Chapter 1
Introduction

1.1 Why study rocks?

I am a petrologist and I study rocks. Petrology is the study of how rocks form (rocks being defined as a natural assemblage of individual minerals). More specifically, petrologists seek the physical and chemical conditions necessary for the formation and modification of certain types of rocks. On one level, petrology involves the art of identifying and classifying different minerals and rock types into some sort of petrogenetic context. By petrogenesis, what we mean is some sort of model or theory about the physical and chemical processes by which a particular rock or group of rocks formed or evolved. This leads naturally to the second level on which petrologists operate, that is, petrologists must bring together principles from physics and chemistry to develop petrogenetic models that are consistent with observation. The predictions resulting from any such models should then be tested by returning to the rocks or by reproducing certain processes in laboratory experiments. It is through this culmination of physical and chemical principles in the context of rocks that we as Earth scientists can eventually unravel larger scale processes, such as ore formation, continent formation, evolution of the Earth and other planets, and so forth.

The approach we will take in these lecture notes will be to focus first on processes and then come back to rock classification in the context of processes. This is of course the reverse of the historical development of petrology. I have chosen this approach because all too often I have heard from my non-petrology colleagues that petrology seems like an exercise in somewhat arbitrary classifications accompanied by unnecessarily dense terminology. There is of course some truth to these statements, but there is actually some method to this madness and it is this method that I would like to get across here. Perhaps by developing an initial appreciating of the processes by which rocks are formed and how they relate to the origin and evolution of our planet, we can then develop a better appreciation for the rock record itself.

So let's begin with what rocks might tell us. A few simple definitions are in order. Rocks can be classified as igneous, metamorphic and sedimentary origin. *Igneous rocks* form by the direct solidification of magma (*molten rock*). *Metamorphic rocks* represent the integrated end-product of all processes that impart a mineralogical, textural and/or chemical transformation (e.g., metamorphosis) of an igneous, sedimentary or metamorphic rock into a different rock. *Sedimentary rocks* are the lithified products of sediments. Sediments represent all loose solid materials that accumulate at the surface of the Earth, with the loose particles making up sediments being derived from physical or chemical weathering of pre-existing rocks, direct chemical precipitation from water, and/or organisms' skeletal remains.

Dividing the field of petrology into these sub-categories helps to simplify the content for teaching purposes. In reality, however, each of the rock-forming processes - igneous, metamorphic, and sedimentary- is intimately linked as they are all components within the rock cycle. To illustrate, we can simplify the Earth into a number of sub-reservoirs, i.e. large, coherent regions of mass characterized by a distinctive composition. The most obvious Earth reservoirs are the atmosphere, oceans, and the solid part of the Earth, which itself can be subdivided into continental crust, oceanic crust, mantle and metallic core (Fig. 1.1 and Table

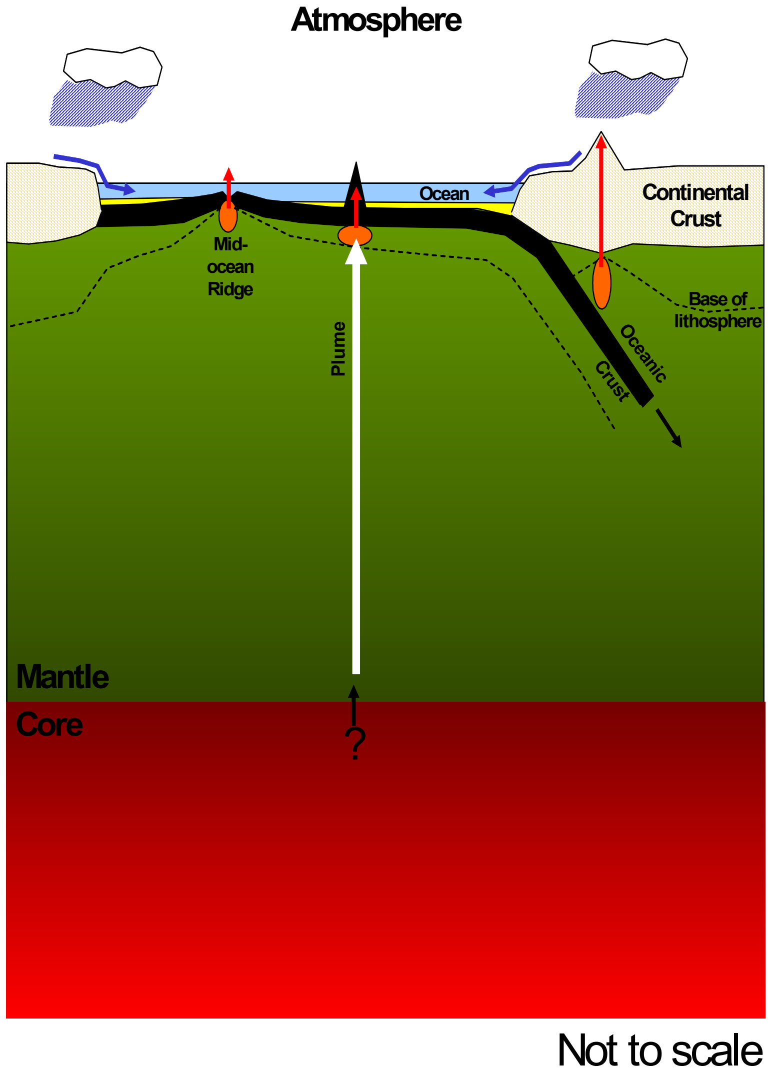

Figure 1.1. Schematic diagram showing some of the major reservoirs in the Earth (atmosphere, ocean, continental crust, oceanic crust, mantle, core) and how they may interact. Red ovals represent regions where solid mantle is being partially melted.

1.1). That each of these reservoirs exists reflects the fact that the Earth has been unmixing (differentiating) ever since it was formed. Each of these reservoirs is continually evolving in size and composition as each reservoir interacts with each other or in other words mixes back together. For example, there is a net contribution of material from the mantle to the atmosphere, ocean, continental crust and oceanic crust: degassing at mid-ocean ridges or arcs releases magmatic gases to the atmosphere and oceans, hydrothermal fluxing at mid-ocean ridges results in the net flux of certain chemical components into the ocean, and magmatism at mid-ocean ridges and arc volcanoes result in the addition of new mass to oceanic and continental crust (Figs. 1.1 and 1.2). In addition, physical and chemical weathering of continental crust results in the transport of sediments to the seafloor, some of which are ultimately recycled along with oceanic crust back into the mantle at subduction zones. These examples represent only part of

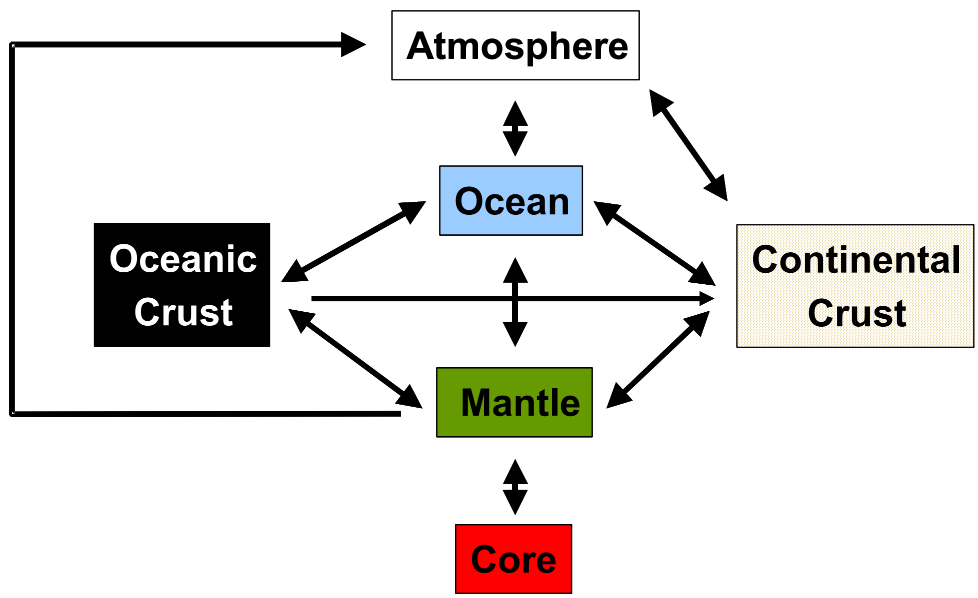

Figure 1.2. Earth systems cycles. A simplified subdivision of the Earth into its major reservoirs and how they interact with each other. Arrows reflect direction of

all the different ways in which these reservoirs can interact with each other (Figs. 1.1 and 1.2). In trying to decipher how the nature of the Earth has evolved through time, it is necessary to quantitatively describe the physical and chemical processes represented by each of the arrows, and moreover, how the system as a whole or its subsystems have changed.

In this treatise, we will cover the fundamental principles of igneous, metamorphic and sedimentary petrology in the broader context of the rock cycle and the driving forces that allow for chemical and physical differentiation of the Earth into its major rocky reservoirs. We will

Table 1.1. Some physical properties of the Earth

Mean radius (km)	6371.01		
Total surface area (km^2)	5.10×10^8		
Oceanic surface area (km^2)	3.62×10^8		
Volume (km^3)	1.08×10^{12}		
Mass (kg)	5.97×10^{24}		
Mean density (g/cm^3)	5.52		
Core radius (km)	3485		
	Mass (kg)	% of Whole Earth	% of Bulk Silicate Earth
Total atmosphere	5.14×10^{18}	8.65×10^{-5}	
Hydrosphere	1.66×10^{21}	0.0279	0.0413
Total Crust	2.37×10^{22}	0.3951	0.585
Continental crust	1.52×10^{22}	0.25	0.38
Oceanic crust	8.45×10^{21}	0.14	0.21
Mantle	4.01×10^{24}	67.08	99.37
Bulk Silicate Earth	4.03×10^{24}	67.51	
Core	1.94×10^{24}	32.5	

Data are from [*Lodders and Fegley*, 1998]

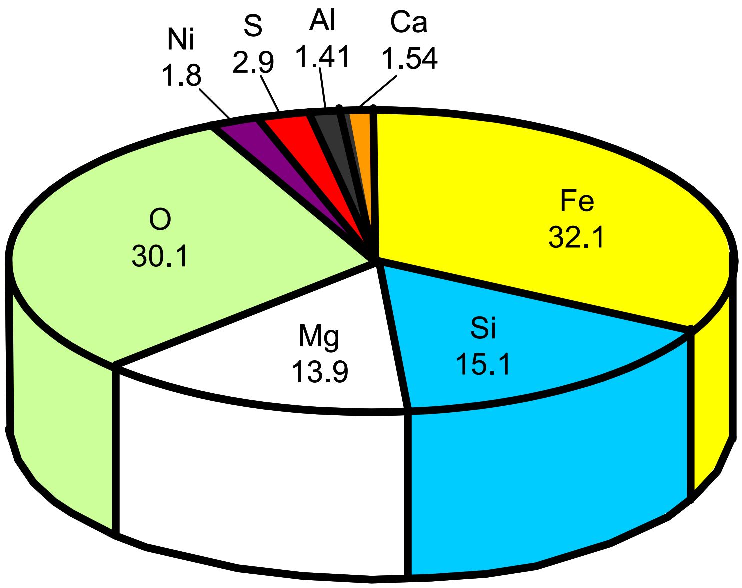

Figure 1.3. Relative atomic abundances of the eight most common elements that comprise 99% of the mass of the solid Earth.

not discuss processes that are internal to the ocean and atmosphere, but we will end our journey through petrology by ultimately linking the solid Earth reservoirs to the ocean and atmosphere.

1.2 Composition of the solid Earth

The solid part of the Earth (e.g., the metallic core and rocky crust and mantle) is predominantly made up of O (~30 wt. %), Fe (~32%), Mg (~14%), Si (~15%), Ca (1.5%), and Al (1.4%), as shown in Figure 1.4. Most of the Fe is in the core, which makes up ~32.5 wt. % of the entire Earth. This leaves the rocky outer part of the Earth with ~44% O, ~23% Mg, ~21% Si, 2.5% Ca and 2.4% Al. Thus, the outer part of the Earth is dominated by Si and O and is accordingly dominated by silicate minerals[1].

It is likely that the outer part of the Earth, the *bulk silicate Earth*, was more or less compositionally homogeneous during its earliest history because the oldest known rocks are only ~4.1 Ga old (the age of the Earth is 4.55 Ga), indicating that whatever rocks were created at the Earth's surface during the first ~0.5 Ga were efficiently rehomogenized in the bulk silicate Earth. It is clear now that the bulk silicate Earth has subsequently differentiated into various subreservoirs. The most striking subreservoirs are the continental crust and oceanic crust. Continental crust is on average ~35 km thick and makes up ~0.38 % by mass of the bulk silicate Earth. Oceanic crust averages ~7-10 km thick and makes up ~0.2 % by mass of the bulk silicate Earth. The remaining 99.4 % of the bulk silicate Earth is the mantle, which consists of that part of the mantle that is residual to the extraction of continental and oceanic crust, any primitive (e.g., undifferentiated) parts retaining bulk silicate Earth compositions, and any crustal components that may have been recycled back into the mantle by subduction or related processes.

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[1] Silicates are compounds based on the silica tetrahedron, in which a Si is surrounded by four O atoms.

Table 1.2 shows the average compositions of these differentiated reservoirs. A striking difference is that continental crust is considerably enriched in Si and Al compared to the bulk silicate Earth. This difference is manifested in contrasting mineralogies: the continental crust is principally made up of quartz (SiO_2), feldspars ($KAlSi_3O_8$) and some mafic (mafic = Mg, Fe, and Ca-rich) minerals, such as hornblende and/or biotite; the mantle being enriched in Mg and Fe relative to continental crust is accordingly dominated by mafic minerals, such as olivine ($(Mg,Fe)_2SiO_4$) and pyroxene ($(Ca,Mg,Fe)_2Si_2O_6$), but no quartz. As we will learn later, an even more striking observation is that the continental crust, despite its extremely small relative mass, contains most of the highly incompatible trace elements in the Earth, such as Cs, Rb, Ba, U, Th, and K (estimates range from 50 to 90%). U, Th, and K are radioactive elements, and their high abundance in the continental crust indicates that internal heat production in the Earth associated with radioactive decay has been gradually migrating towards the surface of the Earth in direct correlation with the rate of continental crust formation.

We are thus left with a myriad of questions. What processes dictate how the Earth unmixes, i.e. differentiates? In this context, we are concerned with not only the processes that control chemical mass transfer during differentiation, but also the physical processes of differentiation. How does the mantle melt? What conditions dictate the composition of the melt and how does the melt evolve in composition as it cools? How does tectonic environment control the nature of differentiation? How is melt transported from its source region to the crust? What processes allow for recycling of crustal material back into the mantle? What chemical changes occur during crustal recycling? Can we quantify how these differentiation processes and their rates have evolved through time?

There is no doubt that answering these questions have and will continue to improve our understanding of problems in global tectonics, mantle and crustal geodynamics, Earth systems history, planetary geology, and even, economic geology (i.e., the study of economic ore deposits). However, in order to pursue these questions from a petrologist's viewpoint, it will be necessary to master a number of tools in the modern petrologist's toolbox (Figure 1.3). This

Table 1.2. Comparative chemistries of different Earth reservoirs

	Bulk Silicate Earth[1]	Continental Crust[2]	Oceanic crust: Most primitive basaltic glass[3]	Oceanic crust: Calculated primary composition[4]
MgO (wt. %)	36.33	4.4	10.1	17.81
Al_2O_3 (wt. %)	4.73	15.8	16.4	12.08
SiO_2 (wt. %)	45.56	59.1	49.7	47.85
CaO (wt. %)	3.75	6.4	13.0	11.22
FeO^T (wt. %)	8.17	6.6	7.9	8.98
Mg#	0.888	0.543	0.695	0.780

[1][*O'Neill and Palme*, 1998]
[2][*Rudnick*, 1995]
[3][*Frey et al.*, 1974]
[4][*Elthon*, 1979]
Mg# = molar Mg/(Mg+Fe); FeO^T is total Fe^{2+} and Fe^{3+}

treatise will cover the fundamentals of chemical thermodynamics, mineral physical chemistry, trace-element geochemistry, radiogenic and non-radiogenic isotope geochemistry, kinetic theory, and basic fluid dynamics pertinent to magma transport. Every attempt will be made to link theory to observations, such as petrographic and field examples, insofar as it is possible to write about field observations!

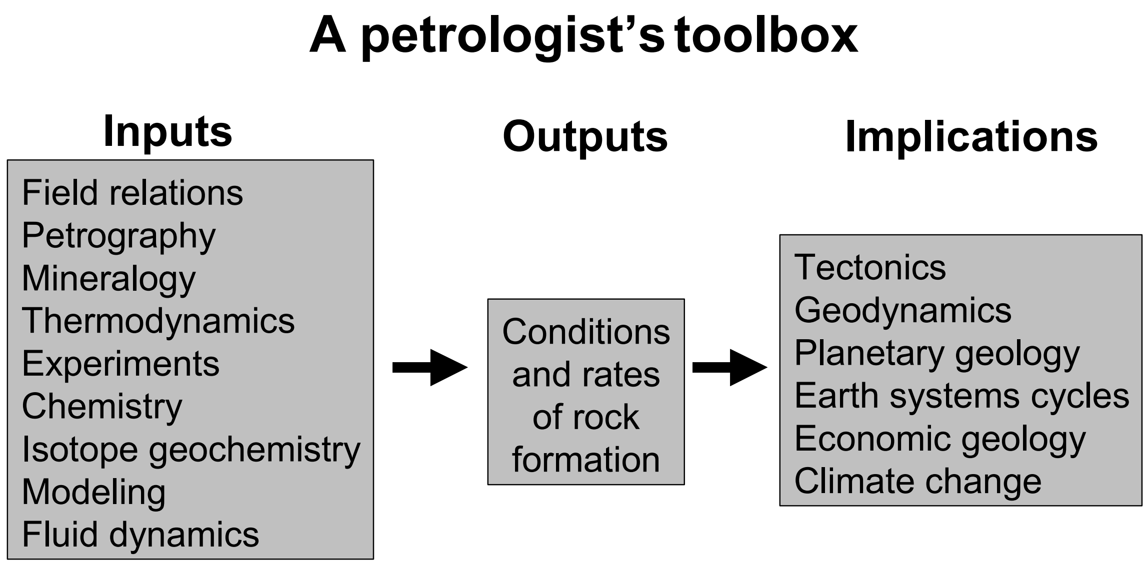

Figure 1.3 Diagram showing the tools of a modern petrologist in relation to the goals of petrology.

Chapter 2
From accretion of the Earth to plate tectonics

2.1 Making the bulk Earth

2.1.1 The Earth has an iron core

In Chapter 1, we briefly touched on the composition of the bulk Earth, but how is the bulk Earth composition determined? There is presently no rock sample that has a representative bulk Earth composition as the Earth has been grossly differentiated. The first hint that the Earth might be radially heterogeneous came as early as the 18^{th} century when Lord Cavendish of England determined the mass of the Earth[2], and hence established the density of the bulk Earth to be approximately 5.45 g/cm^3 (the presently accepted bulk Earth density is 5.25 g/cm^3). The implications of these calculations were far-reaching. Most of the rocks on the Earth's surface have densities between 2.5 and 3 g/cm^3, and thus somewhere in the Earth's interior must contain a denser component. These calculations have been further verified and refined by seismological studies and studies of the Earth's moment of inertia. In particular, in 1906, Richard Oldham of the Geological Survey of India, used seismic waves (elastic waves in the Earth's interior created by earthquakes) to show that the Earth was largely partitioned into two major reservoirs: the central core, with a radius of ~0.4 that of the Earth's radius, and the surrounding mantle. The boundary between the core and the mantle was found to coincide with a distinct density and seismic velocity jump (Fig. 2.1), confirming earlier suggestions of an Earth with a dense interior.

The combination of gravity, inertia, and seismic measurements constrained the Earth's core to have a density of ~10 g/cm^3, but density alone cannot provide unequivocal information on what the core is made of as many materials have the same density. For example, heavy metals such as silver, iron, nickel, copper, tin, molybdenum, cadmium, ruthenium, and palladium, all have densities within the range of 7 to 13 g/cm^3. In 1961, Francis Birch of Harvard University constructed an elegant experiment that resolved this amibiguity. He empirically determined how a material's seismic velocity, density, temperature and chemical composition are related (Fig. 2.2). Importantly, he showed that not only were solids more compressible than believed at that time, but most materials have an unique density-seismic velocity relationship associated with compression. In fact, at the very high pressures characteristic of core conditions, Birch not only showed that the density of iron metal increased to ~10 g/cm^3 compared to 7.9 g/cm^3 at atmospheric pressures, but also showed that the velocity-density relationship of iron metal fit very closely that inferred from seismic studies. The combination of experimental mineral physics and seismology established that the core must be predominantly made up of compressed Fe, whereas the mantle is made up primarily of silicates[3].

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[2] In 1798, Cavendish determined the gravitational constant using a torsion balance. He then used this number in a pendulum experiment and was able to determine the mass of the Earth from the period of the swinging pendulum. Even earlier, however, Pierre Bourguer of France had visited the Andes in 1748 in order to study the Earth's magnetic field. Instead, Bourguer made the interesting observation that his plumb line was deflected from vertical towards the Andes. The English astronomer Nevil Maskelyne used this observation to determine the mass of the Earth, yielding a density of 4.5 g/cm^3.

[3] In detail, the velocity-density relationship of Fe metal does not fit exactly that inferred from seismic studies of the core: seismic studies predict a slightly lower density than pure Fe (Fig. 2.2). It appears that although the core must be made up predominantly of Fe (and Ni), there must be a small proportion of a light element in the core in order to explain its slightly lower density. Elements, such as O, S, Si, and K, have all be suggested, but there is presently no consensus.