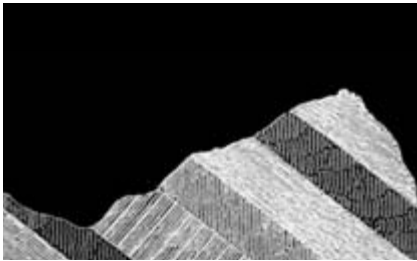
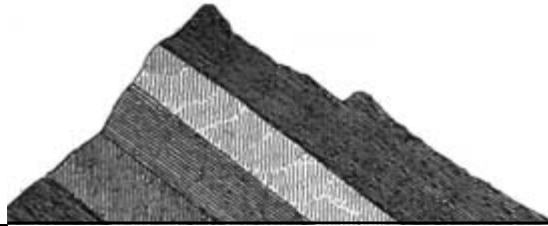
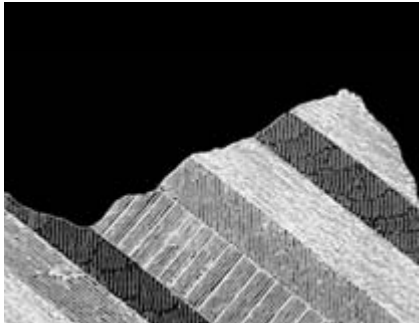


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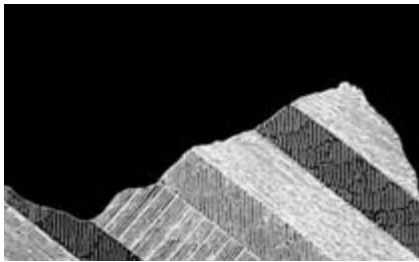
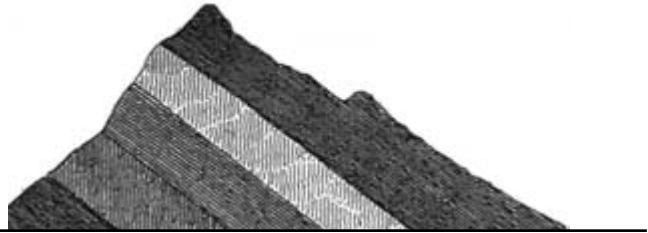


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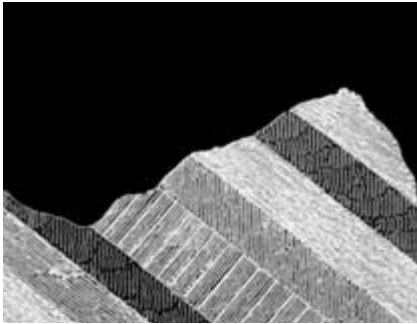


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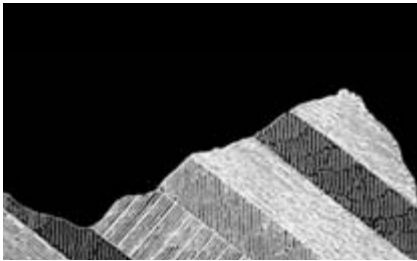
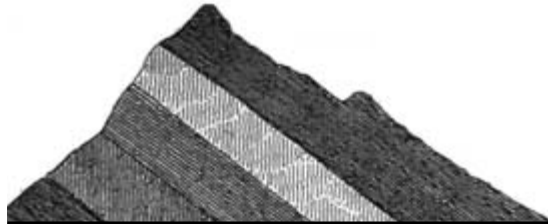
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Myron G. Best

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Editorial Offices:

350 Main Street, Malden, MA 02148-5018, USA
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First published 2003 by Blackwell Science Ltd

Library of Congress Cataloging-in-Publication Data

Best, Myron G.
Igneous and metamorphic petrology / Myron G. Best.—2nd ed.
p. cm.
Includes bibliographical references and index.
ISBN 1-40510-588-7 (alk. paper)
1. Rocks, Igneous. 2. Rocks, Metamorphic. I. Title.
QE461 .B53 2002
552'.1—dc21

A catalogue record for this title is available from the British Library.

On the cover: Photographs of rock from the upper mantle and deep continental crust. Beneath the lettering is a photomicrograph of peridotite viewed in cross-polarized light. The other rock lies in an outcrop in Swaziland along the Ngwempezi River and is Archean mafic gneiss that was probably derived from a basaltic protolith initially metamorphosed at 3.5 Ga, making it one of the oldest rocks exposed on the African continent. The gneiss was subjected to at least three episodes of deformation and nine intrusive events, the youngest product being a 2.6 Ga felsic pegmatite seen here. Photograph courtesy of Cees Passchier.

Set in 10/12pt Simoncini Garamond
by Graphicraft Limited, Hong Kong
Printed and bound in
Italy by G. Canale & C. S.p.A., Turin

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PREFACE



Igneous and metamorphic petrology in the last decades of the twentieth century exploded into a broad, multifaceted, increasingly quantitative science. Advances in physical and field petrology and geochemistry have forever changed our thinking about the origin and evolution of magmas, their dynamic behavior, and the way in which they are intruded and explosively extruded. Developments in geochronology, quantitative evaluation of the role of heat and fluid transfer in crustal rocks, and new field discoveries have impacted our understanding of the evolution of metamorphic systems and their dynamic interaction with tectonic processes. Geophysics and mineral physics have provided new insights into the nature of the convecting mantle and its role as a giant heat engine driving magmatic and metamorphic processes. New tools of all kinds allow new ways of gathering petrologic data, while phenomenal developments in computers and computer software permit data to be stored, processed, and modeled in ways unimaginable as recently as a couple of decades ago.

It has been a very daunting challenge to create within one book of reasonable length a balanced comprehensive coverage of topics that embodies the classical as well as the new advances. I hope that this textbook will provide a foundation for future geologists that not only informs but furnishes the intellectual mindset enabling them to pursue higher levels of professional endeavor. I have attempted to emphasize controlling petrologic processes in the formation of rocks, while not sacrificing basic descriptive information about them, on which interpretations of their origin must be firmly based. The organization of chapters is essentially by process rather than by rock type or association. The overarching themes of this textbook are the dynamic interactions between matter and energy and the ways in which transfers and transformations of gravitational and thermal energy drive changes in rock-forming systems.

This textbook has been designed as a balanced instructional tool for the college sophomore or junior. It is assumed that the student is acquainted with basic chemistry, physics, mineralogy, and physical and historical geology. A background in optical mineralogy is desirable. As for mathematical background, a course in calculus will be helpful but not essential. The mathematical inclination and capability of geology students vary widely and so as to avoid intimidating some at the outset, I have generally limited the more quantitative material to certain chapters, set-aside boxes, and problems at the end of most chapters. The intent is not to minimize the growing importance of the quantitative facets of petrology. Some of the problems are amenable to attack by computers and spreadsheets. I assume that

instructors have their own favorite computer-based teaching exercises. “Fundamental questions considered in the chapter” provide a brief preview of each chapter. “Critical thinking questions” at the end of each chapter provide an incentive for the student to think about the chapter contents. A comprehensive glossary and index are included at the end of the textbook together with a list of references cited. Space limitations permit citation of only the most crucial references or recent lucid summaries and select early classical works.

I am indebted to many persons who assisted in crafting this textbook over its five and one-half year gestation. Before starting, evaluations of the first edition and constructive advice on the course to take in the second were provided by William D. Carlson, Colin Donaldson, Michael O. Garcia, Edward Ghent, Scott S. Hughes, Douglas Smith, and Ron Vernon. Eric H. Christiansen thoroughly critiqued Chapters 1–13, greatly improving their clarity and accuracy, and in addition provided many hours of beneficial discussion and computer assistance. Stephen T. Nelson and Ron Harris offered valuable help. Bill Carlson and Douglas Smith also provided an opportunity to study the metamorphic rock collection at the University of Texas at Austin and helped in capturing images of thin sections. Dan Barker, Fred McDowell and especially Doug and Jean Smith were gracious hosts. Individual chapters were constructively reviewed by Katherine Cashman (University of Oregon), Mihai Ducea (California Institute of Technology), Michael Garcia (University of Hawaii), Charles Lesher (University of California at Davis), Calvin Miller (Vanderbilt University), Raj Sharma (Western Michigan), Suki Smaglik (Metro State College of Denver), Douglas Smith (University of Texas at Austin), Marian B. Holness, G. T. R. Droop. Without their input this textbook would be much less than it is. At Blackwell Science, Simon Rallison first contacted me about doing a new edition and Jane Humphreys followed up. Nancy Duffy saw the work almost throughout its lengthy gestation; her good-natured patience and forbearance and positive interactions and advice will forever be appreciated. Jill Connor, Rosie Hayden, and Delia Sandford were always cheerfully available for help and information. Irene Herlihy patiently responded to endless queries and tactfully coordinated the illustrations and manuscript for the igneous chapters. I am also indebted to Ian Francis, Lisa Flanagan, Manufacturing Manager, Nancy Whilton, Publisher/Science Books, and Graphicraft Ltd, who labeled the illustrations and typeset the text.

Myron G. Best
Provo, Utah

SI Base units

PHYSICAL QUANTITY	SI UNIT	SYMBOL
Length	Meter	m
Mass	Kilogram	kg
Time	Second	s
Temperature	Kelvin	K
Amount of substance	Mole	mol

SI Derived Units

PHYSICAL QUANTITY	SI UNIT	SYMBOL	IN TERMS OF BASE UNITS	IN TERMS OF OTHER SI UNITS
Area			m^2	
Volume			m^3	
Density			kg/m^3	
Specific volume			m^3/kg	
Velocity (speed)			m/s	
Acceleration			m/s^2	N/kg
Force	Newton	N	$kg\ m/s^2$	J/m
Pressure (stress)	Pascal	Pa	$kg/(m\ s^2)$	N/m^2
Energy	Joule	J	$kg\ m^2/s^2$	N m
Power	Watt	W	$kg\ m^2/s^3$	J/s
Viscosity			$kg/(m\ s)$	Pa s
Heat flux			kg/s^3	W/m^2
Heat capacity (entropy)			$kg\ m^2/(s^2\ K)$	J/K
Specific heat capacity			$m^2/(s^2\ K)$	J/(kg K)
Thermal conductivity			$kg\ m/(s^3\ K)$	$W/(m\ K)$
Molar heat capacity			$kg\ m^2/(s^2\ mol\ K)$	$J/(mol\ K)$

SI Prefixes

FACTOR	PREFIX	SYMBOL
10^9	Giga-	G
10^6	Mega-	M
10^3	Kilo-	k
10^{-2}	Centi-	c
10^{-3}	Milli-	m
10^{-6}	Micro-	μ

(From Le Système International d'Unités; see National Bureau of Standards Special Publication 330, July, 1974.)


Other units—including the “CGS system” of centimeters, grams, seconds, and calories—used in the geologic literature.

PHYSICAL QUANTITY	UNIT	SYMBOL	IN TERMS OF OTHER UNITS
Length	Angstrom	Å	10^{-10} m
	Micrometer (micron)	μm	10^{-6} m
	Centimeter		10^{-2} m
	Mile		1.61 km
Mass	Gram	g	10^{-3} kg
Time	Minute	min	60 s
	Hour	h	3600 s
	Day	d	86,400 s
	Year	y	3.15×10^7 s
	Million years	My	10^6 y
	Annum (years before present)	a	
	Mega annum	Ma	10^6 a
Giga annum	Ga	10^9 a	
Temperature	Degrees Celsius	°C	K + 273.15
	Degrees Fahrenheit	°F	$1.8^\circ\text{C} + 32$
Force	Dyne		10^{-5} N = 1 g cm/s ²
Pressure (stress)	Bar	bar	10^5 Pa = 0.1 MPa = 10^6 g/(cm s ²) = 10^6 dynes/cm ² = 0.1 J/cm ³
			Kilobar
	Standard atmosphere	atm	101,325 Pa = 1.01325 bar = 14,695 pounds/inch ²
Volume		cm ³	10^{-6} m ³ = 0.1 J/bar
Density		g/cm ³	10^3 kg/m ³
Viscosity	Poise	poise	0.1 Pa s
Energy	Calorie	cal	4.184 J

FUNDAMENTAL PHYSICAL CONSTANTS

Acceleration of gravity	9.8 m/s ²
Avogadro number	6.022136×10^{23} /mol
Molar gas constant (R)	8.31451 J/(K mol)
Boltzmann constant	1.380658×10^{-23} J/K

(From Anderson, 1996.)



Overview of Fundamental Concepts

FUNDAMENTAL QUESTIONS CONSIDERED IN THIS CHAPTER

1. What role is played by energy in its various forms to create magmatic and metamorphic rocks?
2. What is the source of internal thermal energy in the Earth; and how does it function as a giant heat engine to drive rock-forming processes?
3. What is the role of the mantle of the Earth in rock-forming processes?
4. In what way does mantle convection focus rock-forming processes in specific tectonic settings?
5. What do changes in geologic systems have to do with the formation of rocks?
6. What are the most significant properties of rocks, and what specific information does each property provide about rock-forming processes?
7. How does a petrologist study rocks to determine their nature and origin?

INTRODUCTION

This book is about rocks that were once hot. **Magmatic rocks**, also called **igneous rocks**, form by cooling and solidification of **magma**, which is mobile molten rock material whose temperature is generally in the range of 700–1200°C (about 1300–2200°F) near the surface of the Earth. **Metamorphic rocks** form by reconstitution of pre-existing rocks at elevated temperatures well beneath the surface of the Earth. Both classes of rocks possess textures, structures, and mineral constituents

indicative of their high-temperature ancestry. When sampled and studied by geologists, these rocks not only have cooled, but in many cases have been brought by geologic processes to the surface from some considerable depth in the crust or mantle. Obviously, the origin of these once-hot rocks, and their exposure at the surface, involves flow of heat as well as movement of rock mass in the gravitational field of the Earth. Thus, interactions between heat and gravity are involved in their creation. Understanding the nature of magmatic and metamorphic rocks and the related interactions between matter and energy is to understand, in a major way, how the planet Earth works.

With the development of concepts of plate tectonics in the 1960s all concepts of a static Earth became obsolete. Plate motion—a basic facet of the way the Earth works—manifests the interaction between gravity and outward flow of heat from the hot interior of a cooling, dynamic Earth. Oceanic lithosphere that is cooler and therefore denser than the underlying asthenosphere sinks at subduction zones. Plumes of hot mantle rock rising from near the core–mantle boundary and upwelling mantle beneath oceanic spreading junctures constitute the return circuit in the global mantle convection system. Seafloor spreading from the oceanic junctures maintains a constant global surface area, compensating for subduction.

Most magmatism and metamorphism in planet Earth occurs along the two linear tectonic regimes of plate convergence and divergence because that is where most interactions between energy and matter take place. Active volcanism and related but hidden intrusive magmatism is focused over less than 0.6% of the surface of the Earth, assuming a modest width of 100 km along the boundaries of converging and diverging plates (Plate I). Hidden from view beneath the sea,

about three-fourths of global magmatism is estimated to occur along the world-encircling system of oceanic spreading ridges (Figure 1.1). Hot submarine magmatic rocks interact with seawater at oceanic ridges, become metamorphosed, and, through subsequent plate motion, can be emplaced on overriding plate margins at converging plate junctures alongside other metamorphic rocks created by heating and tectonism.

Localized and volumetrically minor magmatism far removed from plate boundaries is commonly related to mantle plumes ascending through the mantle. Such intraplate activity is manifest, for example, as volcanism in the Hawaiian Islands.

*1.1 ENERGY AND THE MANTLE HEAT ENGINE

Without a critical amount of thermal energy within a planetary body there can be no movement of lithospheric plates or rise of mantle plumes and hence no magmatism, metamorphism, or tectonism. The geologically dead Moon, for example, has been too cold for billions of years for any such geologic activity. However, throughout its approximately 4.5-billion-year existence, the Earth has acted as a giant heat engine, powering all kinds of geologic processes. In this engine, the mantle of the Earth reigns supreme as the major source of driving energy. It is by far the most voluminous part (84%) of the planet, has the most mass (68%), and

stores the most thermal energy. Ultimately, in one way or another, most magmatic rocks and magmas trace their ancestry to the mantle.

To understand how the Earth works as a heat engine driving rock-forming processes it is important to understand the various forms of energy, the ways they are transferred and converted into other forms, and the sources of thermal energy within the Earth.

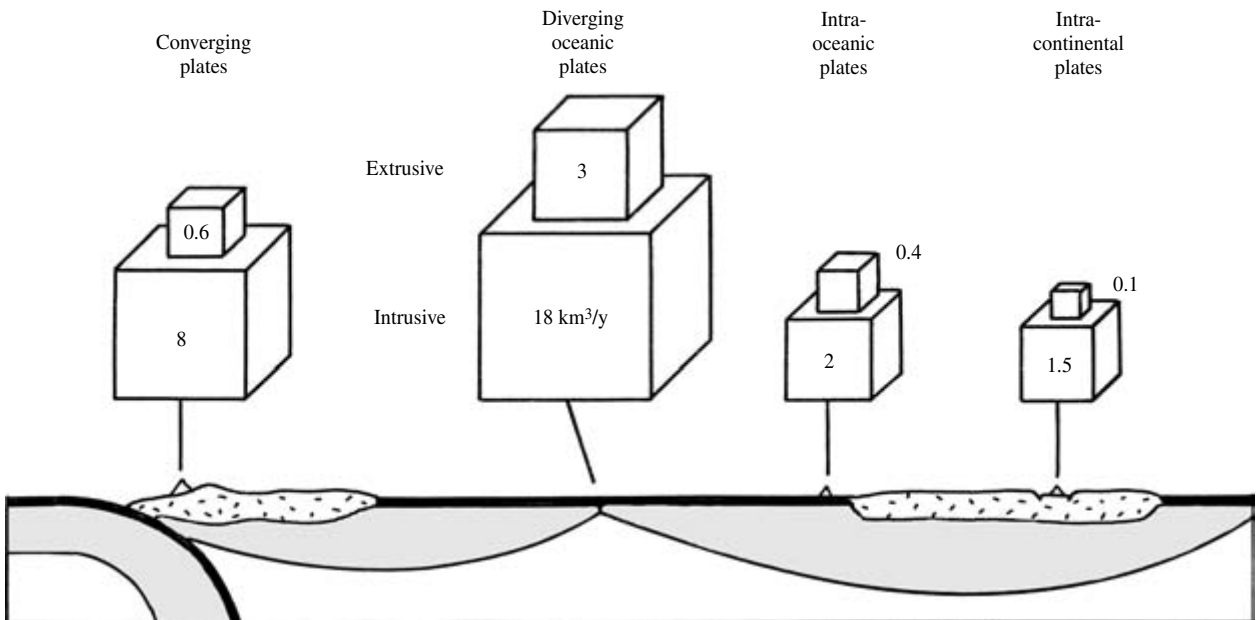
1.1.1 Forms of Energy

Energy exists in various forms and is manifest in terms of motion, or potential for motion, and by the temperature of matter. An asteroid approaching the Earth, a high mountain from which boulders can be rolled downhill, an exploding volcano, and a hot lava flow all have energy, but in different forms. (Some forms of energy, such as magnetic energy, are important in man-made machines, but the geomagnetic field of the Earth is too weak to cause geologically significant movement of matter.)

Energy is commonly defined as the capacity for doing work. **Work**, w , is defined as the product of force, F , times a displacement over a distance, d , in the direction of the force

$$w = Fd \quad 1.1$$

Thus, for example, energy is required to perform the work of shoving a thrust plate in an actively growing compressional mountain system or of throwing a frag-



1.1 Global inventory of magma production in different plate tectonic settings (numbers in cubic kilometers). Estimates of the ratio between erupted magma and magma lodged as intrusions in the crust vary, depending on geologic factors and considerable uncertainties in the interpretations of the geologist. Production of basaltic magma predominantly in oceanic settings and mostly along ocean ridges far exceeds that of any other magma composition in any tectonic regime. (From Schmincke H-U. *Vulkanismus*, 2. Darmstadt, Wissenschaftliche Buchgesellschaft, 2000. With permission of Wissenschaftliche Buchgesellschaft, Copyright © 2000.)

Worked Problem Box 1.1 How much energy is required to lift this textbook 1 meter above the table?

Assume the book weighs 1 kg and the acceleration of gravity is 9.8 m/s^2 . The increase in gravitational potential energy equivalent to the work, w , is $E_p = w = mgz = 1 \text{ kg} \times 9.8 \text{ m/s}^2 \times 1 \text{ m} = 9.8 \text{ J}$. (See the inside cover for units and conversions between them.) For comparison, one beat of the human heart consumes about 1 J and a small cup of water, $3.7 \text{ cm} \times 3.7 \text{ cm} \times 3.7 \text{ cm} = 50 \text{ cm}^3$ in volume, heated by 1000 J = 1 kJ of thermal energy raises its temperature by 5°C .

ment of rock from an explosive volcanic vent. An important type of work in geologic systems is called **PV work**, where P is the pressure, such as possessed by a volcanic gas, and V is the volume of the gas. Expansion of pressurized gas does work in displacing magma out of a volcanic vent, creating an explosive eruption. Because **pressure** is defined as a force divided by the area over which it acts, $P = F/\text{area}$, and because volume, $V = \text{area} \times d$, then $PV = Fd = w$.

Kinetic energy is associated with the motion of a body. A body of mass, m , moving with a velocity, v , has kinetic energy

$$E_K = \frac{1}{2}mv^2 \quad 1.2$$

A moving lava flow, ejecta thrown from an exploding volcano, and agitating molecules in a gas all have kinetic energy.

Potential energy is energy of position; it is potential in the sense that it can be converted, or transformed, into kinetic energy. A boulder cascading down a hill slope gains velocity and, therefore, kinetic energy as it loses potential energy. Potential energy can be equated with the amount of work required to move a body from one position to another in a potential field, in this instance, the gravitational field of the Earth. In lifting a boulder of mass m through a vertical distance z in the gravitational field of the Earth, whose acceleration is g , the amount of work equivalent to the **gravitational potential energy** is

$$E_P = mgz \quad 1.3$$

The distance z is measured outward from the Earth above some reference level. Thermal energy within the Earth is expended to do the work of uplifting a mountain range, which imparts increased gravitational potential energy to the mountain mass.

Operating a bicycle tire pump demonstrates that mechanical work can be converted, or transformed, into thermal energy. As the pump handle is repeatedly depressed, the pump piston's rubbing on the inside of the cylinder produces frictional heating of the pump cylin-

der; in addition, the work of compressing the air in the cylinder heats the air. The increased temperature of the tire pump is a manifestation of an increase in the thermal energy internally within the metal parts of the pump. The **thermal energy** of a body resides in the motions—kinetic energy—and the attractions—potential energy—of the atomic particles within it. An increase in the internal thermal energy of a solid is associated with greater kinetic energy via faster motion of the atoms and is manifest in a greater temperature, T . This motion can become sufficiently vigorous to break atomic bonds momentarily so that the solid becomes a flowing liquid, or, if bonds are fully broken, a gas. The term *heat* is sometimes used synonymously with *thermal energy*, but, strictly speaking, **heat** is transferred thermal energy caused by a difference in temperature between bodies. For example, the thermal energy of a magmatic intrusion is reduced as heat moves into the surrounding cooler wall rocks, heating them to a higher T .

The *joule*, J, is the fundamental unit of energy (see the inside cover for units used throughout this textbook).

1.1.2 Flow and Transformation of Energy

In nature, energy moves, is transferred, or flows from place to place. Energy is also exchanged, converted, or transformed, from one form into another. Thus, decay of an unstable radioactive U nucleus emits high-speed smaller particles whose kinetic energy is transformed into thermal energy that heats the mineral hosting the U atom. As rocks adjacent to a magmatic intrusion are heated, they expand and exert an increased pressure on adjacent rocks, displacing them outward and doing PV work on them. Thermal energy and work are, therefore, interconvertible. And work can be converted into thermal energy—such as in a tire pump. PV work is a transfer of energy due to a difference in pressure; heat is a transfer of thermal energy due to a difference in temperature, T . In all such flows and transformations of energy the total amount is rigorously and quantitatively conserved in agreement with the law of conservation of energy, also called the first law of thermodynamics.

This law claims that the total amount of energy and mass in the universe is constant. The total amount of energy is not added to or subtracted from; it only moves about and is converted to other, perhaps less obvious, forms. In all such flows and transformations we are concerned with *changes* in the amount of energy. In contrast, the total, or absolute, amount of energy residing in a system is difficult to evaluate and generally is unimportant.

1.1.3 Heat Flow in the Earth

Within Earth systems the transfer of thermal energy, or flow of heat, is especially important and is therefore

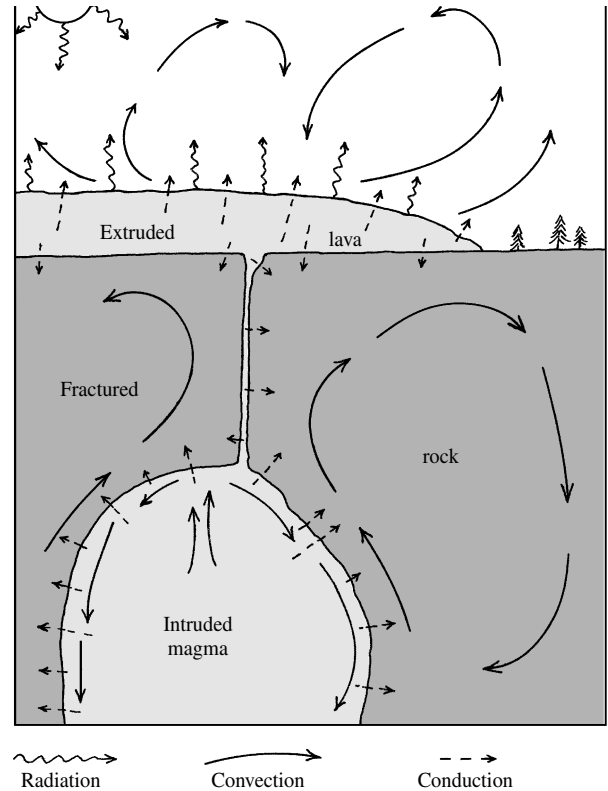
considered further here. Movement of thermal energy is obviously involved in magmatic rock-forming processes, such as heating solid rock so it melts, forming magma. On a larger scale, cooling oceanic lithosphere becomes denser and sinks as subducting slabs into the hotter, less dense upper mantle. Without heat, the Earth would be geologically dead.

An increment of heat, Δq , transferred into a body produces a proportional incremental rise in its temperature, ΔT , given by

$$\Delta q = C_p \Delta T \quad 1.4$$

where the proportionality constant, C_p , is the **molar heat capacity**, with units of J/(mol degree). The subscript in C_p indicates the heat capacity is for a condition of constant pressure, a common geologic situation, as for example, when rock is being heated by a nearby magmatic intrusion at a particular pressure in the Earth. The heat capacity based on mass is the **specific heat**, with units of J/(g degree). Because 1 calorie of heat (1 cal = 4.184 J) raises the T of 1 g of water 1 degree (Celsius or Kelvin), the specific heat of water is 1 cal/g degree. However, rocks generally have specific heats of 0.25–0.3 cal/g degree, which means that a given amount of heat, Δq , can raise the T of a mass of rock three to four times more than it can an equal mass of water. In other words, water absorbs relatively large amounts of heat per unit of mass for a particular T increase, ΔT ; it is an effective thermal transfer agent and moderator. Use of water in building heating systems and in automobile radiators is serendipitous because water is also inexpensive and readily available. In geologic systems, water absorbs considerable heat from nearby magmatic intrusions and as it moves through cracks can effectively transport this heat to distant rock, changing its T .

Heat can be transferred in four different ways: radiation, advection, conduction, and convection. Commonly, two or three of these act in unison, as in the cooling of the lava flow in Figure 1.2. **Radiation** involves emission of electromagnetic energy from the surface of a hot body into transparent cooler surroundings, such as the Sun into surrounding space or a hot lava into the atmosphere. In a vacuum, this energy moves at 277,800 km/s, the speed of light. Radiation is insignificant in cool rocks because they are opaque, but the effectiveness of radiative transfer increases exponentially with T as rocks become more transparent above about 1200°C. **Advection** involves flow of a liquid through openings in a rock whose T is different from that of the liquid. Because all rocks near the surface of the Earth are fractured on some scale and because these fractures are, at least partly, filled with water, advection is a significant heat transfer process. For example, hot water heated by a nearby magmatic intrusion advects through cracks in cooler rock, heating it while moderating the T of the water. The greater heat



1.2 Schematic diagram (not to scale) showing four modes of heat transfer. Heat from an intrusive body of magma, in which convection may occur, conducts into the cooler wall rock, where heat is further transferred away by advective flow of heated groundwater through interconnected cracks. Heat is mainly dissipated from the top of the lava by conductive and radiative transfer into the overlying air, which expands and buoyantly convects upward so that cooler air descends, is heated, expands, and ascends.

capacity of water than of rock makes advective heat transport more effective. Advective heat transfer is also important where magma penetrates cooler rock.

Conduction. Transfer of kinetic energy by vibrating atoms in any material is called **conduction** of heat. Heat cannot be conducted through a perfect vacuum because of the absence of atoms. Imagine a box filled with rigid balls (representing atoms) all interconnected by springs (representing atomic bonds). If a ball in one corner of the box is set into motion (i.e., is given kinetic or internal thermal energy) all of the balls in the box eventually will be set into motion and given kinetic energy, but the motion and energy of any individual ball are less than those for the ball in the corner because the initial energy input is dissipated throughout the box. Internal kinetic energy moves throughout the box, manifest in heat conduction. Heat always flows from a hotter region, where atomic motion is greater, to a cooler region, where motion is less. A cool metal pan on a hot stove becomes hot as a result of conduction

through the metal. Heat from a magmatic intrusion conducts into the enclosing cooler rocks, which become hotter and may be metamorphosed, while the magma cools. (In this instance, conduction acts in concert with advection of water moving through cracks in the wall rocks.) For a given volume, a hot body conducts heat away faster if its enclosing surface area is larger; this is why air-cooled engines have attached fins to dissipate the heat faster.

The difference in T between adjacent hotter and cooler masses, called the **thermal gradient**, is reduced and may eventually be eliminated over sufficient time, provided heat is not restored to the hotter mass. The rate at which heat is conducted over time from a unit surface area, called the **heat flux** or **heat flow**, is the product of the thermal gradient and the thermal conductivity, or

$$\text{heat flow} = \text{thermal conductivity} \times \text{thermal gradient} \quad 1.5$$

Because of their extremely low **thermal conductivity**, compared with that of familiar metals, rocks are considered to be thermal insulators. All other factors being equal, copper conducts heat nearly 200 times faster than rock. Because of the low thermal conductivity of rock and the large dimension of the Earth (radius about 6370 km), little heat has been conducted from the deep interior over the lifetime (4.5 Ga) of the Earth (Verhoogen, 1980).

Thousands of measurements all over the planet since the 1950s reveal that the surface heat flow from the hotter interior averages about 0.09 watt/meter² (W/m²). If one recalls the wattage of a common incandescent light bulb, say 60–100 W, this is an extremely small quantity of heat! Significant variations in the heat flow and corresponding geothermal gradient depend on the plate tectonic setting. The **geothermal gradient**, or **geotherm**, expressed as the change in temperature divided by the depth interval over which it occurs, or $\Delta T/\Delta z$, has been found to vary from hundreds of degrees per kilometer beneath oceanic spreading ridges to about 20–30°C/km in active orogenic belts alongside convergent plate junctures to as low as 7°C/km in the nearby deep-sea trench. These variations in gradient might reflect lateral variations around the globe in thermal conductivity of rocks, in their radiative transparency, or in heat transferred by another mechanism. As the first two possibilities involving lateral variations in rock properties are unreasonable, the possibility of another mechanism of heat transfer should be considered.

Another argument illustrates that global heat flow may not be solely by conduction. If a modest geotherm of 20°C/km is extrapolated to a depth of, say, 200 km, the temperature there would be 200 km \times 20°C/km = 4000°C. This is an impossible T because it exceeds the

Special Interest Box 1.2 Experimental petrology of the deep interior of the earth.

Laboratory devices that can create the P - T conditions prevailing to the center of the Earth provide important information on which minerals might constitute the deep interior. All of the devices produce high pressures in basically the same manner by squeezing a sample between opposing pistons or anvils to which force is applied. Various gasket materials are used to contain the sample between the anvils. Bench-top piston-cylinder devices can create pressures as much as 40 kbar (4 GPa) equivalent to depths of roughly 120 km. These consist of two opposing hard-metal pistons, one driven against the other by a hydraulic jack, and the sample is constrained within a cylinder into which the pistons move. The sample is heated by a furnace that surrounds the cylinder and can be opened at the end of the experiment so a blast of cool air can be directed at the sample assembly to “quench” the high P - T run products. Multianvil presses can achieve pressures to as much as 35 GPa (roughly 850-km depth), weigh several tons, and consist of four or six hydraulic pistons that move from their rigidly supported cylinders and converge symmetrically upon a tetrahedral or cubic sample assembly. This can be a block of soapstone that serves as an extrudable gasket between the carbide anvils at the ends of the pistons. Inside the block is the sample surrounded by a small electrical resistance furnace that is destroyed during the experiment. The third and most intriguing device consists of two small faceted diamonds held between the jaws of a hand-size “nutcracker” device (Jayaraman, 1984). The mechanical leverage exerted by the nutcracker arms pulled together by a large screw and the small diameter (about 100 micrometers = 100 microns) of the faces of the opposing gem diamonds can create pressures exceeding those at the center of the Earth. Not only are the diamond anvils very strong, they are also transparent to a laser beam for heating the sample, to light used for directly observing the sample, and to X rays to do diffraction analysis of the sample while under high P and T . Yet another high- P - T technique employs shock waves created by firing a projectile at a fixed target.

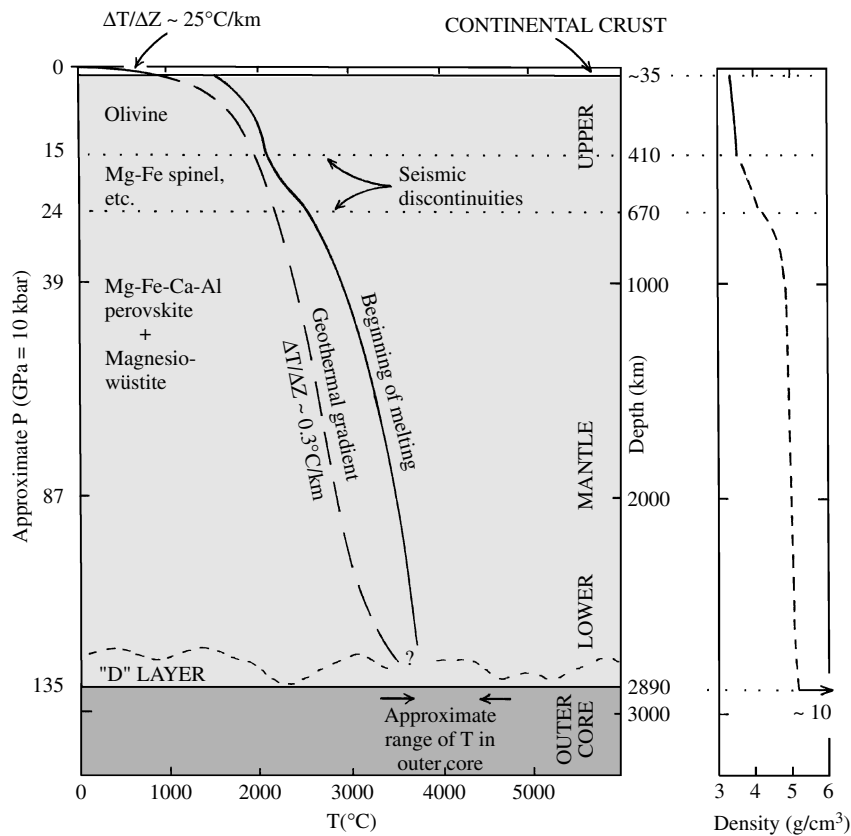
Information provided by high- P - T devices has revolutionized understanding of the mineralogical composition of the interior of the Earth and has answered questions posed by seismic data. Although much remains to be learned, the upper mantle appears to consist of peridotite—a rock made mostly of Mg-rich olivine and subordinate monoclinic and orthorhombic pyroxenes, and Mg-rich garnet, or at shallow depths a complex Cr-Fe-Mg-Al spinel. At a

depth of 410 km, orthorhombic olivine begins to transform into a denser cubic Mg-Fe silicate and other complex dense phases; pyroxenes transform into denser cubic garnetlike minerals. Below 670 km, the lower mantle consists of still denser cubic Mg-Fe-Ca-Al silicates, whose atomic structure is like that of the mineral perovskite, CaTiO_3 , plus cubic magnesiowüstite, $(\text{Mg,Fe})\text{O}$, whose structure is like that of halite, NaCl .

melting T of mantle rock at that depth (Figure 1.3; as measured in the laboratory) and because seismic shear waves, which cannot pass through a liquid, are propagated throughout the mantle. Obviously, the measured near-surface geotherm cannot be extrapolated far into

the interior of the Earth to obtain the T : That is, the geothermal gradient is not constant with respect to depth. There are at least two possible reasons for a substantially reduced geotherm at depth in the Earth so wholesale melting does not occur. One is that another mechanism of more efficient heat transfer prevails in the deep mantle, and the other is the presence of a concentration of heat producing rock nearer the surface. Both turn out to be true.

Convection. Movement of material having contrasting temperatures from one place to another is **convection**. Movement is caused by significant differences in density of different parts of the material so that, under the influence of gravity, less dense expanded material rises and more dense sinks. For example, soup in a pan on a hot stove convects as it warms and expands at the bot-



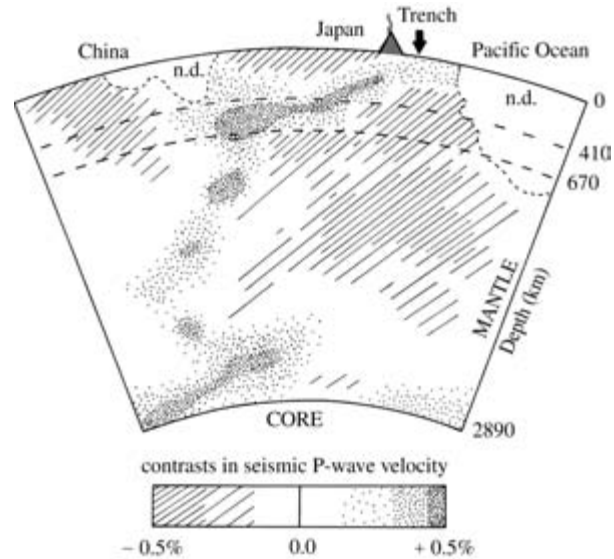
1.3 Relations among pressure, temperature, mineral composition, density, and melting conditions with respect to depth in the mantle and outer core of the Earth. Beginning-of-melting temperatures of mantle silicate rock and core Fe alloy have been determined experimentally in the laboratory (Special Interest Box 1.2). The geothermal gradient, or geotherm (dashed line) must lie below melting temperatures in the *solid* mantle and also pass appropriately through the 410- and 670-km phase transitions of olivine to spinel and spinel to Mg-Fe-Ca-Al perovskite plus wüstite, which cause discontinuities in seismic velocity. Note that the geotherm has a more or less constant slope through the convecting mantle of only about 0.3°C/km. The geotherm in the *D'* layer and lithosphere is much greater because of less efficient conductive heat transfer in these lower and upper thermal boundary layers, respectively. Note the exaggerated thickness of the continental crust, which averages about 35 km. Pressures from Stacey (1992).

tom, becomes less dense, and buoys upward, displacing cooler, denser soup at the top of the pan, which sinks toward the bottom to complete the circuit. Density contrasts driving convection can also be related to contrasts in composition. For example, surface evaporation of water in saline lakes in hot arid regions increases the surface salt concentration, making the water more dense and causing it to sink, even though it may be warmer than underlying less saline water. It should be emphasized that, unlike heat transfer by radiation and conduction, convection depends upon gravity. Without gravity, there is no buoyancy force to act on density contrasts that move matter.

Because convection is generally associated with fluid bodies, such as soup in a pan and bodies of water but also the gaseous atmosphere, it may seem surprising that this mode of heat transfer is possible in the solid rock mantle. Indeed, the reality of mantle convection was not a part of geologists' thinking until the acceptance of the lateral motion of plates and continental drift in the 1960s. The paradox, on the one hand, of a solid mantle that transmits seismic shear waves and, on the other hand, of a fluid mantle capable of convection is resolved by a consideration of the factor of time in viscous bodies. **Viscosity**, a measure of the resistance to flow, is illustrated by tar (asphalt). On an average 24°C day a body of viscous tar, like a brittle solid, can be broken into sharp-edged fragments by a hammer blow. But over a period of several hours this same body of tar flows under its own weight into a flat blob. Tar at 24°C is more viscous than honey, which is more viscous than water. Because hot mantle rock is about a billion billion times— 10^{18} , or 18 **orders of magnitude**—more viscous than 24°C tar, the rate of convective flow in the mantle is measured not in centimeters/hour, as for tar, but at most in centimeters/year—the speed at which lithospheric plates move. One way of defining the difference between a fluid and a solid is the time scale of their measurable flow.

Slabs of cooler oceanic lithosphere, mostly mantle rock, too viscous to convect within themselves, sink into the underlying hotter and relatively less dense underlying mantle. Computed tomography of the Earth using seismic waves (instead of X rays, used in scanning a person's body) has shown that at least some subducting lithospheric slabs sink all the way through the mantle and come to rest on top of the dense metallic core (Figure 1.4). Convectively upwelling hotter mantle at oceanic ridges and associated seafloor spreading complement subduction of lithosphere.

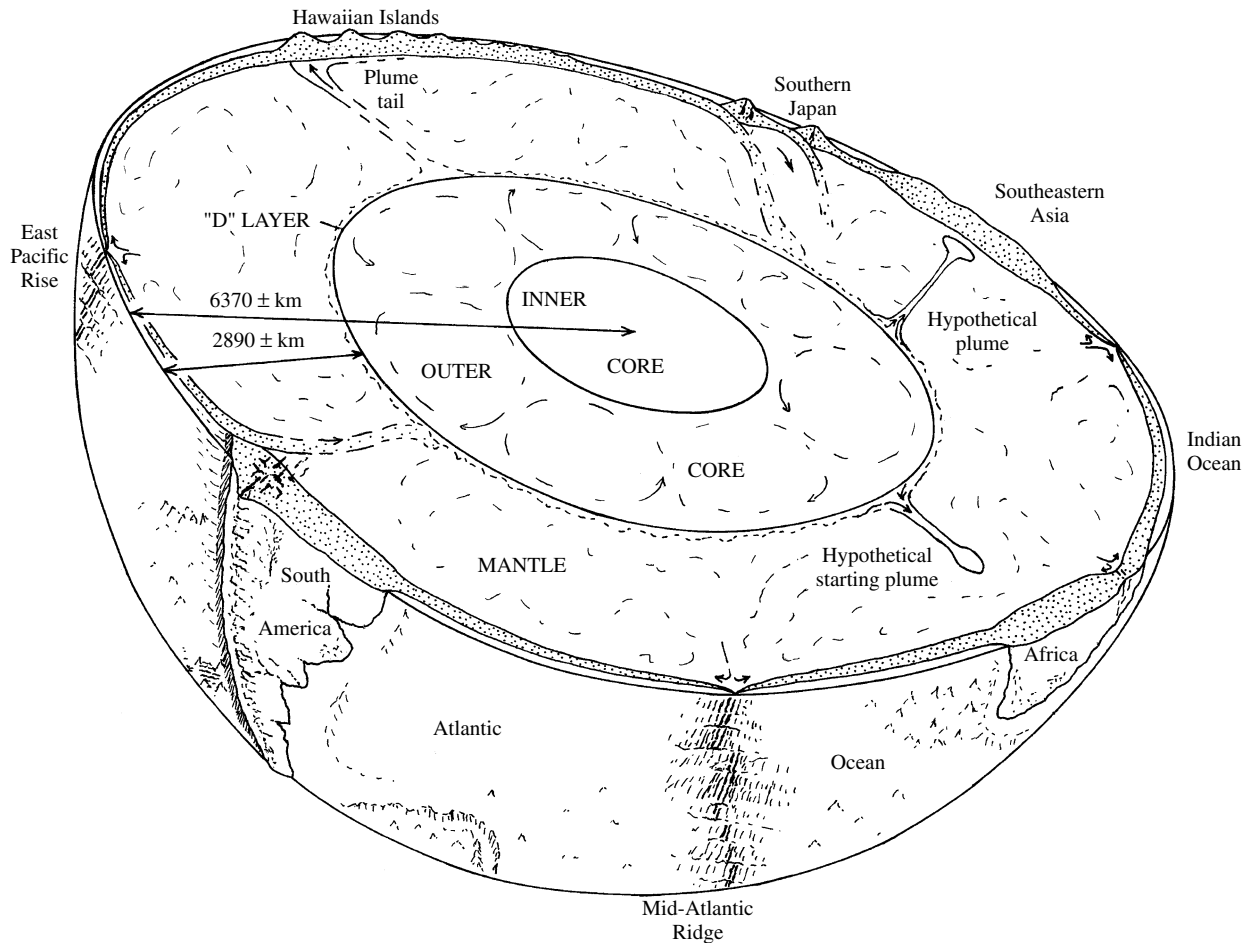
Even though the rate at which the mantle convects, on the order of a few centimeters/year, seems minuscule, it is far greater than the rate of heat transfer



1.4 Seismic tomography cross section of the mantle beneath the Japan subduction zone. The most densely stippled regions have seismic primary-wave velocities as much as 0.5% greater than average mantle and delineate the cooler, denser lithospheric slab, which has been subducting for most of the Cenozoic and has not conductively heated up to the T of the surrounding mantle. Note the segmented nature of the slab at midmantle depths and crumpled deeper slab that rests atop the core. Diagonally ruled regions represent hotter mantle where velocities are as much as 0.5% less than average. n.d., regions of no data; dashed lines, seismic discontinuities at depths of 410 and 670 km (Figure 1.3). (From a color diagram created by Rob van der Hilst of the Massachusetts Institute of Technology in Levi [1997; see also Grand et al., 1997]. Reproduced here as a modified black and white version with his kind permission.)

by conduction. The fact that lithospheric slabs can sink convectively over tens of millions of years about 2800 km through the hotter mantle but still maintain a recognizable cooler T throughout their approximately 100-km thickness demonstrates how slowly they are conductively heated. Conversely, if heat conduction were more rapid than convection, the subducting slabs would absorb heat and lose their density contrast and identity before sinking very far.

A second style of convection in the mantle, which can apparently operate independently of plate motion, consists of columns of relatively hotter mantle a few hundred kilometers in diameter that are rising vertically toward the base of the lithosphere. First proposed by Morgan (1971), **mantle plumes** had been confirmed to exist by the end of the century through seismic tomography imaging (see, for example, articles in the March 19, 1999, and May 14, 1999, issues of *Science*). The source of these plumes is apparently at the base of the mantle in a so-called D'' layer (Figure 1.5). Perhaps 10–20% of the heat driving mantle



1.5 Northern part of the Earth sliced off to reveal the convecting interior. Thickness of lithosphere (stippled pattern) is exaggerated to show details. Note descending oceanic lithospheric slabs in subduction zones and mantle plumes, two of which are hypothetical, rising from the bumpy D' layer at the base of the mantle.

convection comes from cooling of the core and perhaps all of this drives plumes. Seismological investigations reveal the D' layer has considerable relief (Jeanloz and Romanowicz, 1997), and from time to time and place to place, a thick bulge becomes sufficiently buoyant to move upward as a hotter, lower viscosity, but still solid-rock plume, drawing nearby D' layer with it. The decorative "lava lamp" in some homes is a colorful model. Many geologists believe the head of the plume partially melts in the shallow mantle, producing massive outpourings of basaltic magma onto the crust, forming huge continental and oceanic basalt plateaus. The plume tail, which can persist for tens of millions of years, is believed to be responsible for volcanic island chains, such as the Hawaiian.

In conclusion, heat transfer mainly in the upper mantle is manifest in movement of lithospheric plates, whereas plumes transfer heat from the core and lowermost mantle.

1.1.4 Implications of Mantle Convection

No apologies need be made for an extended discussion of mantle convection in this introductory chapter. Mantle convection is of critical importance in understanding how the Earth works as a gigantic heat engine driving geologic processes that create magmatic rocks. Descending cooler, denser lithospheric slabs and complementary upwelling of hotter mantle at spreading ridges and ascending deep hot mantle plumes constitute a whole-mantle convective system, one consequence of which is a small geothermal gradient of only a few tenths of a degree per kilometer throughout most of the mantle (Figure 1.3). Another consequence is substantial lateral variations in surface heat flow, which would not be expected if heat flow were wholly governed by conduction. Rock-forming processes, especially the creation of magmas and magmatic rocks, are strongly focused near the surface of the Earth by mantle convection. Figure 1.5 (see also Plate I) shows that

magmatism is localized in relatively narrow belts along convergent and divergent plate boundaries and in so-called hot spots above mantle plumes, virtually to the exclusion of any other surface area of the Earth.

Associated with focused magmatic activity is the concept of **petrotectonic associations**: that specific types of rocks are found together in specific tectonic regimes. Although basaltic rocks composed mostly of plagioclase and pyroxene are created in most of the tectonic settings diagrammed in Figure 1.5, there are significant differences from one tectonic regime to another in their chemical compositions, particularly in so-called trace elements such as Sr, Ba, Ta, and Nb. Also, the types of associated rocks are different. For example, andesites are common in convergent plate subduction zones but are essentially absent at oceanic spreading ridges. Rhyolites and their plutonic granitic rock counterparts are widespread along continental margin subduction zones but rare where two oceanic lithospheric plates converge, as in the island arcs of the western Pacific. These petrotectonic associations are discussed further in Chapter 13.

1.1.5 Energy Budget of the Earth

With all of this heat within the Earth one cannot but wonder, What is its origin? Is the thermal energy in the Earth the result of a one-time investiture, or is it being replenished somewhere as it is being expended elsewhere? Are **energy sources** being exhausted, or are they still operative to compensate for **energy sinks**? Countless volcanic eruptions mainly from oceanic ridges over eons of geologic time have dissipated heat from the interior of the Earth into the oceans and atmosphere, from which it is radiated into outer space, the ultimate heat sink. So why are volcanic eruptions still occurring?

The largest source of energy driving terrestrial processes, roughly 50,000 times all other sources, is radiant thermal energy from the Sun. The 70% trapped in the atmosphere drives the global hydrologic system of moving masses of air, water, and sediment. Radiant solar energy does not conduct very far into the ground, perhaps only a few meters in sunny areas. Although the surface heat flow from the interior of the Earth is minute compared to the solar influx, it is perhaps 20 times greater than all of the energy dissipated in magmatism, metamorphism, and tectonism.

A major source of internal heat within the Earth is the radioactive decay of the long-lived isotopes ^{238}U , ^{235}U , ^{232}Th , ^{40}K , and ^{87}Rb , which have half-lives of billions of years. Most investigators (e.g., Stacey, 1992; Verhoogen, 1980) calculate that this heat source is probably at least half and possibly approaching 100% of the total for the Earth. The uncertainty stems from the fact that concentrations of these isotopes are highly variable in different types of rock and where and in

what quantity these isotopes occur are poorly known. Overall, concentrations are greatest in the continental crust in granites, lower in basalt, in minute but uncertain amounts in the much more voluminous peridotitic mantle, and probably nonexistent in the core. Because of radioactive decay over eons of Earth history, the thermal energy produced when Archean rocks were created, 2.5–4.0 Ga, was roughly three times that of today; at 4.5 Ga, when the Earth was born, the rate was six times greater. Additionally, in that youthful Earth, decay of short-lived radioactive elements, such as ^{26}Al (half-life of 0.7 My), may have been significant.

Other important sources of internal thermal energy in the Earth (Verhoogen, 1980) are due to tides and to “original” heat. Tidal deformation of the solid Earth and oceans due to the gravitational pull of the Sun and Moon is dissipated as thermal energy, but this contribution is estimated to be an order of magnitude less than that of radioactive decay. In addition to current heat production by radioactive decay and dissipation from tides, some original heat inherited from the formation of the Earth at 4.5 Ga remains. Formation of the Earth is now generally believed to involve accretion of solid particles from a hot but cooling solar nebula of condensing gas and dust. As these particles and larger bodies (planetismals), themselves formed by collection of dust in the nebula, accreted into a proto-Earth, their gravitational potential energy was transformed into kinetic and then into thermal energy. Compression of these particles by additional accretion of more solids on top added more thermal energy. Compression does work on rock in the interior of the Earth, which is transformed into thermal energy, raising the rock T . Once the rock is compressed, no more thermal energy is created because no more work is done. Continuing capture of Sun-orbiting debris and impact of these fragments as asteroids onto the Earth for about 600 million years raised T further. The total energy in this growth process is estimated to have been sufficient to raise the T of the Earth tens of thousands of degrees. But the actual T increase was less, by some unknown amount, because heat was radiating and convecting away in the primitive atmosphere during accretion.

If, as is generally believed, the accreted Earth was initially chemically homogeneous, a large amount of thermal energy was generated during formation of the core as dense iron particles segregated from the molten silicate mantle by gravity settling. The calculated thermal energy gained from the loss of gravitational potential energy in core segregation is more than sufficient to produce the current surface heat flow, throughout the history of the Earth. Yet another source of heat related to the core is the ongoing solidification of the liquid outer core, releasing latent heat of crystallization. In other words, the core is currently heating the mantle.

Obviously, there are ample sources of internal thermal energy to drive the mantle heat engine.

It should not be forgotten that most geologic processes depend not only on thermal energy but also on gravity. Without gravity, matter would be dispersed indefinitely by the thermal processes of expansion, melting, and even vaporization. But on the other hand, gravity pulls matter together, compressing it. Interacting thermal energy and gravity constitute a push-pull in global geologic processes.

*1.2 GRAVITY, PRESSURE, AND GEOBARIC GRADIENT

Thermal energy is manifest in temperature, T ; the rate at which T increases into the interior of the Earth is the geothermal gradient, or $\Delta T/\Delta z$. In parallel fashion, the force of gravity acts on mass to produce pressure, P , in the interior of the Earth; the rate at which P increases into the interior of the Earth is the **geobaric gradient**, or $\Delta P/\Delta z$.

As previously defined, pressure is the force acting over a particular area. Within the Earth, pressure caused by the weight of overlying rock is called load, lithostatic, or **confining pressure**. It is denoted throughout this text by P ; the context will indicate whether confining pressure or the chemical element phosphorus, P , which is not italicized, is being discussed. There are many units of pressure (listed inside the cover). In the International System (SI) of units, the unit of pressure is the *pascal*, denoted Pa, but the most commonly used unit in petrologic work is the *bar*: 1 bar = 10^5 Pa (0.1 MPa) and 1000 bar = 1 *kbar* = 10^8 Pa (0.1 GPa). Also, 1 bar = 0.9869 atmospheres (atm) = 14.504 pounds per square inch, the mean atmospheric pressure at sea level.

The magnitude of P increases with depth in the Earth in a predictable manner. To evaluate this increase, we note that bodies of rock more than several kilometers beneath the surface are hot and, over long periods of geologic time, behave as viscous fluids. Like water, which seeks its own level because it has no intrinsic strength, hot rocks have low strengths and flow readily, particularly over long periods of geologic time. This concept is implied in mantle convection and isostasy. The confining pressure, P , at the base of a vertical column of rock, considered as fluid, with cross-sectional area, A , equals

$$P = \frac{F}{A} = \frac{mg}{A} \quad 1.6$$

from Newton's first law, $F = mg$, where m is mass and g the acceleration of gravity. In terms of the **density**, $\rho = \text{mass/volume}$ of the rock, equation 1.6 becomes

$$P = \frac{\rho V}{A} = \rho gz \quad 1.7$$

where z is the height of the rock column, considered positive downward here. In the Earth, the geobaric gradient is

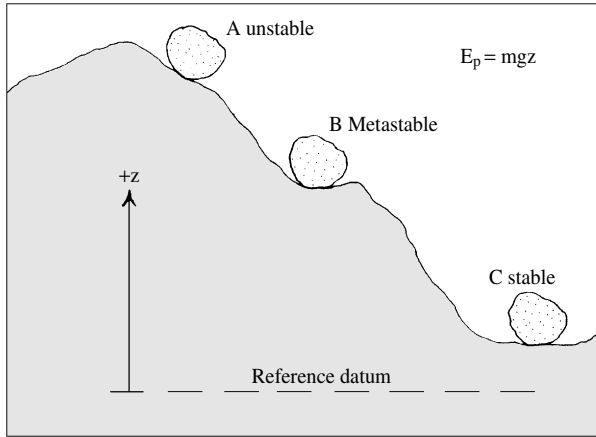
$$\frac{\Delta P}{\Delta z} = \rho g \quad 1.8$$

ρ being the mean density of the column of rock. In the continental crust, where the mean density $\rho \sim 2.7 \text{ g/cm}^3$, $\Delta P/\Delta z = 270 \text{ bar/km}$ ($\sim 27 \text{ MPa/km}$). Actual densities in the continental crust range between 2.2 and 3.0 g/cm^3 . (The less familiar SI unit of density is $\text{kg/m}^3 = 10^{-3} \text{ g/cm}^3$.) In the basaltic oceanic crust, $\rho \sim 3.0 \text{ g/cm}^3$. In the upper mantle, where $\rho \sim 3.3 \text{ g/cm}^3$, $\Delta P/\Delta z = 330 \text{ bar/km}$. These values of the geobaric gradient are useful in converting P into an approximate equivalent depth in the crust or mantle, or vice versa. For example, at a depth of 30 km in the crust, $P = 30 \text{ km} \times 270 \text{ bar/km} = 8100 \text{ bar} = 8.1 \text{ kbar}$.

*1.3 ROCK-FORMING PROCESSES AS CHANGING STATES OF GEOLOGIC SYSTEMS

All geologic processes in the dynamic Earth, including rock-forming processes, involve energy changes and interaction between energy and mass. Some geologic processes are driven wholly, or in large part, by changes in thermal energy and involve heat flow or transformation of heat into other forms of energy, or the reverse; cooling of an intrusive magmatic dike is a thermal process. Other geologic processes involve work, which can be considered as a mechanical or physical process in which, for example, rock is crushed into smaller pieces or magma is expanded by internal gas pressure into a greater volume. Still other geologic processes involve chemical reactions and movement of atoms, such as their organization into a well-ordered crystal of feldspar as a silicate melt cools. Most geologic processes are a combination of changes in several forms of energy. However, changes in thermal and gravitational potential energies dominate on a global scale. Hence, T and P are important characterizing variables in changing geologic systems.

All natural changes in a system move it toward a state of lowest possible energy, which is the most stable of all possible states. An example, familiar to any mountain hiker, illustrates these fundamental concepts. In Figure 1.6 are three hypothetical identical boulders; two are positioned on a hill slope and the third is in a lower valley. Each boulder has a different vertical position above some reference elevation and hence has a different gravitational potential energy. If boulders A and B are dislodged from their positions of rest, gravity causes them to fall down into the valley alongside boulder C, where their gravitational potential energy is the lowest possible. The potential energy given up is



1.6 Concept of stability. Three boulders possess different amounts of gravitational potential energy, $E_p = mgz$. The tendency of changing natural systems to move toward a state of minimum energy may be referred to as the principle of parsimony, or laziness!

transformed into kinetic energy and ultimately into heat and mechanical work of breaking rock.

In this example, the three boulders initially possessed three different gravitational potential energies representing three possible energy states of a geologic system. A **system** is simply a part of the universe that is set aside in one's mind for the purpose of study or discussion. All else, the remainder of the universe, is the **surroundings**. As the hillside boulders were dislodged, their **state**—the particular conditions defining their properties or energy—changed. Once boulders A and B join C in the valley, all three are at the same lowest possible state of gravitational potential energy.

Equilibrium is a state that has no tendency to change spontaneously. The net result of forces acting on an equilibrium system is zero. Atoms may move about in a chemical system but at equilibrium nothing happens over time. Any slight disturbance will not result in any permanent change, as the system will return to its original condition. Boulder A satisfies some of the equilibrium conditions but not all. It is **unstable**, because disturbed even slightly it will cascade downhill to the lowest energy level alongside boulder C, which lies in a state of **stable equilibrium**. Boulder C in the valley may be rocked from side to side, say, by an earthquake or by a hiker but will readily return to its original, lowest-energy, position. Boulder B represents a state of **metastable equilibrium** whose energy is more than the lowest state but is prevented from moving to that more stable, lower-energy state by an energy barrier, called the **activation energy**, represented by an elevated lip on the terrace on which the boulder rests. A metastable state can persist indefinitely; if subject to only a small disturbance—represented, for example, by the hiker's slightly rocking boulder B on the terrace on the hill slope—the metastable state will return to its original

configuration. Only a more forceful shove by the hiker will send the boulder over the lip of the terrace downhill to stable state C.

Glass and virtually all high- P and - T magmatic minerals are metastable under atmospheric conditions. Their thermodynamic energy (Chapter 3) is not the lowest possible under atmospheric conditions. **Glass** is a solid aggregate of atoms more or less randomly arrayed as if it were liquid: that is, it is amorphous. Whether of human or natural volcanic origin, glass eventually crystallizes into an aggregate of stable crystals under atmospheric conditions. Most volcanic glass is of late Cenozoic age, little glass is early Cenozoic, and so on. Diamond, formed at depths of at least 150 km in the mantle, has a much larger activation energy barrier to overcome than glass, and this characteristic prevents it from converting into stable graphite of minimal energy for hundreds of millions of years while lying in an African kimberlite pipe in the shallow crust. However, if a diamond is heated to several hundred degrees in the oxygen-rich atmosphere, the provided thermal energy supplies the necessary activation energy so that the metastable diamond decomposes into stable CO_2 .

Without metastability there would be no magmatic rocks exposed at the surface of the Earth, only minerals such as quartz, calcite, gypsum, and clays that are stable at atmospheric conditions.

These concepts apply to chemical equilibria where chemical reactions involving movement of atoms are occurring, to thermal equilibria where heat is being transferred to parts of a system at different temperatures, and to other types of equilibria. In evaluating the stability and equilibrium of any system, one must carefully examine the nature of all possible changes in state. Thus, a particular system, such as boulder C, would be chemically unstable if it were limestone subject to acid rain. When asking the question, What state is most stable?, the next question should always be, Under what conditions?

In the dynamic Earth, rock-forming geologic systems are dislodged from states of higher energy and move naturally to a more stable state of lowest possible energy. For example, lava extruded at the summit vent of a volcano loses gravitational potential energy and gives up heat to its cooler surroundings as it flows down slope and solidifies into solid rock. Both forms of energy change to a new lower-energy level. An energy gradient is available through which to move. The direction of transfer is always from a higher to a lower energy level, in the one case elevation and in the other T .

* 1.4 ROCK PROPERTIES AND THEIR SIGNIFICANCE

Thus far, our approach to understanding rock-forming geologic processes has been deductive, starting with

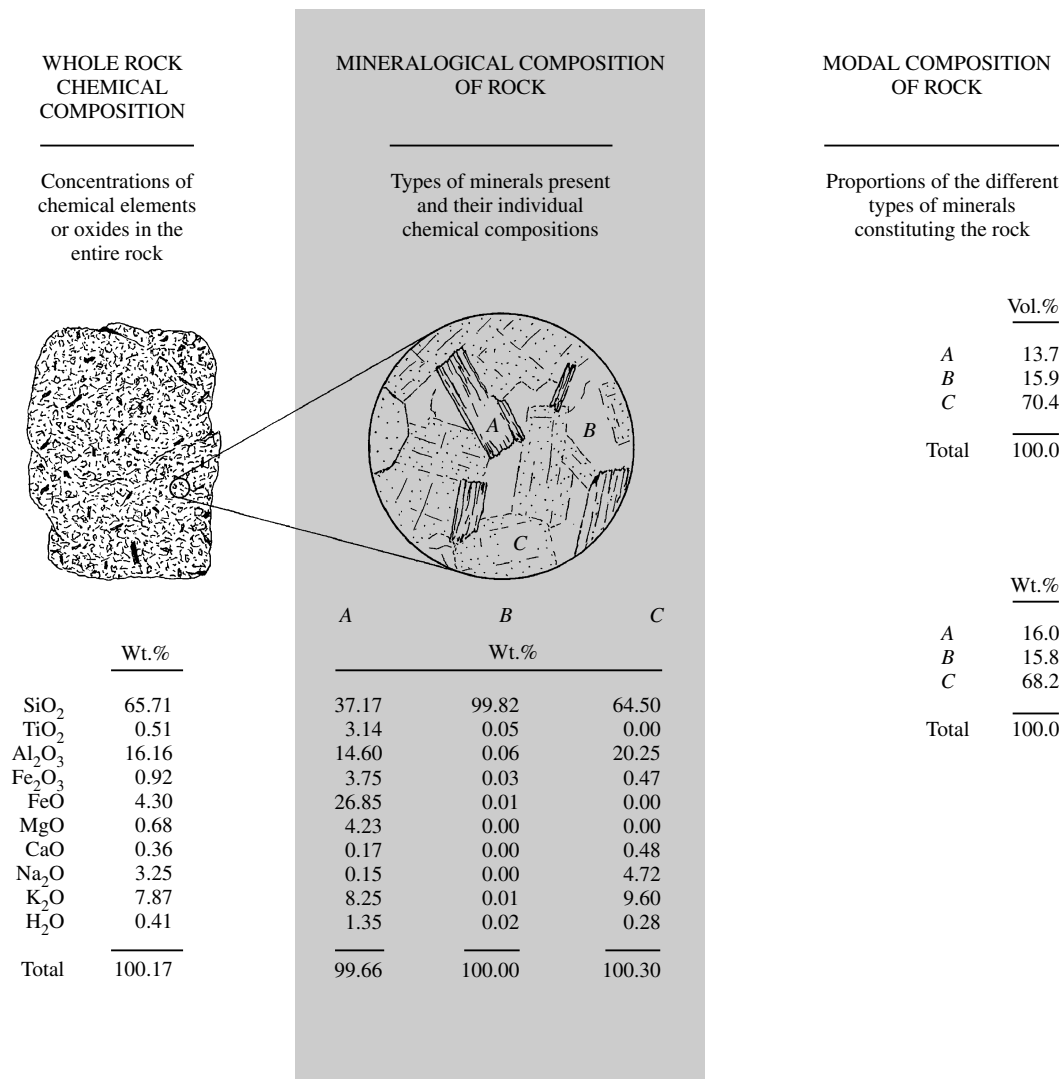
general principles and illustrating them by specific examples. In practice, however, the geologist is usually faced with the opposite, inductive problem. Given a particular mass of rock and its observable properties, the geologist asks, What was the state of the past geologic system in which it was created? What geologic processes of energy transfer and transformation and movement of matter were involved? What caused the changes in state of the rock-forming system, perturbing a previous state of equilibrium and producing a new state of stable equilibrium? From the only tangible record of the system—the rock itself—the geologist must work backward, trying to comprehend the rock-forming system in which it was created. So, what properties of the rock are most significant? And what specifically do they tell us?

Among many rock properties—such as aesthetic, electrical, magnetic, and mechanical—the rock properties of most significance for the geologist are composition, fabric, and field relations.

1.4.1 Composition

Rocks consist of minerals and locally occurring glass (actually, an amorphous solid, not a mineral) that are made of atoms of the chemical elements. Three basic compositional properties can be recognized in any rock: the concentrations of chemical elements in the bulk or whole rock, the character of the minerals and glass in which they reside, and the amounts of the different minerals and glass (Figure 1.7).

Whole-Rock Chemical Composition. An analysis of a rock for its chemical elements, irrespective of mineral



1.7 The three compositional aspects of a rock. The modal composition is expressed on a volume percent volume basis, and from this, using mineral densities for biotite (A), quartz (B), and alkali feldspar (C), the mode in weight percentage (wt.%) can be calculated.

constituents, yields its **whole-rock chemical composition**, also referred to as its **bulk chemical composition**, or sometimes simply as its chemistry. The bulk chemical composition is expressed in terms of weight concentrations on a percentage basis of chemical species such as SiO_2 , Al_2O_3 , H_2O , and CaO . Thus, in a 100-gram sample of the hypothetical rock (Figure 1.7), 65.71 grams, or 65.71 wt.%, would be SiO_2 ; 16.16 g, or 16.16 wt.%, would be Al_2O_3 ; and so on. The SiO_2 and Al_2O_3 actually occur as Si, Al, and O atoms in the minerals making up the rock.

Rocks contain virtually all of the approximately one hundred chemical elements in continuously varying concentrations. The nature of such variations, why they occur, and how they can be used to elucidate the origin of rocks are major topics explored in this textbook. The chemical composition of magmatic rocks reflects the conditions of creation of magma from solid rock and its composition and the subsequent evolution of the magma. The chemical composition of metamorphic rocks reflects the nature of the pre-existing rock and conditions of metamorphism.

Mineralogical Composition. The types of minerals constituting the rock *and* their chemical compositions are the **mineralogical composition** of the rock. The types of minerals—such as biotite, plagioclase, or olivine—and perhaps glass, in the case of a volcanic rock, can be identified by naked eye or a magnifying lens in hand samples, a petrographic microscope using thin sections, and by X-ray diffraction and other laboratory techniques. Chemical compositions of minerals can be roughly approximated by optical techniques using a petrographic microscope but are most accurately determined by an electron microprobe analyzer.

Of the thousands of known mineral species, only seven major minerals or mineral groups make up most magmatic rocks; these include quartz and solid solutions of feldspars, micas, amphiboles, pyroxenes, olivine, and Fe-Ti oxides. Metamorphic rocks contain in addition minerals such as Al_2SiO_5 polymorphs and solid solutions of carbonate minerals, zeolites, garnets, chlorites, and epidotes.

The minerals that constitute magmatic and metamorphic rocks depend on the P and T of the last equilibrium state of the rock-forming system and its chemical composition. In most cases, this final equilibrium state is “frozen into” the rock and persists metastably and indefinitely. For example, whether cristobalite or quartz or some other polymorph of silica (SiO_2) crystallizes depends on the prevailing P and T of the system that has a sufficient concentration of silica.

Modal Composition. The volumetric proportions, on a percentage basis, of the different minerals in a rock are referred to as the **modal composition**, or simply the **mode**. It can be roughly estimated through examination by eye or in thin section under a microscope.

Further Comments. Modal, chemical, and mineralogical compositions of a rock are correlated, but not always in an obvious manner. For example, in two rocks of the same bulk composition containing, say, 63 wt.% SiO_2 , one may contain quartz but the other not, depending on what other minerals are present. A granite containing 75 wt.% SiO_2 in its bulk chemical composition must not only have quartz as a constituent mineral, but in substantial amounts in the mode because no other common rock-forming silicate mineral has more than about 66 wt.% SiO_2 . Rocks with abundant biotite in the mode will have relatively large concentrations of K_2O in the bulk rock chemical composition. Some minor or trace chemical constituents in the whole rock chemical composition, such as Sr, will be found in solid solution in major minerals, in this case feldspar.

A chemist investigating a rock would be expected to go no further than an analysis of composition and perhaps theoretical interpretations based on thermodynamics and kinetics (Chapter 3). However, rock-forming processes reflect changing states of geologic systems that operate in a context of *space and time*. This context is especially manifest in the field relations and fabric of the rock body. The task of the geologist interpreting rocks, therefore, goes far beyond simple chemical study.

1.4.2 Field Relations

Field relations are the larger-scale features of a mass of rock discerned in exposures (outcrops) in hills and mountains and man-made road cuts, quarries, and mines and seen in drill core and on geologic maps. The study of field relations in layered volcanic rock bodies is called **stratigraphy**. Field relations include, but are not restricted to, such characteristics as

1. the nature of contacts of a rock mass with neighboring bodies, whether sharp or gradational, conformable or cross-cutting;
2. the relative chronology to neighboring bodies, whether older or younger;
3. the spatial aspects of the fabric and composition, whether they are uniform over the extent of the mass or vary in some way, particularly near contacts or borders;
4. the dimensions of the mass.

Field relations provide insights into the causes of changes in state of the rock-forming system, or why transfer or transformation of energy occurred to drive the rock-forming geologic process. Field relations of magmatic rocks reflect the physical dynamics of the body of magma from which it was created: for example, whether it was extruded onto the seafloor or on top of dry ground, or intruded deep in the crust. Understanding the evolution of long-lived extrusive magma systems depends on a thorough investigation of the stratigraphic features of the volcano.

1.4.3 Fabric

In this textbook, the term **fabric** is used for all of the noncompositional properties of a mass of rock discernible on scales of observation from the outcrop and hand sample to the microscopic. Fabric encompasses both texture (sometimes called *microstructure*) and structure. **Texture** refers to grain characteristics—including grain size and shape, intergrain relations, and amount of glass—generally seen at the scale of a hand sample or smaller. **Structure** refers to features seen at a scale of a hand sample or larger, such as bedding in a pyroclastic deposit from an explosive volcano, columnar joints in a lava flow, or chunks of foreign wall rock within a magmatic intrusion. Tectonic and structural geologists also refer to folds in metamorphic rocks as structures. A sharp distinction between texture and structure is not always possible to make, hence the utility of the all-inclusive term **fabric**. Also, there is no sharp distinction between some structures and field relations.

Fabric chiefly reflects the time-dependent path of change in state of the rock-forming system. The rate of shearing in a deep crustal fault zone strongly influences the fabric of the metamorphic rock. How fast magma rises in the crust and boils can determine whether it erupts placidly as vesicular lava or violently explodes to create far-flung pyroclastic particles. Successive changes in state are commonly recorded in overprinted fabrics. In detail, the variety of time-dependent paths and the resulting fabrics are virtually infinite.

* 1.5 HOW PETROLOGISTS STUDY ROCKS

As a subdiscipline of geology, **petrology** seeks to understand the nature of rocks and their origin. Past rock-forming *processes* in dynamic geologic systems are recorded in the observable rock *product*—its field relations, fabric, and composition. Sometimes, petrologists approach the cause-effect, process-product correspondence in the opposite way by creating models or doing experiments simulating rock-forming, or petrologic, systems and then making comparisons with properties of real rocks.

In their attempts to understand the origin of rocks and explain how the mostly inaccessible Earth has worked through billions of years of geologic history recorded in rocks, petrologists theorize, using paper-and-pencil and computer models, and experiment, using simple model magmatic systems in the laboratory. In his presidential address to the Geochemical Society in 1969, James B. Thompson, Jr. (1970) had this to say regarding petrology:

We who are concerned with such [petrologic] matters fall mainly in one or more of three categories: (1) those who collect rocks and study them; (2) those who try to duplicate rocks in the laboratory; and (3) those who worry

about rocks. All three facets, the observational, the experimental, and the theoretical should clearly be integrated in a healthy science. Description as an end in itself is sterile, experiments that answer unasked questions are irrelevant, and theory unchecked by fact is useless. Theory, at its best, is in many ways a link between the other two. Good theory can thus be used to interpret or reinterpret the results of observation in order to ask more significant questions of experiment and vice versa.

Petrologists study rocks as interpretive historians who try to “solve” or explain geologic problems and puzzles. They ponder and try to answer fundamental questions, such as the following:

1. Where and how did a particular magma originate?
2. How was the magma transported from its source and emplaced where the magmatic rock is now found?
3. What physical, chemical, and thermal processes operated on the rock-forming system during crystallization?
4. What was the nature of the rock prior to metamorphism and its history of deformation and recrystallization?
5. How do petrologic processes control evolution of the crust and relate to global tectonic setting?
6. How can modern petrotectonic associations be used to infer tectonic processes and regimes in ancient rocks?
7. How did the planet Earth originate and evolve?
8. What is the effect of petrologic processes on our society and life in general?

The intent of this textbook is to provide a background sufficient to *begin to answer* these petrologic questions, and many others of a more specific nature. Interpretations of the origin of rocks, or **petrogenesis**, must be based on accurate **petrographic** data (modal composition and fabric) as well as field, chemical, isotopic, experimental, and theoretical data.

Petrology has become a global science that is finding increasing integration into other fields of Earth science. An understanding of petrology leads to a better understanding of how the Earth works.

In his presidential address to the Mineralogical Society of America, Peter Robinson (1991, p. 1781) said:

In petrologic studies . . . the observations that are possible are conditioned by the experience and training of the observer. Petrologic research is an attempt to define the sequence of events in the formation of the rocks, the conditions under which the events happened, and why they happened. Petrology can never be totally quantitative, i.e., “getting numbers out of rocks” . . . observational and descriptive petrography from the atomic through the microscopic and macroscopic to the petrologic map scale [are essential in] providing the framework for interpretations.

Careful observation and innovative ways of thinking about rocks can lead to new discoveries and overturn established dogma. That an interpretation is espoused by esteemed experts is no guarantee of its validity. Continual reassessment and well-documented criticism of established truths make for better science.

Petrology at the dawn of the 21st century is a vibrant, dynamic discipline in which unsolved problems and controversies abound. Numerous puzzles remain to be explained and resolved. Many have not even been discovered!

SUMMARY

The central themes of this textbook developed in more detail in subsequent chapters are these: On a global scale, rock-forming geologic processes are driven by interactions between matter and energy, chiefly thermal and gravitational. These interactions are strongly focused along convergent and divergent plate boundaries and above ascending mantle plumes, the two expressions of a convecting mantle, which is the major heat engine driving geologic processes. Without thermal energy in the Earth it would be geologically dead. Major sources of thermal energy in the interior of the Earth are radioactive decay and “original” heat generated at the time of its creation. Rock-forming processes focused by mantle convection create characteristic petrotectonic associations of magmatic rocks that are more or less unique to each tectonic setting in which they occur.

In the dynamic Earth, transfers and transformations of different forms of energy and movement of matter cause changes in states of geologic systems. The direction of change tends toward the lowest possible energy state of stable equilibrium. However (by good fortune!), metastable states can persist indefinitely, allowing the petrologist to study magmatic mineral assemblages stabilized at high P and T and thus gain insight into rock-forming processes. Other clues to petrogenesis are found in the chemical, isotopic, and modal composition of the rock body and its fabric and field relations. Aspects of the time-dependent path taken during changing states of the rock-forming system are recorded in the rock fabric, such as the rate of cooling of the magma. Field relations provide insights into the flow of matter and energy in the system, for example, whether the magma was extruded explosively from a volcano or intruded into the deep crust.

Petrology seeks to understand the nature and origin of rocks and the way the whole Earth works by making critical observations, asking relevant questions, and integrating all available information into a self-consistent interpretive petrogenetic model.


CRITICAL THINKING QUESTIONS

These questions may not have definite, specific answers. More than one answer may be valid.

- 1.1 Describe examples of interaction between energy and matter that create magmatic rocks.
- 1.2 Where do rock-forming processes occur in relation to global tectonics?
- 1.3 Describe and give examples of the different forms of energy.
- 1.4 Cite an example of how energy is transferred, transformed, and conserved in a geologic system.
- 1.5 Characterize four modes of heat transfer and indicate their significance in different geologic systems and for the whole Earth. Why is less heat transferred into the substrate beneath a lava flow than out of the top of the flow?
- 1.6 In what way is the Earth a heat engine capable of doing work?
- 1.7 Why is the mantle important in creation of magmatic rocks?
- 1.8 Describe two modes of convection in the mantle and indicate one example that provides specific evidence for each.
- 1.9 Contrast the heat capacity of water and that of rock. Why is this contrast important?
- 1.10 Account for the inflections in the geotherm from the surface to the mantle-core boundary.
- 1.11 Account for the paradox of fluidlike convection in the solid rock mantle.
- 1.12 What is a petrotectonic association?
- 1.13 What are sources of internal heat in the Earth?
- 1.14 Why does T increase with depth in the Earth? P ?
- 1.15 Characterize a geologic system that is in an unstable state. In a state of stable equilibrium. Of metastable equilibrium. How can these states be changed?
- 1.16 Briefly describe the significant properties of rock masses and tell how each property provides particular information about the rock-forming process.
- 1.17 Contrast the three different compositional properties of a rock.
- 1.18 Discuss the goals of petrology and the way a petrologist accomplishes them.

PROBLEM

- 1.1 What is the confining pressure, P , at the base of the oceanic crust 4 km below the seafloor? What is P 30 km below the base of this crust, assuming the density of peridotite is 3.3 g/cm^3 ? Assume that the ocean is 5 km deep and the density of seawater is 1 g/cm^3 . Express your answers in bars, kilobars, megapascals, and gigapascals. (*Answers:* 1700 bars = 1.7 kbar = 170 MPa = 0.17 GPa; 11,600 bars = 11.6 kbar = 1160 MPa = 1.16 GPa)



Composition and Classification of Magmatic Rocks

FUNDAMENTAL QUESTIONS CONSIDERED IN THIS CHAPTER

1. How are rocks sampled in the field and analyzed in the laboratory to determine their chemical, modal, and mineralogical composition?
2. What do these analyses disclose regarding the composition of magmatic rocks?
3. How can analytical data be presented to reveal petrogenetically significant similarities, patterns, and contrasts?
4. How are magmatic rocks classified so as to convey meaningful petrogenetic information concerning the origin and evolution of the magmas from which they solidified?
5. What special petrogenetic information do the trace element and isotopic compositions of magmatic rocks provide?

INTRODUCTION

Magmatic rocks possess a seemingly endless variety of chemical, modal, and mineralogical compositions from which the petrologist must extract information about how rocks form. Samples can be collected, laboratory analyses made, similarities and contrasts in composition sought through graphed data and classification, compositional patterns found, and interpretive models hypothesized.

Compositions of magmatic rocks, together with their fabrics and field relations (Chapters 7, 9, and 10), are the essential “nuts and bolts” from which interpretive models (Chapters 11–13) can be constructed. After a

brief introduction to sampling and analytical techniques for obtaining compositional data on rocks the chapter focuses on three compositional attributes of rocks:

1. Mineralogical, modal, and major element compositions that are the bases for different classifications of magmatic rocks
2. Trace element compositions
3. Isotope ratios

All of these compositions provide crucial information regarding the sources and evolutionary paths of magmas.

*2.1 ANALYTICAL PROCEDURES

Modern petrology relies heavily on quantitative, or numerical, data on rock and mineral compositions obtained by a variety of analytical devices and techniques. Data can be stored and processed and interpretive models created by using computers.

2.1.1 Sampling

The validity and usefulness of any compositional data are crucially dependent upon the quality of the sample collected in the field. “Garbage” samples collected in the field are liable to yield “garbage” analytical results. The importance of this crucial first step cannot be ignored or minimized. If the analysis of the sample is to be truly representative of the body of rock, the sample should be as free as possible of overprinting effects of alteration and weathering. This is especially important for chemical and isotopic analyses. Virtually all primary, high-temperature (high- T) minerals and glasses are metastable near-atmospheric conditions. Except for quartz, they are subject to replacement by secondary minerals more stable at near atmospheric temperatures such as carbonates, clays, and ferric oxides. This **weath-**