its rise. Important parameters are transport properties (e.g., viscosity, electrical conductivity, and cation and volatile diffusion), physical properties (e.g., density, elastic properties), and thermal properties (e.g., thermal conductivity, heat capacity, thermal expansivity). Considerable experimental work on these subjects has been done, but suffers from lack of information concerning initial conditions and the ability to simulate the high pressure-temperature environment. Another major problem arises from the differences in behavior of a melt, pure liquid phase which can be more readily produced in a laboratory, and a magma, which includes gas and solid phases as well.

The migration of magma upward through a channel is driven by hydrostatic pressure (Fedotov, 1975). This pressure is the result of: 1) the increase in volume of the magma during melting, 2) buoyancy of magma, 3) the change in viscosity of the material and subsequent response to mantle pressures and/or boiling of the magma. These pressures increase as the volume of melt increases, until a critical driving pressure is reached. The diapiric rise model is driven by the difference in density of the liquid-crystal mush and the country rock. Whether or not the diapir remains detached is debatable. Magma production rates would certainly be an important consideration. Diapir detachment is suggested by episodic volcanism, distinct chemical characteristics, and variations in volumes. This implies rate of rise is greater than rate of production in most

tectonic environments. The rate of rise is controlled by the density ratio of the magma and host rock as well as the relative viscosities of the two materials.

The ascent of magma through the lithosphere involves many of the same mechanisms as found in the mantle, although the relative importance of processes varies. Ramberg (1970) experimentally determined that the relative viscosities of the crust and magma dominate control over magma motion. Where this ratio is high, the motion is controlled by the structures and stresses of the crust. Where it is low, the path of the rising magma will be unaffected. Implications of this model are that basalts will be more likely to be controlled by crustal properties than silicic magmas.

Viscosity of the magma, however, is not simply a function of composition. The role of volatiles in the ascending magma will greatly affect viscosity. Two offsetting situations occur. Volatiles in a liquid state will decrease viscosity, while those in a vapor state will increase it. Volatiles in vapor state, however, will increase the hydrostatic pressure of a magma body which would offset the increase in viscosity as it controls the magma motion. Figure 7 shows the feedback process that can occur in a conduit. The level at which volatiles vaporize will be important, as it could either stagnate the magma mass or accelerate its motion (Anderson, 1973). The thermal history of ascending magma is discussed by Marsh (1978). Magma moving in a blob or sphere could

Figure 7 Dynamic process of magma in a volcanic conduit.

Shimozuru, 1978

ascend much slower than that moving upward by crack propagation. In either case, the magma is expected to remain superheated through much of its upward migration. A thermal feedback mechanism such as that due to adiabatic heating is thought likely.

Emplacement of magma chambers

The controls on magma emplacement are essentially the same as those controlling the ascending processes. Magma properties and crustal stress dominate this process. In this study we are interested principally in the emplacement of magma in the mid- to upper crust, and as such concentrate on this situation.

Roberts (1970) considered the intrusion of magma into a brittle crust as a function of the relative horizontal and vertical stresses of the crust and the magmatic pressure. Figure 8 shows how these factors can predict the nature of the intrusion. In the case of a sill, Roberts' model suggests intrusion is normal to the least principal stress when the magma pressure exceeds the host rock's minimal tensile strength and stress.

Another hypothesis (William and McBirney, 1979) argues that sills form along planes where the hydrostatic pressure of the magma and lithospheric pressure of the crust are maximum. Where buoyancy is the principal component producing the hydrostatic pressure, sill formation can be described as occurring when the density of the magma exceeds the density of the host rock.

Figure 8. Relations of vertical and horizontal stress, σ_z and σ_x , tensile strength, T_0 , and magmatic pressure, P , to the form and orientation of intrusions.

Obviously, if this were the sole mechanism of sill formation, there would be few volcanoes.

Geologic field investigations of exposed intrusions is our principal means of understanding their history. Major mafic intrusive bodies are relatively few compared to those of the calc-alkaline series. The Bushveld complex in Africa and Skaergaard in Greenland are perhaps the best studied mafic bodies (see Wager and Brown, 1967). These occur predominantly in extensional tectonic environments. Large silicic intrusions, such as the Sierra Nevada Batholith, occur in compressional tectonic regions. Table 7 compares the major features of these two types of major intrusive complexes.

Magma evolution

The diversity of igneous rocks found at the Earth's surface is indicative of processes which modify primary magmas. Given that we must assume the nature of the primary magma, understanding the modifying processes is highly model dependent. There are clues, however, in the igneous rocks, such as bulk composition, mineral properties and texture, which provide considerable constraints on temperature of formation, rate of crystallization, and other important parameters. Thus, from field observations, petrologic analysis, and experimental work, three basic processes of magma evolution can be defined; magma differentiation, magma mixing, and assimilation. These processes are not mutually exclusive and undoubtedly occur in nature in a complex interplay.

**TABLE 7. COMPARISON OF THE FEATURES OF MAJOR
DIFFERENTIATED COMPLEXES**

	GABBROIC SERIES	CALC-ALKALI SERIES
STRUCTURAL FORM	STRATIFIED SILL, LOPOLITH, OR FUNNEL INTRUSION	COMPLEX BATHOLITHS OF MASSIVE STEEPWALLED PLUTONS, DOMES, OR MUSHROOM-SHAPED PLUTONS
STRUCTURAL ENVIRONMENT	NONOROGENIC : FAULTING DOMINANT	OROGENIC; BOTH FAULTING AND FOLDING PROMINENT
NUMBER OF INTRUSIONS	ONE OR FEW	MULTIPLE
WATER CONTENT OF THE MAGMAS	LOW	HIGH
MEAN TEMPERATURES OF DIFFERENTIATION	HIGH TO MODERATE	MODERATE TO LOW
DOMINANT CHARACTER OF MINERALOGY	ANHYDROUS GABBROIC	HYDROUS QUARTZO- FELDSPATHIC
TYPE OF DIFFERENTIATION	ABSOLUTE IRON ENRICHMENT, NO SILICA ENRICHMENT	RELATIVE IRON ENRICHMENT; STRONG SILICA ENRICHMENT
CONTACT EFFECTS	MODEST THERMAL	MODEST TO PROFOUND THERMAL AND METASOMATIC

MUELLER AND SAXENA, 1977

Magma differentiation considers the variations within a magma during crystallization. As crystals form in a magma they can separate from the liquid, by gravity settling or flotation, flowage separation or fluid removal, or remain in suspension and continue to react with the cooling liquid. The separation of the crystals, fractional crystallization, results in an evolution of the remaining liquid. The Skaergaard intrusion in Greenland (Wager and Brown, 1967) has been extensively studied as an example of this process. As fractional crystallization continues, the residual liquid becomes increasingly silica-rich. Where crystals remain in suspension they will equilibrate with the magma, eventually forming a rock with a chemistry similar to that of the initial magma. As a magma cools it can also differentiate due to the formation of immiscible liquid phases. The phases could theoretically separate and form separate magma bodies.

Magma mixing involves the production of hybrid magmas as a result of the mixing of two compositionally differing magmas. Evidence for this process occurring can be found in mafic inclusions within dacite and andesite lavas (Eichelberger, 1975), as well as mafic glasses within silicic tephras (Anderson, 1976). Whether or not this process is dominant in the generation of andesites and dacites is hotly debated (see Eichelberger, 1978a, and McBirney, 1980). Eichelberger, 1978a, proposes that basaltic magma, produced in the mantle, is emplaced as sills in the lower crust resulting in crustal melting and the production of rhyolite magma. The rhyolite ascends but subsequent basalt ascends along the same path, intrudes

into the rhyolite, mixing to form dacite or andesite (Figure 9). When this mixing occurs near the surface, rapid vesiculation can trigger explosive eruptions (Sparks et al., 1977).

McBirney (1980) argues that magma mixing, while feasible, is less likely than assimilation of felsic wall rock or absorption of water as a means of producing andesites.

Assimilation is the modification of a magma due to interaction with the wall rock it is in contact with. The degree to which this process occurs depends on the melting temperatures of minerals in the country rock and temperature of the magma. Minerals may become partially or completely melted and incorporated, or incorporated as a solid. Xenoliths are formed in this fashion. The release of volatiles from the wall rock can also greatly change the physical characteristic of the magma, as we have already discussed.

Plutonism and Volcanism

In this final section on magma processes, we consider the activity of magma within shallow chambers and its relation to volcanic processes. Magma chambers have been directly observed only after they have crystallized and been exposed; however, inference to the physical nature of magma processes within chambers can be deduced. Volcanic eruptions also give clues as to the processes within a chamber, but detailed observations are few. The nature of volcanic products, their fragmentation and dispersion, provides further evidence of activity. Finally, geophysical data such as

Figure 9. A model for evolution of silicic igneous complexes: (a) Basalt from the upper mantle heats the crust. (b) Partial melting in the lower crust produces rhyolite liquid which gathers into diapirs and ascends. In a strongly block faulted region (top) the diapirs reach the upper crust rapidly, and little mixing occurs (e.g., Yellowstone plateau). In areas lacking extensional tectonics (bottom), mixing is thorough (e.g., Mount Baker). The case in the middle is intermediate; mixing occurs, but some unmixed rhyolite reaches the surface (e.g., Lassen Volcanic National Park). Large chambers in these centers are compositionally zoned as shown (e.g., San Juan Volcanic Field). Activity may cease with eruption or emplacement of silicic bodies if the basalt flux declines. (c) If rhyolite is not replenished due to depletion of the lower crust but basaltic activity continues, magma in the chamber becomes more mafic, and basalt vents encroach on the silicic center (e.g., Long Valley).

seismic profiles and activity, gravity anomalies, and heat flow, present another line of evidence.

Three multidisciplinary studies of magma chambers have been made and reported in the literature: Long Valley (see Bailey et al., 1976) and Coso (See Duffield et al., 1980) in California, and Yellowstone, Wyoming (see Smith et al., 1977).

Thermal models of magma chambers have been discussed by Jaeger (1964, 1968), Kolstad and McGetchin (1978), and Spear (1980), among others. Convection processes have been discussed by Bartlett (1969) who deduced a relationship between viscosity and size controls convection (Figure 10).

We have briefly discussed the effect of magma mixing on a chamber. Sparks et al., (1977) suggest that as the mafic magma intrudes, local superheating of the silicic magma induces vigorous convection and vesiculation which increases the fluid pressure of the system resulting in an eruption (Figure 11). Eichelberger (1980) suggests that water in the mafic magma plays a key role in the vesiculation process.

Smith (1979), in an excellent review paper of ash-volcanism, relates surface volcanic deposits and caldera size to the volume and depth of the magma chamber (Figure 12).

Figure 10. Magma Convections - Stability Limits in Plutons

Figure 11. Schematic model of the mechanisms of triggering explosive eruptions by magma mixing. a. Basic magma is intruded into the base of acid magma. b. Pillows are formed and local superheating of the acid magma induces vigorous convection and vesiculation. c. The combination of intrusion of basic melt and vesiculation results in an increase in the fluid pressure of the magma system on the confining walls. Fractures form above the magma and an eruption is triggered.

Figure 12. Model showing relationships among volume, depth, periodicity, and composition.

Smith, 1979

Conclusion

This review of igneous geology has hopefully helped the reader appreciate the complexity of magma processes. At the same time, some of the models may stimulate thoughts and ideas for new research. The prime objective of this report has been to provide a basic understanding of the variety of magma process and facilitate future research.

The major breakthrough in our understanding of magma systems in the coming years will be in widely diverse fields. Models of magma behavior will need a better data base of basic magma properties as a function of temperature, pressure, and composition. Improving our understanding of lithospheric stress regimes and plate tectonics will aid new magma intrusion models. Field data on individual igneous bodies and regional histories will also continue to play a major role in our development of ideas.

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