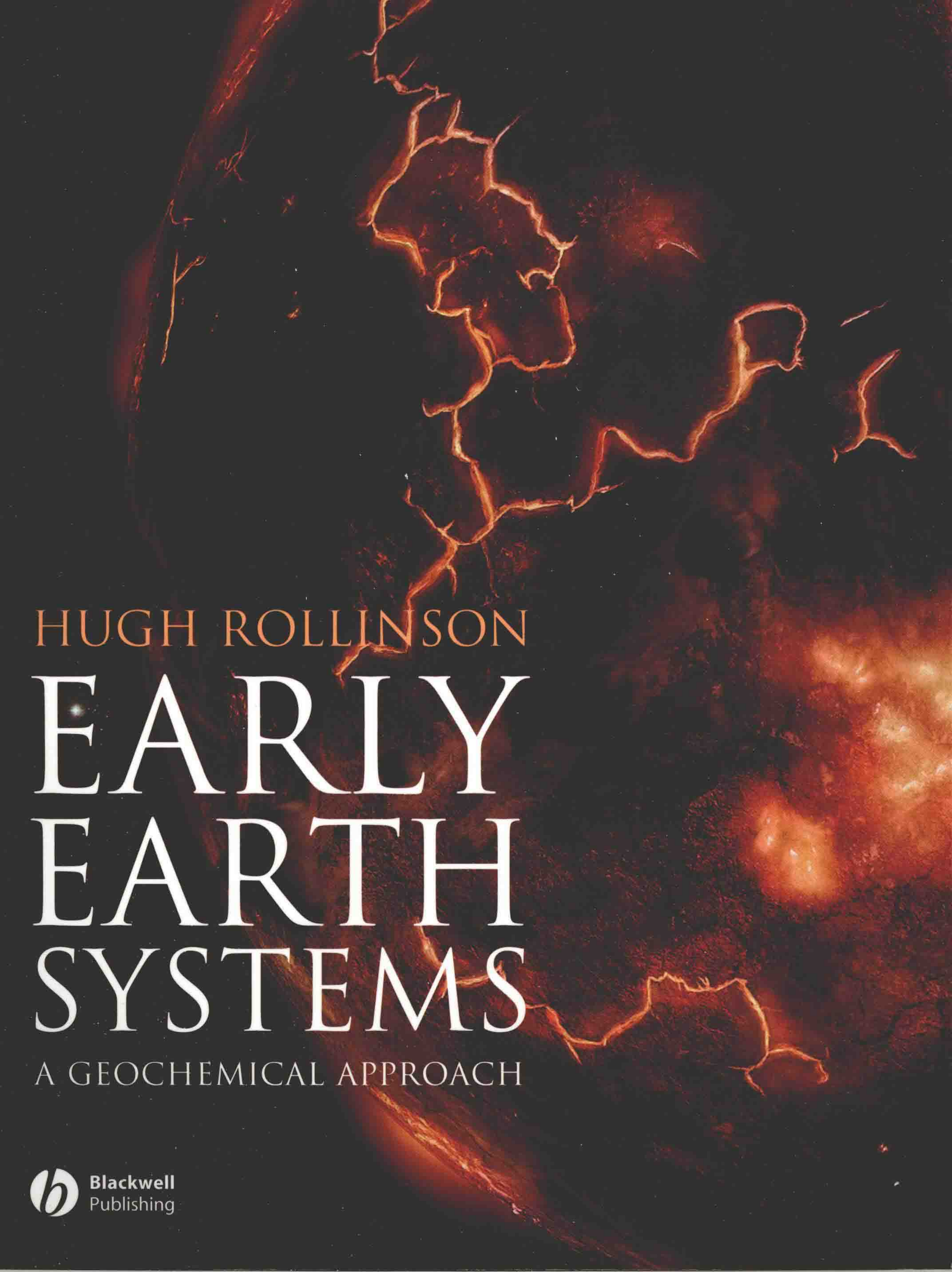

HUGH ROLLINSON

EARLY EARTH SYSTEMS

A GEOCHEMICAL APPROACH

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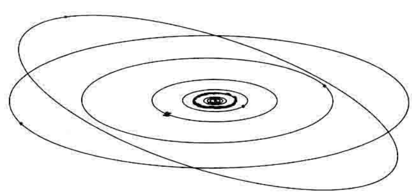

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1

THE EARTH SYSTEM

1.1 Introduction

The Earth is not the planet it used to be. It has changed substantially since it first formed and at the present time is still changing. The quest of this book is to explore the way the Earth was in its earliest history, in particular during the first 2 billion years of its 4.6 billion year life. This knowledge is vitally important, for it provides essential clues in charting how the Earth has arrived at its present physical, chemical, and biological state and leads to an understanding of the large-scale processes which bring about change in our planet.

Over the past decade, there has also been a major change in the way we think about the Earth. The Earth Systems paradigm is now strongly influencing contemporary thinking in the Earth Sciences. Thus, a distinctive feature of this text is an emphasis on the insights which the "systems approach" offers into processes on the early Earth. In particular we will explore the linkages that exist between the different parts of the Earth System. So we will discuss in some detail the interactions that take place between the Earth's crust and mantle; its oceans and atmosphere; its oceans, ocean crust, and mantle; and between the "Earth" and life. These are topics of considerable importance for the modern Earth System, and in seeking to apply this approach to the early Earth, we shall discover how the early Earth System operated.

The findings of planetology imply that during the first 100 Ma of its life the Earth must have experienced a number of extreme changes unlike anything in the recent geological past (see Chapter 2, Section 2.4). These processes shaped the primordial Earth and yielded a "starting point" for the Earth System. What exactly this primordial Earth looked like is one of the subjects for this book and it plunges us directly into some of the biggest scientific questions we might ask. For example, exactly when and how did the Earth form, and is it different from the other "terrestrial" planets? What was the composition of the earliest atmosphere and oceans? Did they come from inside the Earth, or were they acquired from elsewhere within the solar system? To what extent had the Earth melted in its first 100 Ma? What sort of imprint did this leave on the Earth's mantle and can we still see this imprint today? When did the first crust form? What did it look like, and is it still preserved, either within the continents or in the mantle? What were the conditions which permitted life to form, and how exactly and where did life begin? These questions have been asked for a long time, but in recent years we have begun to obtain some answers.

This book also tracks the changes in the Earth from its primordial state to the time at which "modern" processes were established. Whilst this time is not exactly known there is evidence to suggest that some modern processes were established by the end of the Archaean, 2.5 billion years ago, maybe earlier. The change from the primordial Earth toward the modern Earth over a period of 2 billion years requires significant interactions between all the major reservoirs of the Earth system and understanding these requires insights from the Earth Systems way of looking at the Earth.

1.1.1 Earth system science

1.1.1.1 The new paradigm

Over the past two decades Earth Scientists have begun to realize that a reductionist paradigm provides an incomplete picture of the Earth. Segmenting the Earth Sciences into fields such as "igneous petrology" and "carbonate sedimentology" with firm boundaries demarcating these subdisciplines provides inadequate answers to whole Earth problems. The new insights of the past two decades have been prompted by two disparate areas of research. First, comparative planetology arising out of the NASA space missions, has forced us to think about the Earth in the context of our planetary neighbors. This in turn has contributed to a more "global" view of the Earth. At the same time scientists studying climate and the atmosphere have become increasingly aware that they could not fully explain these systems by studying them in isolation. They found that they needed contributions from other parts of the Earth System to understand the parts that they were considering. In particular the scientific elements of the Gaia hypothesis of James Lovelock have been a stimulus to this process (e.g. Lovelock, 1988). These new approaches have caused Earth Scientists to think beyond the traditional "boxes" of their specialist disciplines and to consider the links that exist between their particular "box" and other parts of the Earth System.

Forty years ago, our understanding of how the Earth works was revolutionized by the theory of Plate Tectonics. This theory, however, only described the workings of the solid Earth, with a particular emphasis on the origin of ocean basins and active mountain belts. In contrast, Earth System Science is about much more than simply the solid Earth. Its scope encompasses the whole of the Earth System – the deep Earth and surficial processes; it includes the oceans, the atmosphere, and the Earth's diverse ecosystems. It also offers a new level of integration between the Earth Sciences, (traditionally the "solid" Earth), Environmental Science and Physical Geography (traditionally surface processes and ecosystems), Oceanography, and Atmospheric Science. Furthermore, Earth System Science has implications which reach far beyond the way in which we do science, extending into the realms of environmental management and policy (Midgley, 2001).

Hence the philosophy of Earth System Science is holistic. It argues that we need to explore the Earth System as an integrated whole, rather than simply as a series of separate entities. As stated above, the reductionist approach has now been shown to be wanting, for it is incapable of tackling large-scale issues such as global change. The new paradigm encourages interdisciplinary thinking and an exploration of the processes which *link* the different "boxes" (see Fig. 1.1), in order first to describe, and then to quantify the exchanges that take place. There is a much greater emphasis than before on defining reservoirs, residence times, fluxes, transport mechanisms, and transport rates (Table 1.1). The Earth System Science perspective therefore, sets a new research agenda for Earth Scientists, Environmental Scientists, Atmospheric Scientists, Oceanographers, and Geomorphologists, for suddenly the interesting science is that which is happening at the *boundaries* of the once traditional disciplines.

1.1.1.2 The Gaia hypothesis

One area of thought which has had a profound influence on the evolution of Earth System Science thinking has been the Gaia hypothesis, popularized by James Lovelock (Lovelock, 1979, 1988). Lovelock, a planetary scientist, recognized that the Earth is unique amongst the terrestrial planets in that it possesses an atmospheric blanket, which, in the words of the classical Goldilocks narrative, was not too hot (as is Venus) and not too cold (as is Mars) but "just right" for life to exist. Lovelock, following the work of planetary scientists in the 1950s and 1960s, argued that the uniqueness of the Earth's atmosphere, as evidenced by its disequilibrium state, was because the planet was the host to life. Thus he turned on the head conventional arguments that state that the Earth is the home to life *because* it had a suitable atmosphere. In other words the atmosphere which we have today was produced by and is controlled by "life."

This idea that there is an interaction between living organisms and the Earth has given rise to the concept of "self-regulation" in the Earth System. In his early work Lovelock likened this to the process of homeostasis – the process whereby a living organism regulates its internal environment in order to maintain a stable condition. This caused some misunderstanding of his ideas, for many believed that he was arguing that the Earth *was* a living organism. The scientific community began to back off at this point and a number of quasi-religious groups weighed in, hijacking the Gaian concept, taking it away from its scientific roots in a completely different direction. Sadly, this diversion of the Gaian concept has harmed its scientific credibility. Nevertheless, the ideas promoted by James Lovelock have had an important impact on the development of Earth System Science. Whether or not they are given the Gaian label is relatively unimportant.

Lovelock's great insight was to recognize that life affects the global environment and profoundly affects surficial processes on the Earth. Thus when the Earth System is perturbed, either from outer space or from the interior of the Earth, the Earth's self-regulatory systems come into play, eventually restoring the system to its original conditions. This makes the Earth Surface System resilient to all but the most extreme perturbations. Lovelock and subsequent workers have illustrated this process with a simple model which they called "Daisyworld."

Daisyworld is a computer model of a simplified Earth in which variation in the Earth System is described by one parameter, surface temperature, which in turn is affected by a single property of living matter – its reflectivity to solar radiation – its albedo (Watson & Lovelock, 1983). There are a number of versions of Daisyworld, for it is an evolving model. Here the version of Kump et al. (1999) is used. Daisyworld is a world in which there are vast tracts of white daisies. They grow on grey soil, which is the only other surface feature of this Earth model. Daisyworld is subject to a sun with an increasing luminosity. It might be thought that an increase in solar luminosity would lead to an increase in surface temperature and the daisies would die. What in fact happens in the model is that, in response to the increased surface temperature, the white daisies increase in abundance. The greater proportion of white daisies leads to an increased reflectivity (albedo) of the Earth's surface, leading eventually to a lowering of the surface temperature. Thus Daisyworld is an example of a negative feedback loop (Table 1.1) and illustrates the interconnectedness between life and the Earth in a manner which regulates the Earth's surface temperature. It is therefore a simplified model for how the Earth's climate system works. The important general points are that first, there can be a realistic coupling between life and the physical environment and second that the Earth can behave as a self-regulating system.

1.1.2 A systems approach to the modern Earth

The application of systems theory to the Earth recognizes that rather than separating the Earth into its component parts, there is a focus on the whole. There is a recognition that natural systems are open and interact with other systems outside of themselves and that through this interaction they acquire new properties and evolve over time. Earth System Science views the Earth as a synergistic physical system, governed by complex processes involving the solid Earth (crust, mantle, and core), including its land surface, its atmosphere, hydrosphere (oceans, rivers, and lakes), cryosphere (ice caps, sea-ice, and glaciers), and biosphere. The relevant interactions operate over timescales ranging from milliseconds to billions of years and on spatial scales from microns to millions of kilometers. The Earth

Systems approach not only includes solid Earth processes such as the origin and motion of continents and the creation and destruction of ocean basins, but also the origin and evolution of life and the changing pattern of the Earth's climate through time.

The language of systems is briefly summarized in Table 1.1 and these terms are now increasingly used to describe Earth processes. Earth System Science places a strong emphasis on quantifying processes as a prelude to mathematical modeling. For example, models of the Earth's modern climate system are becoming increasingly mathematically sophisticated with the intention of producing three-dimensional global climate models. Similarly, a major goal of modern geochemistry is to quantify elemental fluxes into and out of major Earth reservoirs.

This type of approach is possible for processes which we can quantify now and in the recent past, but as we go further back in time quantification becomes much more difficult. It is clear that there is a trade off between the timescale of the model and its mathematical complexity. For long timescales, of the type considered in this book, Earth system processes can only be satisfactorily defined by zero-dimensional, box models.

1.1.2.1 Identifying reservoirs and fluxes – box models for the Earth

Box models are diagrams in which the different components, or reservoirs in the Earth system are represented as boxes. These are particularly useful for illustrating processes which need to be viewed over a long timescale. Currently there are a number of box models relevant to the Earth's surface systems, the most detailed of which are being developed by climate modelers. One example is the Grid ENabled Integrated Earth system model (GENIE), which seeks to develop a grid-based computing framework for the Earth System on a timescale of tens of thousands of years, to include the atmosphere, ocean, sea-ice, marine sediments, land surface, vegetation, soil, and ice sheets (http://www.genie.ac.uk/).

The importance of timescale in developing box models is illustrated well by Kump et al. (1999) in their discussion of the carbon cycle. They show that the organic and inorganic carbon cycles operating on timescales of less than a century include the atmosphere, biosphere, the oceans, soil, and marine sediments, but on a longer timescale (thousands to millions of years) the additional reservoir of sedimentary rocks also becomes important, since it contains up to three orders of magnitude more carbon than the Earth's surface reservoirs.

Probably the most comprehensive box model for the whole Earth is the GERM model (Geochemical Earth Reference Model, Staudigel et al. (1998), http://www.earthref.org/), Fig. 1.1. The purpose of this model is to provide a geochemical reference model for the Earth, similar to the Preliminary Reference Earth Model (PREM) used in geophysics (see Chapter 3, Section 3.1). In detail the model

- divides the modern Earth into a complete set of geochemical reservoirs, (Fig. 1.1 and Table 1.2); and will
- provide an internally consistent set of data describing their chemical and isotopic composition; and seeks to
- establish values for the chemical fluxes between these reservoirs over relevant timescales (ca. 1000 yr to 1 billion yr).

It is this type of model which is most relevant to the systems of the early Earth. For at this time period the only processes which can be discerned are those operating over long timescales and the Earth reservoirs are only broadly defined.

1.1.2.2 Quantifying fluxes – geochemical cycles

Where there are exchanges between different reservoirs in a geochemical box model of the Earth, and these can be quantified, then these exchanges may be described in terms of geochemical cycles. The basic conditions for identifying an elementary geochemical cycle are that the "essential reservoirs must be identified, their contents estimated and the fluxes between the reservoirs must be evaluated over a sufficient length of time compared to the relaxation time (homogenization time) of the system" (Albarede, 2003). The basic transport equations for modeling geochemical cycles are given by Lasaga and Berner (1998).

TABLE 1.1 A summary of some of the key terms used in Earth System Science, many of which were first used in the field of chemical engineering.

System	An entity made up of different parts, but which are related. Together the different parts function as a whole. Individual parts are called components
Reservoir	A part of a system, defined in terms of the amount of material contained (either as mass units or volume). Usually one of the "boxes" in a box model
Flux	The amount of energy or matter that is transferred into/out of a reservoir in unit time. A more general term for this process is *Mass Transfer*
Feedback loops	A linkage between two or more components of a system so that there is a self-perpetuating mechanism of change. A set of actions produces automatic reactions within the system. Feedback may be positive, amplifying change, or, negative, diminishing the effects of change.
Steady state	The condition when a system is unchanging in time. A reservoir is in a steady state when the inflow and outflow are equally balanced
Perturbation	A temporary disturbance to a system
Forcing	A long-term, persistent influence on a system bringing about a disturbance to the system
Residence time	The average length of time a substance spends within a reservoir in a steady state with respect to the abundance of that substance

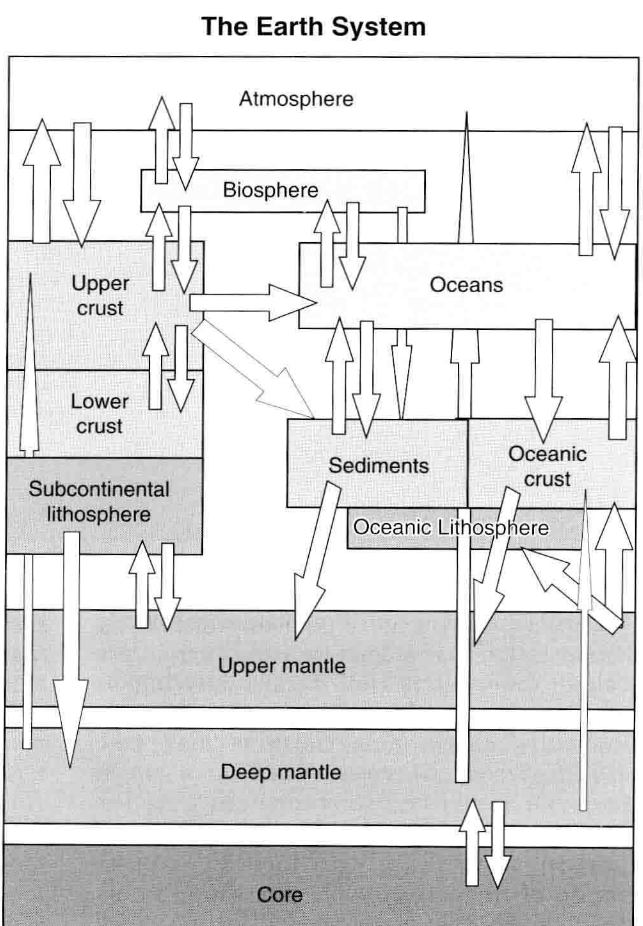


FIGURE 1.1 A box model for the Earth System, showing some of the major reservoirs and the interactions (white arrows) between them.

TABLE 1.2 The principal reservoirs in the GERM Earth model (Source http://www.earthref.org).

Atmosphere	*Stratosphere*				
	Troposphere				
Hydrosphere	*Cryosphere*				
	Lakes				
	Rivers				
	Seawater	Deep water			
		Surface water			
		Hydrothermal vents			
Solid Earth	*Silicate Earth*	Continental crust	Lower crust		
			Middle crust		
			Upper crust		
		Oceanic crust	Mafic crust	Fresh MORB	E-MORB
					N-MORB
				Hydrothermal systems	
				Komatiites	
				Mature oceanic crust	Extrusive rocks
					Dykes
					Gabbro section
					Transition zone
			Marine sediments	Subducted sediment	
			Oceanic plateaux		
			OIB		
		Mantle	Continental mantle		
			Depleted mantle		
			Enriched mantle	EMI	
				EMII	
				FOZO	
				HIMU	
	Core				

This definition of a geochemical cycle demonstrates that recognizing the appropriate scale of the cycle is vital. Particularly important is the timescale. It is clear that on short timescales catastrophic transfers may take place between reservoirs which on a longer timescale would be "smoothed out." As has already been noted, the appropriate choice of timescale is particularly pertinent to a consideration of the carbon cycle (see Kump et al., 1999), for whether or not the sedimentary rock reservoir is included, completely alters the mass balances in the system. The choice of timescale will also vary for the geochemical cycle of different elements in the same set of reservoirs. This is because elements have different residence times in the same reservoir. An unreactive element will have a long residence time, whereas a reactive element will be quickly removed from the reservoir and thus has a short residence time. Defining the length scale of geochemical reservoirs is also an important factor in developing models of geochemical cycles. In some cases linked reservoirs should be merged and at other times separated. This may be linked back to the timescale of a geochemical cycle.

The simplest geochemical cycles are based around the concept of maintaining a steady state to satisfy the constraints of mass balance. More complex cycles incorporate the dynamic response of the system to perturbation (Lasaga & Berner, 1998). In this case, the resultant feedback processes are normally quantified using linear models, but, depending on which geochemical cycle is being considered, more complex, nonlinear responses may be more appropriate.

The key point for an Earth Systems approach to the Earth is that models of geochemical cycles provide the means whereby exchange processes can be quantified. This quantification is element-specific and is a function of the timescale under consideration. It provides the basis for calculating the steady state abundances of particular substances such as oxygen or carbon dioxide in the atmosphere, or sulfate in the ocean. Since these abundances are related to fluxes which relate to biogeochemical and physical processes, we can gain an understanding of the controls on key variables in the Earth system.

An issue of particular interest in the early Earth is the rise of atmospheric oxygen (Chapter 5, Section 5.3). Figure 1.2 shows a box model for the modern steady state oxygen cycle. This model can be used to show how a modern geochemical cycle might be used as the basis for understanding fluxes in the Archaean. For example we can explore how oxygen levels might have been controlled in the early Earth – through say a smaller amount of organic carbon or a greater volume of reducing gases released from the mantle. The model also illustrates the importance of "Earth systems thinking," since it shows how mantle processes can influence the composition of the atmosphere. Other examples of whole-Earth geochemical cycles – for water, carbon and nitrogen – are given in Chapter 5 (Section 5.1). These geochemical cycles are characterized by reservoirs that are large and

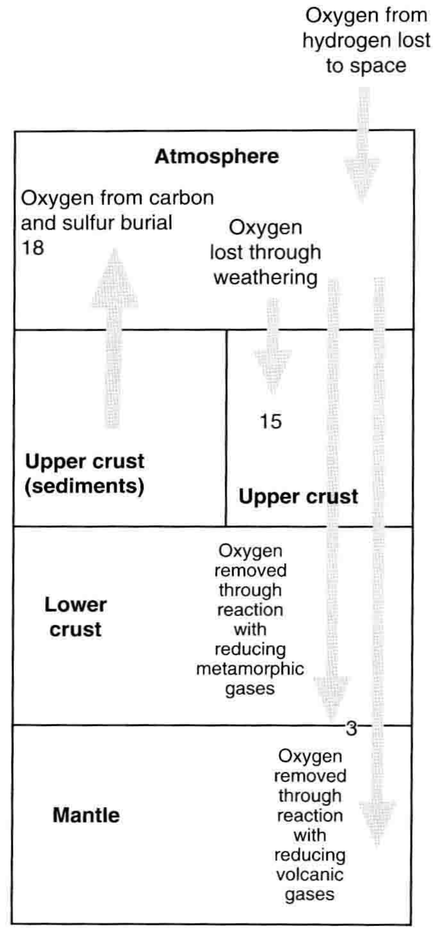


FIGURE 1.2 Box diagram for the modern oxygen cycle (after Catling & Claire, 2005). The direction of the arrows show the contributions to and removal from the atmosphere. The value of the fluxes is in Tmol/yr (10^{12} mol). The principal contributor of oxygen to the atmosphere is the burial of organic carbon and pyrite. Oxygen is lost through continental weathering and through reaction with reducing gases from the deep Earth.

broadly defined, and a timescale of mass transfer which is between 10^6 and 10^9 years.

1.1.3 Early Earth Systems

If the defining character of modern Earth System Science is "the need to study and understand the between-component interactions" (Lawton, 2001), it is now time to enquire how this might be applied to the early Earth. During its first 100 Ma, perhaps even within the first 30 Ma, the Earth differentiated into some of its principal reservoirs – the core, mantle, oceans, and atmosphere (see Chapter 2, Section 2.4, and Chapter 5, Section 5.4). Also at this time a crust formed, initially basaltic in composition (Section 4.5, this volume), followed some time later by a felsic, continental crust (Section 1.4). And then, at some stage, life appeared.

In these very first stages of Earth history it is likely that the interactions between the reservoirs were dynamic and extreme. For example it is possible that there was a strong interaction between the mantle and a primitive atmosphere during the Earth's magma ocean stage whereby the mantle became hydrated (see Chapter 2, Section 2.4.1). These likely interactions demonstrate that in some ways the interconnectedness of the Earth System was *most* apparent, and the Earth Systems approach is most relevant, during the Earth's earliest history, in this case within the first 30–100 Ma.

A central question however, is the extent to which we can access the Earth system at the earliest stage of its history. How much information is there? It is difficult enough grappling with the modern Earth system, to which we have good access. Is it possible to know enough about the early Earth to attempt to describe the early Earth system? It is the claim of this book that such an approach is possible. In the following sections of this chapter we consider the nature of the geological record for the early Earth and examine the data on which much of what is written here rests. Furthermore, modern isotope studies show that the Earth's mantle preserves a remarkable record of its earlier history and that events in the very early development of our planet have left their fingerprint in the modern mantle.

Inevitably we can only understand the early Earth in a qualitative rather than quantitative manner. The best we might hope for is a semi-quantitative understanding and so, in this sense, our understanding of early Earth system will never compete with our understanding of the modern Earth. Nevertheless, even a better qualitative understanding of the large-scale processes and interactions in the early Earth system is vitally important, for many of the basic parameters of the early Earth are, at present, very poorly defined. For example, in the case of the first continental crust, we are not clear on how it formed. In the language of systems we have not even agreed on the inputs and outputs. We are still constructing the box model. The same can be said for most aspects of the early Earth. Compared with our knowledge of modern geochemical cycles, this seems rather elementary – but constructing basic models for the components of the early Earth is a necessary prelude to understanding the earliest Earth system. This book does not therefore provide a sophisticated quantification of geochemical fluxes in the early Earth, but what it does do is set the agenda for such studies in the future. If we can begin to agree on the box models, then in the future we have some hope of beginning to quantify the mass transfers between them. And if we know what the "boxes" are, then we shall know where to look and what to measure. Part of this agenda is to demonstrate that in the Archaean the "solid Earth" has an important contribution to make to our understanding Earth systems processes, contrary to the emphasis of much modern Earth systems science.

Along the way we will discover discrepancies between what observational approaches and theoretical approaches can tell us about the early Earth and also many examples of contested interpretations of the rock record. Controversy of course is the life-blood of science and normally leads to an improved understanding of a topic. However, where controversial interpretations are still unresolved we must tread carefully not to build these data into our models.

This approach requires insights from all branches of the Earth sciences, although the reader will detect a bias toward geochemistry, for this is a particularly powerful tool with which to explore the early Earth and the author's prime area of expertise. An exploration of early Earth Systems is different from the climate modeling of modern Earth System Science with its strong societal implications, but is still of value, for an "Early Earth perspective" raises profound philosophical questions about origins, which we do well to ponder.

1.2 The nature of the early geological record

How can we know about the early history of the Earth? What evidence do we have for this period of Earth history, and how reliable is this evidence? In this section we review the nature of the early geological record, enquire about the types of data that might be extracted from it and assess the quality of the information available. Of course in addition to the geological record, there is also much information to be gained about the early Earth from the planetary record. This is discussed in Chapter 2.

The earliest period of Earth history is subdivided into the *Hadean* and the *Archaean*. Here, the Hadean (literally – the hidden period of Earth history, or for some, the "hellish period"), is taken to be the period of time between the formation of the Earth at about 4.6 billion years and the beginning of the Archaean. However, the beginning of the Archaean (literally – the beginning of the rock record), has not been defined by the International stratigraphic commission (Gradstein et al., 2005). Here, following common usage, it will be taken as 3.8 Ga, and the Archaean as the period of time between 3.8 and 2.5 billion years ago (Fig. 1.3). In this text billions of years are abbreviated to gigayears (Ga – 10^9 years ago) and millions of years to mega-years (Ma – 10^6 years ago).

There are few terrestrial materials older than about 3.8 Ga and so our knowledge of the Hadean, the first 750 Ma of Earth history, depends upon a small number of terrestrial samples, supplemented with inferences from the study of meteorites and planetary materials and deductions based upon the distribution of some radiogenic isotopes in the Earth's mantle. In contrast Archaean rocks abound and regions with rocks dated between 2.5 and 3.8 Ga are found in most major continents. A map showing their distribution is given as Fig. 1.4.

Regions where Archaean rocks are extensively preserved are known as *cratons*. This term

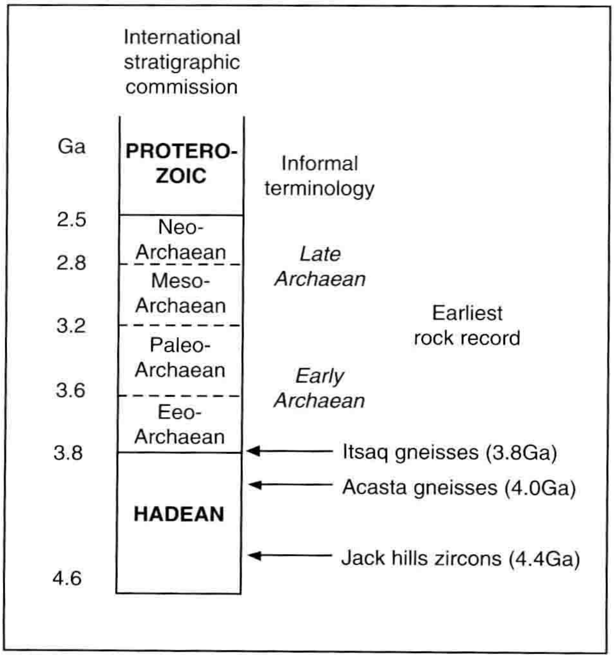


FIGURE 1.3 A geological timescale for the earliest part of Earth history (partially after Gradstein et al., 2005).

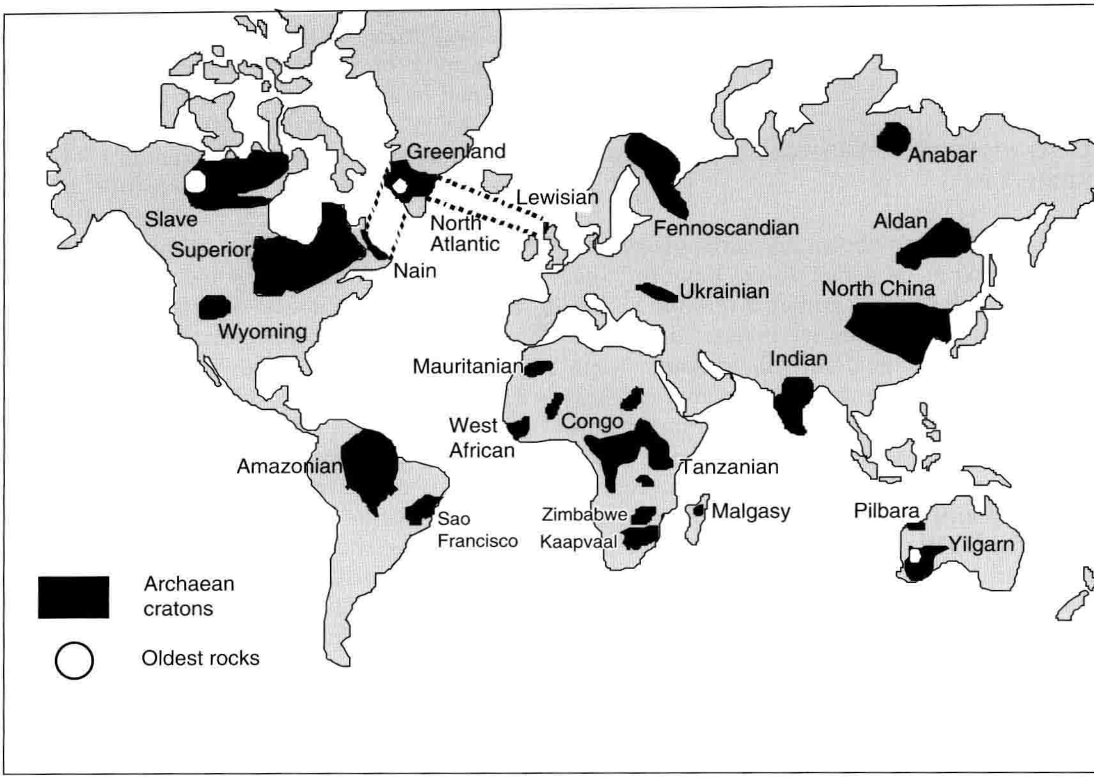


FIGURE 1.4 The distribution of Archaean rocks worldwide, showing the principal cratons and regions where the oldest rocks are preserved. Not all these areas contain exposed Archaean rocks, for some cratons are now partially covered by younger sediment, or have been reworked during later orogenic events. Not shown is the Archaean Enderby Land Craton in Antarctica. The dashed lines indicate the rocks of the North Atlantic Craton now separated by the creation of the Atlantic Ocean.

implies an area of continental crust, normally made up of crystalline rocks, which has been stable for a very long period of time. The term is synonymous with the older term *shield*. The names of some of the major cratons are given in Fig. 1.4 and a typical craton, the Zimbabwe Craton is illustrated in Fig. 1.5.

1.2.1 Types of Archaean crust

Traditionally Archaean crust has been subdivided into two types – granite–greenstone belt terrains and high-grade gneiss terrains. This subdivision reflects a major contrast in both lithological association and metamorphic grade. Typical granite–greenstone belt terrains would be the Zimbabwe or Superior Cratons and a typical high-grade gneiss terrain would be the North Atlantic Craton. In this book a slightly different approach is taken to that previously used in the literature, and Archaean crust will be classified as one of three types of lithological association – Archaean greenstone belts, late Archaean sedimentary basins, and Archaean granite–gneiss terrains.

Of these, late Archaean sedimentary basins are the most "modern" in their form and can be clearly recognized as upper crust. Similarly, the relatively low metamorphic grade of many Archaean greenstone belts indicates that they too represent upper Archaean crust. In contrast, Archaean granite–gneiss terrains more commonly belong to the deeper continental crust and variously represent middle to lower continental crust. There are only a few places

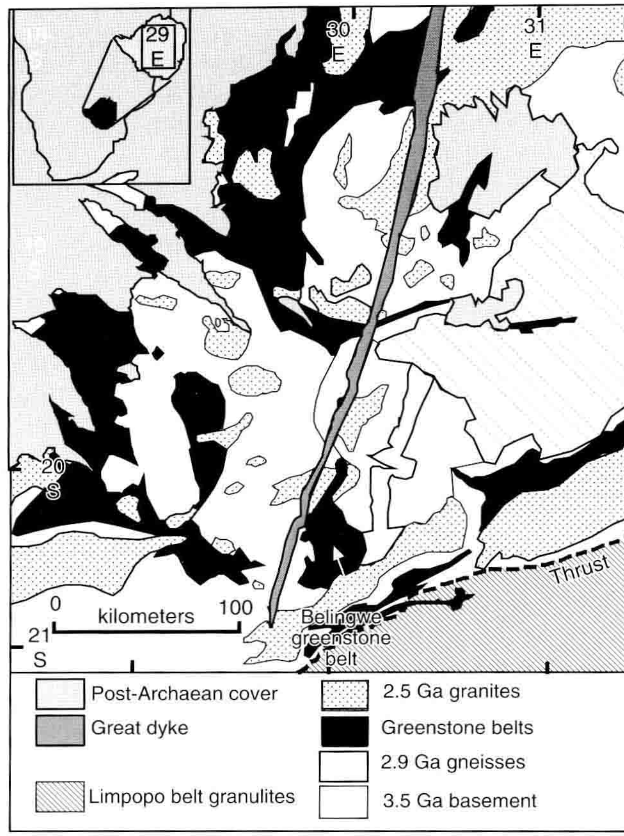


FIGURE 1.5 The central part of the Archaean Zimbabwe Craton, showing the principal geological components: granitoid gneisses of various ages, greenstone belts, younger granite intrusions (2.5 Ga granites) and the late Archaean "Great Dyke." To the south the Archaean Limpopo Belt granulites are thrust onto the Zimbabwe Craton. In the northwest the craton is partially covered by younger sediments. The area of the map within Zimbabwe is identified on the inset map.

in the world where these relationships are made clear, one of which is the Kapuskasing structural zone in the southern part of the Superior province in Canada. This region shows a west to east transition from a shallow level of erosion (<10 km) to a deep level of erosion (20–30 km), over a distance of about 150 km, exposing a cross section through Archaean continental crust. With increasing depth of erosion the lithologies change from the metavolcanic rocks of the Michipicoten greenstone belt, through the midcrustal Wawa granitoid gneisses, into the granulite facies gneisses of the Kapuskasing zone, showing a transition from Archaean upper to lower crust (Percival & West, 1994).

Later in this chapter, each of the major Archaean lithological associations – greenstone belts, late Archaean sedimentary basins, and granite–gneiss terrains – is examined in detail, in order to clarify which type of information each preserves about the early Earth. However, before we turn to this subject it is worth reflecting in more detail on the precise nature of the Archaean geological record.

1.2.2 The Archaean geological record

The Archaean geological record presents particular problems and poses particular challenges. Archaean rocks have commonly experienced long deformation histories. Frequently they are either metamorphosed and/or altered, and recovering their original character requires patient fieldwork, sometimes in very remote locations, and thoughtful geochemistry. When we come to the earliest history of the Earth we find that our knowledge base is very thin and the only information is preserved in single mineral grains – zircons (zirconium silicate grains, $ZrSiO_4$) of Hadean age, protected in younger Archaean sediments (see Section 1.4.3).

A very specific challenge in the Archaean geological record are those rocks which are without modern counterparts. One example is the ultramafic rock komatiite, first discovered as a lava in the Barberton area of South Africa. At the time of their discovery, in the early 1960s, ultramafic lavas were thought to be an impossibility. Over the past 30 years we have learned a great deal about komatiites and they are discussed in detail in Chapter 3 (Section 3.2.1.2). Nevertheless, why they are more common in the early history of the Earth than at the present is an important scientific question.

Similarly, banded iron formation (BIF), a sedimentary rock produced by chemical precipitation, is extremely rare in the Phanerozoic but common in the Archaean and Proterozoic record. Explaining its origin in terms of the atmospheric or ocean chemistry of the early Earth is an important part of recovering the history of early Earth. This is discussed in Chapter 5 (Section 5.4.3.2).

1.2.3 Advances in geochronology

A repeated theme, which will be recognized in the later parts of this chapter, is the importance of U–Pb zircon geochronology in unraveling the early history of the Earth. It is impossible to overemphasize the importance of the recent advances in this field for the study of the early Earth.

Zircon is an important mineral in geochronology for it contains a sufficient quantity of the trace element uranium to provide a workable chronometer and it is a robust "time capsule," which can survive even under extreme conditions. In the past conventional zircon dating methods relied on studying groups of zircon grains, but the method was unable to take account of the complexities that may exist within a single zircon grain. A major breakthrough came with the development, in the early 1980s of a single grain method, using the Sensitive High Resolution Ion Microprobe (SHRIMP). The SHRIMP instrument, developed first at the Australian National University, allowed the measurement of very precise U–Pb ages from within individual zircon crystals (Compston et al., 1982). In fact individual zircon grains, or areas of grains, as small as 20–30 μm across can now be dated with a precision of 1–2 Ma, in rocks as old as 3–4 Ga.

This method is now widely used and is hugely important in stratigraphy of all geological ages. The detailed examination of individual zircon grains now permits the recognition of multiple events in metamorphic rocks; the recognition of different populations of detrital zircons in Archaean clastic sediments, and of inherited, xenocrystic zircons or zircon cores in Archaean granitoids. In fact it is single zircon grains, surviving through several cycles of erosion, sedimentation, and metamorphism, which are the most ancient terrestrial materials so far discovered (see Section 1.4.3).

1.3 Archaean lithological associations

Archaean Cratons contain three different types of lithological association, each of which provides important information about the early stages of the Earth System. These associations are Archaean greenstone belts, late Archaean sedimentary basins, and Archaean granite–gneiss terrains. We will consider each in turn.

1.3.1 Archaean greenstone belts

Archaean greenstone belts are one of the most important primary sources of information about surface processes in the early Earth. Greenstone belts are sequences of volcanic and sedimentary rocks, in varying states of deformation, which occur as sublinear belts, tens to hundreds of kilometers long in Archaean Cratons (Fig. 1.5). They tend to form a striking physiographic and geological contrast to the granitoids plutons and gneisses, which most commonly surround them. Their name comes from the color of the lightly metamorphosed

TEXT BOX 1.1 Radioactive dating methods applicable to the early Earth

1. Isochron calculations

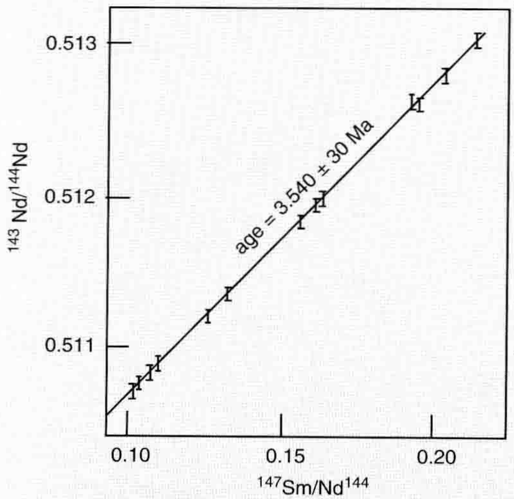
An isochron diagram is an bivariate plot of the ratio of a radioactive parent isotope to a reference stable isotope, plotted on the *x*-axis of the diagram, to the ratio of the radioactive daughter isotope relative to the reference isotope plotted on the *y*-axis, for a suite of cogenetic samples. For example, in the Sm–Nd (samarium–neodymium) isotopic system ($^{147}Sm \rightarrow {}^{143}Nd$), an isochron diagram plots the $^{147}Sm/^{144}Nd$ isotope ratio on the *x*-axis and the $^{143}Nd/^{144}Nd$ isotope ratio on the *y*-axis. In a geochemical system that has been closed to outside interferences, a suite of samples which formed at the same time, but which have slightly different Sm/Nd ratios, will plot on a straight line, the slope of which is proportional to the age of the samples (see Box 1.1 Fig. 1). The age of the sample suite is calculated from the equation

$$t = 1/\lambda \ln(\text{slope} + 1)$$

where t is the age of the rock in years and λ is the decay constant (the decay constant for the Sm–Nd system is 6.54×10^{-11} yr^{-1}). The calculated age has an error, based on the statistical fit of a straight line to the sample points and the statistical errors in the isotopic measurements. The isochron age of an igneous rock therefore, is the time at which the rock crystallized and the isotopic system became isolated from that of neighboring rocks.

Isochron age calculations are commonly made for the Rb–Sr (rubidium–strontium), Sm–Nd (samarium–neodymium), and U–Pb (uranium–lead) radioactive systems. They are most commonly applied to whole-rock systems, that is, a suite of samples thought to have formed at the same time, such as an igneous pluton or a suite of lavas. Isochron age calculations may also be made for a suite of minerals in a rock, in which case they date the time at which the minerals "lost isotopic contact" with each other, that is, became closed systems. This approach can be useful in dating metamorphism.

Isochron age calculations are as good as the assumptions behind them – namely that the samples all formed at the same time and that they have been geochemically undisturbed since that time. The best isochron ages will have errors of a few tens of millions of years on an age of 2000–3000 million years.



BOX 1.1 FIGURE 1 Isochron diagram for the Sm–Nd system showing a suite of 13 samples which formed at the same time, but which have slightly different Sm/Nd ratios. The calculated age for the 13 samples is 3,540 ± 30 Ma.

2. Model age calculations

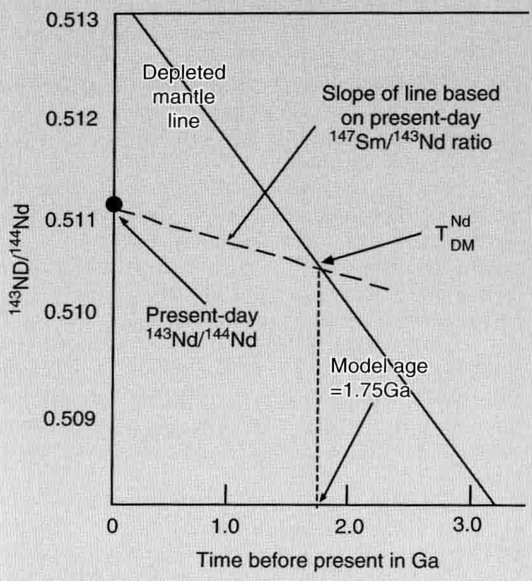
Model age calculations are based, as their name implies, on a particular model of mantle isotopic evolution. They are a measure of the time when a particular sample became separated from its mantle source. They are most commonly used for the Nd-isotopic system, but increasingly are also being used in the Hf-(hafnium) isotope system.

A model Nd age is based upon the present-day measured $^{143}Nd/^{144}Nd$ and $^{147}Sm/^{144}Nd$ isotope ratios in a rock sample. The present-day $^{143}Nd/^{144}Nd$ isotope ratio is extrapolated back in time, using the measured $^{147}Sm/^{144}Nd$ isotope ratio, to the time when it was the same as that of the mantle. This is effectively "undoing" the radiogenic ingrowth in that sample of ^{143}Nd over time. One of the most frequently used mantle models is that of the depleted mantle (DM).

Graphically the process is illustrated in Box 1.1 Fig. 2. Normally, however the result is calculated from the measured isotopic ratios in the sample compared to reference ratios for the depleted mantle.

Model Nd-ages are as good as the assumptions they are based on, that is, the mantle model used and the assumption that the sample came from a depleted mantle source. They work quite well for igneous rocks. For sediments they can measure the age of the sedimentary source.

TEXT BOX 1.1 *(Cont'd)*



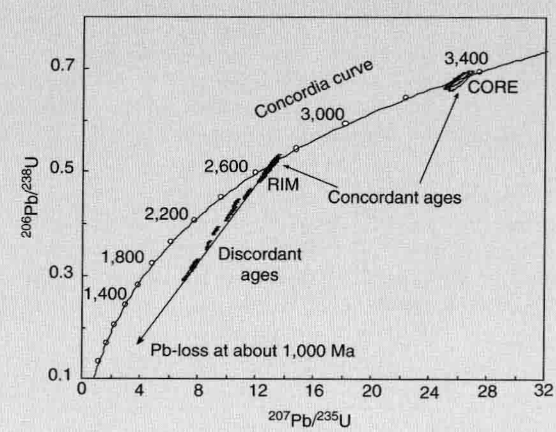
BOX 1.1 FIGURE 2 The evolution of the depleted mantle with respect to $^{143}Nd/^{144}Nd$ over geological time. The model depleted mantle neodymium age (T^{Nd}_{DM}) is the time at which the $^{143}Nd/^{144}Nd$ ratio of the sample is the same at that in the depleted mantle.

3. U–Pb (uranium–lead) zircon ages

There are a number of ways of obtaining U–Pb ages on zircons, but in the last two decades the use of the ion microprobe has become the method of choice. Zircon (zirconium silicate, $ZrSiO_4$) is a common mineral in granitic rocks, often abundant as a detrital mineral in clastic sediments and a rare accessory mineral in mafic rocks and some metamorphic rocks. The frequent occurrence of the mineral zircon, and its resilience to later thermal events that would reset other isotopic systems are the reasons for its frequent use in studies of the early Earth.

Zircons are analyzed in an ion probe using the technique of secondary ion mass spectrometry in which a small area (20 μm in diameter) of a zircon grain is bombarded by high-energy ions and this leads to the ejection, or sputtering, of both neutral and charged species from the zircon grain. Of interest are the charged ions of U and Pb and these are filtered and analyzed according to their mass. Of particular interest are ^{206}Pb, the decay product of ^{238}U and ^{207}Pb, the decay product of ^{235}U.

The results are plotted on what is known as a Concordia diagram (see Box 1.1 Fig. 3), a graph of $^{206}Pb/^{238}U$ and $^{207}Pb/^{235}U$ ratios, showing the locus of agreement of U–Pb ages obtained in the two U-Pb decay systems (^{238}U and ^{235}U). Analyses which plot on the Concordia curve indicate the time of crystallization of the grain. Analyses which plot off the Concordia are interpreted in terms of late Pb-loss from the grain, most normally as the result of a subsequent thermal event which affected the rock.



BOX 1.1 FIGURE 3 U–Pb Concordia curve showing the compositions of individual analyses as error ellipses. The data, which are for a single grain, show that the grain has a core which is about 3,350 Ma old and a rim which is about 2,700 Ma old. The rim has experienced subsequent lead loss, giving rise to a series of discordant points on the Concordia diagram. This lead loss reflects a later thermal event at about 1,000 Ma.

U–Pb ages on zircons yield a wealth of data. Granitic igneous rocks can be dated for their time of crystallization, metamorphic rocks for the time of zircon growth, and sediments provide information about their igneous sources. In addition, the crystallization history of individual grains in igneous rocks can yield information about source materials or mixing processes involving older rocks, as illustrated in Box 1.1 Fig. 3. In some cases the precision is excellent, yielding error limits of a few mega-years on zircons 2000 to 3000 Ma old.