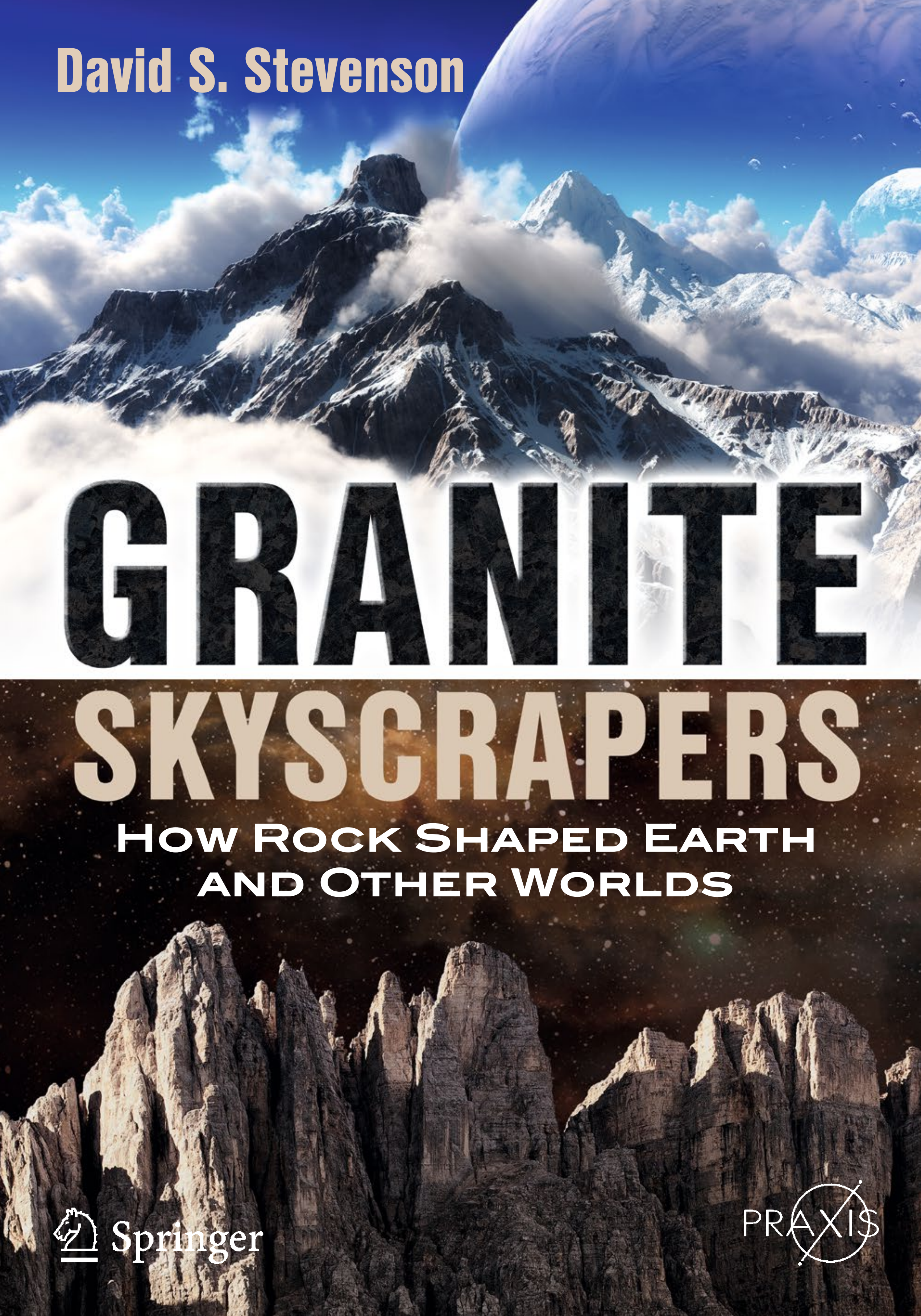

David S. Stevenson

GRANITE SKYSCRAPERS

How Rock Shaped Earth and Other Worlds

Springer

PRAXIS

Granite Skyscrapers

How Rock Shaped Earth and Other Worlds

David S. Stevenson

Granite Skyscrapers

How Rock Shaped Earth and Other Worlds

Springer

Published in association with
Praxis Publishing
Chichester, UK

PRAXIS

David S. Stevenson
Nottinghamshire, United Kingdom

SPRINGER PRAXIS BOOKS IN POPULAR ASTRONOMY

Springer Praxis Books
ISBN 978-3-319-91502-9 ISBN 978-3-319-91503-6 (eBook)
https://doi.org/10.1007/978-3-319-91503-6

Library of Congress Control Number: 2018944146

© Springer International Publishing AG, part of Springer Nature 2018
This work is subject to copyright. All rights are reserved by the Publisher, whether the whole or part of the material is concerned, specifically the rights of translation, reprinting, reuse of illustrations, recitation, broadcasting, reproduction on microfilms or in any other physical way, and transmission or information storage and retrieval, electronic adaptation, computer software, or by similar or dissimilar methodology now known or hereafter developed.
The use of general descriptive names, registered names, trademarks, service marks, etc. in this publication does not imply, even in the absence of a specific statement, that such names are exempt from the relevant protective laws and regulations and therefore free for general use.
The publisher, the authors, and the editors are safe to assume that the advice and information in this book are believed to be true and accurate at the date of publication. Neither the publisher nor the authors or the editors give a warranty, express or implied, with respect to the material contained herein or for any errors or omissions that may have been made. The publisher remains neutral with regard to jurisdictional claims in published maps and institutional affiliations.

This Springer imprint is published by the registered company Springer Nature Switzerland AG
The registered company address is: Gewerbestrasse 11, 6330 Cham, Switzerland

To my wonderful family, both close and far away, for without whom none of these endeavors would be possible. I am a very lucky man.

Preface

If one could strip away the Earth's oceans, it would reveal a surprising dichotomy. While the ocean basins form the deepest depressions on the planet's surface, the continents stand proud at some 3–4 kilometers higher. Rising sheer from the deep abyss below them, these skyscrapers effectively form unique structures in the Solar System. While both Mars and the Moon also have a crustal dichotomy, the origins of the formations found on our planet are distinct. Venus, too, displays some interesting divisions in surface topology, but nowhere is it as extreme as what we see on Earth.

Why is this so?

The answer lies with a most unusual rock: granite. We see granite all around us, but its terrestrial abundance is misleading: the silicate rock is rare elsewhere in the universe. While all solid planets, asteroids and satellites contain large amounts of granite's dark and distant cousin basalt, granite is conspicuously absent. That is not to say that there exists no granite off the Earth's surface, but merely that if it is present elsewhere, it is so uncommon as to be nigh-on invisible.

The mineralogy of granite is fairly simple. It is mainly an assemblage of three silicate minerals: quartz, feldspar and amphibole, with an ad hoc mixture of others. Despite its inherent simplicity, granite is utterly and complexly entwined with our existence. Adorning the walls, floors, and surfaces of some of our grandest buildings and forming many of the headstones that mark our demise, granite is the bedrock that made complex, intelligent life possible on our world.

So, why is granite so uncommon in the first place, and why is so important for intelligent life – or, if you prefer, our particular planet's intelligent life? The answer lies deep in granite's physical properties. Through these, we can link our existence to the presence of water and an oxidizing atmosphere that has just enough of everything to keep our world safe and secure enough for intelligence to prosper. For this reason, apart from its already well-known aesthetic qualities, granite deserves a book all to its own.

Central to granite's cultural and physical importance is its resilience to change. While all rocks eventually succumb to the rigors of nature, granite's intrinsic strength guarantees it a place in our culture and our architecture. Without this, Mount Rushmore would already be a weathered lump, while Yosemite would have been reduced by glaciation to low-lying rubble. Where would Scotland be without its 430-million-year-old Highlands, or Cornwall without its bleak torrs or Scilly Isles? None of these would exist without this gritty rock at their core.

As I was completing this book, I came across a rather unusual Christmas gift set. It consisted of two small bottles of Scotch whisky, accompanied by two mottled cubes to keep the liquid cool. Naturally curious, and Scottish, I bought one of these packs. It turned out that the cubes were granite – the idea being that a true Scot will not drink whisky diluted with melting ice. What better alternative than small, refrigerated granite cubes? (There is irony in that the granite came from China; that Scotland has more than its fair share of the crystalline rock seemed not to impact the manufacturer.)

Granite is so entwined with our psychology that were it to suddenly vanish, our world would be dramatically impoverished and submerged beneath dark oceans. No other rock has such an impact on the habitability of our planet. This book, then, is a celebration of one of Earth's most pervasive but unusual rocks: it links granite's place in our psychology with its role in making our planet a suitable place to live. Welcome to our granite skyscrapers.

Nottinghamshire, UK David S. Stevenson

Acknowledgements

I'd like to thank Eric Wolf, Ravi K. Kopparapu and Jacob Haqq-Misra for helpful discussions regarding the atmospheric models that are discussed in chapters 6 and 8. Eric in particular was very supportive of my application of their models to habitability.

Contents

Preface ..... **vi**

1 Our Planet's Torrid Heart **1**
- Introduction 1
- From Dust to Planet and Back Again 2
- The Early Crust 8
- Igneous Rocks 9
  - Basalts and Their Kin 10
  - The Intermediate Rocks 14
- A Tale of Three Metals 19
- Carbon and Other Planets 26
- Conclusions 29
- References 30

2 The Formation of Granites & Plate Tectonics **32**
- Introduction 32
- A Diverse Family of Related Rocks 32
- The Formation of Granite – General Processes 34
- Geological Reductionism 38
- The Geographical Locations of Granites and Granodiorites 41
- The Origin of Continents 42
- The Real Game of Thrones – The Rise of the Caledonian Mountains 48
- Scotland's Munros 63
- The Ups and Downs of Mountain Building 64
  - King Arthur and Cornwall's Rich Mineral Heritage 66
- Granite as an Abode for Life 73
- Granite, Bacteria and a Trip Abroad 77
- Headstones and the "Edstone" 77
- Conclusions 78
- References 79

3 The Evolution of Modern Continents 83
Introduction 83
How Subduction Built our Modern World 84
The Growth of North America – A 4-Billion-Year-long Story 85
The Plume-Shield Connection 86
  The Proterozoic 89
  The Grenville Orogeny 97
Pangaea 103
The Laramides 104
Africa 111
Asia 116
Indonesia – An Illustrated Continent Factory 119
Drips, Diapers and Slabs: Backdoor Restructuring of Continental Crust 123
  The Izu-Bonin Arc 123
  The Andean Puna 125
  The Sierra Nevada, California 127
  Hoggar – Central Sahara 129
Plumes Then and Now 132
Roots, Plumes and Diamonds 136
  Diamonds Trace the Thickness of Continental Crust 136
  The Chemical Formation of Diamonds 137
  Emplacement of Diamonds – Kimberlites 139
  Other Diamond-Bearing Rocks 141
  Diamonds as a Tracer of Plate Tectonics 143
Oldoinyo Lengai – Not Diamond, But Soap Suds 144
Conclusions 145
References 147

4 Plate Tectonics, Planetary Magnetism and Life 155
Introduction 155
Protection by an Invisible Cloak? 155
  The Gas Giants, Jupiter and Saturn 157
  The Twisted Tale of the Ice Giants 157
  The Terrestrial Planets 161
How Might Plate Tectonics Stir the Global Pot? 162
Atmospheric Stripping 168
  Jeans escape 168
  Charge Exchange 169
  Electronic Force Field Acceleration 169
  Hydrodynamic Escape 170
Sequestration 172
Magnetism and Mars 174
Sunburn – The loss of Atmospheres Irrespective of Magnetism 179
A Young, Magnetic Moon? 185

Special Circumstances – Tidally Locked Planets 187
Conclusions 190
References 190

5 Jotunheim – In the Realm of Giants 193
Introduction 193
The Birth of the Himalayas 193
The Rise of Tibet 200
Volcanism in Tibet – A Clue to its Ongoing Evolution 205
How Tibet Kept Its Edge 208
How Man is Making Mountains Out of Molehills 209
Did a Dam Cause the Death of 80,000 People? 210
Antarctica's Gamburtsev Mountains 212
The Alpine Front and the Cradle of Western Culture 214
  Delamination and the Central Italian Volcanoes 219
  Vesuvius and Its Surrounding Region 221
  Subduction and the Southern Italian Volcanoes 226
  The Balkans and Eastern Europe 228
  The Ever-Youthful Sierra Nevadas of America and Spain 229
Decline and Fall – How Mountains Come Apart 232
Erosion Happens Mostly in the Plain 234
Mountain, Weather and Climate 235
  Gap Winds 237
  Snow-Eaters 239
  Flying Saucers and Venusian Gravity Waves 240
Rossby Waves 241
The Asian Monsoon 243
Conclusions 245
References 246

6 Mountains, Atmosphere and Long-Term Habitability 251
Introduction 251
Volcanism and Our Early Atmosphere 251
The Real Gaia – The Carbonate-Silicate Cycle in a World Without Granite 254
A Spoonful of Sugar: Granitoids and Precipitation 261
The Dry Monsoons and Storms of Mars 261
Tidal Locking and the Carbonate-silicate Cycle 266
Superrotation 267
Super-Earths and the End of Planetary Gaia 269
Super-Earths, Plate Tectonics and Oceans 279
Conclusions 282
References 283

7 Our Island Earth: Granite Here, Granite Everywhere? 287
Introduction 287
The Moon 287
  Granite on the Moon 292

Io 295
  The Nature of Io's Mountains and Lithosphere 295
  Volcanism on Io 297
  How Hot are Io's Lavas? 298
Mars 299
  Martian Volcanism 303
  Tharsis Montes and Olympus Mons 304
  Rootless Cones, Tholi and Highland Paterae 307
Ceres and Vesta 313
Venus 313
Venusian Crust, Tectonism and Volcanism 318
  The Nature of Venus's Crust 318
  Venusian Volcanism and Tectonism 323
Conclusions 331
References 332

8 Exoplanets, Granitoids and Evolutionary Potential 338
Introduction 338
Plate Tectonics, True Polar Wander and Tidal Locking 338
Thick Lids and Swiss Cheese 343
Stagnant Lid Worlds and Life 348
Aquaplanets, Species Dispersion and Biodiversity 352
A Biodiversity Model for Habitability 353
Niche Filling, Tidal Locking and the Fermi Paradox 358
Conclusions 359
References 362

Glossary 365

Index 371

1

Our Planet's Torrid Heart

Introduction

We have a peculiar and very misguided view of our planet as a solid lump of rock. This misconception is based on our everyday experience of standing on something that seems eternal and largely unchanging. If it were not for periodic earthquakes or occasional, somewhat inconveniencing eruptions, we might forget entirely that in truth, we live on a thin skin of cool rock floating on a torrid sea of malleable material.

If one could bring the planet's interior to atmospheric pressure, most of the planet (over 99% by mass) would be molten. The inner core is solid and the mantle a kind of silly-putty-like solid, aberrations that are a result of the enormous pressures contained within. Below a depth of 100 kilometers, temperatures exceed the melting points of all terrestrial rocks and metals. The inner core is so hot that if it were to suddenly be exposed to the vacuum of space, its iron-nickel mass would boil away. At over 6,000 °C, the inner core is hotter than the surface of the Sun. Under these conditions, most of the surface minerals we are familiar with at the surface are mulched into unfamiliar brews of ions – metal and nonmetals that have lost or gained electrons from one another. The rocks we know condense out of this stew of mineral ions as conditions become less hostile closer to the exterior. Fig. 1.1 illustrates some of the more common igneous rocks that make up the bulk of our crust.

Take the mantle – the single most massive component of our planet – a multi-layered, complex, evolving mass of silicates. Its layers primarily consist of olivine crystals, pyroxene, garnet and oxides of iron and magnesium near the top. Below 400 kilometers, these evolve through the iron and magnesium silicates spinel and another dense silicate, wadsleyite. Still further down, spinel and wadsleyite morph into denser and more obscure silicates and metal oxides such as calcium silicate, perovskite or ferropericlase at depths in excess of 650 km. While the bulk

D. S. Stevenson, *Granite Skyscrapers*, Springer Praxis Books,
https://doi.org/10.1007/978-3-319-91503-6_1

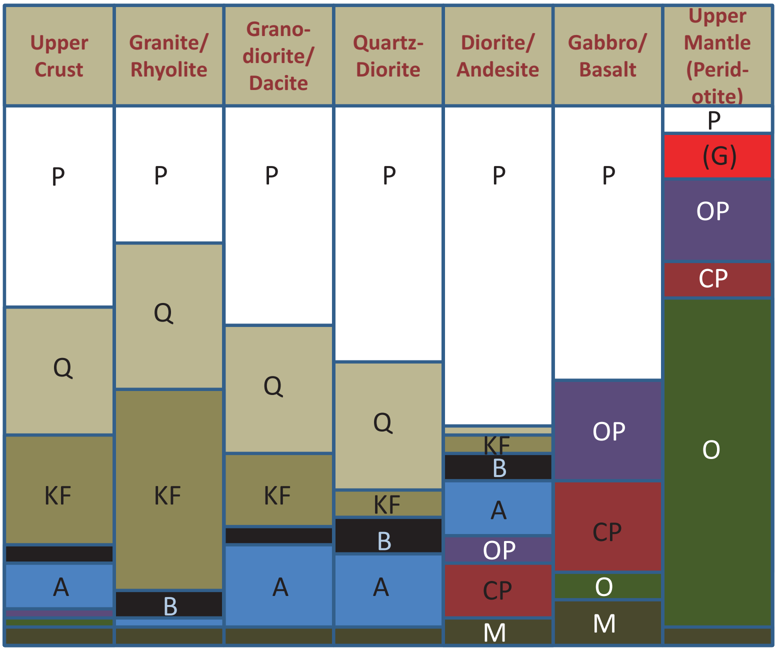

Fig. 1.1 The composition of common rocks compared with the upper crust and upper mantle. The composition of the upper mantle is "averaged." The uppermost mantle is peridotite, which contains up to 10% plagioclase. However, below about 70 kilometers, garnet increasingly replaces plagioclase and clinopyroxene at 1.5 GPa (1.5 billion Pascals of pressure). P – Plagioclase; Q – Quartz; KF – Alkali Feldspar; A – Amphibole; B – Micas; OP – Orthopyroxene; CP – Clinopyroxene; O – Olivine; M – Magnetitite, ilmenite and apatite; G – Garnet.

chemistry remains similar, the change in pressure and temperature forces minerals to realign with one another, with ions forming new allegiances depending on the structures that the intense pressures permit (Fig. 1.1).

From Dust to Planet and Back Again

How did our world and all its chemical glory evolve from the gaseous firmament of the galaxy? The beginnings of an answer came over 300 years ago with the work of Emanuel Kant and Pierre Simon Laplace. Both imagined that the Sun and its retinue of worlds were formed from a spinning cloud of gas and dust. Over time, gravity brought the material together to form our planet and its siblings. Fast forward three hundred years, and their vision has been beautifully demonstrated in a multitude of Hubble images. More recently, ALMA (Atacama Large Millimeter Array) has spied youthful worlds coalescing around nearby newly born stars.

Although many of the details remain to be worked out, we are now certain that all planetary systems form in a similar manner. First, gravity brings the galaxy's gas and dust into large agglomerations called giant molecular clouds. Within these multi-parsec-wide shrouds, gravity works with interstellar magnetic fields to break the cloud into a number of denser cores. Temperatures are initially only a handful of Kelvin above absolute zero. However, as gravity begins the process of collapse, frictional forces gradually raise the temperatures to several thousand Kelvin. For the most massive cores, the full collapse from cloud to star may take less than 50,000 years. For cores with masses similar to the Sun, the process takes several million to a few tens of millions of years. The most diminutive cores may take a few billion years to collapse, completing their in-fall long after the parental giant molecular cloud has dispersed around it. As the core shrinks inwards, the material conserves a property known as angular momentum – the momentum of spinning matter – by rotating faster. Over time, process organizes matter into a spinning accretion disc around the protostar.

In the case of the Sun, 10 million years separated the initial collapse from the ignition of the first nuclear fuel. The final stages of collapse are confounded by interstellar magnetism. As the core collapses, magnetic field lines initially offer little resistance. Dust grains and molecules of hydrogen and other gases simply slip effortlessly between the lines, forming an increasingly torrid lump known as a protostar. However, once the object's temperature exceeds a few thousand Kelvin, gases begin to ionize – losing first one, then several electrons to the increasingly intense radiation field around them. At this point the gases form a plasma, which conducts electromagnetic fields with ease. The ionized gas is then driven along field lines rather than across them. Field lines weave through the disc of material surrounding the star and through the protostar. While the interstellar field is too weak to resist the force of gravity completely, it exerts increasing amounts of pressure. This magnetizes the material and begins to drive some of it – perhaps one hundredth to one tenth of the protostar's mass –outward along its magnetic axis.

The protostar is now a fast-spinning orb of hydrogen, helium and a smattering of other elements. In low mass stars, like the Sun or red dwarfs such as Proxima Centauri, the star's outer layers bristle with convection currents that thread the trapped interstellar field through them. Twisted further within the rapidly rotating star, field lines periodically twist until they snap. Such events unleash a barrage of energetic radiation into the disc surrounding the star. This helps catapult material outwards into the outer disc, carrying both energy and redistributing momentum from the struggling protostar. These eruptions are of critical importance, because they explain peculiarities in the materials out of which the planets condensed. By the time the Sun ignited its fuel, radiation from its increasingly hot surface was blowing much of the hydrogen and helium away from the star. Given that radiation intensity decreases exponentially with distance from a star or any other hot object, the material closest to the Sun should consist of compounds that can avoid evaporation at high temperature: these are referred to as *refractory*.

Further out, materials with correspondingly lower boiling points should persist. These materials are referred to as *volatiles.* Such fractionation should leave the planetary building blocks with a compositional gradient that reflects this temperature gradient. Alan Boss (Carnegie Institute) carried out simulations of the early protoplanetary disc and demonstrated that early in the life of the disc, mixing processes are relatively inefficient. This allows localized regions of differing composition to persist for thousands of years once they are established. Interestingly, over this timescale, dust will coagulate into small, pebble-sized "lumps," which could explain some differences in the distribution of elements in certain meteorites, known as chondrites.

Chondrites are meteorites that show odd aggregations of materials that have clearly been molten (chondrules). Aluminum and calcium oxides with high boiling points (refractory elements and compounds) can be found as "blobs" within a matrix of more volatile carbon-rich compounds. They could only have become combined in this manner if the "blobs" had been molten, sprayed and mixed into the more volatile matrix. Taishi Nakamoto and colleagues (University of Tsukuba) demonstrated that with their accompanying shockwaves, stellar eruptions known as X-flares seem the only viable mechanism that could cause such mixing. In these eruptions, young Sun-like stars (known as T Tauri stars) undergo magnetic storms caused by their rapidly rotating surfaces stretching and then breaking their strong magnetic fields. In a scaled-up version of the solar cycle, such periodic eruptions generate powerful pulses of x-rays and highly charged particles that surge outward across the surfaces of the disc. These pulses of very hot gas generate shockwaves within the cooler disc that melt and bring very refractory metal oxides into intimate contact with the more volatile materials. A combination of mixing and settling then gives rise to the well-mixed but still somewhat patchy mesh of materials from which the planets formed.

As a result of all of this mixing, the inner terrestrial planets were left with broadly similar compositions. Early geochemists referred to this as "chondritic" – a term that is gradually being eroded as observations improve. The assumption is that "chondritic" reflects the composition of the early proto-Solar nebula – hence, "chondritic" really means "solar." Within the literature, the term "chondritic" persists, referring to a group of stony asteroids that populate a sizable fraction of the inner asteroid belt. These rocky asteroids have compositions very similar to the terrestrial planets and are therefore assumed to be relics of planetary formation. The compositions of the 27 known asteroids have been accurately sampled, and while there are variations in composition, these are relatively minor. Moreover, the ratio of so-called RLE (refractory lithophile elements) is consistent between these samples – and between the rocks on the Earth and Moon. RLE refers to 28 elements that have high boiling points and are not miscible with (do not dissolve in) iron or sulfide. These refractory elements include magnesium, calcium and aluminum, the radiogenic elements uranium and thorium and the rare earth elements (REEs), which include samarium and neodymium. The consistency of the ratio is presumed to reflect the origin of the Earth and Moon's materials within smaller bodies that are themselves preserved as the current population of stony asteroids.

Yet, the picture is not quite what it appears. The REEs show variations in the proportion of certain elements such as neodymium and samarium. Samarium-146 decays to form neodymium-142 with a half-life somewhere between 68 and 103 million years, effectively meaning that the Earth's entire original inventory of this isotope of samarium is gone. Meanwhile, samarium-147 decays into neodymium-143 with a vastly longer half-life of 106 billion years – meaning that very little has decayed since the Earth accreted from its source materials. Neodymium-143 is stable. Examination of all three isotopes tells you how much neodymium and samarium were present in the Earth when it formed. The problem is that when you plug the numbers in, the ratio of ^{142}Nd to ^{143}Nd (and hence samarium to neodymium) in terrestrial rocks is about 6% above the average chondritic value. This doesn't sound like much, but it is enough of a difference to conclude that something fairly dramatic has happened with the Earth.

Some geochemists have suggested that there may be a deep reservoir of these depleted elements that separated from the bulk of the mantle, a theory contradicted by evidence provided by mantle plumes. These ascend from the hot base of the mantle and would bring these REEs with them: yet, this isn't seen. Other possibilities exist. For one, while unlikely, the Earth may have condensed in part from non-chondritic material. Perhaps the Earth happened to form in a corner of the solar neighborhood where the composition of materials varied from the average. However, given the violent mixing of planetesimals that characterizes planet formation, it seems unlikely that the Earth would have managed to form such pristine material.

Alternatively, during the violent process of planet formation, some of the early terrestrial crust may have been blasted off into space by early collisions. This would deplete the Earth of REEs, which may have preferentially concentrated in the early crust relative to the mantle through processes we will discuss in Chapter 2. Or, we may underestimate how well such elements dissolve in molten iron under the brutal conditions found inside the planet, leaving much more of these elements in the core than we otherwise believe. Complications of this kind appear to explain some peculiarities in the abundance of gold, which we will discuss shortly.

Further hitches to the convenient pattern are suggested in the chondritic model. While a compositional gradient exists from dense Mercury to more volatile Mars, there is a more subtle set of problems embedded in the structure of the solar system's rocky bodies. Mars is unduly small, and the asteroids do not follow the smooth trend in composition one might expect as we travel away from the Sun. To address these and other issues, the so-called Grand-Tack hypothesis has been put forward. Here, the orbits of Jupiter and Saturn have varied considerably, throwing materials around the infant solar system. Calculations by Kevin Walsh (Nice Group) indicate that 150,000 years after Jupiter's inception –long before the Earth or other rocky, terrestrial planets had formed – the young Jupiter migrated inwards.

It was driven by frictional and gravitational forces exerted by and on Jupiter by the remaining disc of debris that lay inwards of its orbit. Over a very short period, Jupiter spiraled in until it lay approximately where Mars lies today. As Jupiter migrated to around 1.5 AU, it spat out a large amount of refractory material that would have contributed to the formation of the Earth, Mars and the other rocky worlds. Jupiter was only stopped from spiraling all the way into the Sun through a stealthy pursuit by embryonic Saturn.

As Jupiter moved inwards, Saturn swallowed much of the material left in its wake, causing it to also spiral inwards. When Saturn drew closer to Jupiter, it created an orbital resonance with its larger sibling. As Saturn made three orbits of the Sun, Jupiter made two. In this arrangement, Jupiter stole momentum from Saturn, and the mighty Jove then began to spiral outwards, kicking Saturn ahead of it. Although these two worlds were to eventually settle into orbits similar to those of present day, the carnage they left behind was to catastrophically alter the fates of the inner planets. Starved of much of their rocky material, all of them would end up with less mass than would otherwise have been the case. Mars came off the worst. Left to coalesce from the scarce remains left by Jupiter, this world was unable to collect more than one tenth the mass of the Earth. The end result was a small dry and vulnerable world that would never compete with the Earth or Venus. Mercury, meanwhile, appears to have suffered a catastrophic collision that removed much of its outer layers.

Things weren't all bad for the Earth, Venus and other terrestrial worlds. While Jupiter and Saturn removed much of the birthing material for the inner worlds, they also delivered a cargo of rocky debris from further away. These would include carbon-rich compounds, water and other materials that would have otherwise been seriously depleted by heat from the tempestuous infant Sun. By the time the Earth was largely formed, some 4.55 billion years ago, it had a sizable mass of volatile materials atop a dense core of iron and nickel that would later offer some protection from our young Sun. Our fledgling Earth would be completely unrecognizable, but by then it had most of its ingredients in place to form the rocks we recognize today. Save for two "minor" encounters, the modern Earth would emerge from this rocky chaos and become the world we know. Those two "minor" encounters? In the first, a Mars-sized protoplanet smashed into the planet around 60 million years after it formed. This effectively reset the clock on the planet and bequeathed it with a tilt, a rapid spin and a disproportionately massive iron-nickel core, all of which would be critical for the evolution of life several hundred million years later.

The more controversial second encounter would play out over the next half a billion years and once again involve shenanigans in the outer solar system. We will leave the details of this until later: it is enough to note that these events may have had a critical influence on humanity's love for granite.

Leaving our planet's origins behind for now, we can begin to look at what its formation left us with. The infant Earth had an abundance of energy reserves. Heat was liberated from various sources: the gravitational potential energy of impacting and coalescing rocks, the compression of rock under gravity as crystals realigned their structures, the formation of the massive iron-nickel core and radioactive decay of various elements from which the planet formed. Of these, radioactive decay and core formation easily released enough energy to melt the outer portion of the planet and the entire metallic core. While the lower mantle would likely have remained solid, the lower density upper mantle would have been heated well beyond its melting point, forming a deep magma ocean.

The Moon shows the leftovers of this process. Whereas the more massive and turbulent Earth has recycled all of the material produced in this phase of its evolution, the Moon's crust still bears witness to this phase in out planet's evolution. The thick, light-grey lunar highland crust consists of a single rock type, anorthosite (see also Chapter 7). This consists primarily of one mineral, plagioclase feldspar. While this is very common on Earth, its dominance on the Moon's crust suggests that the Moon was molten through a considerable period early in its life. Why? Plagioclase feldspar forms abundantly where the material is hot and fairly dry. Under such dry conditions, plagioclase floats upwards, leaving a denser assemblage of more iron and magnesium-rich minerals behind. The entire lunar highland crust is thus a magma ocean fossil, one which – at least in the lunar case – was largely devoid of water. The Earth's outer mantle must also have been heated this strongly, but there is no evidence of an anorthosite crust on Earth. There are two possible reasons for this. In the first, the mantle of our planet retained a substantial fraction of water. Under these conditions, plagioclase sinks into the magma and won't form a crust. Alternatively, an anorthosite crust did form, but tectonic processes on Earth removed it or buried it under large volumes of other rocks until any anorthosite crust was remelted and recycled into the mantle. There is no evidence that such a crust remains buried today. Anorthosite does occur in some terrestrial locations, but it is a particularly rare rock, and outcrops appear to represent subterranean intrusions that have now been uncovered by erosion (Chapter 3). Most show layering, with other common minerals present beneath layers of less dense anorthosite.

There is good evidence that some form of surface tectonics – if not modern day plate tectonics – was operational by 4.2 billion years ago. The evidence, while sparse, is fairly conclusive. Small, hardy crystals of a mineral called zircon (zirconium oxide) have been found within other younger rocks. Zircons are formed in the outflows of cool, giant stars as well as in the chemical processes that form granites. Zircons are particularly useful, as once formed, such minerals are hard to destroy, and their crystal structure can retain chemical clues to their origin.

4.404-billion-year-old zircons were found in the Jack Hills of Australia. The rocks comprising the bulk of these hills are themselves an ancient 3.6-billion-year-old metamorphic belt. Upon closer examination, they were found to have relics of even older crust. Zircons in particular provide evidence of two main features of a planet's surface. Firstly, they demonstrate that the planet created granite, as zircons contain an abundant secondary mineral alongside the silicates that make up the bulk of the rock. Granites generally need water to form, implying the second main feature – that the Earth had oceans (or at least running water) as early as 4.4 billion years ago and was therefore temperate enough to form and host life.

The Early Crust

Before we examine the rich menagerie of igneous rocks that comprise the modern crust of the Earth, we'll make a quick pit-stop at the first long-lived crust of the planet (Fig. 1.2). From here, we can move on to examine the rocks that subsequently formed as a result of the various processes that turn mantle into crust.

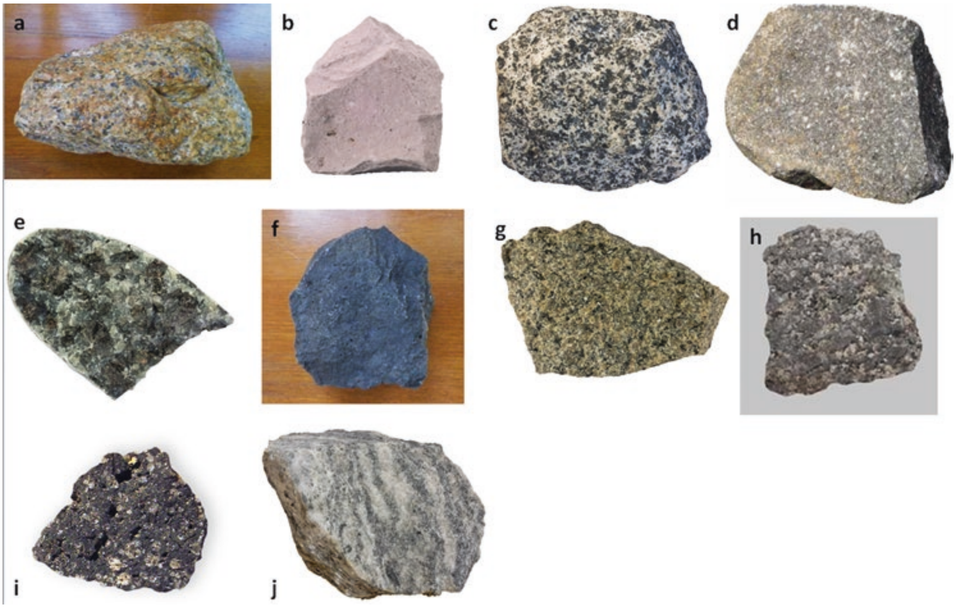

Fig. 1.2 Common igneous rocks. Highland Granite (**a**), Liparian Rhyolite (**b**), Diorite (**c**), West Highland Andesite (**d**), Skye Gabbro (**e**), Elie Basalt (**f**). Less common rocks, such as syenite (**g**), nephelinite (**h**), Southern Highland phonolite (**i**) and East African carbonatite (**j**) are shown below. Images (**h**), and (**i**) courtesy of Wikipedia commons. Remaining images author.

By the time the Earth cooled sufficiently to form oceans, anorthosite, if it ever existed as terrestrial crust, would have given way to a dark, dense basalt or an even denser rock known as komatiite (Table 1.1). The problem with anorthosite is that even when the planet was hot and had a magma ocean, this turbulent sea of rock was likely overlain by a very dense, torrid atmosphere of carbon dioxide and water vapor. Most models suggest the Earth had an atmosphere comparable in mass to that seen around Venus today. At a pressure of 100 atmospheres, such a hot, humid environment would likely allow oceans to form, despite the temperature being high enough to boil them dry. Consequently, with an abundance of hot water, most magmas reaching the surface would have been grossly contaminated with water, making plagioclase sink and mix with other minerals in a hot sea of rock. Although the nature of the tectonic processes that sculpted this early Earth are unclear, the information gathered from present Earth rocks is far more direct.

Igneous Rocks

In this section, we delve into the wealth of igneous rocks that are found on the Earth today. We begin our journey with the planet's most ubiquitous volcanic rocks – the basalts and their ultramafic parents – those that contain the least silicon dioxide (silica). Then, we move onto rocks that are progressively richer in silica. Without suggesting any evolutionary relationship between these, you will see how our planet has produced an interesting wealth of rocks from a rather monotonous parent.

In the years following the First World War, geologist N. L. Bowen carried out a series of experiments to investigate how the wealth of igneous rocks came about. There are far more rocks than could be reasonably covered here, but the principles are fairly straightforward. The densest rocks are mostly found in the mantle and are low in silica and rich in iron and magnesium (ferromagnesian-rich). These rocks, along with the somewhat less dense, feldspar-rich basalts of the ocean floor are lumped into a silicate collective known as mafic rocks. Mafic rocks comprise the most silica (silicon dioxide)-poor igneous rocks. "Mafic" is an abbreviation of magnesium and ferrous, indicating that these rocks are rich in these elements and compounds of them. Once again, anorthosite is the exception, but as we've already seen, in most cases this odd rock only forms in very limited circumstances not found on Earth or the other terrestrial planets. The mafic rocks comprise the rocks of the upper mantle, the oceanic crust and the lower crust of the continents.

In the upper mantle rocks, olivine is the dominant mineral, with smaller quantities of pyroxene, making these rocks mafic as well. (Sometimes these are referred to as ultramafic because they are so rich in iron and magnesium).

As we descend 100 kilometers into the mantle, the aluminum-rich silicate plagioclase feldspar dominant in the crust gives way to the denser silicate garnet. The change is due to increasing pressure, which destabilizes the crystal structure of the feldspar and rearranges its elements into the denser structure of garnet. This has one very important consequence for our planet and likely all other sizable terrestrial worlds: with increasing depth, basalt morphs into another rock called eclogite. The transition makes the resulting rock denser than the surrounding mantle, so it sinks under its own weight. We'll look at this in more detail in Chapters 2, 3 and 5, as it has profound implications for the evolution of the planet as a whole.

Basalts and Their Kin

Table 1.1 Common iron-magnesium-rich silicate rocks. Peridotites are common in the upper mantle, while tholeiites make up most of the ocean floor. Rarer alkali basalts are erupted in places like the Azores, while anorthosite is largely found on the Moon, making up its Highlands. Kimberlite is very rare but of extreme economic importance, as it often contains diamonds.

Basic (Mafic) and *Ultrabasic* Igneous Rock	**Principle chemical compounds**	**Principle minerals**
Peridotite/Komatiite	$(Mg, Fe)_2SiO_4$ $(Na, Ca)(Mg, Fe, Al)(Al,Si_2)O_6$ $(Na,Ca)(Si,Al)_4O_8$ TiO_2	Olivine; Pyroxene Less than 5% plagioclase feldspar Ilmenite
Kimberlite	$(Mg, Fe)_2SiO_4$ $[M]CO_3$ $Mg_3Al_2(SiO_4)_3$ $(Na, Ca)(Mg, Fe, Al)(Al,Si_2)O_6$ C	Olivine Various carbonates Garnet (pyrope) Pyroxene Diamond (minor)
Gabbro/Basalt (Tholeiite)	$(Na,Ca)(Si,Al)_4O_8$ $(Na, Ca)(Mg, Fe, Al)(Al,Si_2)O_6$ $(Mg, Fe)_2SiO_4$ SiO_2	Plagioclase feldspar; Pyroxene A few percent olivine (silica saturated Olivine Tholeiites) or (less than 5%) quartz (quartz oversaturated tholeiites)
Alkali Basalt	$(Na,Ca)(Si,Al)_4O_8$ $(Na, Ca)(Mg, Fe, Al)(Al,Si_2)O_6$ $(Mg, Fe)_2SiO_4$ $Na_3KAl_4Si_4O_{16}$	Plagioclase Feldspar Pyroxene Limited olivine Less than 5% feldspathoids
Anorthosite	$(Na,Ca)(Si,Al)_4O_{8,}$ $(Na, Ca)(Mg, Fe, Al)(Al,Si_2)O_6$	>90% Plagioclase feldspar (with very small amounts of other minerals – mostly pyroxene, olivine and ilmenite)

Magma, or molten rock, is produced by the melting of parent rocks in the mantle. This occurs at very narrow ranges of temperature and pressure. On Earth, nowhere near the crust or upper mantle is it hot enough to melt all of the available rock.

Instead, a reduction in pressure or rise in temperature causes rocks to partially melt. During partial melting, some but not all of the minerals within the rocks are able to melt. This process can be enhanced by the presence of a suitable solvent, such as water or carbon dioxide; of the two, water is by far the most abundant on Earth. Typically, the magmas that are produced this way are less dense than the remaining rock, so they rise towards the surface. Thus, partial melting fractionates the rock, separating those minerals with lower melting points from those that are more refractory. Depending on the extent of the melting process, the resulting magma will either closely resemble or more likely differ from its parent rock.

Most present-day magma is basaltic and is produced by partial melting within the upper mantle, which is primarily composed of a dense rock known as peridotite. This dark-greenish rock is very low in silicon dioxide (less than 45%) and rich in magnesium oxide (Table 1.1). Peridotite constitutes most of the upper mantle of the four terrestrial planets, Jupiter's satellite Io, as well as many asteroids. When solidified, the elements arrange into two principle compounds: olivine and pyroxene, of which there are different sub-types. Basalt and gabbro are the most common rocks in oceanic crust and the surfaces of the other terrestrial planets. Dense and dark in color, these are less iron and magnesium rich, and they have higher proportions of silicon dioxide (45–52%). Basalts are erupted either as blocky flows known as "aa" or more fluid, ropey lava known as "pahoehoe". Titanium oxide (as ilmenite) and iron oxide (as magnetite) are often common accessory minerals in basalts and peridotites. The basalts are themselves sub-divided according to how much silica they contain. Where partial melting of mantle peridotite occurs at very shallow depths, it produces more silica-rich quartz tholeiites. Olivine tholeiites are produced by partial melting at higher pressures and/or with smaller degrees of partial melting. These olivine tholeiites dominate the rocks erupted at constructive plate margins and in ocean islands but are also common in ocean island arcs related to subduction.

The alkali basalts are a rarer group of rocks that are poorer in silica and contain additional minerals known as a feldspathoids. They never contain quartz and are less oxidized than normal tholeiites – this means they contain less ferric iron (Fe_2O_3) than ferrous iron (FeO). Their origin appears to trace back to plumes that rise from the base of the mantle and carry entrained material from subducted oceanic crust (Chapters 2 and 4). These alkali basalts are produced at greater depths or with lower degrees of partial melting. Common in some hotspot locations, they are also regularly found in areas of continental rifting.

Basalt is made by very modest amounts of partial melting of the source rock peridotite (Table 1.1 and Fig. 1.1). Typically 3–5% of the peridotite melts in this way. The remaining mantle material is depleted in plagioclase feldspar and to a lesser extent the more magnesium-rich and denser mineral, pyroxene. Most of the leftover material is a dense slush of olivine crystals. These don't make it past the lowermost crust, forming a rather attractive rock known as dunite. At slightly lower temperatures, pyroxene-rich slushes form a darker rock called pyroxenite. Meanwhile, lighter basalt rises upwards, carrying some olivine with it but mostly consisting of plagioclase feldspar and pyroxene. The separation of the denser ferromagnesian minerals like pyroxene and olivine from plagioclase leaves basaltic crust less dense than the mantle underneath. Consequently, the oceanic crust floats on top of the mantle.

The pyroxenes in basalt are of two different kinds. The first, orthopyroxene, is marginally lower in density (3.2 $g.cm^{-3}$) and richer in magnesium than the other form, known as clinopyroxene (3.2–3.6 $g.cm^{-3}$). The latter is richer in calcium and iron. The distinction is important. Orthopyroxenes are the dominant pyroxene in peridotite, the source rock in the upper mantle, while clinopyroxene is dominant in eclogite, the dense rock produced by high-pressure metamorphism of basalt and its water-altered form, amphibolite. Consequently, clinopyroxene is dominant in the lowermost crust under the continents, and its ability to morph into garnet under pressure can make the lowermost crust of the continents denser than the underlying peridotite mantle. While Chapter 2 explores some of the consequences of this behavior, there are some profound implications of this subtle change in chemistry that are discussed at the end of this chapter.

Beneath the 6-km-thick basaltic veneer lies a layer of frozen upper mantle. This lid, with crust on top, is collectively called the lithosphere. Near the ocean ridges where the crust is generated, the lithosphere is relatively thin – perhaps 10 kilometers, all of which is comprised of crust with no underlying layer of frozen mantle. However, as one moves away from the ridges, this layer thickens as the mantle underneath loses heat by conduction to the oceans above. The lithosphere thickens with age, becoming 50–140 km thick once it has matured to 150 million years or so.

Early in the Earth's history, the mantle was substantially hotter at shallower depths because more energy was available from the formation of the planet. At higher temperatures, greater degrees of partial melting become possible. Along with the abundance of basalts, another denser rock, komatiite, probably contributed to a significant portion of the crust. I use the word "significant" advisably, as there is still considerable controversy over exactly how abundant komatiite was. We certainly know that it existed, because it is preserved in blocks of crust known as greenstone belts. These odd assemblages of rock appear to represent slivers of oceanic crust that were trapped and heavily intruded near subduction zones, regions where old oceanic crust was slipping back into the mantle.

While it now seems more likely the komatiite was erupted primarily at the subduction zones, it is likely that some also erupted in regions where hot mantle plumes were rising towards the surface. Indeed, the youngest such terrain erupted in the Cretaceous and is preserved on Gorgona Island in the Caribbean. These are clearly associated with a mantle plume. Here, the hotter mantle can melt more extensively and generate more ferromagnesian-rich lavas. A combination of subduction-like settings and plumes probably led to eruptions of komatiite that were extensive enough to produce sizable areas of oceanic crust. Interestingly, the preponderance of komatiite in these settings implies that while the mantle was hot, it may never have been hot enough to erupt komatiite at the ocean ridges, where the majority of ocean crust is made today. Instead, it took a healthy dose of water or an extra injection of heat to produce the 30–60% partial melting required to generate komatiite.

The ocean crust of the Hadean might then have consisted of two broad rock types. Although the majority might have been basaltic (olivine tholeiitic basalts) – albeit richer in olivine and pyroxene than current basalts – komatiite might have been erupted along the margins of the ocean basins, where enough water was introduced by subduction or processes like it.

Komatiites are effectively extinct igneous rocks on Earth, with the vast majority of komatiites dating to the Archaean. Our upper mantle is about 200 °C too cool to melt enough rock to liberate komatiite, and its density easily exceeds that of the surrounding continental rock. Both of these factors effectively rule out eruptions on present-day continental crust. At present, Io is the only body in the solar system hot enough to produce this dense lava. No other terrestrial body remains hot enough to melt enough of its mantle. Io only achieves this feat because of its constant stretching and kneading under the gravitational pull of Jupiter and the other Galilean satellites (Chapter 7).

Venus may also have hosted komatiites in the recent past (Chapter 7). Winding across its surface are many long rivers of frozen rock. The length – up to hundreds of kilometers – and pattern of these fossil flows suggests the magma was particularly fluid. Komatiite lava is one possibility, as its high eruption temperature (1,560–1,600 °C) would allow it to flow with low viscosity over lengthy distances. Moreover, Venus's torrid surface would delay the cooling and crystallization of the lava, allowing it to flow over longer distances than happens on Earth. In this regard, komatiites have some spectacular properties. Io's komatiites are erupted at ferocious temperatures in excess of 1,500 °C, detected using long-range sensing. Terrestrial komatiites appear to have been similarly hot and extremely fluid when vented. With viscosities two orders of magnitude lower than basalts, these molten rocks would have flowed with high velocities from their vents and travelled several kilometers before solidifying. They may also have been hot enough to melt the rocks onto which they were erupted and physically erode the underlying bedrock. Such channels are preserved in a few Archaean shield rocks, such as at Kambalda in Australia. Komatiites are a fossil of the early Earth – a hotter, more plastic and mobile world than it is now. Although we are unlikely to see such ferromagnesian rocks liberated at volcanic vents today, komatiites provide an interesting and at times unexpected portrait of our planet in its turbulent youth.

Komatiite *may* have formed the relatively recent flows on Venus, but a more likely candidate is an even more unusual volcanic rock known as carbonatite. While the hot Venusian atmosphere will delay the cooling of the planet's interior, it has also dried out much of the crust and likely the upper mantle, too. This makes melting more challenging, so perhaps Venus cannot melt enough of its mantle rocks to form komatiites. Instead, Venus's wealth of carbon dioxide and high surface temperatures could explain the likelier presence of carbonatite (Table 1.3) – an odd, carbon dioxide-rich rock that may form the sinuous channels.

The Intermediate Rocks

Table 1.2 Intermediate and alkaline igneous rocks. While andesite is by far the most common intermediate igneous rock, alkaline igneous rocks are important contributors to volcanic eruptions in Africa's rift valley, Vesuvius and Antarctica's Mount Erebus. Silica-poor carbonatite is almost pure sodium carbonate and calcium carbonate and is the most alkaline volcanic rock. This has virtually no silicon dioxide and is extremely rare on Earth. It may contribute to recent lava flows on torrid Venus (Chapter 7). Andesite may also be present on Mars (Chapter 7).

Intermediate and *Alkaline* Igneous Rock	**Principle chemical compounds**	**Principle minerals (in descending order of importance)**
Andesite/Diorite	$(Na,Ca)(Si,Al)_4O_8SiO_2$ $NaAlSi_3O_{8,}$ $KAlSi_3O_8$ $(Ca_2Mg_5)Si_8O_{22}(OH)_2$ $K(Mg,Fe)_3(AlSi_3)O_{10}(OH,F)_2$ $(Na, Ca)(Mg, Fe, Al)(Al,Si_2)O_6$ SiO_2	Plagioclase Feldspar Alkali feldspar; Amphibole; Biotite mica Pyroxene; with limited quartz
Trachyte/ Syenite	$NaAlSi_3O_8$, $KAlSi_3O_8$ $(Na, Ca)(Mg, Fe, Al)(Al,Si_2)O_6$ $(Na,Ca)(Si,Al)_4O_8SiO_2$ SiO_2	Alkali feldspar Pyroxene; Plagioclase feldspar; with limited quartz
Phonolite/Nepheline Syenite	$(Na,Ca)(Si,Al)_4O_8SiO_2$ $K[AlSi_2O_6]$, $(Na,K)AlSiO_4$ $(Na, Ca)(Mg, Fe, Al)(Al,Si_2)O_6$	Alkali feldspar; Feldspathoids (nepheline), Pyroxene
Nephelinite/Ijolite	$Na_3KAl_4Si_4O_{16}$ $(Na, Ca)(Mg, Fe, Al)(Al,Si_2)O_6$ $(Mg, Fe)_2SiO_4$	Feldspathoids (mostly nepheline) Pyroxene Olivine
Carbonatite	$Na_2Ca(CO_3)_2$ $Na_2Al_2Si_3O_{10}$ $Na_8(Al_6Si_6O_{24})Cl_2$	Nyerereite or other carbonates (>50%); Natrolite; Feldspathoids (e.g. Sodalite)

In this section, we leave the ancient Earth behind and examine more contemporary rocks that are richer in the elements silicon and oxygen. Many of these rocks may still be considered unusual, but we begin to encounter those rocks that constitute the ground upon which we stand. These silica-rich rocks are fundamental to evolution of our planet and life itself.

While the bulk chemistry of the Earth reflects the materials from which it coalesced, the rocks we observe near the surface are more a reflection of the processes that occur within our still-hot planet. In the previous section, we saw how the processes that produce magmas rarely melt the whole rock from where they came. Instead, a process known as partial melting generates these. Another more limited process is known as gravitational fractionation, wherein new magmas form through the gravitational settling of denser minerals. This is largely the source of anorthosite, which is common on the Moon. Meanwhile, a third process known as chemical fractionation allows minerals that have crystallized within magmas to react with the leftover melt. All of these processes contribute to the rocks that ultimately comprise the magmas that erupt on the Earth's surface or which crystallize beneath it.

Table 1.3 Acidic (felsic) igneous rocks. These rocks are the most silicon dioxide (silica)-rich rocks in the Solar System and make up the continents of Earth and some of the highlands of Venus. These are very poor in magnesium and impoverished in iron. Dacite and its intrusive form, granodiorite, contain 63–68% silicon dioxide (silica), while true granite contains more than 68% silica. Obsidian is a silica-rich volcanic glass, which had huge cultural significance on prehistoric Earth.

Acidic (Felsic) Igneous Rock	Principle chemical compounds	Principle minerals
Granite/Rhyolite	SiO_2,	Quartz,
	$NaAlSi_3O_8$,	Alkali feldspar
	$KAl_2(Si_3Al)O_{10}(OH,F)_2$	Muscovite mica,
	$K(Mg,Fe)_3(AlSi_3)O_{10}(OH,F)_2$	Biotite mica
	$(Na,Ca)(Si,Al)_4O_8$	Plagioclase feldspar
	$(Ca_2Mg_5)Si_8O_{22}(OH)_2$	Amphibole
Granodiorite/Dacite	SiO_2,	Quartz;
	$NaAlSi_3O_8$, $KAlSi_3O_8$	Alkali Feldspar;
	$(Na,Ca)(Si,Al)_4O_8$	Plagioclase feldspar;
	$K(Mg,Fe)_3(AlSi_3)O_{10}(OH,F)_2$	Biotite
	$KAl_2(Si_3Al)O_{10}(OH,F)_2$	Muscovite mica;
	$(Ca_2Mg_5)Si_8O_{22}(OH)_2$	Amphibole
	$(Na, Ca)(Mg, Fe, Al)(Al,Si_2)O_6$	Rarely, but also limited pyroxene

As their name suggests, intermediate rocks (Table 1.2) have middling proportions of silica compared to the basalts and their more silica-rich cousins, the granites. By far the most famous of these is andesite, named after the Andes, where they were first identified. Andesites have around 50–55% silica, and when erupted form thicker and more viscous, blocky lava flows compared to basalts. These lavas are only produced in significant quantities at subduction zones, where the ocean crust descends into the mantle.

Andesites were thought to be a true magma in their own right. However, some doubt has been cast upon this idea. Instead, research by Susanne M. Straub (Columbia University) and others suggests that andesites form immediately above the subducting slab through a complex mechanism. As the wet ocean crust subducts, very silica-rich magmas (dacite and rhyolite in composition) emerge and infiltrate cracks and spaces in the overlying hot mantle. Here, mixing and chemical reactions produce a mixture of magmas from magnesium-rich basalts through andesites to dacites. Whereas the more fluid basalts and andesites can rise through the bulk of the overlying continental crust (or thickened oceanic crust), the most viscous magmas (those with rhyolite compositions) get stuck. This allows a mixture of basalts and andesites to erupt at the surface.

Unsurprisingly, andesite is an averaged rock – a mixture of two extremes, rather than a unique one in its own right. Whatever the truth of the matter, andesites are very common, and their high viscosity leads to their association with the kinds of explosive eruptions that generate mass casualties. Andesites grade imperceptibly into increasingly silica-rich rocks known as dacites, and eventually the granites (Table 1.3). Coarse-grained andesites, which crystallize underground, are known as diorites and contain a mixture of feldspar, quartz and pyroxene. As their intermediate nature suggests, olivine is rare or absent, but quartz is also only present in small amounts. As we advance into the dacites (with 63–68% silica) then the granites (with 68–75% silica), any olivine is lost, followed soon after by mafic pyroxene, while quartz becomes increasingly abundant.

Several key minerals emerge that hint at the formation of the intermediate and silica-rich acidic or felsic rocks (felsic being a forced acronym of feldspar and silica-rich). These are amphibole and the micas. Both of these minerals contain water and often chloride or fluoride, the origin of which lies within the oceans. The melting point of these igneous rocks progressively lowers from around 1,200 °C for gabbro and basalts to 700 °C for granites and rhyolites. Water is present in the chemical structures of some of the minerals in these rocks, and it is also abundant as a volatile – gases that are trapped within the magma. In part, this is because molten basalt is much less viscous than granite, so gases can escape more readily than in granite. However, the presence of water also reflects their origin, which usually requires a lot of water. (See Chapter 7 for some exceptions.)

The presence of amphibole is a critical indicator of how granite and andesite form. As chapter 2 will explore, water is critical in the formation of this mineral and the rocks that contain them.

Amphibole is interesting for other reasons. One of these is the conversion of basalt to a metamorphic rock: the formation of eclogite, which Chapter 2 explores, helps drive plate tectonics. Water's presence in this mineral transformation also helps lubricate plate motion (Chapters 2, 7 and 8). Moreover, while amphibole is officially a mineral of intermediate rocks, when basalts erupt under or onto the seafloor, most of the pyroxene is altered to amphibole, and basalt becomes amphibolite. When this rock is subducted into the mantle, amphibolite decomposes and releases its store of water back into the mantle. This helps drive abundant volcanism around the edges of many of our oceans. There is a darker side to amphibolite: the ferromagnesian silicates that dominate this rock are arranged in long chains. One incarnation of amphibole is the once commercially important silicate, asbestos. This infamous mineral is avoided today, but until the 1970s it was used as a thermal and electrical insulator and still exists in buildings constructed up until this time.

In this capacity, asbestos was an extremely effective and cheap material. My parents' garage and other outbuildings had roofs made of this noxious material. Indeed, many public buildings in the UK built from the 1910s to 1960s still contain a potentially lethal morass of this material locked up in their walls. Mining asbestos was one of the most toxic professions one could have had. The fatal cancer mesothelioma, emphysema and other cancers plagued those who made a living from it. Asbestos is no longer added to buildings, yet estimates place the annual US death toll from exposure at around 12,000 per annum, with roughly 2,500 killed per year from mesothelioma in the UK. Although asbestos is lethal on direct exposure, its role (with the other amphiboles) in sustaining plate tectonics is critical to the long-term survival of life on Earth (Chapters 6 and 8).

The solar system's most abundant felsic rocks, the granodiorites and their fine-grained equivalents the dacites, often contain the ferromagnesian mineral pyroxene alongside amphibole and quartz. In general, these are pale-colored rocks dominated by iron and magnesium-poor minerals. As one moves into the most silica-rich rocks, the micas evolve from more iron and magnesium-rich forms to ones poorer in these elements; biotite is increasingly replaced by muscovite as the rock's iron and magnesium content declines. Muscovite mica has a lovely origin to its name. Linked to Russia's capital, mica was frequently used as a substitute for glass from Medieval times until fairly recently. This was particularly true in Siberia, where Mica's superior insulating properties made it an effective substitute for glass. Muscovite mica forms flat, often transparent, peelable sheets that can on occasion be a meter or more across. As a flexible insulator, mica has found itself used in a variety of demanding circumstances. It is used for peepholes on furnaces, electrical capacitors, an additive in fracking solutions, a plastering additive and a component of some lasers. Mica is a remarkable mineral that has had huge social impacts. However, like asbestos, when ground into a powder for use in plaster or other pastes, the dust can cause damage to respiratory surfaces, albeit not to the extent of its more lethal chemical cousin.

Branching off from this generic series of increasingly silica-rich intermediate rocks is a roughly parallel series that is particularly rich in potassium and/or sodium. This series is called the alkaline rocks (Table 1.2). While they never attain the same level of silicon dioxide as granites, there are alkaline basalts, alkaline intermediate rocks known as syenite, and an even more alkaline series of rocks that contains silica-poor minerals called feldspathoids. There are even alkaline versions of the granites, known as peralkaline granites. These still contain quartz, but they are more enriched in potassium than normal granites.

The feldspathoids, which comprise many of these alkaline rocks, are an unusual collection of very attractive semiprecious minerals. While they are very similar to the feldspars, they are poorer in silica (Table 1.2). Syenite may contain up to 5% feldspathoids, as well as quartz, but as the quantity of these minerals increases and silica decreases, syenite evolves into nephelinite, a rock dominated by feldspathoids and pyroxene. Nephelinite and a somewhat less alkaline rock, basanite, form the ferromagnesian-rich end of this alkaline series.

Nephelinite is a fairly rare volcanic rock with a good degree of notoriety. Nephelinite volcanoes are generally restricted to areas where continental crust is splitting in two. The Earth’s most obvious rifting occurs in East Africa. In Tanzania, the volcano Nyiragongo contains one of the few permanent lava lakes on Earth. It is also the only volcano where the lava lake is composed of nephelinite. While this lava is cooler than basalt, it is far more fluid, and Nyiragongo’s slopes are steeper than typical shield volcanoes. Nyiragongo’s lava lake rises and falls with varying magma supply. Decades may pass between eruptions, but periodically, the magma column rises so high in the volcano’s caldera that its walls collapse and magma spills out into the surrounding countryside. On January 10th, 1977, Nyiragongo’s walls faltered while people were on their way to the morning market. Commuters were overwhelmed by a torrent of nephelinite moving at up to 60 kilometers per hour, and 70 people lost their lives, along with a herd of elephants. They were incinerated in their tracks, and the cooling lava left casts around their bodies as a grim reminder of what happened.

Nyiragongo increased in notoriety after a further eruption in 2002 saw highly fluid nephelinite lavas flood through the capital of Rwanda, Gomo. A 13-km-long fissure opened on the southern flank of the volcano, from 2,800–1,500 meters above sea level. A 2-meter-high wave of lava poured into the outskirts of Gomo along a 200–1,000-meter-wide front. Lava reached Lake Kivu, prompting warnings that it might trigger the catastrophic release of carbon dioxide from deep, anoxic waters near the lake bed in a similar manner to the carbon dioxide eruption from Lake Nyos in 1985. Although Gomo had prepared for the arrival of the lava, a large number of people were left homeless by the lava’s invasion. Worse still, the death toll rose to 147 people – not from the lava itself, but from carbon dioxide gas that seeped through fissures in the volcano’s flanks and pooled in deeper hollows or in buildings.

Interestingly, Venus shows similar evidence of highly fluid lavas. The Ammavari volcano shows 350-km-long flows that emerge from vast fissures in the side of the cone, where the flank has collapsed. Nephelinite or carbonatite lavas may be the ideal candidates (Chapter 7).