

Historical Geology

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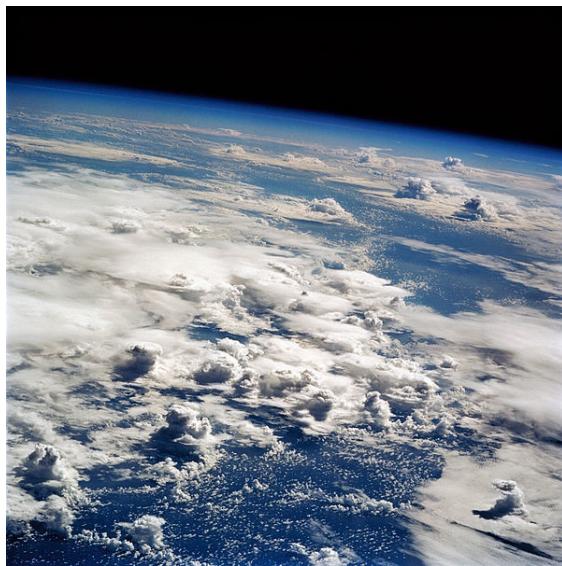
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0.1 Introduction



The Earth from space: thunderstorms brew over the Pacific Ocean

In this introductory article I shall explain what this textbook contains and why I wrote it the way I did.

0.1.1 What is historical geology?

Geology can roughly be divided into physical geology, which studies the materials of the Earth and the processes operating in it, and historical geology, which aims at a reconstruction of the history of the Earth.

Historical geology requires some knowledge of physical geology for its elucidation. (Imagine, by way of analogy, forensic scientists diagnosing cause of death as a gunshot wound, which is a historical question. It would obviously be necessary for them to know something about the behavior of guns, which would be a physical question.) However, the aim of historical geology is to understand the past, and knowledge of physical geology is merely an adjunct to this aim.

We may also speak of applied geology: for example, finding and extracting oil would fall under this category. This depends on both physical and historical geology: when petroleum geologists extract oil, it is certainly their understanding of the physical nature of rocks that allows them to extract it; but when they locate oil, it is their understanding of historical geology that makes them able to find oil with a success rate better than that which would be achieved just by guessing.

That is one application of historical geology. Here is another example: suppose someone wants to build a structure such as a major dam or a nuclear power plant at a certain site, and it is discovered that a geological fault runs under the site. It would then be crucial to discover

when last there was movement along the fault: if it was a hundred or even a thousand years ago, then the proposed location is dangerous; if it was ten million years ago then it is probably safe.

However, historical geology is by no means confined to facts about the past that are presently useful: it is what is called a “pure” science, in which knowledge is sought for the sake of knowledge itself, whether it turns out to be useful or is merely interesting.

0.1.2 Pre-requisites for reading this book

This is an introductory work, and there are no pre-requisites for reading it except perhaps that the readers should be able to remember the chemistry they learned in high-school; for those who cannot, or who have not yet attended high school, an article on chemistry for geologists is provided in [Appendix C](#).

In writing this textbook, I have not assumed that the reader knows any physical geology, and so have introduced concepts from physical geology where necessary. I have tried not to introduce *more* physical geology than is necessary, although occasionally I may have let my enthusiasm get the better of me.

0.1.3 The purpose of this book

This book has a single purpose: to explain how it is even possible to reconstruct the history of the Earth from data available in the present. The emphasis of this book is therefore methodological: rather than explaining *what* is known, we shall look at *how* it is known.

While these issues are sometimes mentioned in introductory geology textbooks, none that I know of addresses such questions systematically: one textbook will explain how geologists originally determined that granite is an igneous rock, but devote only a cursory page to radiometric methods; another, conversely, will provide us with lots of interesting information about the isochron method, but take the igneous nature of granite as read. I, on the other hand, have tried to be thorough in explaining how we know what is known.

There are a number of reasons why I have taken this approach.

Firstly, because other textbooks do not, and it seemed to me that this was a gap that I could usefully fill.

Secondly, because it seems to me that this is where the interest in geology really lies. Geology is not a set of dead facts: it is a detective story in which the application of reason takes a thousand diverse clues and unifies them into a single narrative that makes sense of it all. It is not, for example, particularly interesting that the core of the Earth is made mostly of iron — at least, it would be more interesting if it was made of cotton candy — but the fact

that we can find out what it's made of should inspire awe in any reader not too jaded to feel that emotion.

Thirdly, because the lay reader may feel an understandable skepticism when hearing an expert expound on what happened 100 million years ago. Skepticism, when it is honest, is an admirable attitude, and deserves an honest reply.

Finally, the average reader with only a basic scientific education will probably have ended up with a better grasp of scientific facts than of the scientific method. This is probably an inevitable consequence of the demands of a broad science curriculum, and in making this observation I intend no criticism of science teachers; nonetheless, it appears to be the case. It is my hope that for readers in that situation, this textbook will serve as a worked example of the scientific method. Geology is ideally suited for this purpose, since it requires clarity of thought but not (at least at the introductory level) advanced mathematics.

0.1.4 The introductory nature of this book

This is an introductory work: it has been my aim to keep it short. When I inform the reader that the geologists Chilingar and Wolf managed to write a book of 808 pages on the single topic of the compaction of coarse-grained sediments, you will appreciate how much more I have left out than I have put in — and will probably be grateful.

Its brevity, however, means that it will probably leave some questions of some readers unanswered. It is my hope, however, that it will at least give readers the concepts and vocabulary to ask *intelligent* questions, and to understand the answers when they find them.

0.1.5 Selection of material

The material in this textbook tends to reflect my own interests, in that it dwells more on what can be learned from sedimentary rocks and fossils than from igneous and metamorphic rocks, and more on the history of the Earth from the origin of complex life onwards than on the history of the early Earth. This emphasis is a matter of personal preference: other authors have other preferences, and have written other books.

Except for a brief summary of the geological column in the article devoted to that concept, I have not attempted to give an account of what the history of the Earth actually is. Instead, as explained above, I have concentrated on methodological questions. It is not difficult, after all, for someone with internet access to discover what happened in (for example) the Ordovician period; but accounts of what happened will typically not explain how such things can be known: it is this gap that I have attempted to fill.

Readers who have read other textbooks on historical geology will notice one unusual omission. It is customary for such works to provide at least a brief account of the

theory of evolution and the evidence for the fact of evolution. I have not done so. According to the plan of this book, with its emphasis on asking: "How do we know?" if I were to deal with evolution I would have to review evidence which is not only copious in quantity but also extraordinarily diverse in kind, more so than in any other branch of science. I estimate that to give even an outline sketch of the evidence would increase the length of this book by half again. A brief review would be too brief; an adequate review would be too long.

Even so, I would undertake this task if it was of critical importance, but it is not: all it would do for the reader is that, understanding evolution, the reader would understand why the principle of faunal succession ought to be true. But to practice geology, it is enough to know that it *is* true; and for the purposes of this book we need go no further than investigating the evidence that it is true, without an exceedingly lengthy digression explaining its underlying causes. In the end, in the hierarchy of ideas it is geology that supplies evidence for evolution, and not *vice versa*.

0.1.6 Arrangement of material

The material in this textbook is badly arranged. This is because the material in every geology textbook is necessarily badly arranged. The problem arises from the necessity of arranging it at all: of presenting it to the reader in a certain order. There is no right order because all the concepts in geology fit together to form a unified whole.

So with a few obvious exceptions it would be best if every article in this textbook, or any scientific textbook, came last in the book, so that the reader could read it in the light of all the other articles; but this is not possible. The reader does, however, have the option of reading through the whole book twice, and I recommend this course of action to any serious reader of this or any other similar textbook.

0.1.7 Vocabulary in this book

New terms will be marked by the use of **boldface**. This may not mark the absolute first use of a word, but the place in which it is first fully defined.

All words so introduced will also be defined in Appendix A.

0.1.8 Note on references

I have thought it unnecessary to provide references to facts which can be found in all or most introductory works on geology, and which may be considered common knowledge among geologists. I have, however, provided references to support specific pieces of data which

are not so widely known, and for which a practicing geologist might wish to see a reference.

A bibliography is provided in Appendix B.

0.1.9 Note to potential editors

By contributing this book to the Wikibooks project, I am making it possible for others to edit it. It would be courteous to contact me before making any major changes or additions to this book. I would particularly ask that no-one should try to expand this book in a way that digresses from its stated purpose, since that would increase the length of the work without helping to achieve its aim.

Minor corrections of fact, grammar, spelling, etc, are of course welcome. Note however that where I have consistently used some stylistic feature this is deliberate. In particular I have purposely used only standard scientific units such as meters, grams, degrees Celsius, etc, without translating them into feet, ounces, degrees Fahrenheit, etc, as I feel that any reader who wishes to study geology should get used to the standard units, and the sooner the better.

0.1.10 Acknowledgements

I would like to thank the people, too numerous and usually too anonymous to mention, who have pointed out errors or infelicities in early versions of this text.

I should particularly like to thank Louis Kirstein for his patience and diligence in reviewing the entire work. This was undertaken in his personal capacity and so should not be construed as endorsement or approval by the distinguished body that employs him.

Any remaining errors or omissions are of course my own fault.

Tim Hardcastle

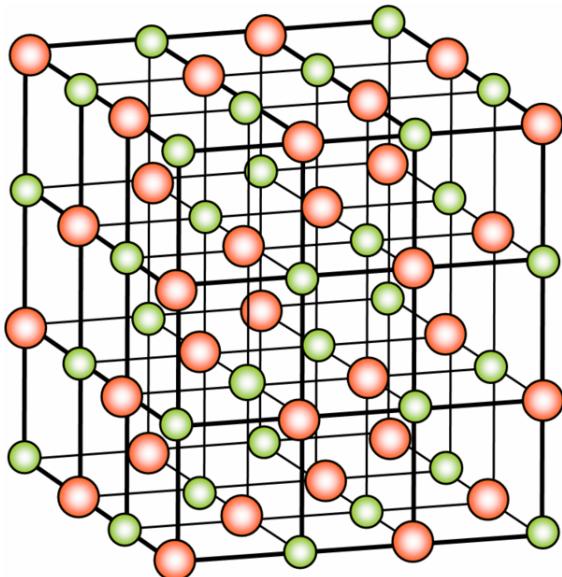
Las Vegas

April 2013

Chapter 1

Minerals and rocks

1.1 Minerals



The chemical structure of halite.

In this article we shall introduce some useful basic definitions: in particular, we will look at the definition of a mineral and see how it is different from the definition of a rock.

1.1.1 Minerals

A **crystal** is defined as a solid in which the constituent **atoms** are arranged in an orderly repeating pattern. The diagram to the right, for example, is a ball-and-stick model of the molecular structure of halite. The large red balls represent atoms of chlorine; the small green balls represent atoms of sodium; the sticks represent the bonds between atoms.

A **mineral** is a naturally occurring inorganic solid which is defined by a chemical formula and a particular crystal structure.

Take for example the mineral halite (more familiar to you as table salt). It has the formula NaCl , because it is formed from units consisting of one atom of sodium

(Na) and one atom of chlorine (Cl), and it has a cubic structure as shown in the diagram above. The formula and structure define the mineral.

It is perfectly possible to have two different minerals with exactly the same formula but different crystal structures. One commonly used example of this are the minerals diamond and graphite. Both consist entirely of atoms of carbon, and so have exactly the same chemical formula, but graphite has a hexagonal crystal system and diamond has an cubic crystal system; as a result their physical properties are very different. Two minerals having the same formula but different molecular structures are known as **polymorphs**.

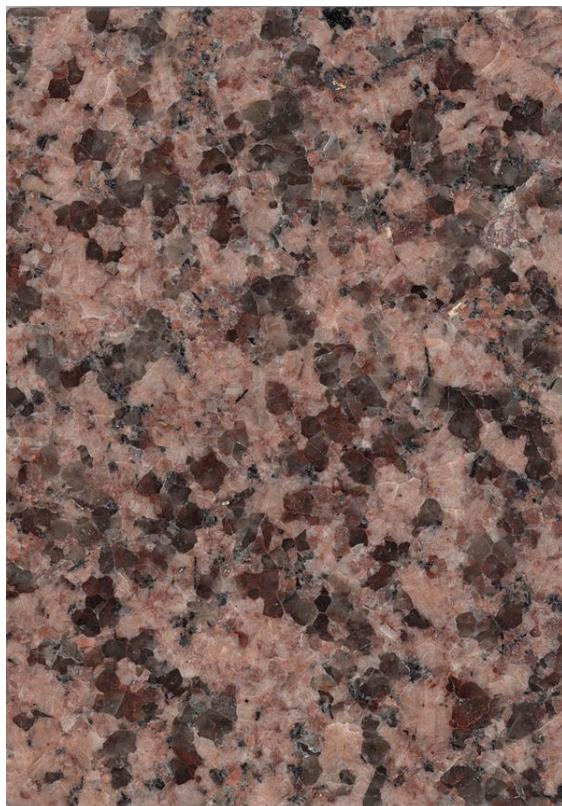
Some minerals are what is known as **solid solutions**. Take for example the mineral olivine. This has the formula $(\text{Mg},\text{Fe})_2\text{SiO}_4$. The part of the formula which says (Mg,Fe) indicates that there are positions in the crystal structure each of which can be filled by an atom of *either* magnesium (Mg) *or* iron (Fe). What proportion of these positions is filled by iron and what proportion is magnesium will vary from sample to sample of olivine. It is convenient to treat these as varieties of the same mineral.

There are thousands of known minerals, most of which are of interest chiefly to collectors, and which we can largely ignore in an introductory text such as this. Of those that we shall mention, far and away the most important class are the **silicate minerals**, and the next article will be devoted to their physical and chemical properties.

1.1.2 Mineraloids

A **mineraloid** is a substance which is similar to a mineral in some respects but does not fulfill all the criteria necessary to be a mineral. There does not appear to be a complete consensus on what is a mineraloid and what is just a plain non-mineral, and the term is little-used.

One important sort of mineraloid is naturally occurring glass, such as **obsidian** or **pumice**. **Glass** by definition does not have a crystal structure, but is a confused mess at the molecular level. Such a structure, or rather lack of structure, is described as being **amorphous**.



Granite.

1.1.3 Rocks

A **rock** is an aggregate of one or more minerals or mineraloids. For example **granite** consists mainly of the minerals **quartz**, **mica** and **feldspar**. In granite the crystals of the different minerals are actually visible to the naked eye, giving it its characteristic speckled appearance, as in the picture to the right.

A rock *can* consist of a single mineral: for example the semi-precious stone known to dealers in gemstones as “rock crystal” consists entirely of **quartz**.

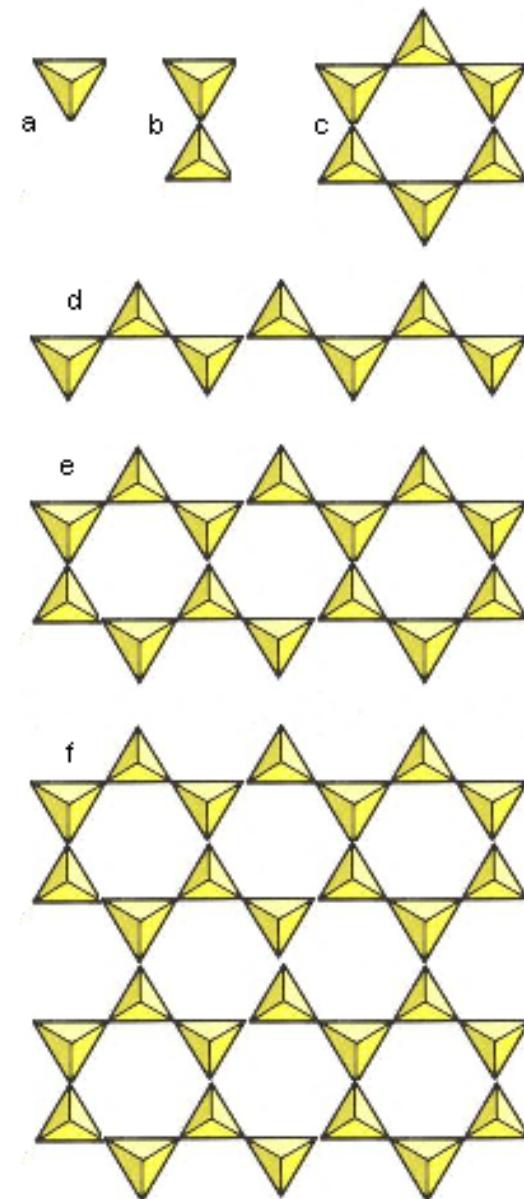
1.2 Silicate minerals

In this article we shall look at the important class of minerals known as **silicates**.

1.2.1 Silicate structures

By a **silicate tetrahedron** we shall mean an atom of silicon bonded with four equally spaced atoms of oxygen forming the four corners of a tetrahedron: that is, a pyramid having a triangular base.

Each tetrahedron can share each one of its oxygen atoms with one other tetrahedron, so that two tetrahedra can join together corner-to-corner (but not edge-to-edge or



Ways in which silicate tetrahedrons can join together in silicate minerals. Key: (a) an isolated tetrahedron; (b) a pair; (c) a six-member ring; (d) a chain; (e) a double chain; (f) a sheet.

face to face). Hence each tetrahedron can be linked with up to four other tetrahedra, one for each corner of the tetrahedron; or three, two, one, or none. This gives us a wide variety of structures that can be built out of tetrahedra: three-dimensional lattices; two dimensional sheets, chains, double chains, rings, et cetera. The diagram to the right shows some of the possible structures. Note that the chain and double chain can be extended indefinitely in one direction and the sheet in two directions.

A **silicate mineral** (or **silicate** for short) is a mineral containing silicate structures; so silicate minerals can be classified according to their silicate structures as **lattice silicates**, **sheet silicates**, **chain silicates**, and so forth. Note

that with the exception of **quartz** and its **polymorphs**, a silicate mineral will not consist entirely of such structures. Atoms of other elements must necessarily be involved so that the rings, chains, sheets or whatever form part of a three-dimensional crystal.

Most silicate structures can be described either by a descriptive English name such as “sheet silicate” or by a name which describes the same thing in Greek such as “phyllosilicate”. Where a plain English term exists, I shall employ it in this text. The table below shows how the structures relate to the English and Greek names and gives examples of **minerals** important to this textbook.

Because tetrahedra link together by sharing the oxygen atoms at their corners, the structure formed by the tetrahedra is reflected in the chemical formula of a silicate. For example, **quartz** consists of nothing but tetrahedra linked together in a three-dimensional lattice. This means that every tetrahedron is linked to another at all four corners; which means that every oxygen atom is shared by two silicon atoms; which means that quartz has the formula SiO_2 . Similarly, if you look at the formula for zircon (ZrSiO_4) you can see that it must be a nesosilicate.

1.2.2 Aluminosilicates

In some silicates, the structure is based not just on silicate tetrahedra but also on tetrahedra with a central atom of aluminum rather than silicon. Silicates which incorporate these aluminum-based tetrahedra as well as silicate tetrahedra are known as **aluminosilicates**.

Aluminum-based tetrahedra have different chemical properties from silicate tetrahedra. This means that you cannot have an aluminosilicate which differs from an ordinary silicate only by the substitution of atoms of aluminum for some of the atoms of silicon; there must necessarily be other differences. For example, it is chemically impossible to build a lattice just out of these two kinds of tetrahedra that is analogous to **quartz**; other atoms are required to balance the charge of the chemical formula. Hence lattice aluminosilicates such as **feldspars** do not have the formula $(\text{Si}, \text{Al})\text{O}_2$, which is impossible, but have more complicated formulas such as KAlSi_3O_8 and $\text{CaAl}_2\text{Si}_2\text{O}_8$.

1.2.3 Felsic and mafic silicates

Silicate minerals which are high in silicon are called **felsic** minerals; the opposite of felsic is **mafic**; minerals which are very mafic are known, sensibly enough, as **ultramafic**. Note that this term only applies to minerals which *are* silicate minerals and so contain *some* silicate tetrahedra; no-one would call **calcite** (for example) an ultramafic mineral on the grounds that it contains no silicate tetrahedra at all.

Some generalizations can be made about the differences

between felsic and mafic minerals: as we progress from felsic to mafic the minerals are more dense (because the atoms in them which aren't oxygen or silicon are heavier elements such as magnesium or iron); they have higher melting points; and when they do melt they are less **viscous** (that is, they flow more easily).

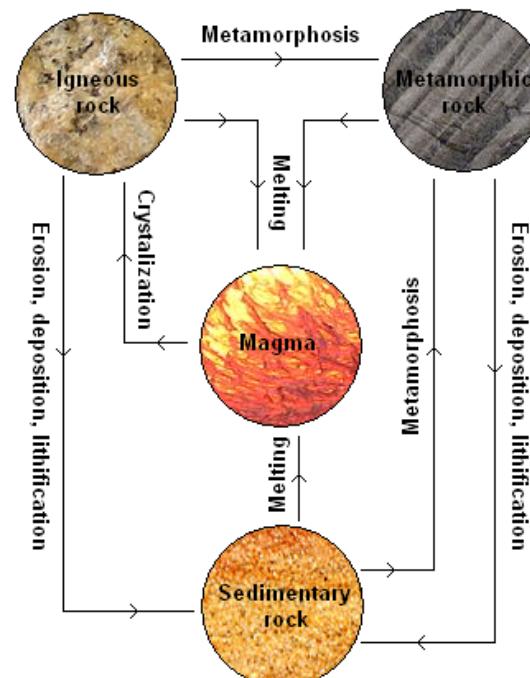
1.2.4 Note on vocabulary

In some texts, particularly older texts and British texts, you may see the words **acidic**, **basic**, and **ultrabasic** used instead of felsic, mafic, and ultramafic. These terms refer to an obsolete hypothesis of mineral formation, and I shall not use them; I mention them only for the benefit of those readers who might come across them in further reading.

1.3 Rocks

In this article we shall look at the most significant way in which geologists classify rocks. The reader should recall from the article on minerals that a rock is an aggregate of one or more minerals or mineraloids.

1.3.1 Igneous, sedimentary, and metamorphic rocks



The rock cycle.

There are all sorts of ways that we might classify rocks. We might, for example, divide them up according to chemistry: and indeed the distinction between silicates

and carbonates is a useful one. We might also classify rocks according to whether they contain **felsic** or **mafic** minerals, and as we shall see this is a good way to classify certain rocks. But the most fundamental way in which geologists classify rocks is to label them as igneous, sedimentary, or metamorphic.

- **Igneous rocks** are rocks formed by the cooling and setting of molten rock.
- **Sedimentary rocks** are formed by sediment (for example, **sand** or **mud**) turning into rock (such as sandstone or mudrock).
- **Metamorphic rocks** are formed when rocks are subjected to heat, to pressure, to chemical reactions, or to any combination of these three, in such a way as to change the properties of the original rock in some way.

1.3.2 The rock cycle

The three types of rock can be converted on into the other by geological processes. Metamorphism can turn igneous and sedimentary rocks metamorphic; erosion can turn igneous and metamorphic rocks into sediment which can then become sedimentary rock; and sedimentary and metamorphic rocks can be melted down into molten rock which can then cool to form igneous rock.

The relation between the various kind of rocks can be summarized by a diagram of the **rock cycle**. One representation of it is shown to the right.

1.3.3 How do we know?

At this point we are beginning to touch on the main theme of this textbook. For to classify rocks as igneous, sedimentary, or metamorphic is implicitly to classify them not by their directly observable properties such as color or density or chemical composition, but by their *history* as inferred from their present-day properties. The reader will therefore want to know how this history is inferred. This question will be answered in separate articles on igneous, sedimentary, and metamorphic rocks.

In the meantime, let us observe how intrinsically historical geology is. After we've got past the most basic of considerations such as defining a **mineral** and defining a **rock**, we are plunged into historical considerations. And this is not just because this course is about historical geology: it would be the same if it was an introduction to how to find oil. Geology is so intrinsically a historical science that if we tried to do without historical inferences we might as well classify rocks by how pretty they are for all the good it would do us.

1.4 Igneous rocks



Eruption of lava, Hawaii.

In this article we shall look at igneous rocks: what they are and how they are classified.

1.4.1 Intrusive and extrusive igneous rocks

Igneous rocks are rocks formed by the cooling and solidification of molten rock. They fall into two main categories:

- **Intrusive** rocks are those which are caused by the cooling of molten rock underground. Subterranean molten rock is known as **magma**.
- **Extrusive** rocks are those formed from molten rock on the surface, which is known as **lava**.

Igneous rocks can be further identified and classified by their texture and their chemistry, as will be described in the following two sections of this article.

1.4.2 Texture

It is a universal law that fast crystallization makes small crystals and slow crystallization makes large crystals. This is because crystallization is a kinetic process: for a molecule to join onto a crystal it must bump into it and then align with it.

The thermal properties of rock are such that magma cooling underground will cool slowly as compared to lava cooling above ground. Hence, by looking at the texture of the rock, we can find out how it cooled: an intrusive rock will be **coarse-grained**; an extrusive rock will be **fine-grained**.

Sometimes lava is ejected from a volcano with such force that it goes shooting high up into the air, causing it to solidify so quickly that it doesn't have time for crystals to form at all, making an **amorphous** solid known as a **glass**. The glass in windows is an artificial glass produced by the

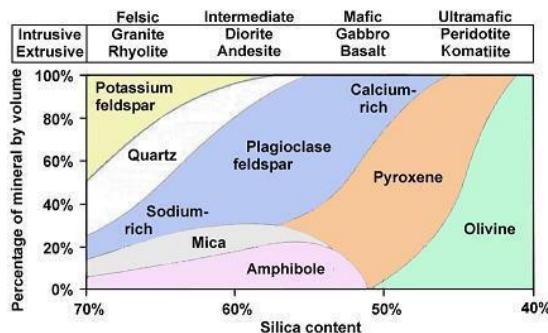
rapid cooling of molten silica; examples of natural glasses are **obsidian** and **pumice**.

Occasionally magma will begin to cool below the surface and then be ejected on to the surface; in this case it will have a **porphyritic** texture, with a few larger crystals (**phenocrysts**) embedded in a finer-grained **ground mass**.

1.4.3 Chemistry

The simplest way to classify the chemistry of igneous rocks is by the amount of **silica** they contain.

An igneous rock with a high silica content is said to be **felsic**, and an igneous rock which is low in silica is said to be **mafic**. You will recall that these are the same terms used for high-silica and low-silica **minerals**; and in fact it is the case that felsic rocks will contain felsic minerals and mafic rocks will contain mafic minerals.



Composition of igneous rocks

Classifying rocks by their silica content is convenient because typically the chemistry of igneous rocks lies on a continuum such that if you know the proportion of silica in an igneous rock, you can say what minerals it contains. The rules for doing so can be represented by the diagram to the right. Note that this applies only to igneous rocks, and not to **sedimentary** or **metamorphic** rocks.

To read the diagram, look along the bottom of the graph for the silica content of the rock: then a line drawn directly upwards from that point cuts through the minerals it will contain in their relative proportions. So, for example, if we tell you that a certain rock contains 50% silica, then you can see from the chart that it contains about 5% olivine, 75% pyroxene, and the remaining 20% will be calcium-rich plagioclase feldspar.

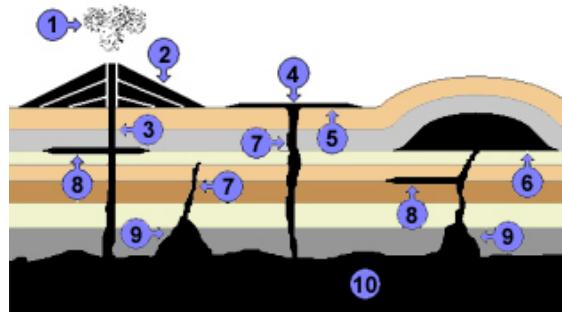
This diagram divides the rock types into fairly coarse divisions. It is possible to make finer distinctions: we could, for example, have put granodiorite between granite and diorite, as a rock type having a silica content lying between granite and diorite; or we could have placed dunite to the right of peridotite, to denote those rocks which consist of pure olivine. The divisions we have proposed are, however, sufficient for our present purposes.

It is more important that the reader realizes that whatever divisions we impose on the diagram, they are arbitrary: there is a *continuum* between felsic and ultramafic rocks.

Also, as we look along the continuum from felsic to ultramafic, the rocks are progressively denser; they have a higher melting point; and they have a less viscous flow when molten. This is the same progression as we see as we pass from felsic to ultramafic minerals, and is a natural consequence of the fact that felsic rocks consist of felsic minerals and mafic rocks of mafic minerals.

We should perhaps add a note on the presence of komatiite (extrusive ultramafic rock) in our diagram, as some textbooks omit it entirely from such diagrams. Komatiite is never observed forming today: as ultramafic magma rises from the hot interior of the Earth to its cool surface, it will fall below its melting point before it gets near to the surface, forming peridotite, komatiite's intrusive counterpart. Consequently komatiite is found only in rocks dated to over 2.5 billion years ago, consistent with geologists' belief that the Earth was hotter at that time.

1.4.4 Igneous structures



Some igneous structures. Key: (1) Volcanic ash. (2) A volcano. (3) A volcanic conduit. (4) A fissure. (5) A lava flow. (6) A lacolith. (7) Dikes. (8) Sills. (9) Stocks. (10) A batholith.

The diagram to the right shows some of the structures formed by igneous rocks. The black represents igneous rock; the other colors represent sedimentary rocks.

As this is a cutaway diagram, it may be slightly misleading. The reader should bear in mind that a fissure is a crack in the surface; we have shown it end-on. Similarly, the lava flow which emerges from a fissure will be a sheet of lava; and a dike is not a spike of rock, but a vertical or near-vertical sheet of rock. And a sill, again, is a horizontal sheet of rock.

That last statement needs a little qualification. In the diagram, we have shown the layers of rock lying flat, except around the lacolith (item (6) on the diagram) and so we have shown the sills as horizontal structures. However, layers of rock can be folded by tectonic activity. When a sill intrudes into rocks like this, it intrudes between the layers of rock (this is the definition of a sill) and so will itself be contorted.

We shall have more to say about igneous structures when we consider **stratigraphy** and cross-cutting relationships, but for now this brief introduction is sufficient.

1.4.5 How do we know?

How do we know that igneous rocks are igneous? Like everything else in geology, this had to be proved at some point: indeed, there was once a body of thought known as “Neptunism” which asserted (amongst other things) that granite was sedimentary.

In the case of extrusive rocks, the answer is obvious: we can see **basalt** (for example) forming when lava flows cool: so it certainly *can* form as an extrusive rock. But it could not also form as an intrusive igneous rock, because under such circumstances, being thermally insulated, it could not cool quickly enough to produce a fine-grained structure, and the physics of the situation would dictate the formation of **gabbro** instead.

Since we can actually watch the formation of basalt, we can make further deductions about it. When basalt cools underwater (as observed by divers), it forms the distinctive shapes known as **pillow basalt**, which is not the case when it is observed forming on dry land. This criterion allows us to distinguish between basalt formed on land and on the sea floor; a deduction confirmed by the association of pillow basalt with marine sedimentary rocks.

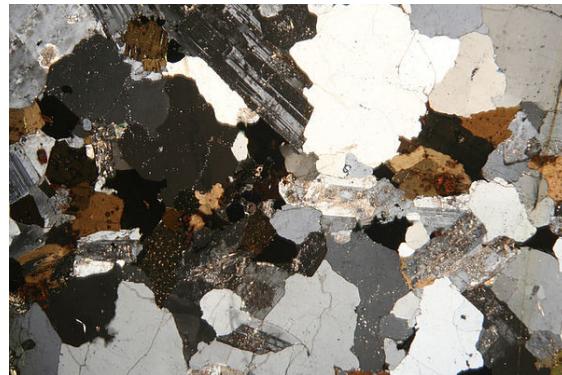
But what about intrusive rocks? Take granite, for example, since it is the commonest intrusive igneous rock. If we are absolutely right about how it forms, we should never see it forming. So how do we know how it forms?

As a matter of fact, the fact that we never see it forming is one of the predictions of the theory that it is an intrusive igneous rock, and so tends to confirm the theory. We do not see granite or granite-like **sediment** forming by surface processes; what else can we conclude but that it is formed underground?

In the second place, as we have observed, granite has the same chemical composition as rhyolite, differing from it only in its **texture**. Now, as we know that larger crystals form when cooling is slower, and as the thermal properties of rock as opposed to air or water will lead to slower cooling underground, we must conclude that granite is exactly what we should expect to see if the magma that forms rhyolite when extruded onto the surface was to cool below the surface instead.

A close look at its texture through a microscope confirms the igneous nature of its formation. The picture to the right is a photomicrograph of granite. Note how the crystals, however bizarre their shape, fit together perfectly. We may compare this with the texture of **sedimentary rocks** such as **sandstone**, which are clearly made of non-interlocking particles cemented together.

Then we may consider the structures formed by intrusive



Photomicrograph of granite.

rocks. It is difficult to see how something such as a dike, which, as explained above, is a vertical or near-vertical sheet of rock, could form by any process except the intrusion of magma into a crack in pre-existing rocks.

Finally, we may note that the rocks into which granite intrudes are typically changed in ways we would expect if they had been subjected to great heat; for example, when granite intrudes through a layer of **limestone**, the limestone immediately adjacent to the granite will be turned to **marble**. This suggests that the granite was itself once at a high temperature and has subsequently cooled, consistent with the theory that it is an intrusive igneous rock.

For these reasons, we may conclude that granite is an intrusive igneous rock; similar remarks might be made about the other rocks classified as igneous intrusive.

1.4.6 Note on vocabulary

Igneous rocks are sometimes called **primary rocks**, extrusive rocks are sometimes called **volcanic rocks**, and intrusive rocks are sometimes called **plutonic rocks**. We shall not use these terms in this text, and mention this only for the benefit of those readers who wish to pursue a course of further reading.

The rocks that we have described as fine-grained and coarse-grained are also known by the terms **aphanitic** and **phaneritic** respectively. These terms are rather commonly used by geologists, but I shall stick to the more self-explanatory terms.

Finally, just as silicate minerals are sometimes referred to (erroneously) as “acidic”, “basic” and “ultrabasic” rather than felsic, mafic, and ultramafic, the same is true of igneous rocks; as in the case of minerals, I do not intend to use these terms, as they are obsolete and misleading.

1.5 Sedimentary rocks

This article is a brief introduction to the various kinds of sedimentary rocks. Further information about the



Sandstone, The Wave, Arizona

sources of sediment, its transport, and its deposition, will be covered in further articles; indeed, in much of the rest of this textbook.

1.5.1 Types of sedimentary rocks

Sedimentary rocks can be divided into three main classes:

- **Clastic** sedimentary rocks are formed from sediments created by rocks being broken down into small particles (**clasts**).
- Chemical sedimentary rocks are formed from sediments created by dissolved chemicals being precipitated out of the water they're dissolved in.
- Biochemical sedimentary rocks are formed from sediments consisting of dead organisms, or parts of dead organisms.

In some schemes of classification, biochemical sediments are treated as a sub-class of chemical sediments, but this leaves one with the awkward question of what to call chemical sediments which aren't biochemical. For this reason I shall treat them as two non-overlapping classes.

Before we review the main types of sedimentary rocks, it is worth mentioning the process by which they turn into rock: this is known as **lithification**. In some cases, such as shale, mere compaction, along with the resulting loss of water, is sufficient. Coarser sediments, such as sandstone, are both compacted and **cemented**, as can be seen under a microscope. The cements are minerals precipitated out of the water in which they are dissolved: **silica** and **calcium carbonate** are the commonest forms of cement, with iron oxides and hydroxides coming a distant third.

In the sections below we shall list the main types of sedimentary rock.

1.5.2 Clastic sedimentary rocks

As defined above, clastic sedimentary rocks are formed from broken pieces ("clasts") of pre-existing rocks.

- Sediment: rounded gravel. Rock: conglomerate.

Gravel is defined as clasts with diameter 2mm or more. We should note that when geologists speak of **rounded** clasts, they do not necessarily mean that they are round like a ball, but merely that the sharp corners and edges have been worn off them by erosion. **Conglomerates** are rocks formed mainly from rounded gravel which has been compacted and cemented together.

- Sediment: angular gravel. Rock: breccia.

Breccia is like conglomerate except that the gravel is **angular**: that is, it has not been rounded. This reflects a different history, since gravel that has been transported any appreciable distance by water, or which has been rolled about by waves on a beach, will quickly have its corners and edges worn away.

- Sediment: sand. Rock: sandstone

Sand is defined as clasts less than 2mm and more than 1/16mm in diameter. **Sandstone** is sand that has been cemented together.

Most sandstone is **quartz sandstone**; that is, it consists of grains of **quartz**. This is because the process known as **chemical weathering** dissolves many rock-forming minerals, or, in the case of **feldspar** minerals, converts them to **clay**, leaving behind only the quartz from the original rock. We shall look at the process of chemical weathering in a subsequent article.

Arkose sandstone is sandstone with an appreciable proportion of **feldspar** minerals in it. This reflects a somewhat different history to quartz sandstone, in that it must have been formed when **mechanical weathering** (the physical process of breaking rock into clasts) has predominated over the **chemical weathering** that would otherwise have converted the feldspar minerals to **clay** minerals.

Greywacke is sandstone that, in addition to **quartz** and **feldspar**, also contains sand-sized fragments of **igneous** or **metamorphic** rocks. Similar remarks apply to greywacke as to arkose sandstone.

- Sediment: silt and/or clay. Rock: mudrock.

Silt is defined as clasts between 1/16mm and 1/256mm in diameter.

The term "**clay**" is a little ambiguous. In the classification of sediments, it is defined as particles less than 1/256mm in diameter. However, in mineralogy, clay is a class of minerals (technically, **hydrous aluminosilicates**). In practice, this need cause no confusion, because what is clay by size will be overwhelmingly clay by composition.

Mudrocks can then be divided into **siltstone** (formed from silt sediments); **mudstone** (from sediments that are a mixture of silt and clay); and **claystone** (from clay sediments).

- Sediment: bedded mud or clay. Rock: shale.

Most mudstone and claystone is **bedded**. When it is, it is referred to as **shale**.

1.5.3 Chemical sedimentary rocks

- Sediment: salt. Rock: halite.

Halite, also known as rock salt, is an **evaporite**, formed by the evaporation of salt water.

It can be formed by complete evaporation of salt water, as seen, for example, in desert salt flats. However, complete evaporation is not necessary; it is sufficient that enough water should evaporate that the remaining water can't hold all of the salt in solution; so halite can also form in shallow seas or salt lakes in a hot environment.

- Sediment: hydrated calcium sulfate. Rock: gypsum.

Calcium sulfate is another substance to be found dissolved in salt water, and **gypsum**, like halite, usually forms as an evaporite under pretty much the same circumstances.

- Sediment: silica. Rock: chert.

Dissolved silica can precipitate out of the water in which it's dissolved to form **chert**. Note, however, that chert is more usually formed as a biochemical sedimentary rock.

- Sediment: calcium carbonate. Rock: limestone.

Calcium carbonate, like silica, can precipitate out of water to form **limestone**. Sometimes it forms tiny spheres called **oids**, which form around grains of sand or fragments of shell, which are then cemented together by the further precipitation of calcium carbonate; such limestone is known as **oolitic** limestone.

Most limestone, however, is biochemical sedimentary rock, formed from shells or coral.

1.5.4 Biochemical sedimentary rocks

- Sediment: shells of calcium carbonate. Rock: limestone.

Most limestone is formed from tiny hard parts of marine creatures which build their shells out of calcium carbonate; these settle on the sea floor to form **calcareous ooze**. Chalk is an example of such a rock: the tiny fossils that compose it can be clearly seen and identified under a microscope.

The hard parts of coral **reefs** are sometimes preserved intact, giving us **reef limestone**.

- Sediment: shells of silica. Rock: chert.

While calcium carbonate is the most popular substance to make shells out of, some organisms such as **diatoms** and **radiolarians** build their shells out of silica; these settle on the sea floor to form **siliceous ooze** which, when compacted and cemented, forms chert.

- Sediment: peat. Rock: coal.

Peat is plant material laid down in oxygen-poor conditions, so that it doesn't entirely decompose. Pressure, and the higher temperatures which come with deep burial, can then convert it into **coal**.

1.5.5 Modes of deposition

In the sections above we have principally divided sedimentary rocks by their composition. We can also classify them by their modes of deposition: for example, **aeolian** (deposited by the wind), or **fluvial** (deposited by rivers) and so forth. We shall have a lot more to say about this in subsequent articles.

1.5.6 Bedding



Cross-bedding, Dry Fork Dome, Utah.

Sedimentary rock often exhibits **bedding**: that is, the rock has distinct layers in it and is **fissile**: that is, it splits more easily at the divisions (**bedding planes**) between the layers. In **cross-bedded** rocks, the layers are not flat but lie at an angle to the horizontal, as a result of the original sediment being formed into dunes or ripples by the action of wind or water.

The picture to the right shows a particularly large-scale example of cross-bedding in sandstone.

1.5.7 How do we know?

How do we recognize sedimentary rocks as sedimentary? How do we recognize the sediments that compose them and the manner of their deposition?

Such questions will be answered later in this course one type of sediment at a time. At present we shall content ourselves with sketching out a general answer.

In the first place, the rocks look just like we would expect if sediments became lithified; for example sandstone *looks* like it's made of sand: everything about the size, the composition, and the erosion of the grains of which it's composed is in agreement with the idea that what we're looking at is grains of sand cemented together.

Secondly, we can drill down and take *cores* of sediments, and we can see, as depth increases, how (for example) sloppy muddy ooze on the surface *grades* into hard mud-stone with no sharp dividing line between them; similarly we can see calcareous ooze grade into solid limestone.

Then again, all types of sedimentary rocks can contain *fossils* (including, as we have remarked, those rocks which consist of fossils, such as chalk). This is consistent with the processes of burial of organic remains in sediment which we can see going on today.

Trace fossils are also a strong argument: when we find, in shale, the recognizable fossil footprints of land animals, it is hard not to conclude that what we are looking at once lay on the surface and was soft enough to take impressions such as we can see being made in mud today.

These considerations also allow us to figure out where the sediments were deposited: on land, in fresh-water, or in the sea.

The sedimentary structures within the rocks, such as bedding and cross-bedding, can be seen today in oozes forming on the sea floor; in sand-dunes; in ripples caused by tidal action, and so forth. Again, consideration of these structures will allow us to make the deduction, not merely that the rock is sedimentary, but also about the method of its deposition; if, for example, we find sedimentary structures such as can only be formed by tidal action, we are forced to infer that we are looking at lithified nearshore sediments.

Also, we may observe the topographic patterns of deposition. For example, when we see sedimentary rocks which because of their structures and fossils we identify as *terrestrial* (i.e. associated with the land) divided from sedimentary rocks which because of their structures and fossils we associate with the sea by a long thin strip of sedimentary rocks which because of their structures and fossils we associate with the *nearshore*, then this observation confirms our identification of the rocks as being terrestrial, marine, and nearshore rocks. If, on the other hand, we found alternating bands of marine and nearshore rocks, this would tend to falsify our theories. The fact that the topography of sediments is always consistent with our theories is therefore a point in favor of their correctness.

So the conclusion that sedimentary rocks are, indeed, sedimentary in origin, is a safe one; and we are certainly not without clues as to the manner of their deposition.

1.5.8 Note on vocabulary

Conglomerates and breccias are sometimes called **rudaceous rocks**; sandstones are sometimes called **arenaceous rocks** or **arenites**; and mudrocks are sometimes called **argillaceous rocks**.

The rocks which we have called clastic are sometimes called **detrital**.

As usual, I shall employ a consistent vocabulary in this text; these terms have only been supplied for the benefit of the reader who wishes to pursue a course of further reading.

1.6 Metamorphic rocks

Metamorphic rocks are those in which a pre-existing rock (the **parent rock**) has been altered chemically or texturally by heat and/or pressure. In this article we shall look at metamorphic processes and their effects of the resulting rocks.

1.6.1 Types of metamorphism

There are two main types of metamorphism; contact metamorphism and regional metamorphism.

In **contact metamorphism**, the heat of **magma** intruding through rocks causes metamorphism to the surrounding rocks, leaving an **aureole** of metamorphic rocks around the igneous rocks formed from the magma.

Regional metamorphism is caused by tectonic events involving both heat and pressure, and will affect large, elongated regions of rock.

Besides these broad classifications, we can also class metamorphic rocks according to their **grade** of metamorphism; the temperature to which they have been exposed. As we shall see, this determines the chemical changes that metamorphic rocks undergo. Grades of metamorphism range from very low (below 300°C) to low (300°C - 500°C) to medium (500°C - 600°C) to high (600°C and upwards).

1.6.2 Chemical changes

Metamorphism often causes chemical changes to the minerals of which the affected rocks are composed. For example, at low temperatures clay minerals will be converted into chlorite; at higher temperatures the chlorite will itself be transformed into other minerals. Hence, if we find a rock with chlorite in it, we know that it has undergone low-grade metamorphism. Minerals such as this, which reveal the grade of metamorphosis, are known as **index minerals**.



Garnets in metamorphic rock.

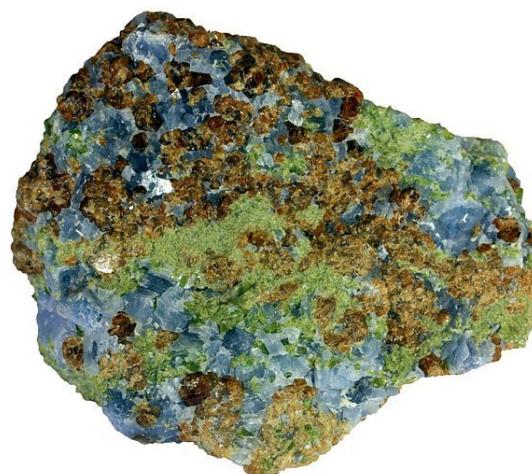
Geologists can correlate index minerals with grades of metamorphism because it is simple enough to repeat the processes of metamorphism in the laboratory; that is, they can take a piece of non-metamorphic rock, subject it to various regimes of temperature and pressure, and see which characteristic minerals form at which temperatures. So in shale, for example, we see a sequence (as temperature increases) from unaltered shale to rocks containing chlorite; then biotite; then garnet; then staurolite; then kyanite; and finally sillimanite. The image to the right shows some particularly large garnets embedded in metamorphic rock.

We see this same sequence arranged spatially as we approach the center of an area of metamorphism: from unaltered shale through the “chlorite zone”, the “biotite zone”, the “garnet zone”, the “staurolite zone”, the “kyanite zone”, and the “sillimanite zone”. We may not get all the way up to sillimanite, that depends how intense the metamorphism was at the center of metamorphism.

The sequence, of course, depends on the parent rock; the sequence given above is specific to shale, and we would see a different sequence if we were looking at (for example) mafic igneous rocks.

1.6.3 Metasomatism

In the section above on chemical changes we dealt with the case where the parent rock reacts with itself as a result



Skarn.

of heat or pressure acting on the rock. But in the case of contact metamorphism, we can also see **metasomatism** taking place: the parent rock mixes and/or reacts with the **intrusive igneous rock** and the hot fluids associated with its eruption.

The picture to the right shows a particularly attractive example of **skarn**, a very distinctive product of metasomatism.

1.6.4 Textural changes

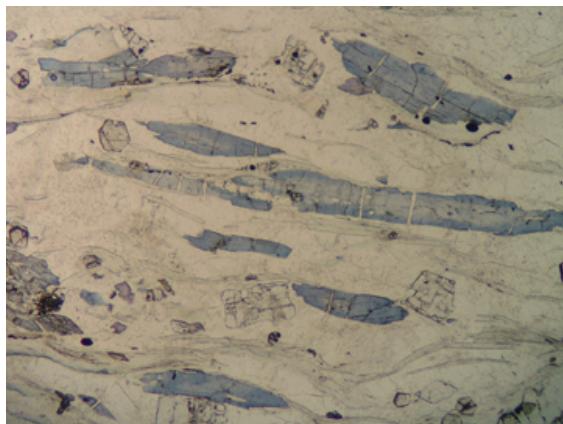
Besides chemical changes, rocks that undergo metamorphism suffer textural changes, such as recrystallization, foliation, and lineation.

In **recrystallization**, the original texture of the rock is lost as the minerals, under the effect of high temperatures, reform as a collection of interlocking crystals of similar size. The effect of this is seen most dramatically in sedimentary rocks.

So, for example, **quartz sandstone** loses its sedimentary structure of cemented grains to become **quartzite**, with a smooth texture consisting of interlocking **crystals**. As a result, except at very low grades of metamorphism, any **bedding** of the rock will be destroyed, as will any **fossils** that the rock contains. Similar textural changes produce **marble** from **limestone**, and **hornfels** from **mudrock**.

When rocks are metamorphosed by pressure as well as heat, they undergo **foliation**, in which **sheet silicates**, if they are present in the rock, rearrange themselves so that the sheets are at right angles to the direction of pressure. The picture to the right shows a view of foliation under a microscope.

Lineation is a similar phenomenon affecting silicates with the structure of a chain or double chain; the direction of the chain ends up, again, at right angles to the direction of pressure.



Foliation in schist, as seen under a microscope and illuminated by plane polarized light.

Not every metamorphic rock will display foliation or lineation: some rocks simply don't contain any sheet or chain silicates: an example would be **limestone**, which metamorphoses to marble. Also some metamorphic rocks are formed by heat without any significant pressure, as is usually the case with contact metamorphism; so, for example, mudrocks, which will form foliated slate or schist under pressure, will produce non-foliated hornfels without pressure.

Foliation comes in several varieties:

Slatey foliation is caused by the alignment of sheet silicates such as clay minerals and chlorite (which is produced by chemical changes to clay minerals). It results in rocks which cleave easily into thin layers.

Schistosity is caused by sheet silicates such as biotite and muscovite. Not only do they align, but they tend to separate out from the non-sheet silicates such as quartz, producing a rock that breaks easily into thicker leaves than those found in slatey rocks.



Gneiss.

Finally, we come to **gneiss**. At high grades of metamorphism, sheet silicates tend to break down, and dark-colored **chain silicates** such as **hornblende** and **pyroxene** begin to appear. These are separated out into dark bands, again at right-angles to the direction of pressure, giving gneiss a distinctive streaky appearance, as shown in the

photograph.

1.6.5 How do we know?

How can we recognize rocks as metamorphic?

First of all, as we have observed, we can reproduce metamorphic processes in the laboratory. Marble, for example, is what we get if we heat limestone; quartzite is what we get if we heat quartz sandstone; schist is what we get if we heat mudrock and apply pressure. It would seem downright perverse to maintain that metamorphic rocks should have been produced by other processes not as yet discovered. (Note that the textures of metamorphic rocks exclude the possibility that they are sedimentary rocks, and their chemical composition usually excludes the possibility that they are igneous rocks.)

We can also look at the patterns we find in the rocks. I shall give some examples of the kind of predictions we can make from the theory of metamorphism; the reader will doubtless be able to think of other examples.

To take a simple example: when we look at an aureole of marble, we should expect to find it embedded in an outer ring of limestone, and not of (for example) sandstone, which would go with an aureole of quartzite.

Then again, according to our notion that metamorphic rocks are indeed produced by metamorphosis, we should not (and we do not) find, looking horizontally at sequences of rocks, alternating bands of unaltered rocks and high grade metamorphic rocks. Instead, as we have noted above, we find concentric zones of rocks with high-grade metamorphic rocks at the center of metamorphism, progressing to lower and lower grades of metamorphism until we reach unaltered rocks.

If we find a foliated rock like schist, then according to our interpretation of schist as produced by temperature and pressure, we should find other evidence of the pressure; we should expect to find the beds of rock buckled and deformed. And this is indeed what we see.

If, on the other hand, we find hornfels, which, laboratory experiments show, requires temperature *without* significant pressure (otherwise it would be foliated) then we expect to find (and do) that it forms an aureole around igneous rock, with progressively lower grade metamorphism in concentric zones around the igneous rock.

Other patterns are discernible: for example, we would not expect to find schist overlying limestone, because the events that created the schist would also have turned the limestone into marble.

In summary, the chemical composition and texture of the rocks that we have classed as metamorphic, together with their arrangement and relation to other rocks in the geological record, is just what we should expect to see if they are indeed produced by metamorphosis.

Chapter 2

Erosion and deposition

2.1 Mechanical weathering and erosion



Bryce Canyon, Utah, a spectacular example of the effects of erosion.

In this article we shall present a brief overview of erosion and of mechanical weathering. We can be brief because the erosional processes involved will be discussed at length in subsequent articles, so there is no need to do more than sketch out the topic and its vocabulary.

2.1.1 Definitions

Geologists make a distinction between weathering and erosion: **weathering** breaks rocks but leaves them in place; whereas the processes of **erosion** are capable both of breaking rocks and transporting the broken fragments (clasts). Most, though not all texts will make this distinction. Those that do will still use “erosion” as a catch-all term: that is, when you see a geologist saying that a rock has been “eroded” s/he does not mean to imply that it has not also been weathered.

2.1.2 Mechanical weathering

Mechanical weathering is sometimes referred to as physical weathering. In both cases, the purpose of this nomenclature is to distinguish it from the processes collectively known as **chemical weathering**, in which chemical action

breaks down the rocks. Chemical weathering will be covered in the next article.

Mechanical weathering involves mechanical processes that break up a rock: for example, ice freezing and expanding in cracks in the rock; tree roots growing in similar cracks; expansion and contraction of rock in areas with high daytime and low nighttime temperatures; cracking of rocks in forest fires, and so forth.

Mechanical weathering is probably the least important process we shall mention in this text, in that the history of the Earth and the resulting geological record would probably have been very similar if there was no such thing as mechanical weathering.

2.1.3 Mechanical erosion

The main agents of mechanical erosion are: gravity; aeolian processes (i.e. those caused by the wind); ice in the form of glaciers; and water in the form of rivers, waves, turbidity currents, and runoff caused by rainfall.

The reader will be familiar with most of the processes described, but we should provide a brief introduction to the concept of **turbidity currents**. A current of water that is turbid (that is, which contains a lot of sediment) is denser than clear water, and will flow along the bottom of a lake or the ocean, often over large distances and at high speeds, before failing and dispersing its load; such currents occur when a turbid river discharges into the clear waters of a lake, or they can be initiated by a mudslide on a continental shelf. A dust storm may be considered the aeolian equivalent of a turbidity current.

2.1.4 Modes of erosion

Abrasion of rocks is caused by the sediments carried by wind and water: waves, for example, can hurl their **sealload** of sand and shingle against a cliff; sandstorms can literally sand-blast rocks; the sand and silt carried by rivers or turbidity currents have the same effect.

Attrition is the effect these same forces have on the sediments themselves, breaking them into smaller fragments or rounding the clasts into smooth pebbles or rounded

grains of sand. The efficiency of this process can be observed anywhere you can find beach glass, which originates as sharp-edged shards; the process of tumble-polishing semi-precious stones artificially emulates this process and will render most pebbles well-rounded in a matter of days.

The simple mechanical force of water or ice can break off chunks of rock, as when glaciers quarry rocks from the surfaces they move over, or when the pounding of waves hammers against a cliff.

Gravity can break off the overhang of a cliff undercut by abrasion and wave pounding, when the rock at the top of the cliff is unable to bear the mechanical strain. It is also instrumental in causing such things as rockslides and mudflows; such downhill motion is known collectively as **mass wasting**.

2.1.5 Transport of sediment

Erosional processes, as we have said, are defined by their ability to transport sediment as well as to create the clasts of which it is composed.

Currents of wind or water can transport sediments in three ways: in **suspension**, where light particles are carried along in the current above the ground, sea bed, or river bed; by **saltation**, where particles too heavy to be carried in suspension are bounced along the ground (or river-bed, or whatever); and **creep**, where particles are rolled along the ground. The size of the particles susceptible to these processes will depend, of course, on the velocity of the current.

By contrast, the size of the clasts that can be carried by glaciers is under no such limitation. One of the characteristic results of glacial action is the transport of huge boulders, up to the size of a house, known as erratic boulders. Gravity, too, is obviously under no such limitation; it is possible for entire layers of rock to slide down a hillside.

2.1.6 How do we know?

In this particular case, asking "how do we know?" seems almost superfluous, for the processes involved are neither hidden nor subtle. We can observe a sandstorm; we can see how the head of a waterfall shifts year on year, or how a river shifting its course scours out a new bed; we can see how cliffs crumble and the effects of landslides. The fact that glaciers carry boulders is evident, and the distance they travel each year can be measured; as can the quantity of sediment discharged at the mouth of a river.

A more interesting question is, how do we know that these processes have happened when the agency that caused them is no longer present? How, for example, do we identify the courses of glaciers long since melted, or of rivers that have dried up or shifted their beds? These are questions that we shall review in subsequent articles.

2.2 Chemical weathering

Chemical weathering is the breakdown of rocks into sediment by chemical processes.

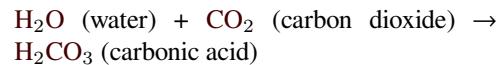
The reader who does not appreciate chemistry may skip the chemical formulas in this article and simply note the results of the reactions, as described in the summary section below. The results themselves cannot be skipped over: understanding chemical weathering is essential to answering such basic questions as: "Why is sand mostly made of quartz?" and "Where does clay come from?"

2.2.1 Agents of chemical weathering

The main agents of chemical weathering are:

Water. Some minerals, such as rock salt, will dissolve readily in water; others such as pyroxene will also do so, though at a much slower rate.

Carbonic acid. Rainwater and groundwater are not pure water; some of the molecules of water react with the carbon dioxide in the atmosphere (in the case of rainwater) or produced by bacteria and plant roots (in the case of groundwater) producing carbonic acid, as follows:



Oxygen. This is a highly reactive chemical, and the only reason that there's so much of it about in the atmosphere is its constant production by biological processes and a shortage of things that it hasn't reacted with already.

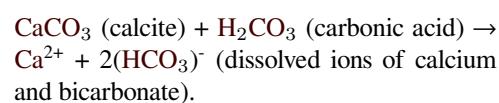
2.2.2 Chemical weathering of common minerals

In this section we shall look at how some common minerals are affected by chemical weathering. We have arranged the list more or less in order from the minerals most susceptible to chemical weathering to the most resistant.

Halite. Salt, of course, dissolves in water. This is why you are unlikely to see rock salt on the surface except in desert environments.

Gypsum. This, like halite, is soluble in water; similar remarks apply to it.

Calcite. This, you should recall from previous articles, is the mineral forming limestone and its metamorphic counterpart, marble. It can just dissolve in water; it also reacts with carbonic acid as follows:

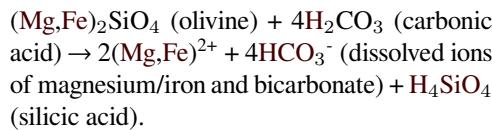


The ease with which limestone dissolves (relative, at any rate, to other minerals) produces the distinct topography of a region built on limestone rocks, with underground caves full of stalactites; sinkholes where the land has subsided; streams disappearing into the ground or rising out of it as springs. This is known as **karst topography**.

This is also the reason why a marble tombstone, though handsome in appearance, is not a good long-term investment.

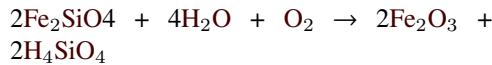
Silicate minerals. Of **silicate minerals** in general we may observe that the more mafic minerals with higher melting points (those to be found to the right of our diagram in the article on igneous rocks) are more susceptible to chemical weathering than felsic low-temperature minerals. This will be reflected in the order in which we list them below.

Olivine. This mafic mineral has the formula $(\text{Mg},\text{Fe})_2\text{SiO}_4$. Recall that the (Mg,Fe) in the formula means that it is a solid solution in which varying quantities of magnesium or iron can play the same chemical role. It reacts with carbonic acid as follows:



As with limestone, the constituent parts of the mineral are now entirely dissolved in water, leaving no residual mineral.

On the other hand, iron olivine can react with water and atmospheric oxygen like this:

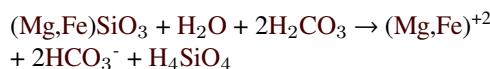


And the hematite can further react with water as follows:

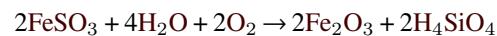


Hematite and goethite are both very insoluble in water: they remain as **residual minerals**. It is these iron oxides that give many soils their reddish or yellowish color.

Pyroxene. The typical rock-forming pyroxines have the formula $(\text{Mg},\text{Fe})\text{SiO}_3$. This can react with carbonic acid as follows:



Again, as with olivine, the constituent parts of the mineral are dissolved in water. However, as with olivine, iron pyroxene can react with oxygen and water to produce the residual mineral hematite:

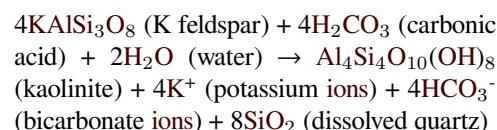


Again, the hematite can turn to goethite.

Mica and **amphibole**, minor constituents of felsic and intermediate rocks, undergo rather more complicated reactions. (Details may be found [here](#) for biotite, a mica, and [here](#) for amphibole.)

In summary, the residual minerals produced are clay minerals; iron oxides; and, in the case of biotite, the mineral gibbsite ($\text{Al}(\text{OH})_3$), which is usually found in association with clay.

Potassium and sodium feldspars produce residual clay minerals. Here, for example, is the reaction by which potassium feldspar produces kaolinite (plus various dissolved substances):



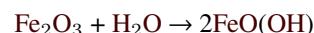
As sodium is chemically almost indistinguishable from potassium (see the article on [Chemistry for geologists](#) for further details) potassium feldspar reacts in a similar way.

Quartz is the most stable of the silicate minerals. This is why quartz sand is so common as a sediment; when all the other constituents of [igneous rocks](#) have either dissolved or been converted to clay, grains of quartz will remain. This is why there is no such thing as a large grain of sand: the maximum size of such grains is limited by the size of the quartz crystals that form in [granite](#) and similar rocks.

Note, however, that **quartz sandstone** is vulnerable to chemical weathering, because although the grains of quartz themselves are resistant, the minerals cementing them together may not be.

Clay minerals are very resistant to chemical weathering, because they are, as we have seen, a product of chemical weathering, and, like all minerals, they are stable under the conditions under which they were formed.

Iron oxides. These, as we have seen, are a product of chemical weathering of iron-bearing forms of such mafic minerals as olivine and pyroxene. We have noted that hematite can be converted to goethite by oxidation:



Once iron oxide has formed, there is very little that can happen to it except conversion to another sort of iron oxide; these are regarded as the most stable of all common classes of minerals.

2.2.3 Summary

The residual minerals left after chemical weathering has done its work are **quartz**, **clay minerals**, a scattering of

iron oxides, and sometimes a little gibbsite. The other constituents of minerals are dissolved; their usual fate is to be carried by rivers to the sea, where they contribute to the dissolved mineral content of seawater.

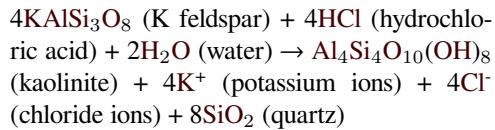
We may note that most land-formed sediments are in fact quartz sand, clay, or a mixture of the two. This demonstrates the predominance of chemical weathering over mechanical weathering and erosion. If sand or mud were produced simply by mechanical crushing of granite, then they would be 60% feldspar; but they are not. When we find any appreciable amount of feldspar in sand (as in arkose sandstone), we may infer that there has been a higher than usual ratio of mechanical to chemical processes.

2.2.4 How do we know?

The processes of chemical weathering are sufficiently slow that it is reasonable to wonder how we know that they take place at all.

In the first place, we know that they ought to take place. According to the theory of chemistry, in the chemical equations given above, under the conditions in which chemical weathering takes place, the situations described by the right-hand side of the equation are more stable than the situations described on the left hand side; so the reactions described *should* happen.

We can speed up reactions which, in nature, involve weak, highly-diluted carbonic acid, and instead use something stronger, such as hydrochloric acid (HCl). In principle the only difference this should make (besides, obviously, the substitution of chloride ions for bicarbonate ions) is that as hydrochloric acid gives up its hydrogen ions more readily, the reaction will go faster. So, for example, we can use hydrochloric acid to convert potassium feldspar to clay as follows:



And this happens fast enough for us to observe it happening. Alternatively, to experiment under more natural conditions, you can bury samples of minerals in a nice moist acidic soil, where chemical weathering occurs fastest, leave them for a few years, and see what's happened to them; they will not weather completely, but if the process involves conversion of one mineral to another, rather than just dissolution, then the chemical changes are observable on the surface of the mineral.

Without any of these artifices, we can find naturally occurring rocks which appear to be in the process of weathering: for example, the exterior of such mafic rocks as basalt can often be seen to be rusty, as a result of the iron pyroxene being converted to iron oxides. In the same

way, granite boulders can be found with a **weathering rind**, where on the outside of the boulder the feldspar on the outside has been partly converted to clay, while the feldspar on the inside is still relatively intact. In tropical soils, it is possible to find granite boulders in which the feldspar has become so "rotten" with clay that it is literally possible to kick the boulder to pieces. Under a microscope, the feldspar crystals will appear corroded and pitted. Or if a road cut or railroad cut goes through a hill (an event which always delights geologists) then since the rocks near the top are more weathered, we can take a series of samples going up through the vertical section from unaltered rock to completely weathered rock (**saprolite**) as in this study of the weathering of biotite.

We can observe the effect of weathering on old tombstones or on dressed stones used for building; as you would expect, this is most noticeable in the case of limestone or marble. In such cases we can usually be quite certain that the stones in question have suffered little from merely physical processes such as abrasion by sandstorms, tidal action, transport in rivers, and so forth.

Finally, we may note that the processes we have described explain the nature of sediment: they explain why so much of it is quartz sand and clay; they explain, as we have seen, the size of sand grains; they explain the origin of the minerals that cement together the grains in sandstone; and they explain the origins of the minerals found in sea water as dissolved ions.

In summary: these processes ought to happen; we can simulate them happening; and what we see in nature is just what we should see if they did happen.

2.3 Glaciers

A **glacier** is a mass of ice moving on the surface of the Earth. In this article, we shall discuss how glaciers form and move; we shall discuss the geological features associated with glaciers; and we shall show how recognizing these features can allow us to tell where glaciers have been in past ages of the Earth's history.

2.3.1 Formation and motion of glaciers

A glacier forms at an **accumulation point**, that is, a place where more snow accumulates than melts. This snow then piles up and compacts under its own weight to form ice.

Even if this happened on a perfectly level surface, as the ice mounted up it would eventually start to squidge outwards under the pressure of its own weight; and often glaciers will form on mountaintops, where gravity is also a factor. Under the effects of pressure and/or gravity, the ice will flow. A glacier flows in two ways: by sliding along its base, and by "plastic flow" of the molecules of ice within the glacier.



Satellite view of the Nimrod Glacier, Antarctica.

You may recall from science class that ice tends to melt under pressure; this means that the base of a glacier is often lubricated by water.

The overall speed of a glacier can be measured by simple methods: hammer a stake into a glacier, wait a while, come back, and see how far it's moved. The speeds so measured range from centimeters to meters per day, depending on the glacier.

One significant difference between the flow of ice and the flow of water is this: a river is pulled downwards by gravity. This happens to glaciers too, when flowing downhill; but glaciers are also pushed by the pressure behind them: as a result, glaciers can and do flow uphill.

Once in motion, the ice in the glacier will keep flowing until it reaches a point where the ice **ablates**: either it reaches the sea, breaking up into bergs, or it reaches a zone where the climate is warm enough to melt the advancing glacier. In the latter case, the end of such a glacier represents an equilibrium state at which the rate of melting is just sufficient to balance the rate of flow of the glacier.

Now, while this equilibrium is maintained the glacier as

a whole will stay still. The ice in the glacier will move, starting off at the accretion point and ending up at the ablation point, but the glacier as a whole stays in one place: it is like a conveyor belt of ice moving from accumulation to ablation.

The length of the glacier will change with the **climate**: for example, if the climate gets warmer around the ablation end of the glacier, then the glacier won't be able to progress as far before reaching a zone in which the rate of melting equals the rate of flow, so the glacier will retreat (note that the ice in the glacier will still be moving forward while this is going on). Conversely, of course, a drop in temperature will let the glacier get further from the accumulation point. Changes at the accumulation site of the glacier will also affect its length: the more it snows at the accumulation point, the greater the volume of flowing ice, and the further it will get before it melts. It follows that global cooling will cause glaciers to extend further from accumulation points, and global warming will see them extend less far, or vanish entirely if the temperature rises so much that the snow melts at the former point of accumulation.

2.3.2 Types of glacier

A glacier originating where the snow accumulates on a mountaintop and flows down the mountain is known as a **valley glacier** or an **alpine glacier**: the terms are synonymous.

The larger glaciers, such as can be found today covering the surface of Greenland and Antarctica, are known as **continental glaciers**, or **sheet glaciers** or **ice sheets**. Again, these are synonymous and the diversity of terms does not indicate that some sort of distinction is being drawn. Whereas valley glaciers flow downwards from the point of accumulation, the sheet glaciers of Greenland and Antarctica flow outwards in all directions from the point of accumulation.

2.3.3 Erosion associated with glaciers

A glacier moving over a landscape will sweep up topsoil and loose rocks as it goes, transporting them towards the ablation end of the glacier and revealing the **bedrock** beneath. This **bedrock** will be polished by the passage of the ice over it; it will also be grooved and scored in the direction of travel by the rocks contained in the glacier: such grooves are known as **striations** or **striae**, and the rocks are said to be **striated**. The picture to the right shows an example of striation:

The passage of a glacier will produce debris of all sizes, from enormous chunks of rock plucked from the **bedrock** to very fine **rock flour** produced by the grinding action of the glacier. A valley glacier will also carry along any rock fragments that fall from the valley walls. Some



Striation caused by a glacier.

glaciers, known reasonably enough as “rock glaciers”, consist mostly of rocks cemented together by ice.

One erosional feature commonly associated with glaciers is the **roche moutonée**, caused when a glacier slides over a hummock of rock. As the glacier slides up the hill, it polishes and striates it; flowing down the other side, it plucks fragments from the rock, leaving a steeper and more ragged face in its wake.

A valley glacier will create a bowl known as a **cirque** at the accumulation point, with about a quarter of the rim of the bowl missing in the direction in which the glacier leaves the cirque. A mountain eroded by glaciers will have a rugged, jagged topography, with knife-edge ridges where two cirques or glacial valleys adjoin. The valleys carved out by a valley glacier will have a characteristic U-shaped cross-section quite different from that produced by a river, which produces V-shaped valleys.

2.3.4 Deposition associated with glaciers

The rock fragments transported and deposited by glaciers are known as **till**. This is unrounded and unsorted by size. This may not sound very remarkable, until we reflect that the action of wind or water cannot produce sediments like this: such an unsorted jumble of shards is characteristic of glacial action.



Moraines, Glacier Bay National Park, Alaska.

In a valley glacier, there will be concentrations of till along the border of the glacier, where it has been plucked or ground from the valley sides. Such an accumulation of till is known as a **lateral moraine**. Where two valley glaciers meet, lateral moraines will merge into **medial moraines** in the middle of the larger glacier so formed, as shown in the photograph to the right.

Ground moraine is till deposited over a wide area either when the till on the underside of the glacier lodges against something, or when a glacier retreats in response to climatic changes. Ground moraine is often found in small hills, shaped somewhat like the back of a spoon, known as **drumlins**. No-one is really sure how these form, but the fact that they are composed of till and found in conjunction with other signs of glacial action confirm that they are glacial in origin; also, they are invariably found oriented with their long axis in the direction of the flow of the glacier (as determined by study of striations, roches moutonées, etc). At the ablation end of a glacier, the sediments transported by the glacier will be dumped to form an **end moraine**, resulting in a ridge of till in the same convex shape as the lobes typically found at the end of a glacier.

Beyond the ablation zone, where the glacier melts, the water from it will be carried away, typically in a **braided stream** (a term which will be explained in more depth in the article on rivers). This will carry with it the lighter sediments, known as **outwash**, which will be deposited in front of the glacier as an **outwash plain**.

One interesting sedimentary feature can be seen in lakes fed by glacial waters (**proglacial lakes**). In summer, relatively coarse outwash of sand and gravel will be deposited in these lakes; in winter, when the lake freezes over, the calm conditions below the ice allow fine particles of clay and of organic material to settle. The result is the formation of couplets of sedimentary material, one fine, one relatively coarse, repeated over and over, each couplet being known as a **varve**. Because the deposition of varves is an annual event, the study of varves is of interest in dating, as will be discussed in a later article.



Walden Pond, Massachusetts, a kettle lake made famous by the writings of Henry David Thoreau.

Kettles are another feature we associate with glaciers. When a glacier retreats, we often observe that it leaves a large block of ice behind it. The outwash sediment from the retreating glacier will then build up around the orphaned block of ice. When it has melted, which may take many years, the result is a depression in the outwash plain: this is a kettle. If it lies below the water table, it will fill up to produce a small lake with outwash banks.

2.3.5 Former glaciers: how do we know?

It is not difficult to detect the passage of a glacier, even if it is no longer present: for if you take away the glacier, you are still left with the patterns of deposition and erosion that glaciers produce; and these are highly distinctive and cannot be produced by other mechanisms. We may note that we can see glaciers disappearing today: for example, Glacier National Park in Montana has at the time of writing only 26 named glaciers, down from 150 in the year 1850; so our statements about what evidence former glaciers leave behind are by no means hypothetical, but rather are based on direct observation.

With all former glaciers, whether valley glaciers or continental glaciers, we see characteristic patterns of erosion: we see such things as **bedrock** polishing, striation, roches moutonées, and so forth. We also see till. As we have remarked, till is a very distinctive sort of sediment, which cannot be produced by the action of wind and water, as shown by its **unsorted**, **unrounded** nature. The arrangement of till can also be quite distinctive; a pile of till in a crescent-shaped end moraine admits of no other explanation except that a glacier deposited it there.

Erratic boulders, when we find them, present another blatant clue. An **erratic boulder** is one which, in terms of the rock of which it is composed, has nothing in common with the geology of its surroundings, and which must have been transported to its present location over some distance; in some cases, hundreds of kilometers. Ice, as we can observe, can transport such enormous rocks; water and wind do not.

We are therefore left with some unambiguous signs of former glaciers. When we see smoothed bedrock marked with striations leading to a semi-circular moraine of unrounded and unsorted rocks, many of them way out of their original geological context, beyond which is what looks suspiciously like an outwash plain, there is really no other conclusion that we can draw, except that we are looking at where a glacier once flowed and terminated.

Besides the erosional and depositional features already mentioned, glaciers leave behind some highly distinctive landforms. Consider for example the photograph to the right. There is no longer a glacier present, and, indeed, a lake has formed at the former accumulation point. (Such a lake is known as a **tarn**.)

Despite the absence of any actual ice, the reader should



Tarn, North Cascades National Park, Washington.

have no trouble in recognizing the landforms associated with a valley glacier, which are as clear and distinctive as an elephant's footprint. Here, surely, is the great bowl of a cirque, and just where we would expect to find it, near the peak where the temperatures are lowest; in the foreground, where the rim of the cirque is open, we see a valley with the distinctive U-shaped cross-section of a glacial valley, where the glacier once exited the cirque.

Sheet glaciers do not carve out the same forms, but they do leave some large-scale clues behind. They often sweep away soil and other sediment, leaving large expanses of bare rock. In doing so, they also erase the drainage systems that were present in the landscape before their arrival, so that after they retreat, the landscape is poorly drained: such features are a sign to geologists to look for other indications of glaciation.

We should mention a couple more signs of glaciation. The first of these is **isostatic rebound**. Ice is heavy, and for reasons that we shall go into more thoroughly in later articles, the weight of a continental glacier should press the **crust** of the Earth down into the **mantle**, and, when the glacier is gone, the Earth's crust should slowly "bounce" back up. This happens rapidly enough as to leave its marks over mere centuries: so, in parts of Scandinavia, we can see former harbors now standing uselessly distant from the sea. Today, the rate of rebound in post-glacial areas is measured directly by a **GPS** monitoring system called **BIFROST**: the maximum rate of rebound is about 1cm/year. This on its own would not prove the former presence of glaciers, but in combination with the less ambiguous signs of erosion and deposition, the phenomenon of rebound does confirm the hypothesis of glaciation.

There is one more prediction that we can make and confirm. We can use geological dating methods, of which more will be said in later articles, to establish the times at which sheet glaciers covered northern North America and Eurasia. Now, if we are correct in attributing these striations, moraines, and so forth to glacial action, then we ought to find that at the same time, we have other evidence of a colder climate, such as flora and fauna adapted

to colder climates; lower sea levels caused by water being locked up in continental glaciers; temperature-dependent changes in the composition of shells; and so forth. And this is exactly what we do find, providing an independent confirmation of an ice age.

We shall deal more fully with these topics in later articles on geological dating methods, on **paleoclimatology**, and on **ice ages**; for now we shall simply note that these techniques can be used, and that they confirm what we can learn from studying landforms, sediments, and erosional features.

2.4 Deserts



The Painted Desert, Arizona.

In this article we shall discuss the forces that shape deserts, and discuss how geologists can use this knowledge to identify rocks that are the lithified remains of former deserts. We shall use some terms that have been introduced in previous articles on **sedimentary rocks** and **mechanical weathering** and **erosion**; the reader who has not already read these articles may find it useful to go back and do so.

2.4.1 Definition of a desert

A **desert** is an area of low rainfall.

A desert by this definition is not necessarily hot: there are some areas of Antarctica that are considered deserts. Nor does a desert necessarily conform to our stereotype of being sandy: a sandy desert is known as an **erg**.

2.4.2 Causes of deserts

The immediate cause of a desert is, by definition, lack of rainfall. This itself can have a number of causes, which are not mutually exclusive: an area can be a desert for more than one reason.

- *High-pressure deserts.* In zones of high atmospheric pressure, the ability of air to contain moisture is increased, resulting in little rainfall. Examples include the Sahara, Arabian, Thar, and Kalahari deserts, and the desert regions within the Arctic and Antarctic circles.

- *Mid-continent deserts.* Areas in the middle of a continent can receive little rainfall simply because rain, originating from evaporation of seawater, will tend to fall before it can reach the middle of a large continent. Modern examples are the Turkmenistan, Gobi, and Great Australian deserts (the Great Australian Desert is also in a region of high pressure).

- *Rain-shadow deserts.* Rain will tend to break over mountains, so the presence of mountains can prevent rain from the sea from coming inland. Examples of rain-shadow deserts include the Mojave desert in the rain-shadow of the Sierra Nevada, the Patagonian desert in the rain-shadow of the Andes, and the Iranian desert in the rain-shadow of the Zagros mountains.

- *Upwelling deserts.* Finally, a desert may be by the coast and not in the rain-shadow of any mountains, but be adjacent to where a cold current of water rises to the ocean surface, reducing evaporation. Examples include the Atacama desert, the Western Sahara, and the Namib desert; these are all also in high-pressure zones.

2.4.3 Deserts and water

Although rainfall is rare in deserts, its effects are important in understanding the geology and ecology of deserts. It is characteristic of deserts that they exhibit **internal drainage**: that is, the water input to the desert in the form of rain or runoff from mountains does not flow out of the desert in the form of a river, but rather evaporates within the desert: this produces some highly typical depositional features.

Alluvial fans and bajadas. Deserts are often found associated with mountains: indeed, as we have seen, in some cases the mountains are the indirect cause of the desert. Runoff from rain on the mountains erodes them and transports the sediment downslope: when the sediment-rich water loses energy at the foot of the mountains, it forms a **braided stream** which rapidly deposits all but the lightest sediment in an **alluvial fan** (“alluvial” because it is formed by water, and “fan” because this is the shape in which it is deposited). Two or more adjacent alluvial fans can merge to form what is known as a **bajada**.

Playa lakes. These are temporary lakes, which, being shallow and fed only by intermittent rainfall, dry up in



The Grandstand, Death Valley.

the arid climate of the desert, leaving behind a flat, heat-cracked bed of clay, known as a **playa**. One such is pictured to the right.

If the water contains significant quantities of dissolved minerals these will be deposited on top of the clay layer as the water evaporates, leaving an evaporite bed of such minerals as **halite**, **calcite**, **gypsum**, **borax**, or **trona**, depending on the rocks that were their source. Repetition of this process builds up alternating layers of clastic and evaporite sediments.

Oases. **Oases** are small lakes in the desert. These have a number of causes: they can be fed by springs; they can be **deflation lakes**, where erosion has caused a hollow the bottom of which lies below the water table; or they can be intermittent lakes filled by occasional rainfall and runoff.

2.4.4 Deserts and wind

Deserts are the only places where wind is a major factor in erosion and deposition: **soil** bound by moisture and vegetation is harder for the wind to budge than loose, dry particles. The reader should recall from earlier articles that processes associated with the wind are known as **aeolian**; the same term is used to distinguish **sediment** transported by the wind, or **sedimentary rocks** formed from such sediment: so we can talk, for example, of **aeolian sandstone**: i.e. sandstone formed from sand heaped into **dunes** by the wind.

Wind has a number of effects. First, it can remove the sand and other light particles from the surface, leaving bare rock or the stony mosaic surface known as a **desert pavement** as shown in the photograph to the right.

Second, it can **erode** rocks by **abrading** them with the particles it carries.

Thirdly, it can pile sand up into dunes, giving us our stereotypical image of a desert: an **erg**. The dynamics and shapes of dunes vary depending on the regional winds; however, a typical dune is shaped something like a triangular prism with its long axis at right-angles to the direction of the wind. The wind transports the sand up



A desert pavement in the Mojave Desert.

the windward (**stoss**) side of the dune, to build up just over the crest of the downwind (**lee**) side of the dune. It accumulates until it reaches a critical angle at which it must avalanche, at which point the accumulated sand slides down the lee face of the dune to form a **grainflow lamina**.



Dry Fork Dome, Utah.

The result of the transport of grains from stoss to lee is that the dune will move downwind. However, it will not always erode completely on the stoss side, instead leaving behind it a set of cross-beds composed of the bottom of the grainflow laminae. Then the next sand dune to pass that way can deposit another discontinuous set of cross-beds on top of that set (and on top of any sand that may have been deposited by the wind between the passing of the two dunes). By this process, set after set of cross-beds build up: the **lithified** results are shown in the photograph to the right; similar patterns of deposition can be seen by taking a cross-section of modern dunes.

The cross-bedding is informative in a number of ways. First, it shows us that the sand was deposited in dunes.

Secondly, the gradient of a grainflow lamina tends to become shallower at the bottom: this feature can be seen in the cross-beds in the photograph. Now, tectonic events are quite capable of turning rocks sideways or upside-down, as will be discussed in later articles. The curves of the cross-beds are one of a number of features, known as **way-up structures**, that allow geologists to determine which way was originally up: the curves are concave in the up direction, convex in the down direction. In the

case shown in the photograph, we can see that the sandstone is still the right way up.

Thirdly, the crossbeds show us the prevalent direction of the winds, as we shall discuss later on in the article on paleocurrents.

2.4.5 Lithified deserts: how do we know?

There are a number of features that we can use to identify sedimentary rocks that are the **lithified** remains of ancient deserts. In this section we shall review some of them.

Desert sand is **quartz**, and the resulting sandstone will be **quartz sandstone**. This is not a definitive criterion for recognizing aeolian sandstone, since quartz sandstone can be formed under other conditions. We may, however, safely say that **arkose** or **graywacke** is *not* aeolian sandstone.

Desert sands tend to be **well-rounded**, as a result of long abrasion, and **well-sorted** by size, and this is what we find in the grains of aeolian sandstone.

Much, though not all, desert sand is colored, the grains being stained on the outside with **iron oxides** such as **hematite** and **goethite**, giving it a range of colors through red and orange to yellow. However, some desert sands are not stained in this way, so we cannot definitively say that sandstone lacking this feature is not desert sand.

Cross-bedded features reveal transport by a current of wind or water. We can distinguish between the two cases by various features: for example, wind-formed dunes typically result in much bigger sets of cross-beds; also, aeolian sandstone uniquely exhibits **pinstripe laminae**, very narrow stripes only a few grains thick consisting of finer grains than are found in grainflow laminae.

Playa deposits are extremely distinctive, and leave geologists no doubt that what they are looking at was once desert: they can only be formed by repeated episodes of deposition in and evaporation of a playa lake.

Finally, we can look at the **fossils**, which, in a desert should be of land plants and animals, and of freshwater organisms in those rare places where the geology indicates a former oasis. No marine fossils should be found.

These features, especially in combination, allow geologists to identify a lithified desert.

2.5 Volcanic ash

Volcanic ash is the name given to the fine particles of igneous rock produced by a volcano. As such, the name “volcanic ash” is a misnomer, since “ash” really means a residue left after incineration: the name is left over from times when people thought that volcanic ash really was ash.

In this article we shall discuss the origin, transport, and



A volcanic eruption as seen from space.

lithification of volcanic ash, and as usual we shall consider how we can recognize volcanic ash, and its lithified counterpart (tuff) in the geological record. It will be helpful for the reader to have already read the main article on **igneous rocks** before reading this article.

2.5.1 Origin and dispersal of volcanic ash

A volcanic eruption can throw off a fine spray of **lava** into the atmosphere as well as or instead of producing a cohesive **lava** flow. Because it originates as a fine spray, the droplets of lava often cool too rapidly to form a **crystal** structure; instead, they form an **amorphous** solid (a **glass**), although sometimes cooling may be slow enough for minerals to form. Sometimes the ejecta will also include particles of material torn by the eruption from the solid rock forming the volcano. Collectively, these fine air-borne particles are the constituents of volcanic ash.

These particles can be distributed in two different ways. If the volcano is sufficiently forceful, it can throw the volcanic ash high into the air, where it will be distributed by wind. Particles transported in this way can travel great distances: for example, ash from the eruption of Krakatoa was carried all the way to the island of Sumatra.

Alternatively, the ejected material can be transported in a **pyroclastic flow**. In this case, the ash-filled air is denser than the surrounding air because of its load of ash, and for this reason mixes poorly with the surrounding air, and so can travel over great distances (200 kilometers or more) before finally dissipating. The current can travel particularly far if it finds its way into a pre-existing channel such as a river valley.

2.5.2 Tuff

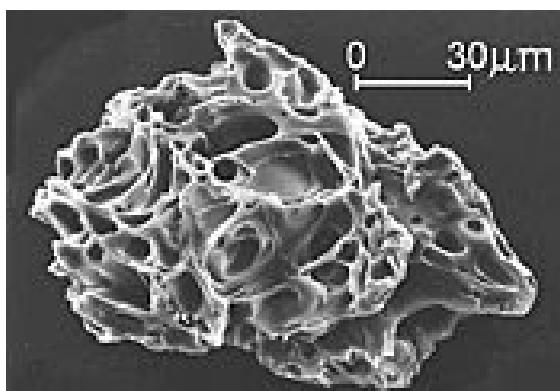
Lithified volcanic ash is known as **tuff**. This comes in two varieties.

Welded tuff has its origins when the beds of volcanic ash, when they form, are hot enough for the **clasts** to compact and weld together.

Cemented tuff is produced by the same cementing mechanism that cements together the **clasts** of more ordinary coarse-grained sedimentary rocks such as sandstone.

2.5.3 Volcanic ash: the igneous sediment

Volcanic ash is, obviously, an **igneous rock**: it comes out of volcanoes, and you can hardly get more igneous than that. As such, you might be surprised to find an article on volcanic ash dropped into a chapter which is otherwise about sediment and sedimentary rocks. However, in many ways volcanic ash behaves like a sediment. As geologists say, it's "igneous when it goes up, and sedimentary when it comes down."



A clast of volcanic ash

For one thing, volcanic ash consists of small **clasts**, just like other **sediments**; also, as we have seen, it can undergo cementation just like the process that cements together sand grains to make **sandstone**. The image to the right is a scanning electron microscope image of a single clast from the eruption of Mt. St. Helens.

Another feature that makes volcanic ash sedimentary in nature is that it can form **graded beds** like a sediment, with smaller particles on top. If the ash is shot high into the air, then the larger particles, the mass of which is better able to overcome air resistance, will settle first. If, on the other hand, the ash is transported by pyroclastic flow, then again the ash will still **grade** upwards from coarse to fine: as the current disperses and fails, the size of the particle that it can transport decreases, and so at any particular point in its path the particles that it deposits will decrease in size. (However occasionally, depending on the behavior of the volcano, the flow can also increase in speed, leading to **inverse grading**.)

Finally, ash beds resemble other sediments because they are a good place to find **fossils**. Inhaling volcanic ash tends to shred the lungs of animals, and if the ash bed is reasonably thick the bodies will undergo rapid burial, leaving them well-preserved.

However, volcanic ash is also **igneous**, and for this reason displays a number of features that make it of special

interest to geologists. A particular volcanic eruption is a single event at a particular point in time: when we see a horizontally continuous layer of volcanic ash, we know that it was all laid down at the same time, and whatever is immediately beneath it was a snapshot of the landscape as it was just before the eruption. Such a geological feature is known as an **event horizon**.

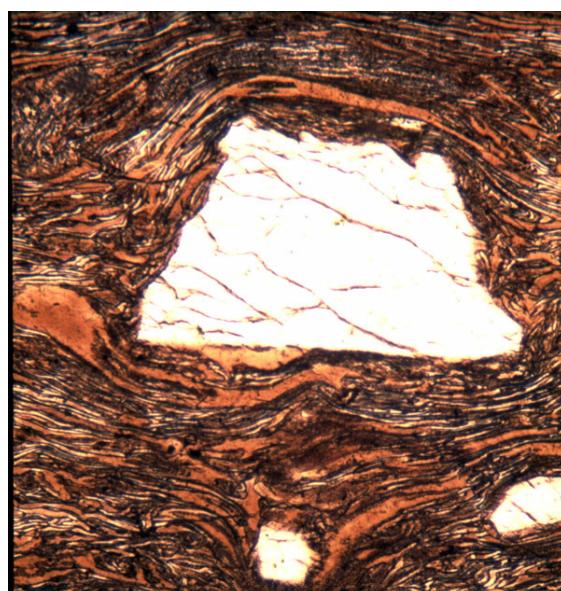
This is by no means true of sedimentary rocks in general. Just because you have a continuous horizontal **stratum** of, for example, sandstone, this does not necessarily mean that the beds at one end were laid down at the same time as the beds at the other end, as we shall discuss in a later article.

Furthermore, each volcanic eruption has its own chemical signature: the exact mix of chemicals in one volcanic eruption will be different from another. This means that even if a layer of volcanic ash is not exposed everywhere, so that we cannot directly observe that it is a single layer, we can still check whether two exposed areas of volcanic ash do or don't belong to the same eruptive event.

Finally, unlike most sedimentary rocks, volcanic ash and the rocks that form from it can be dated directly.

2.5.4 Volcanic ash and tuff: how do we know?

It takes no subtle chain of reasoning to know that volcanic ash or tuff found in the geological record is indeed volcanic ash or tuff. For when a volcano erupts, we can watch volcanic ash beds being formed: and the beds in the geological record look like that.



Photomicrograph of welded tuff.

This appearance is even more evident under the microscope. The photograph to the right shows a photomicrograph of welded tuff, where the **clasts** have been squashed

and welded together (the larger, white fragments are broken **crystals** included in the ash). Note the distinctive shape of the clasts: no-one looking at welded tuff under a microscope could mistake it for anything but welded tuff.

Finally, we may note that since ash and tuff are igneous in origin they have a chemical composition that lies on the usual felsic-ultramafic spectrum of igneous rocks: so we may speak of "rhyolite ash" or "basaltic tuff". So chemically there can be no doubt that volcanic ash and tuff are igneous in origin: but the fact that they are clastic means that we can't confuse them for any other sort of igneous rock.

These considerations mean that geologists can be confident of identifying volcanic ash and tuff when they find them in the geological record.

2.6 Soils and paleosols

In this article we shall discuss soils, their formation, and their preservation as fossil soils (paleosols). It will be useful for the reader to be familiar with the article on chemical weathering before reading further.

2.6.1 Soil: what is it?

The reader will of course have seen plenty of soil. But what exactly is it?

Sediments lying on the surface of the land will undergo chemical weathering as the result of rain falling on them. As this sediment is on the surface it will not have been compacted and lithified, so it will be highly porous as compared to a sedimentary rock, and the rainwater will easily be able to seep through it. This will cause the sort of changes in the composition of the sediment that you would expect if you have read the article on chemical weathering: feldspar minerals, for example, will be converted into clays; soluble minerals will be dissolved, and, depending on the climate and drainage of the local environment, can be deposited lower down in the soil; residual minerals will also be transported.

The other thing that will happen to such sediment is, of course, that plants will grow in it and organisms will live in it. This has various effects: first, it means that decaying organic material will be deposited in the upper layer of the sediment. Second, the actions of plant roots and of burrowing worms and so forth will keep the soil loose and porous. Third, the decaying plant material will release organic acids into the soil, which increases the rate of chemical weathering. In fact, a rock will actually undergo chemical weathering faster buried in soil than it will exposed to the open air.

These processes, known as soil formation or **pedogenesis** cause the sediment to become **soil**. The reader will notice from this discussion that a soil is defined rather differently

from the other **sediments** that we've discussed so far in this textbook. Other sediments are defined chiefly by their origin and mode of deposition. Soil, on the other hand, is defined by what happens to it *after* deposition: a soil is a sediment that has undergone pedogenesis, and this can happen to all sorts of sediments, from volcanic ash to glacial outwash.



Cross-section through soil of the type known as a podzol.

The photograph to the right shows you the results of pedogenic processes on one particular soil. You will observe a number of distinct layers (**horizons**) in the soil. On top, just under the grass cover, is a black horizon which gets its color by being rich in organic material. The white-ish horizon below that has had minerals leached out of it by chemical weathering of the soil. And the pinkish horizon below that takes its color from the **iron oxides** that have been deposited in it by the rainwater seeping through the soil.

This one example is just an illustration of the sort of thing that can happen in soil formation: the horizons will differ from soil type to soil type, and, indeed, identifying the different horizons is the first step towards identifying a soil type. We shall not here discuss all the different types of soil, because this would be something of a digression from the main thrust of this article.

Many determining factors combine to influence how a soil develops and so what type of soil it becomes. These include:

- The original sedimentary material. Obviously, for example, something that starts off with no iron-bearing minerals in it is not going to end up with a horizon rich in **iron oxides**.
- The climate. Chemical weathering acts much more vigorously in warm climates with plenty of rainfall.
- Drainage. For example, in waterlogged soil the decay of organic material is retarded by the fact of being waterlogged, causing the accumulation of such material.

- The vegetation type. For example, pine forests produce particularly acidic leaf litter, accelerating chemical weathering.
- Time. Since pedogenesis takes time, it is clearly going to be the case that sediments of recent origin will not be so well developed as older soils.
- Human activities, such as the addition of manure to fields.

2.6.2 Paleosols: definition

In geology, a paleosol is a fossilized soil. Note that this does not necessarily mean that it has been **lithified**, merely that it has been preserved by burial, perhaps by volcanic ash, or a **lava** flow, or **aolian** sediment, or whatever.

We should note that in pedology (soil science) the word paleosol has a different meaning: in that context it means a soil which developed under a set of conditions that are no longer present, for example a soil which develops under tropical conditions in a country that later acquires an arid climate. In the remainder of this article, we shall be using the word “paleosol” exclusively in the first sense.

2.6.3 Paleosols: how do we know?

As usual in this series of articles, we ask: how can geologists recognize paleosols when they see them?

The identification of rocks as paleosols is not really challenging. For one thing, paleosols, though necessarily buried, are not necessarily **lithified**. So some paleosols can be recognized as once having been soil because they still are soil. When they have been lithified, they often retain a superficially soily appearance. The photograph to the right, for example, shows a paleosol found near Mexican Hat, Utah: note its distinctly soil-ish appearance.

Whether or not a paleosol has been lithified, it will retain the mineralogical changes caused by pedogenesis, and this allows geologists not just to recognize the fact that the paleosol was once a soil, but also to identify the soil type, and so to come to conclusions about the **climate** at the time it formed.

Paleosols will also typically show distinctive signs of biological activity, such as animal burrows and casts, and roots or root casts (these are the white features in the photograph to the right): sometimes one can even find tree-stumps rooted in paleosols, leaving one in no doubt that they were once fertile soil. The exception to this is one that proves the rule: obviously we are not going to find such signs of biological activity in soil which was buried before life evolved to live in it: so when we find paleosols that are dated from the Cambrian (for example) we find no such signs of life, just as one would expect.



Paleosol, Mexican Hat, Utah.

The stratigraphy of paleosols is also consistent with geologists’ theories of their origin. It would be peculiar to the point of inexplicable to find that a paleosol had been buried by (for example) a distinctively deep-water marine sediment such as a **turbidite**, because what would soil, which forms on land, be doing getting itself buried under a **turbidite**? What we expect to find, and do find, is that the sorts of rocks that geologists identify as paleosols are found buried under **volcanic ash**, or **lava** flows, or **coal**, or **wind-borne** sediment, or sediment deposited by rivers; or such sediments as have undergone pedogenesis themselves and become soil.

2.7 Rivers

In this article we shall discuss the action of rivers on the landscape, and show how the characteristic sediments deposited by them can allow us to identify ancient river courses in the geological record. The subject of river deltas will be dealt with in a subsequent article.

The reader should note that there is no qualitative difference between a stream and a river; a stream is simply a small river, or, to put it another way, a river is a big stream. Rather than write “rivers or streams” over and



Horseshoe Bend, a meander of the Colorado River, Arizona, USA.

over again, we shall write about rivers, and the reader may assume that what we have to say applies on a smaller scale to streams.

Things associated with rivers (such as the **erosion** they cause or the **sediment** they deposit), are said to be **fluvial**.

2.7.1 Braided and meandering rivers

Rivers, unless artificially banked, rarely flow in completely straight lines. At low gradients, two characteristic forms they can take are braided and meandering rivers.

Braided rivers, as the name suggests, consist of a number of channels which separate and rejoin around bars of **sediment**. They are formed when a river with a lot of sediment repeatedly deposits the sediment and erodes it; so the braids and **bars** (i.e. the ridges of sediment breaking the surface of the water) are not permanent features, but shift around over time.

Meandering rivers wind from side to side in large loops (**meanders**). As a consequence of the hydrodynamics of this situation, the current is faster and causes more **erosion** on the outside of the loop, while **sediment** will tend to be deposited on the inside of the loop, forming **point bars**. The result of this is that a meandering river will become more meandering over time.

If this tendency goes far enough, the meander approaches a loop doubling back on itself. Eventually the meander may touch itself and the river will suddenly find itself with a new, straighter, shorter path, leaving the meander isolated as an **oxbow lake**.

The shifting of meandering and braided rivers across the landscape as a result of their own deposition of sediment produces a flattened, sediment-rich landscape known as a **flood plain**.

The photograph to the right shows the Rio Negro as seen from space: as you can see, the landscape is scarred with former meanders where the river used to flow before it shifted its bed. Note also the oxbow lake: its peculiar orange color is probably caused by salt-loving bacteria in



The floodplain of the Rio Negro, as seen from space.

the water, and should not be taken as being typical of oxbow lakes.

2.7.2 Sedimentary structures

The sedimentary structures formed by a river at a particular point will depend on its velocity, its depth, and the sediment type. Geologists can discover the relationship between these factors both by observing actual rivers, and by laboratory experiments using **flumes**, allowing them to control variables such as the velocity of flow and the size of the clasts.

For example, consider the effect that a river's velocity has on a bed consisting of average-sized sand grains. At low velocities, the **creep** of sand along the bed will, if anything, tend to smooth out the bed.

At higher velocities, sand **ripples** begin to form: small ridges of sand with the ridge at right-angles to the current. These ripples have a characteristic profile with a shallow slope on the upstream side and a steeper slope on the downstream side. Saltation bounces particles of sand up the shallow upstream side and over the peak of the ripple, eroding the upstream side and depositing sand on the downstream side. This has the effect that the ripples march downstream; it also produces cross-bedding.

At a higher velocity still, dunes (essentially, big ripples) will form; as with ripples, they have a shallow slope on the **stoss** side and a steeper slope on the **lee** side. Dunes formed in this way often have ripples on their shallow stoss side; these are known, logically enough, as **rippled dunes**. As with ripples, transport by **saltation** moves the dunes and ripples downstream (with the ripples moving rather faster than the dunes) and produces **cross-bedding**.

At greater velocities still, especially when the sand is fine, the increased current will flatten out the ripples, resulting in a flat, layered surface known as an **upper plane bed**.

At still higher velocities, **antidunes** form. These have a rounded undulating cross-section. While in dunes sand is eroded from the upstream face of the dune and deposited on its downstream face, in the case of *antidunes*,

sand is eroded from the downstream side of the antidune and deposited on the upstream side of the next antidune downstream. This has the effect that although the sand is moving downstream, the antidunes, being eroded on their downstream sides and built up on their upstream sides, move upstream: this is why they are called antidunes. These may show some slight **cross-bedding**, which, if it occurs, will slope up in the downstream direction; again, the opposite direction to that seen in ripples and dunes.

At greater velocities still, the current is strong enough to carry the sand in **suspension**, moving it downstream, leaving only **gravel**, **cobbles**, or just plain **bedrock**, depending on what other sediments, if any, are present on the river bed.

As I have indicated, the type of sediments involved affect these processes: in fine sediments, which “flow” more easily, dunes will not be formed; in coarser sediments, especially in shallower water, the formation of an upper plane bed is less likely, and the sequence as velocity increases will skip straight from dunes to antidunes.

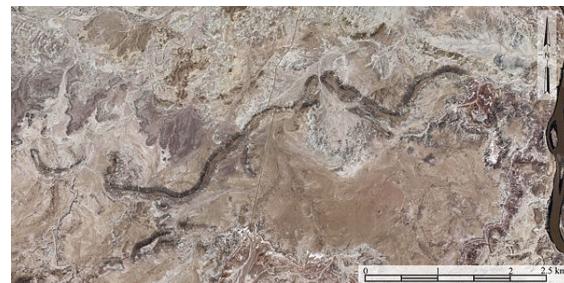
2.7.3 Vanished rivers: how do we know?

Geologists can reconstruct the courses of long-vanished rivers. The method by which they identify them should be obvious and familiar to anyone who has read this far in the textbook. If we take away a river, we are left with its sediments, which will eventually **lithify**. This will leave us with a set of rocks which look just like the lithified sediments of a river. As usual, we apply the rule that “if it looks like a duck and it quacks like a duck, it’s a duck”. In the case of rivers there are some very clear indications in the remaining sediment that allow us to identify what it once was.

In the first place, the sediments will be arranged in the long thin form of a river (what is sometimes called a **shoestring topography**). Note that since rivers shift, the “shoestring” will not necessarily be as narrow as the river was; but it will still be a shoestring. Depending on the depth of river and the rate of flow, the river bed will remain as a shoestring of gravel or cobbles or of duned or rippled sand. The ripples, of course, will cut perpendicular to the direction of the shoestring, making them distinct from beach ripples, which would be parallel to it. As the ripples in rivers are not upstream-downstream symmetric, it also is possible to use them to determine the direction of flow.

The photograph to the right shows a particularly nice example of shoestring topography. In this case, the **sedimentary rocks** formed from the sediment of a former river have proved more resistant to **erosion** than the rocks of its former floodplain, and so the former river stands clear as a ridge of sedimentary rock of different composition from that of the surrounding landscape.

Looking horizontally at the different sedimentary types



Remains of a river, Emery County, Utah

in a line cutting *across* the direction of flow, we will see coarse sand or gravel at the middle, then finer sand representing point bars, and then the mud of the flood plain. Because rivers gradually shift their course and their banks, we will also be able to see exactly the same sequence going vertically as over time river bed is replaced with point bar is replaced with flood plain. This is known as a **fining-up sequence**.

The types of sediment will be consistent with the hypothesis of a river. For example, we will not find gypsum or halite, because these would require totally different depositional environments.

Such fossils as we find will be of freshwater plants and animals, or of land plants and animals, but not marine forms. Similarly on the banks of the river we expect such fossils as are present to be of land animals or their footprints, and of land plants.

In short, we see exactly what we should expect to see as the remains of a former river, and are left with the reasonable conclusion that these features are, in fact, the results of the action of a river now vanished.

2.7.4 Note on superposed and antecedent rivers

In many places in the world, we can find rivers which have cut channels through hills or mountain ranges. This may seem odd at first, since naively one might think that since rivers can't flow uphill, they could never have cut the gorges through which they flow in the first place. However, the thing is quite practicable so long as the rivers were there before the mountains.

In the case of **antecedent rivers**, tectonic uplift slowly raises hills or mountains across the path of the river. So long as the river can erode away the uplifted rock and soil as fast as the rate of uplift, it will maintain its course. As rates of uplift are small compared to the erosional powers of rivers, this should present no problem to a river of reasonable velocity.

In the case of **superposed rivers** (also known as **superimposed rivers**) a river flows over a plain, subject to weathering and erosion, beneath which are upward folds (antisynclines) of rock more resistant to erosion than the

overlying rock. As erosional processes reveal the resistant antisynclines, the river cuts its way through them, resulting in a river that cuts through hills consisting of the exposed resistant antisynclines. An aerial view of the Susquehanna River, Pennsylvania, can be seen [here](#); the exposed ridges of resistant rock are clearly visible.

2.8 Deltas



Delta of the Rapa River, Sweden.

A **delta** is a fan of **braided streams** and **sediment** formed when a river discharges into a larger body of water: a sea or a lake. It will be useful if the reader has already read the article on **rivers** before reading further.

2.8.1 The dynamics of a delta

As a river discharges into a larger body of water, the current disperses and loses energy, and so the river dumps its sedimentary load: naturally, it will shed the heaviest sediments first, where the energy is highest, with progressively lighter sediments being carried further out into the sea or lake.

This means that the river will be constantly blocking up its own course with sediment, causing it to fan out into a web of **distributary streams** interspersed by **bars** of sediment. The process is not unlike the mechanisms which produce **braided streams** along the course of a river (as discussed in the previous article), and which produce **alluvial fans** as mountain streams disperse into a desert (as discussed in the article on **deserts**).

The appearance of a delta in profile will consist of flat **topset beds** of sediment, then **foreset beds** which slope down into the lake or sea; then horizontal **bottom-set** beds deposited on the floor of the sea or lake. Over

the course of time, the delta will build out (**prograde**) into the lake or sea, forming a characteristic sedimentary sandwich of coarser topset beds overlying sloping foreset beds overlying finer bottomset beds.

Because the **distributary** streams of a delta will be constantly dumping sediment in their own path, the pattern of streams and sedimentary **bars** will not be static, but will shift and change, producing a complex pattern of sedimentary deposition in the topset beds.

Indeed, given long enough, the whole course of a river may shift as the delta silts up. The mouth of Mississippi, for example, is known to have shifted several times over the course of the last few thousand years, and it is only by the unstinting efforts of the U.S. Corps of Engineers that the waters of the Mississippi still flow to the sea via the Mississippi River Delta.

2.8.2 Types of delta

Deltas may be categorized as freshwater or marine, depending on whether they discharge into a lake or a sea. Almost everything we have to say about deltas will apply equally to marine deltas and freshwater deltas. However, there is one notable difference in their dynamics. In a marine delta, the river will not be as salty as the sea into which it is discharging, and so the river water will be less dense than the seawater, and so will flow along the surface of the sea, mixing with the sea water in a horizontal layer, resulting in slower mixing and slower dissipation of the current than in a freshwater delta. The practical upshot of this is that the foreset beds of a freshwater delta will slope down at a much greater angle (up to 25° from the horizontal) whereas in a marine delta the foreset beds will have a slope of only a few degrees from horizontal.

Marine deltas may further be categorized by their dynamics as tide-dominated, wave-dominated, or stream-dominated, according to the main factor affecting their form.

- **Stream-dominated deltas**, such as the Mississippi River Delta, have long distributary channels extending seawards.
- **Tide-dominated deltas** have, offshore, long bars of sand parallel to the direction of the tide. Inshore, in the main body of the delta, they have tidal flats: beds of mud deposited by the action of the tide. These exhibit cross-bedding produced by the tidal currents: because tides flow in two directions, they will exhibit **herringbone cross-bedding**, a distinctive sedimentary pattern where alternate sets of cross-beds slope in opposite directions.
- In **wave-dominated deltas**, longshore drift (a current parallel to the shore) smears the deposited sediment across the face of the delta, so that instead of

the tidal bars found in wave-dominated deltas, we get a set of **barrier islands** at right-angles to the direction of the **distributary** streams.

2.8.3 Former deltas: how do we know?

Where a river has only recently shifted from disgorging via a delta, then it is perfectly obvious that the delta used to be a delta: not only does it look just like one in terms of its form and position, but also aerial photographs will reveal the former course of the river and its **distributaries**.

But what of ancient deltas that have been buried and then **lithified**? Well, in that case they will look just like lithified deltas. Perhaps only a fragment of such a delta will be exposed to our examination, and so the entire topography of the delta may not be visible, but we can still look at the patterns visible in the sediment.

So, looking through a vertical section of the rock, we should expect to see coarser horizontally-layered topset beds, with complex patterns of sedimentation caused by the shifting of streams and bars, overlying sloping forset beds, overlying finer horizontally-layered bottom-set beds. This is a very characteristic pattern of deposition produced by no other process.

Looked at horizontally, we expect to see a complex pattern of **interfingering** of land and marine sediments reflecting the complex ragged shape of the edge of a delta.

When we find **fossils**, they will reflect the nature of the beds in which they were deposited: so we expect to see land plants and animals in the topset beds, and the fossilized footprints of marine birds and suchlike fauna. In the bottom-set beds we may, to be sure, find a few remains of land plants and animals carried out to sea by the current: but we would expect the fossils in these beds to be dominated by aquatic fossils and by trace fossils such as the burrows of marine worms; we cannot, of course, find fossil footprints in these beds.

Further indications may be given according to the type of the delta: if, for example, it was tide-dominated, then in the topset beds we will find mudrocks displaying the herringbone cross-bedding characteristic of tidal flats, a phenomenon we shall discuss further in the article on nearshore sediments.

2.9 Lakes

In this article we shall look at the characteristics of lakes and how we can recognize former lakes in the geological record. The reader will find it useful to be familiar with the previous article, on **deltas**, before reading further.



Lake Misasako, Japan

2.9.1 Lakes

A **lake** is an inland body of water fed by rivers, streams, or sometimes by seepage of groundwater, as in deflation lakes in deserts. There is no universally agreed distinction between a lake and a pond, so you may as well think of a **pond** as being a small lake.

A large lake may be influenced by the tide, but even a large lake is small compared to the oceans, and the tides are correspondingly smaller; the highest tides in the Great Lakes of North America, for example, cause variations in water level no greater than five centimeters.

Most lakes are composed of fresh water; a salt-water lake can arise when water doesn't flow out of the lake (in which case it is known as a **terminal lake**) but is instead removed by the water evaporating, leaving behind the **minerals** that were **dissolved** in it: examples include the Great Salt Lake of Utah and the Caspian Sea.

The word **lacustrine** means “related to lakes”, and so the sediments associated with lakes are known as lacustrine sediments.

2.9.2 Lacustrine sedimentary rocks: how do we know?

If enough sedimentary rocks are exposed, then we can recognize a former lake by its shape: we'll find a bullseye of terrestrial sediments and fossils surrounding aquatic sediments and fossils, typically with coarser sediments nearer the shore and finersediments nearer the center.

If, however, only a part of the former lake and its shore are exposed, the **sediments** and **fossils** can still provide us with ample indications. If we can find a fraction of the shoreline showing a transition from a terrestrial to an aquatic environment, then that's a clue. But could we not be looking at a transition from land to sea? Well, since lakes are not tidal, or only barely tidal, the shoreline sediments will not show the same tidal effects that we shall discuss in the article on nearshore sediments.

As discussed in the article on deltas, the foreset beds of a lacustrine delta slope at a different angle to those of a marine delta, because the former are produced by fresh water flowing into fresh water. If we can find a delta with steeply sloping foreset beds, we know that the river was flowing into a freshwater lake.

Usually water will flow through a lake, in by one or more streams or rivers and out through others, but since a lake is so much broader than a river, the rate of flow in the lake itself will be small, and the sedimentary structures we associate with the flow of rivers will be small or absent. **Varves**, on the other hand (as mentioned in the article on **glaciers**) will often form in the relatively still waters of a lake.

When we look at the fossils in the sediment, aquatic fossils will usually be present; often so will some terrestrial fossils of organisms which got washed into a lake: leaves, for example can easily find their way into a lake, and are often preserved in the lake-bottom mud.

What's more, if, as usual, the lake is a freshwater lake, the fossils will not just be aquatic fossils, but freshwater aquatic fossils; when we find these, we're either looking at a lake or a river, and if the sedimentary evidence rules out a river, then we're looking at a lake.

2.9.3 Note on identifying freshwater organisms

The reader may wonder how we can identify freshwater organisms in the fossil record.

If they are recent organisms, then we can just recognize them. But what about extinct freshwater organisms? Well, if they are recently extinct, they will overlap with living freshwater organisms. Looking at somewhat older rocks, there will be some organisms that we don't recognize, but they will overlap in time and location with organisms we do recognize; if we can identify those as freshwater organisms, then the unfamiliar organisms that lived in the same place must also have been freshwater organisms. And we can continue this line of reasoning: when we find *those* organisms in company with still more types of fish or shellfish that we don't know, then they too must be freshwater organisms.

We can also look at the sedimentary environments in which organisms are found; **fluvial** sediments, for example, are very distinctive, and rivers are freshwater except near their mouths, so organisms found in **fluvial** sediments are freshwater, and if the same kinds of organisms are found in what on sedimentary grounds we would identify as a freshwater lake, then it's reasonable to conclude that it is a freshwater lake.

Then again, we can also use similar reasoning to identify marine organisms. As aquatic organisms are either marine or freshwater, those that aren't marine are freshwater by a process of elimination.

2.10 Peat and coal



Vegetable matter in a swamp, southwest France.

In this article we shall look at the deposition of peat in swamps and how it turns into coal.

2.10.1 Some definitions

In this article I shall use the term **swamp** as a catch-all term for an area of waterlogged ground in which the water is shallow enough for land plants to grow: an ecologist might distinguish more carefully between swamps, marshes, bogs, fens and so forth.

From our point of view, swamps become of interest when the swamp plants deposit plant matter faster than it can completely decay: in that case the partially decomposed plant matter known as **peat** will build up in the swamp. This matter will become **coal** on lithification.

The reader should note that there are two types of coal: **humic coal**, produced by the deposition of the remains of land plants in swamps; and the rarer and less economically important **sapropelic coal**, formed by the deposition of algae in lakes. The processes of formation are similar, but what we have to say in this article will refer specifically to humic coal.

2.10.2 Deposition of peat

In peat swamps organic matter accumulates faster than it can decay: this is what makes them peat swamps. But why? On the one hand, peat swamps deposit a lot of matter, but perhaps no more than is deposited in an ordinary forest in the form of fallen leaves, branches, and so forth. The crucial difference is that in swamps the deposited vegetation is waterlogged.

The oxygen content of air is about 20%, and this allows aerobic bacteria and fungi and suchlike organisms of decay to function. On the other hand, the oxygen level in oxygenated water can be more conveniently measured in parts per million, and this is the limiting factor in the decay of organic matter in water. The deposited organic

matter provides the aerobic bacteria with a potential feast, but in the process of metabolizing the available nutrients they must use up oxygen; and when there are so many of them, dining so heartily, that they are using up all the available oxygen, then that is as fast as they can decompose the matter.



Tollund Man, a bog body from the fourth century B.C.

There will be less oxygen available in the water surrounding more deeply buried plant matter, so that once the accumulation of such matter has started, it is likely to continue; in the same way, at a greater depth in the accumulated pile of sediment, the water will be more acidic, which also retards decay. The power of peat swamps to prevent decay is well-demonstrated by the discovery of well-preserved corpses thousands of years old in the peat bogs of Europe; for example the Danish "Tollund Man", dated to the fourth century B.C., shown in the photograph to the right.

Not all swamps will be peat swamps: this depends on factors such as the rate of deposition of vegetable matter and the rate at which oxygen comes into the system, which will vary according to the rate of flow (if any) of the water. We know of no general formula for determining whether a swamp will be a peat swamp: for our purposes, it is sufficient to note that in some swamps this build-up of plant matter does indeed take place.

2.10.3 Peatification and coalification

Peatification is the process of partial decay that we have described above. The action of bacteria destroys the weaker polymers making up the cell walls, such as cellulose, leaving behind mainly lignin, which is tougher. Because cellulose and lignin share a structural role in the cell walls of plants, the removal of the cellulose leaves the cell structures intact: a look at peat through an electron microscope **reveals** that even fine details of cell structure are preserved. The resulting matter is known as peat. The reader should note that this term does not refer only to

gardeners' peat, which is peatified sphagnum moss, but to any plant matter that has undergone the peatification process.



A coal seam.

Coalification is a chemical process in which hydrogen and oxygen are lost from the original peat fool, increasing the ratio of carbon to other elements. This involves alteration to the remaining molecules of the material, in particular the conversion of lignin to vitrinite. Coalification is not an all-or-nothing process: rather it produces coal of various **ranks** having a progressively greater proportion of carbon, from lignite through sub-bituminous coal through bituminous coal to the highest rank, anthracite.

In early coalification the process is carried out by bacterial action; when the material is so compacted that water cannot percolate (and so bacteria cannot penetrate) the later stages of coalification are produced by the action of heat and pressure (both of which are produced by sufficiently deep burial). In practice the processes distinguished as early and late coalification can overlap somewhat.

During the coalification process, pressure removes water from the material: peat has 95% water, anthracite less than 1%. At the same time, of course, the material is compressed, so that it may end up with as little as one-twentieth of its original volume.

Because the coalification process involves modification of the chemistry of the coal, all coal might, in a sense, be considered a **metamorphic rock**; however only anthracite is usually classified as such, since only anthracite approaches the temperatures and pressures that we usually associate with **metamorphism**.

As this text is intended to be read by readers with little knowledge of chemistry, we have treated the details of peatification and coalification as briefly as possible. The subject has been studied in considerable detail: the reader who is interested will find more information [here](#) and [here](#).

2.10.4 Coal from swamps: how do we know?

In the lowest grades of coal, cell structures are still visible under the microscope, revealing their plant origins clearly. As for the higher grades of coal, we should note that from a chemical point of view the various ranks of coal form a continuum: our divisions of coals into lignite, sub-bituminous, and so forth are, as usual in geology, artificial divisions of what is in fact continuous. This is illustrated well by localities where the upper coal beds are lignite and the lower coal beds progress through sub-bituminous to bituminous coal: which fits well with the theory that it is the same substance modified by the increasing heat and pressure associated with burial.

Furthermore, it is possible to simulate coalification in a laboratory. No-one has ever taken a piece of wood and turned it into anthracite coal, since one vital ingredient, that of time, must necessarily be lacking. But it is possible to show that the application of heat and pressure to peat will make it chemically more like lignite, and that similar treatment of lignite will make it more like sub-bituminous coal.

There is, then, no real doubt that coal has its origins as plant matter. We turn now to the question of why geologists ascribe the source of this plant matter to swamps.

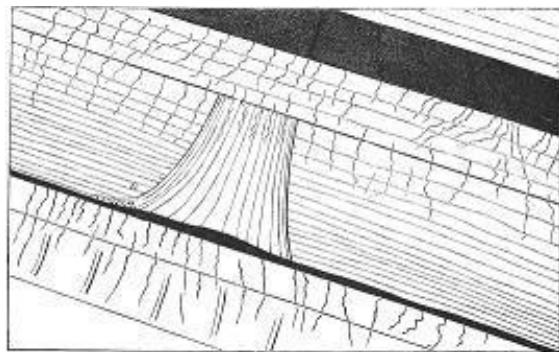
In the first place, coalfields are just what we would expect to see if peat deposits were buried to a sufficient depth, since, as we have observed, heat and pressure, which are both produced by deep burial, cause the chemical changes involved in coalification. Coal fields therefore look like peat deposits should look after sufficient heat, time, and pressure.

We can then ask ourselves: how else can such deposits form? To produce the extent and thickness of coal beds that we observe, we require a lot of plant matter to be deposited over a wide area in anoxic conditions, so that it doesn't rot. No other environment fits the bill. On a forest floor, for example, although plant material will be deposited over a long period of time and a wide area, it will be decomposed fairly rapidly, and never attains any great thickness, as can be easily verified with a trowel. When we see a coal seam ten meters thick, which is not unusual, and when we consider how much it has been compacted down (maybe ten or twenty times) from its original volume, we can see that the original plant matter must have been deposited in conditions where only partial decay took place: i.e., in a swamp.

We can imagine peatification taking place in other environments besides swamps: for example, we can imagine a landslide transporting trees down a hillside into a so-called "dead" lake. The trees then might conceivably peatify and, if buried deeply enough under other sediments, coalify. But once again, we find that this would not account for the great depth and lateral extent of coal beds.

Peat swamps therefore stand out as the one plausible explanation for coal. This is confirmed by examination of the beds of rock underlying and overlying coal beds.

Immediately underlying coal beds, we find **paleosols**, deposits which, as discussed in a previous article, geologists identify as fossilized soils: most obviously, because they have fossilized roots in them. Indeed, the **paleosols** will sometimes have trees or tree-roots rooted in them, projecting up through the coal beds: furthermore, the trees are consistent with the fossil vegetation found in the coal. This fits well with the swamp theory of the origins of coal. The **paleosols** underlying coal beds (which are known as **seat-earth**, or **underclay**) also show the sort of soil characteristics we should expect to find in waterlogged soils.



Coal seams and related strata.

The picture to the right, taken from Dawson's *Acadian Geology*, shows the relation of coal beds to paleosols.

The key from the original text identifies the strata from top to bottom as follows:

1. Shale.
2. Shaly coal, 1 foot.
3. Underclay with rootlets, 1 foot 2 inches.
4. Gray sandstone passing downwards into shale, 3 feet. Erect tree with *Stigmaria* roots (e) on the coal.
5. Coal, 1 inch.
6. Underclay with roots, 10 inches.
7. Gray sandstone, 1 foot 5 inches. *Stigmaria* rootlet continued from the bed above; erect *Calamites*.
8. Gray shale, with pyrites. Flattened plants.

The beds overlying coal beds are also consistent with the swamp theory: they are aqueous deposits, either freshwater or marine, depending on the location and nature of the swamp.

For these reasons we can be confident that humic coal does indeed have its origins in the deposition of plant matter in swamps.

2.11 Nearshore sediments



Marginal Way Beach, Maine.

In this article we shall discuss the sediments of the nearshore, their origin, characteristics, and sedimentary structures; and, as usual we shall discuss how we can identify sedimentary rocks as being lithified nearshore sediments.

2.11.1 Waves and the nearshore

We should first remind ourselves of certain facts about waves and tides. First, the reader should bear in mind that the energy of waves only goes down so far: the rule of thumb usually given is that their action affects the water beneath them to a depth about equal to half the wavelength (where the “wavelength” is the distance between two consecutive waves). Out at sea, the **wave base** (the lowest depth at which the wave has any effect) will be well above the sea-bed, and will move with a regular rolling motion known as swell. However, as waves come into shallower coastal waters, the wave base eventually hits the sea bed. This leaves the energy of the wave with no place to go but up. Therefore the wave rises higher and higher until it becomes unstable and breaks as surf. The consequence of this is that waves will have no effect on the marine sediments of the deep sea.

Tide-generated waves do not always move at right-angles to the beach: the result of this is to generate a **longshore current** which moves parallel to the beach. This is often confused with **longshore drift**, which is also caused by waves approaching the beach at an angle. As they wash up the beach, they travel at an angle, moving the sediment with them; but they tend to roll straight back down the beach, again taking the sediment with them. This means that the sediment on such a beach will be transported along the beach in a zigzag path.

2.11.2 A note on terminology

We can divide the sea and shore up into zones according to the action of the waves on the sea bed. Unfortunately, geologists do not do so consistently, so the same word may mean different things according to which geologist is using it. Indeed, I have seen one textbook in which the set of definitions supplied in the text contradicted the accompanying diagram intended to illustrate them. My advice to readers who wish to pursue their study of nearshore sediments further is that for each book they read they should pay careful attention to how each particular author defines his or her terms.

For the purposes of this article, I shall use the term **nearshore** to describe the zone in which the sea bed is affected by waves; the term **foreshore** to describe the part of the nearshore which is uncovered at low tide; and the term **backshore** to describe the area higher up the beach than the foreshore, i.e. that part of a beach which is above the high-water line. Other writers will differ, especially as to the proper meaning of the term “nearshore”.

2.11.3 Varieties of nearshore environment

Nearshore sedimentary environments are very variable in their nature. The sediments can consist of mud or sand or pebbles, or any combination of the three, which may be mixed or sorted according to the action of the waves.

These sediments can be deposited by rivers; they can have their origin in **clasts** broken from a rocky coast; they can be carried around the coast by longshore currents and longshore drift; they can have their origin as broken fragments of shells or **coral**.

The sedimentary structures will depend on such factors as the energy of the tide in that locality, the slope of the nearshore, and whether or not there is a significant longshore current.

In short, a whole book could be written on this subject enumerating the various characteristics of, for example, a high-energy muddy nearshore with a longshore current; and a low-energy nearshore consisting mainly of **sand** and **gravel** with no longshore current; and so forth.

Because of this, this article can only be a first sketch of the subject.

2.11.4 Some nearshore sedimentary structures

The reader should note that because of the variability discussed in the previous section, we would not expect to find all the structures listed below in one single nearshore environment.

- **Wave ripples.** These are superficially similar to the **ripples** produced in desert sand, but they tend

to be more symmetrical in cross-section, because they are formed by waves going in both directions (**oscillatory flow**).



Interference ripples.

- **Interference ripples.** These are formed when the tide goes out at, or nearly at, ninety degrees from the angle at which it comes in. The photograph to the right shows some interference ripples.
- **Herringbone cross-bedding.** This is one of the most distinctive types of coastal sediment. You should recall that **cross-bedding** is produced by the action of a current, and that the **beds** slope down in the direction of the current. But the current produced by a tide runs in two ways. The consequence of this is that the sets of cross-beds slope in opposite directions, producing a herringbone pattern.



Flaser deposits.

- **Flaser deposits.** Since the action of the tide is weaker at high and low tide than in between, sediments affected by the tide can in effect be in a high-energy environment and a low-energy environment alternately. This can result in a situation where, during the high-energy period of the tidal cycle, the waves shape **sand** into **ripples**, and during the low energy period of the cycle, the waves deposit **mud** in the depressions of the ripples. The resulting pattern of sediment is known as a **flaser deposit**. The photograph to the right shows a **lithified** example of flaser beds.

- **Deltas.** A river disgorging into the sea will often form a **delta**: a fan of shifting bars and channels. There is enough to say about these structures that we have already covered them in a separate article on **deltas**. The reader should recall that we can tell a marine **delta** from the case where a river is flowing into fresh water by measuring the slope of its **foreset beds**.

- **Longshore bars.** At the point on the shoreface where the landward force exerted by the wave base is equal to the seaward force of the backwash of breakers, sand will accumulate in longshore bars: that is, **bars** of sediment running parallel to the beach. On the seaward side these bars typically have layers of crossbeds sloping gently seaward; on the landward side they exhibit ripple-formed **laminae** and trough **cross-bedding** (that is, cross-bedding formed by waves repeatedly scouring out and filling in troughs in the bar). Broken shells are collected in the longshore bar by the same mechanism that accumulates the sand itself, and so longshore bars are typically abundant in layers of broken shells.
- **Sand dunes.** As the wind tends to blow off the sea, a sandy beach will have sand dunes pile up at the back of it. These will be similar to the wind-formed dunes found in deserts. However, unless the beach borders on a desert, there is no reason that it shouldn't get rain, and so beach dunes will tend to have plants growing in them, and often animals such as crustaceans burrowing in them; these will leave root traces and traces of burrows visible in the geological record, if the dunes are preserved. Another difference between these dunes and those of a sandy desert is of course that a desert will be spread across a large area; beach dunes may extend a long way *along* a coastline, but will form only a thin strip.

- **Bioturbation.** The **fauna** that burrows in the shoreface leave burrows of characteristic shapes. Indeed, the inhabitants of different zones of the shoreface leave different traces.
- **Pebbles.** The action of the waves and abrasion by other sediments will rapidly convert **gravel** to smooth pebbles. This can happen in other environments, such as rivers, but there is a tendency for nearshore pebbles to be flat on one axis.

2.11.5 Lithified nearshore sediment: how do we know?

We can identify **sedimentary rocks** that originated as nearshore sediments using a number of criteria.

First, many of the sedimentary structures that occur on the nearshore are unique: there is, for example, no other

way of producing flaser bedding. Others are very rare in any other context: for example, interference ripples can very occasionally be produced by the flooding of rivers, but they cannot be continuously produced except in a tidal environment.



Ripple marks in Moenkopi Formation rock.

The picture to the right, for example, shows a slab of sandstone exhibiting the ripples characteristic of a beach. Certainly it looks exactly like a lithified section of beach to anyone who has ever been to the seaside. The reasonable conclusion is that that's because that's exactly what it is.

Second, there is the fossil fauna of these sediments. The fauna of the nearshore is quite different both from the fauna of the land and the fauna of the deeper sea. For example, on the foreshore we find those creatures which can survive some exposure to the open air, when the tide is out, but cannot permanently survive such conditions.

Another consideration is the topography of the sediments. The coastline is a *line*, so if geologists are right in identifying certain kinds of rocks as **lithified nearshore sediments**, then we should expect to find these rocks in long thin strips running between rocks formed by the sorts of sediments we find on land and the sort of sediments we find further off shore: and this is indeed what we find, confirming the theory that these are nearshore sediments.

Furthermore, we should expect the different sorts of sedimentary structures and fossils to be arranged within each strip as they are in nearshore sediments. For example, if we find sedimentary rocks which we identify as back-shore **dunes**, they should form a still narrower strip on the landward side; on the seaward side we should then find the characteristic **fauna** and **bioturbation** of the foreshore, followed by those of the deeper nearshore.

Finally, there are chemical clues, as explained [here](#), for example. Coastal sediments have consistent chemical differences from those of deeper waters, and these dif-

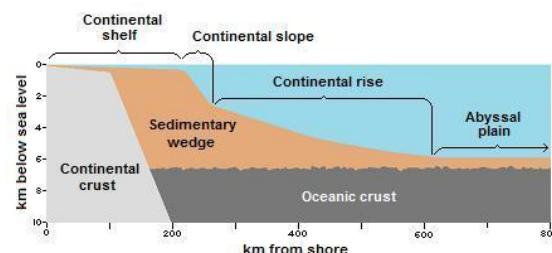
ferences are preserved in sedimentary rocks.

2.12 Marine sediments

This article contains a short general discussion of marine sediments and some important terms and concepts relating to this field.

We need not in this article ask and answer our usual question: "How do we know?" since in the case of the facts given in this article the answer is simply and uniformly: "We looked at the bottom of the sea."

2.12.1 The sea in profile



Profile of the sea.

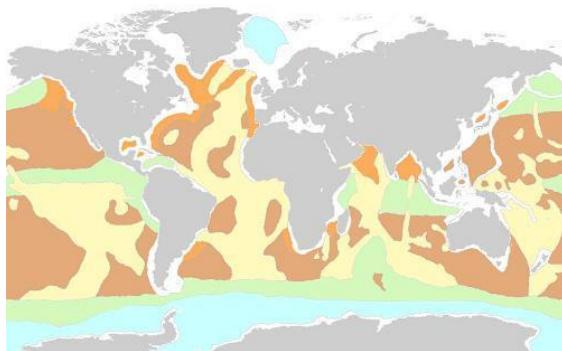
Before we discuss marine sediments, we should introduce a few terms from oceanography. The diagram to the right shows a cross-section of the sea and land. Though it has been simplified and stylized, it is based on actual data. Do note, however, that the vertical scale has been exaggerated by a factor of 25 compared to the horizontal scale, and that this consequently exaggerates all the angles of slope.

Features to note are:

- The **continental shelf**. This may be regarded as the part of the continent which just happens to be underwater. Its true angle of slope is rarely more than half a degree.
- The **continental slope**. As you can see, this is formed of land-derived sediment which has piled up at the foot of the continental crust. Its angle of slope rarely exceeds ten degrees, and is more typically around four degrees.
- The **continental rise**. This has an average inclination of about half a degree from vertical, and flattens out into ...
- The **abyssal plain**. As you can see from the diagram, this tends to be flat (hence the name), because all but the most prominent topographical features are obliterated by sediment.

The shelf, slope, and rise are known collectively as the **continental margin**.

2.12.2 Distribution of marine sediments



Map showing distribution of marine sediments.

The map to the right shows which marine sediments are deposited where. The key is as follows:

- Gray: land.
- White: Sediments of the continental margin.
- Blue: glacial sediments.
- Orange: land-formed sediments.
- Brown: pelagic clay.
- Green: siliceous sediments.
- Yellow: calcareous sediments.

The nature of these sediments will be discussed in more detail below.

There are a few ways in which this map may be misleading. First, note that this is a Mercator projection map, since, for technical reasons, this was the easiest map projection for me to use. All map projections distort reality, since they involve representing a spherical surface as a flat plane. In the case of the Mercator projection, its besetting fault is that it exaggerates north-south distances near the poles.

The second way in which the map does not truly represent reality is that it divides the sea bed into distinct regions of carbonates, siliceous ooze, pelagic clay, and so forth. In fact, the sediments will rarely be pure: the sediments that form pelagic clay, for example, get just about everywhere. So really every region represents a mixture of sediments, of which the color used on the map represents the one that predominates. This inaccuracy will, of course, be especially severe at the borders between regions, where one sediment type will grade into another.

Thirdly, note that this map represents sediments on top of the sea bed: they are not the same all the way down, as we shall discuss in the final section of this article.

2.12.3 Types of marine sediment

- Sediments of the continental margins. These include sand and mud from rivers; material eroded from cliffs; in some latitudes, material deposited by glaciers during ice ages; material carried down to the continental rise by turbidity currents; and, in some cases, calcareous ooze (see below).
- **Glacial sediments.** These are transported out to sea by icebergs calving from continental glaciers.
- Land-formed (or "**terrigenous**") sediments. Technically, both glacial sediments and the sediments of the continental margin can be considered land-formed. However, the areas marked as such on the map are regions where the abyssal plain has been covered by land-formed sediments either borne down to the abyssal plain by turbidity currents, or borne out to sea by river currents powerful enough to carry the sediment beyond the continental shelf.
- **Siliceous ooze.** This is formed from the silica shells of microscopic organisms: diatoms and radiolarians. These are common only in the most nutrient-rich and biologically productive parts of the ocean, such as the polar oceans and upwelling zones near the equator. If **lithified**, this would by definition be **chert**.
- **Calcareous ooze.** This is formed from the remains of tiny organisms such as foraminifera, coccolithophores, and pteropods. Its peculiar pattern of deposition can be explained by the fact that this sediment tends to dissolve in deep cold water. Chemically, it is composed of **calcium carbonate**; hence when **lithified** it would by definition be **limestone**.
- **Pelagic clay** (also called “red clay” or “brown clay”). This consists chiefly of particles fine and light enough to be borne out to sea by currents of wind or water. As such, you might expect these sediments to get pretty much everywhere, and you would be right. The areas on our map showing pelagic clay are not so much areas in which lots of this sediment are deposited as areas in which nothing much else is deposited.

2.12.4 Marine sediments and plate tectonics

We mentioned above that the sediments found in a given region of the sea floor are not the same all the way down.

This would be puzzling if the ocean bed stayed still: for in that case, why should not calcareous ooze (for example) be deposited on the same spot for ever?

But in light of the theory of plate tectonics, this observation will turn out later in this textbook to make perfect sense. Given that the sea floor moves, and has been moving for a long time, we should expect it to become layered as the same bit of continental crust passes under successive different regions of deposition. Indeed, given that we know how the sea floor is moving, and given a map such as the one presented above, we can say what pattern of layers we should expect to see: and this is confirmed by observation.

We shall discuss this point in more detail in the articles on plate tectonics.

2.13 Turbidites

Turbidites are sedimentary rocks caused by the lithification of turbidite sediments, that is, sediments deposited by turbidity currents. In this article we shall review what is known of their sedimentology, and discuss how we know their mode of deposition.

2.13.1 Turbidity currents

When a denser fluid flows through a lighter one, the difference in density prevents them from mixing, so that the denser fluid forms a current within the less dense fluid. In particular, turbidity currents in water are currents which are denser than the surrounding water as a consequence of being **turbid** (loaded with **sediment**). Because the turbidity current only mixes gradually with the surrounding water, its energy only dissipates very gradually into the larger body of water. This means that a turbidity current can flow for great distances (hundreds of kilometers) as a distinct current within the clearer water. Being denser than the surrounding water, it will flow downhill and along the bottom of the surrounding fluid: one might think of such a current as a sort of underwater river, although the analogy is not quite exact in that a turbidity current can flow up and over obstacles in its path.

The turbidity currents of interest to us in this article are those caused by slope failure, where sediment on the continental **slope** begins to slide down it, either as a result of a submarine earthquake or simply as a result of sediment accumulating on the slope until gravity alone is sufficient to start it sliding. This initiates a turbidity current, which flows down the slope accelerating as it goes: also, as it flows down the slope, it churns up more turbidity, increasing the difference in density between the current and the surrounding water.

By the time such a current reaches the ocean floor, it can be traveling at upwards of 100 km/h. As we have noted, the dynamics of a turbidity current ensure that it only loses energy very slowly, and so such a current can travel hundreds of kilometers before giving out.

Because these currents carry their loads of sediment at such high speeds, they must surely have a powerful erosional effect: they are thought to be the main cause of many underwater canyons. However, we are more concerned here with their role in the deposition of sediment, which will be discussed in the next section of this article.

2.13.2 Turbidity sediments and turbidites

The sediments deposited by turbidity currents are known as **turbidity sediments**. The rocks formed from these sediments on **lithification** are known as **turbidites**.

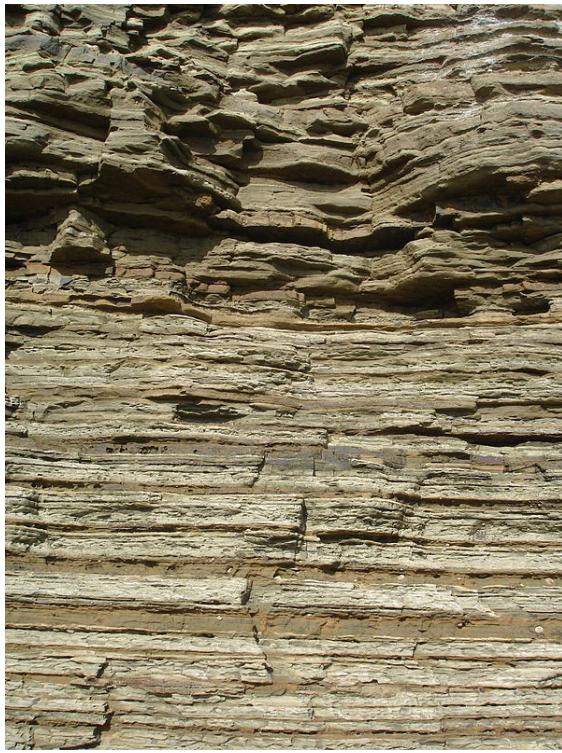
At any particular point over which a turbidity current passes, it will start off strong and gradually weaken until its energy is entirely dissipated. The consequence of this will be that the sediment will **grade** upwards from coarser to finer sediments. How coarse the sediment at the bottom is will depend on the source of the sediment: it may be as coarse as **boulders** and **cobbles**, or as fine as sand. The thickness of the deposit is also variable, from meters to centimeters in scale.

Note that the current fails not only over time, but also spatially, as it loses energy the further it gets from its origin. So at the extreme distance from the origin, only mud will be deposited; closer to the origin than that, we would see silt overlain by mud; and so forth.

After the deposition of the turbidity sediments, there will usually be a more tranquil regime of deposition, during which ordinary marine clay-sized articles will be deposited on top of the turbidity sediments proper. The entire sequence of sediments produced by these two mechanisms is known as a **Bouma sequence**. Note that although the top of the Bouma sequence is not deposited by turbidity currents, the term “turbidite” is used to include the whole Bouma sequence and not just the part of it so deposited.

While the ordinary marine clay in the Bouma sequence will contain organic remains from the deep waters in which they were deposited, the turbidity sediments will typically contain remains from the shallower waters in which they originated, and these remains will typically be fragmented by the violence of the process which transported them. The current-deposited sediments will often display sedimentary structures associated with flow, such as **ripple marks**. When a fresh sequence is deposited on top of the previous one, the force of the turbidity current will erode the layers of fine clay at the top of the previous sequence, producing what are known as **sole marks**.

The typical place to find a Bouma sequence is underneath one Bouma sequence and on top of another; although slope failures are intermittent, they are plentiful, and over a sufficiently long period of time great stacks of them will be deposited. The picture to the right shows part of one of these stacks, in **lithified** form.



Turbidites, Cabrillo Park, California.

2.13.3 Turbidites: how do we know?

How can we recognize the origin of the sediment in these rocks, and conclude that it really was deposited by turbidity currents?

To begin with, offshore drilling on the continental margin finds sequences of unlithified sediments which look just like the sequences of lithified sediment found on dry land. To identify the latter as the **lithified** counterpart of the former is trivial; and so we can be confident that the lithified sediments were marine in origin and were formed by the same processes as the marine sediments sampled from the sea floor.

But how do we know what those processes were? So far as I know, at the time of writing no-one has ever been at the right place at the right time to see a turbidity current depositing its load of sediment; this is unsurprising, since the phenomenon is intermittent and unpredictable, so no-one knows what the right time is; and the right *place* is at the bottom of the sea.

For this reason turbidites were for a long time a puzzle for geologists. But when they started taking turbidity currents into consideration, suddenly everything became clear.

Note first of all that turbidity currents themselves are not hypothetical. They can be produced in the laboratory in tanks of water and their action observed. Furthermore, laboratory experiments confirm that the waning of a turbidity current does indeed result in graded sediments, as

we would expect. Slope failures are also not hypothetical, and turbidity currents have been observed flowing down the continental slope through marine canyons; it is only the actual deposition of the sediments that has so far gone unrecorded.

We know that whatever leaves these sediments flows along the bottom of the sea, because it leaves ripple marks in the sediment and because it leaves sole marks gouged out of the previous layer of sediment. In order for something to flow at the bottom of the sea it has to be denser than seawater — like a turbidity current is by definition.

One frequently cited observation is the aftermath of the Grand Banks earthquake of 1929. In the hours following this, a number of transatlantic cables were severed. Their position was known, as were the exact times when they were cut. It is therefore possible to say that *something* capable of severing cables moved from near the epicenter of the earthquake at a speed of approximately 100 kilometers per hour, and that it moved along the sea floor where the cables were laid. A turbidity current with its abrasive load of sediment would be a highly plausible candidate.

We know that whatever process forms the deposits that we're trying to explain must be happening in the present, because we can see freshly deposited turbidite sediments in the present day. But we also know that the process must be intermittent, partly because we can't see any continuous process forming these deposits on the sea floor, and partly because the sedimentology shows the effects of a high-energy current waning to a low-energy current followed by a period of ordinary marine deposition, followed by the same thing happened over and over again. The turbidity currents generated by slope failure would fit this bill.

Moreover, we know of no other cause that could transport such large **clasts** so far out to sea. This may seem like a mere argument from ignorance, but it gains force when combined with the following argument. We know that there are failures of the **continental slope** causing currents which are by the nature of their origin **turbid**. Therefore, these currents must transport sediment and deposit it in some form. If it is not deposited in the form of turbidity sediments, in what form is it deposited and where is it?

The **fossils** found in turbidites are another important point. The alternation of shallow-water with deep-water fossils was once a baffling mystery. The theory of turbidity currents makes everything clear: the shallow-water fossils are carried by the turbidity current from shallow to deep water, and what was an inexplicable anomaly becomes an expected consequence of the theory.

Perhaps the closest anyone has got to direct observation of turbidite formation is the **events in Lake Brienz in 1996**. The lake showed distinct signs of an underwater landslip, including a sudden increase in the turbidity of the lake waters, a small (half-meter high) tsunami wave, and the release of a 200-year old corpse from the lake bed. Taking sediment cores from the lake revealed that an abnor-

mal layer of sediment, 90cm thick at its thickest part, had been laid down concurrent with this event: the sediment graded vertically upwards from sand through silt to clay: that is, it looked just like turbidity sediment should, apart from not being marine in nature. Further investigation suggested that the 1996 event was caused by accumulated sediment sliding down the slope of the Aare delta.

In the light of all these facts, it seems to be a safe bet that turbidity sediments are indeed caused by turbidity currents.

2.14 Reefs



A blue starfish rests on hard Acropora coral.

In this article we shall consider the formation of reefs, and discuss how we can recognize reefs in the geological record.

2.14.1 What is a reef?

A **reef**, to a geologist, is what you get when macroscopic organisms which secrete skeletal matter (hard **corals**, for example, or oysters) grow on top of one another, forming a mass of skeletal material in relief from the sea bed.

Since the skeletons of reef-forming organisms are invariably made of **calcium carbonate**, reefs are by definition **limestone**.

Note that this definition of “reef” is different from that which would be used by sailors, to whom a reef is any submerged hazard.

Note also that the definition does not just include coral, the main group of reef-producing organisms today, but rather includes any organism with this form of growth, including a number of extinct organisms, some of which we shall list below.

2.14.2 Reef-forming organisms

- **Archaeocyathids** are the first known reef-builders: goblet-shaped organisms typically a few centimeters in size which should probably be classed as calcite-secreting sponges. They flourished in the early Cambrian, began to decline by the middle Cambrian, and appear to have become completely extinct by the end of the Cambrian period.
- **Stromatoporoids** are a group of hard-bodied (i.e. calcium carbonate secreting) sponges which were important reef-formers from the Ordovician to the Silurian periods. While they are not extinct, they have been displaced to marginal habitats by later and more successful groups of reef-forming organisms.
- **Rudist bivalves** are a group of molluscs that flourished in the Jurassic and Cretaceous periods. During the Cretaceous they replaced **corals** in many environments as reef-builders, forming reefs sometimes a hundred meters high and hundreds of kilometers long, until they went extinct, like so many other organisms, at the Cretaceous-Tertiary boundary.
- **Corals**, despite their plantlike appearance, are animals closely related to sea anemones. Many of them secrete a skeleton of **calcium carbonate**, forming the main constituent of most modern reefs.
- **Oysters** are the **bivalves** familiar to gourmets and pearl-divers; they are capable of forming reefs, although not so large and spectacular as those formed by coral.

2.14.3 Ancient reefs: how do we know?



Fossilized tabulate coral.

It is easy to recognize a former reef in the geological record, since they are still clearly formed from the skeletal remains of coral, bivalves, stromatoporoids, etc. Their biological origin is therefore indisputable.

We might, however, if we were exceptionally cautious, ask ourselves whether the reef-shaped deposits of such remains are really reefs. Conceivably, they are piles of debris transported from elsewhere and then deposited in these formations as what we might call “pseudoreefs”.

Like everything else in geology, this question had to be thought about at one point. In the days of Leonardo da Vinci, it was suggested that the oyster reefs and corals discovered inland in his native Italy had been transported there by Noah’s Flood (da Vinci disagreed); and if no sensible person today entertains that conjecture, this is only because it has been considered and found to be wrong. For a number of objections occur:

(1) The positions in which the reef-forming organisms are found are the same as they would have in life. Now this is a powerful objection: there is no reason why any sort of transport or deposition forming pseudoreefs should have deposited the skeletal remains in their natural poses.

(2) There is no known mechanism by which water can pile up debris in the shape of a reef rather than spread it over a wider area (let alone give the transported sediment the configurations found in living organisms).

While the absence of a mechanism is not always a fatal blow to a hypothesis, it is certainly a point against it, especially as we have a perfectly good mechanism of reef formation to back up the theory that the things that look like reefs are in fact reefs.

(3) This hypothetical mechanism would have to be curiously selective in its action. We find apparent reefs built from oysters and corals, which we know from direct observation build reefs. We do not find “pseudoreefs” built out of (for example) the shells of crabs, or the cuttlebones of cephalopods. That is, we never find that this hypothetical mechanism has built reefs out of any organism known not to build reefs. This must cast doubt on the existence of such a mechanism.

2.14.4 A puzzle and a solution

Most varieties of coral are constrained by their biology to be shallow-water organisms, which cannot survive if they are more than a few meters below the surface of the sea. For this reason, living corals of this type are invariably found, not at the bottom of the ocean, but at the margins of continents or islands.

But what initially presents a paradox is this: sometime the islands on which shallow-water coral reefs grow are themselves made of (dead) shallow-water coral, and to a very great depth. For example, when the U.S. Atomic Energy Commission drilled at Eniwetok Atoll they found

1405 meters of shallow-water coral reef before striking basalt.

This seems, on the face of it, biologically impossible. These kinds of coral cannot grow at such depths, and therefore shouldn’t be there. This puzzle was solved by a young man named Charles Darwin, later to achieve greater eminence in the field of biology. His solution seems obvious in retrospect: the islands must have originally been either above the surface of the sea, or within a few meters of it, and must then have sunk beneath the sea at a rate slower than that at which coral can grow. This seems obvious now, as I say, but it was less obvious in Darwin’s time, when the idea that geological phenomena were produced by gradual changes over long periods was a new and revolutionary concept.

In our articles on plate tectonics (chapter 3 of this textbook) we shall explain why we would expect oceanic islands to sink gradually beneath the surface; for now we shall merely observe that by all appearances Darwin was correct.

2.15 Ooids and oolite



Ooids from a beach on Jouster’s Cay, Bahamas.

In this article we will discuss the formation of ooids, and how to recognize oolitic limestone in the geological record.

2.15.1 Ooids and oolite

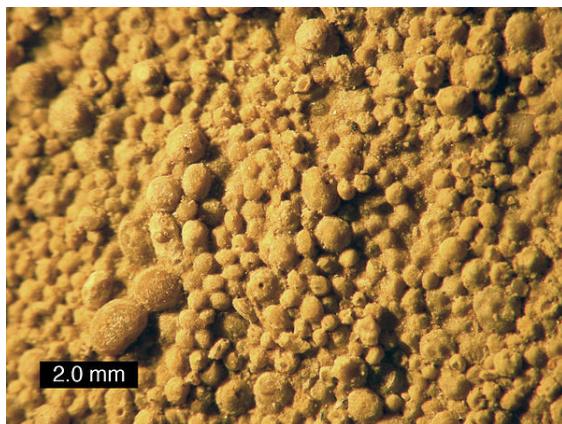
An **ooid** consists of a **nucleus** (a fragment of shell, a grain of sand, or whatever) around which layers of **minerals** are deposited to form roughly spherical grains.

These grains are typically between 0.25mm and 2mm in diameter; in fact, some authors use a different term for ooids of different sizes, but in this article we shall use the word “ooid” as a catch-all term.

Although there are a number of minerals which can form ooids, in this article we are interested in ooids formed

from calcium carbonate, and from now on we shall confine our discussion to them. Such ooids are typically formed in water rich in calcium carbonate (for obvious reasons) and for preference warm shallow water agitated by waves.

Today ooids are to be found in a number of locations with warm shallow water, including the Bahamas, Shark Bay in Australia, and the Arabian Gulf, all of which are marine sites; but they are also sometimes found in inland waters such as the Great Salt Lake in Utah.



Oolite, somewhat magnified.

Like other small grains, ooids can be cemented together to form a kind of rock. Rock formed from Calcium carbonate ooids is, by definition, a form of limestone, and is known as **oolitic limestone** or **oolite**. The term **oolith** may also be used as a term either for the rock or for an individual ooid.

2.15.2 Oolite: how do we know?

It is very easy to distinguish limestone formed from ooids. Magnified, it looks like it is made of ooids, as shown in the top picture to the right. (The limestone in this photograph and the one below is from the Carmel Formation in southern Utah.)



Photomicrograph of the grains composing oolite, in cross-section.

This resemblance might be fortuitous, but any such con-

jecture is laid to rest by looking through a microscope at a cross-section of oolite, as shown in the lower picture to the right.

Not only are the grains the right size and shape to be oolites, but in cross-section one can see the nuclei around which they formed and the growth patterns typical of oolites. Plainly, then, what we are looking at is ooids cemented together to form oolite.

2.16 Calcareous ooze



The Seven Sisters, Sussex, England: a fine example of chalk.

In this article we shall discuss the formation of calcareous ooze by carbonate-secreting organisms; the mechanisms which control the distribution of this ooze (such as the supply of nutrients and the carbonate compensation depth); and we shall look at how rocks are formed from the ooze and how we can recognize such rocks when we see them.

2.16.1 Calcareous ooze: what is it?

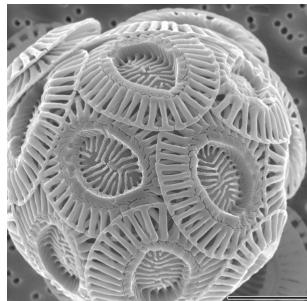
Calcareous ooze is a calcium carbonate mud formed from the hard parts (**tests**) of the bodies of free-floating organisms. Once this mud has been deposited, it can be converted into stone by processes of **compaction**, **cementation**, and **recrystallization**.

The main contributors to the ooze are coccolithophores and foraminifera. **Coccolithophores** are tiny single-celled organisms which cover themselves with tiny plates of calcite known as **coccoliths**. **Foraminifera** are also single-celled organisms. In some species this single cell will grow to be several centimeters in diameter, but most species of foraminifera are less than 1mm in diameter. While some will produce shells by gluing available sediment together, or by secreting shells from **silica** dissolved in seawater, most produce shells of **calcite**.

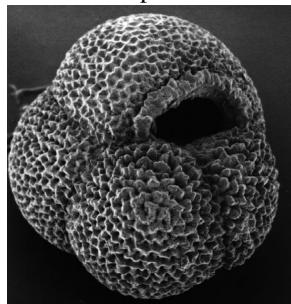
The deposition of foraminifera is generally more common today and in most regions; however, this varies from

place to place and from time to time; coccoliths, for example, are more common in the ooze on the floor of the Mediterranean, and are also more common in rocks dating from the early Tertiary period.

Excepting a few odd species of foraminifera, these organisms float or swim near the surface of the ocean. When they die, they sink. Perhaps “sink” is too strong a word: they are sufficiently small that they drift gently down like dustmotes through air, and can take months to hit bottom. The ooze composed of their hard parts accumulates at a rate of about 10mm - 50mm per thousand years, varying from location to location; which doesn't sound like much, but is actually a faster rate than other marine sediments such as siliceous ooze or pelagic clay.



A coccolithophore.



A foraminiferan.

The images to the right show, on the left, a photomicrograph of a coccolithophore, and on the right a photomicrograph of a foraminiferan.

Since calcareous ooze is formed from the hard parts of the bodies of free-floating organisms, this means that unlike **oids**, which are nearshore sediments, and unlike **reefs**, which require shallow water, calcareous ooze can be deposited over vast swathes of the deep ocean floor.

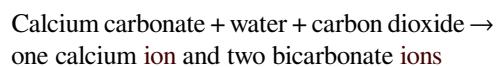
However, calcareous ooze will not accumulate in the very deepest parts of the ocean, even if the surface is teeming with the right sort of organism. The reason for this will be discussed in the next section of this article.

2.16.2 The CCD

Calcium carbonate will **dissolve** in the presence of carbon dioxide and water, as follows:



i.e.



Readers familiar with chemistry will not be surprised to learn that this is the reverse of the process by which calcium carbonate precipitates. The question of which will happen, dissolution or precipitation, depends on the relative abundance of calcium ions, bicarbonate ions, calcium carbonate, and carbon dioxide present. To cut the chemistry short, we may say that where carbon dioxide is scarce, precipitation will take place, and when it is abundant, calcium carbonate will dissolve.

Now, colder and deeper water contains more carbon dioxide than shallower and warmer water. The calcium carbonate compensation depth (or carbonate compensation depth, or CCD) is the depth at which the concentration of carbon dioxide is sufficiently high that calcium carbonate is dissolved faster than it can settle.

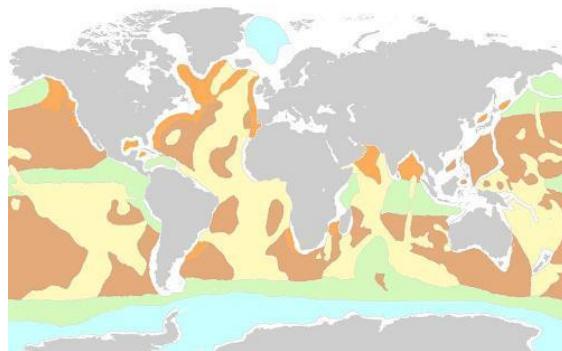
To speak of “the” CCD as though it was one specific depth in the ocean is rather misleading: there are other factors besides depth which affect this issue. First, there is temperature: cold water will hold more carbon dioxide than warm water, and so the CCD will be deeper in warm water. And secondly, there is the fertility of the water. For the reader should always bear in mind that the carbonate compensation depth is not the depth at which calcium carbonate dissolves; rather, it is the depth at which calcium carbonate dissolves faster than it is deposited. We should also note that as **aragonite** is more unstable than **calcite**, it dissolves rather more readily, so the type of calcium carbonate being deposited also plays a role, and we should properly distinguish between the **calcite** compensation depth and the **aragonite** compensation depth.

These caveats aside, we may say that the CCD is about 4500 meters down, give or take a few hundred meters either way.

The fact that this form of chemical weathering takes place has been confirmed experimentally, by scientists who took perfectly machined spheres of calcium carbonate and left them for a year at various depths on the ocean floor. Those in shallow waters showed no signs of weathering; those that were left in deeper waters were found to be pitted and corroded as a result.

The existence of the CCD helps to explain the rather curious pattern of deposition of calcareous ooze on the ocean floor, as seen in the map to the right, where areas where calcareous ooze predominates are marked in yellow. Three things are required for this to be the main sediment: first, there must be sufficient nutrients for calcite-forming organisms to flourish; second, the ocean floor must be above the carbonate compensation depth; third, there must *not* be the right conditions for other sediments to swamp the deposition of calcareous ooze.

There is one point that we should emphasize: calcium



Marine sediments: areas where calcareous ooze predominate are colored yellow.

carbonate below the CCD will not dissolve immediately, like an Alka-Selzer tablet fizzing away in water. The rate at which it dissolves is rather slow. It doesn't need to be fast, it just needs to be faster than the rate at which calcium carbonate is deposited. This point will be significant when we consider the evidence for plate tectonics in a later article.

2.16.3 Rocks from calcareous ooze: how do we know?

We should first sort out a small matter of vocabulary, **Chalk** might be defined as a stone which is, under a microscope, visibly composed of the tests of microorganisms. It differs from calcareous ooze itself by a degree of compaction and cementation that converts it from ooze to rock.

From the nature of its composition, it is by definition limestone. However, many experts on marine carbonate sediments will distinguish between chalk and what they are pleased to call “limestone”, by which they mean a rock which has undergone more extensive recrystallization so that its origin as tests has been largely or completely obscured. We shall continue to regard chalk as a form of limestone, but anyone who wishes to read further on the subject should be aware that the distinction may be made.

If we ask, then, how we know that marine limestone is formed from calcareous ooze, half the question is already answered: we know that chalk is formed from calcareous ooze because it is still visibly formed from tests.

In more completely recrystallized limestone, however, such visible tests may be few and far between. Are we really entitled to say in such cases that the parent material was chiefly calcareous ooze?

The answer is yes. First of all, consider the question of mechanism: we expect time and burial at depth to produce recrystallization in chalk; and we have no alternative mechanism that would explain the production of such limestone.

This is somewhat of a negative argument. A more pos-

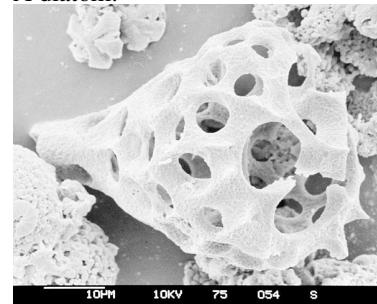
itive argument is produced by deep sea drilling. Geologists have taken core samples which progress from loose calcareous ooze at the top through stiff, compacted ooze, to chalk, with progressively greater dissolution, recrystallization and filling of pore spaces, to limestone in which “all detailed nanofossil morphology is lost near the base as sediment becomes almost totally recrystallized”. (See [here](#) for further details.)

We can therefore suppose either that the transition upwards from limestone to ooze represents a gradual change in the process of deposition, from a process as yet undiscovered to the observable deposition of tests; or that the process of deposition was the same throughout but that the more deeply buried sediments have been affected to a greater degree by known processes, namely compaction, dissolution, and recrystallization. The latter hypothesis, being more parsimonious, is clearly to be preferred.

2.17 Siliceous ooze



A diatom.



A radiolarian.

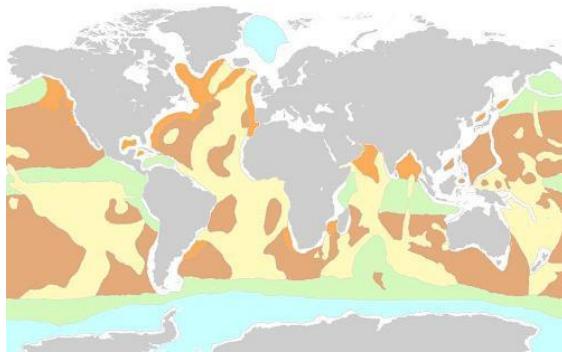
In this article we shall consider the origin, deposition and lithification of siliceous ooze.

2.17.1 Siliceous ooze

As with calcareous ooze, **siliceous ooze** is formed from the tests of microorganisms; in the case of siliceous ooze, the organisms come in two varieties, diatoms and radiolarians. In both cases the **silica** forming the tests is in the **amorphous** form known technically as **opal**. To be classed as siliceous ooze, sediment needs to be composed at least 30% of this material.

The pictures at the top of this article are photomicro-

graphs of, on the left, a diatom, and on the right, a radiolarian. Neither should be considered a typical representative of the type, since there is really no such thing: both diatoms and radiolarians exhibit a wide variety of forms.



Marine sediments: green areas represent siliceous ooze.

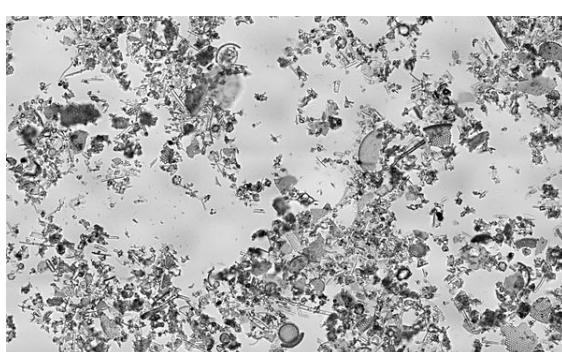
The map to the right shows the distribution of siliceous ooze in green. The tropical siliceous ooze is dominated by radiolarians; the bands at higher latitudes are dominated by diatoms.

It is estimated that the very small **tests** of these organisms would take 20 to 50 years to drift down to the sea floor; however, they can descend more rapidly in the form of the fecal pellets of the organisms which browse on diatoms and radiolarians. The rate of deposition of siliceous ooze is about 10mm per thousand years.

2.17.2 The opal compensation depth

As with calcareous ooze, there is a depth below which siliceous ooze will be dissolved faster than it can be deposited, known as the **opal compensation depth**. However, silica is more resistant to dissolution than **calcium carbonate**, and the depth is correspondingly deeper: approximately 6000 meters.

2.17.3 Rocks from siliceous ooze: how do we know?



Fragments of diatomite under a microscope.

In some cases it is very easy to relate siliceous rocks to siliceous ooze. **Diatomite**, for example, when viewed under a microscope, is quite clearly made of diatoms; it might be described as the **siliceous** equivalent of **chalk**. The photomicrograph to the right shows crumbled fragments of diatomite: it is unmistakably composed of diatoms.

Marine **chert**, however, is at first glance more enigmatic: its structure consists of very fine **crystals of silica**, and hints of its biological origin have largely been lost, raising the question of how we know that it did in fact have a biological origin. The question (and indeed the answer) is similar to the question we raised about marine **limestone** in the previous article. In this section we shall briefly review the evidence suggesting that the silica of which marine chert is composed has its origins as siliceous ooze.

First of all, note that chert is just what we would expect to get if time, pressure, dissolution and reprecipitation caused **recrystallization** of siliceous ooze. Marine chert requires a source of silica; siliceous ooze provides a source of silica. Even if we had no other relevant evidence, we should preferably ascribe the origin of bedded chert to siliceous ooze rather than hypothesizing some other origin for the silica as yet undiscovered.

More direct clues are revealed by the circumstances under which chert is found. Taking samples of rock from the same site at varying depths (an example is given [here](#) on p. 575) then we may for instance find layers of **clay**; of clay rich in radiolarians; of radiolarians filled and cemented with silica; and of true chert. Now it would take a stretch of the imagination to suppose that the silica composing the chert has quite a different origin from the silica in the other layers; especially as there is no good hypothesis as to what that origin could be.

Furthermore, in some cases when chert reveals no visible organic structures, treatment of the chert with hydrofluoric acid reveals radiolarian structures in the chert, presumably because the silica of the **matrix** has a rather more soluble structure than that of the radiolarians (see [here](#) for further details). Now, it would be possible to argue that these radiolarians are “**incidental**”: that they just happened to get buried in some siliceous sediment other than siliceous ooze. However, in the light of the other considerations we have mentioned, this seems unlikely.

Some have argued that there must be at least some alternative origin for chert, on the grounds that Precambrian cherts are known, and these precede the evolution of radiolarians and diatoms. However, study of these Precambrian cherts reveals microscopic spherical ornamented structures (see [here](#) for further details). If some of these are the shells of extinct silica-secreting organisms then we might think it probable that the Precambrian cherts were formed from siliceous ooze composed of these tests.

On the other hand, we should note that if there were no organisms in the Precambrian seas that secreted silica,

then the oceans would have had a much higher concentration of silica than they do today, and perhaps under these conditions non-biological deposition of silica might have been possible by processes which could not operate in modern conditions.

However this may be, there seems little to disturb the consensus that more recent cherts have a biological origin. This is not to say that the topic of chert is without its controversies, but these fall outside the scope of an introductory article such as this one.

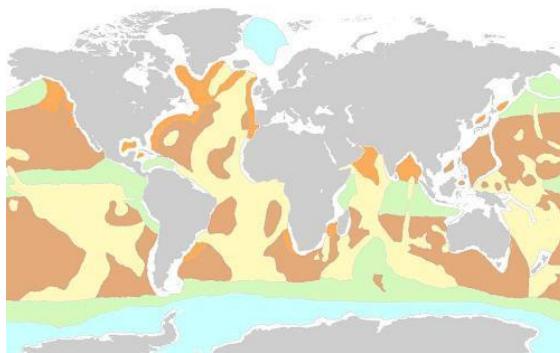
2.18 Pelagic clay

Pelagic clay, also known as **red clay** and **brown clay**, is a fine sediment found in some parts of the abyssal plain. In this article we shall discuss its origin, its distribution, and its **lithification**.

2.18.1 Origin and distribution of pelagic clay

The bulk of pelagic clay has its origin as very fine sediment (which is therefore **clay** in the sense of having small grain size) wind-blown from the land, chiefly fine particles of **quartz** and of **clay** in the mineralogical sense; volcanic ash also contributes, and even **cosmic dust**, i.e. micrometeorites fallen from outer space.

We should not be surprised to find sediment from the land being deposited in the middle of the ocean, since the fine dust raised by duststorms can be carried right across the ocean from continent to continent; dust storms in the Sahara, for example, can trigger smog warnings in Florida, and are also the main source of mineral nutrients for the Amazon Basin.



Marine sediments: brown areas represent pelagic clay.

This dust is spread liberally across the surface of the ocean, but as you can see from the map to the right, not all of the ocean bed is covered with pelagic clay (represented by the brown areas on the map).

This is not because the clay is not deposited in other regions, but because in other areas other sediments

are also deposited to the extent that the resulting sediment is considered to be **nearshore**, **siliceous**, **calcareous**, **terrigenous**, **glacial**, or **turbidite** in origin. Pelagic clay is therefore the “none-of-the-above” sediment; it is what you get where there are insufficient nutrients for **coccolithophores** and **foraminiferans**, or the sea floor is below the **CCD**; and where there are insufficient nutrients for **diatoms** and **radiolarians**, or the sea floor is below the **OCD**; and which is sufficiently far from the **ablation** of glaciers into the sea that **glacial** sediment is not deposited; and ... et cetera, et cetera.

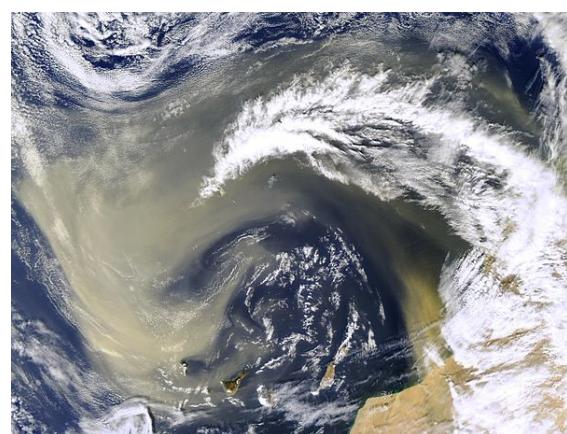
Pelagic clay is the most slowly deposited of all sediments, typically accumulating at less than 10mm per thousand years; hence it is easily swamped by other sources of sediment if these are present.

2.18.2 Appearance

The color of pelagic clay varies from place to place, depending on the source of the dust that composes it. Rich browns or reds are common, these colors being produced by iron-based minerals.

In texture it is of course fine-grained, since only the finest and lightest materials can be wind-borne so far out to sea.

As it is compacted by burial, the flakes of which the **clay** is composed reorient themselves to lie flat and parallel to one another.



Satellite image of a dust storm blowing off the coast of Africa.

Compared with other muds, pelagic clay is remarkable for having very little in the way of sedimentary structure. Because of the extremely gentle mode of deposition, pelagic clay will exhibit no **ripple** marks, **sole** marks, **cross-bedding**, **grading** by size, or other sedimentary structures; the most we may see are the faintest suggestions of extremely fine **laminae**.

2.18.3 Lithified pelagic clay: how do we know?

As with other fine-grained sediments, the **lithification** of pelagic clay is caused chiefly by **compaction**.

The resulting rock can be recognized as having its origin as pelagic clay by virtue of its mineral composition, its extremely fine-grained and dense **texture**, and by its lack of sedimentary structure.

Furthermore, pelagic claystone will be rich in **ichthyoliths**: disjointed parts of fish, usually teeth and scales, testifying to its marine origin. It may also contain the occasional **siliceous** or **calcareous** test; or in some cases it will be interbedded between layers of **limestone** or **chert**, serving as another indication of its origin.

2.19 Deposition rates



A sediment trap.

In previous articles we have mentioned the sedimentation rates for various sediments. Perhaps it is time to ask ...

2.19.1 How do we know?

There are a number of ways we can find out the present, or at least recent, rate of sedimentary deposition.

One of the simplest involves what is known as a **sediment trap**, which is used to measure deposition of marine sediment. One can be seen being recovered from the ocean in the photograph to the right.

Fundamentally, it is a technologically sophisticated bucket. One aspect of its sophistication is that it has a whole set of collection bottles at the base which successively rotate into the collection position at fixed intervals. This allows geologists to measure seasonal variations in the quantity and composition of sediments.

Sediment traps are particularly useful when dealing with very fine sediments with a very low rate of deposition such as siliceous ooze, calcareous ooze, and pelagic clay.

Another method, suitable for when sedimentation rates is higher, is to take a drilling sample in which some layer corresponds to a recent event. For example, the first tests of hydrogen bombs in the 1950s are marked in the sedimentary record by the sudden appearance of cesium-137, an isotope of cesium not found in nature.

Another such time marker consists of the peak in environmental lead that occurs in 1970. Before that point, lead in sediments rose with the use of petroleum; in 1970, the U.S. Congress passed the Clean Air Act, and the lead found in sediments begins to decline.

Other methods involve a rather ingenious use of naturally occurring radioactive isotopes which are constantly being deposited on the sea floor; we shall describe these methods in detail in the article on the U-Th, U-Pa, and Ra-Pb methods of dating.

When we look at sedimentary rocks, and try to figure out their rates of deposition, we may also appeal to more conventional methods of radiometric dating. If we have (for example) a layer of pelagic claystone, itself undatable directly, sandwiched between two layers of **igneous** rock, which is datable by radiometric methods, then by subtracting the date of the lower layer from the date of the upper layer we get a period for the deposition of the claystone in between, and so, given the thickness of the claystone, we can figure out its average rate of deposition over that period. If we wish, we can then use our knowledge of how much more compact pelagic claystone is than the parent sediment to produce a sedimentation rate expressed in millimeters of the original sedimentary material per thousand years.

There is quantitative agreement between the rates of deposition measured in sediments today and the rate of deposition inferred for the corresponding sedimentary rocks; for example the deposition rates calculated for the calcareous material in marine **limestone** are the same as the deposition rates measured for **calcareous ooze**; and similar remarks apply to other rocks and their corre-

sponding sediments.

This agreement is confirmation, if any is still needed, that geologists are correct in their diagnosis of the mode of deposition of the parent sediments of these rocks.

2.20 Glacial marine sediment



An iceberg.

In this article we shall discuss the deposition of glacial till in a marine environment. Readers may find it useful to refresh their memory of the relevant terminology by looking back at the [article on glaciers](#).

2.20.1 Deposition of glacial marine sediment

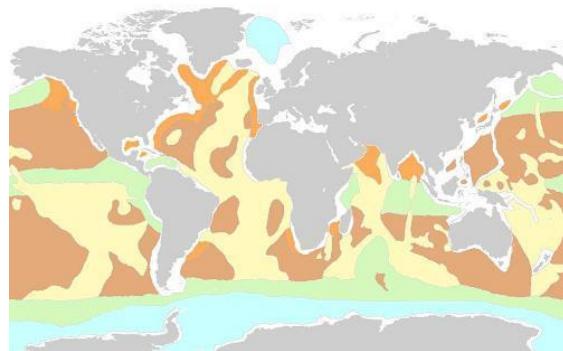
The mechanism by which glaciers form and move has already been discussed in the [main article on glaciers](#).

In that article we discussed the sediment deposited by the [ablation of glaciers](#) on land. Glacial marine sediment is formed when instead glaciers ablate into the sea: the rafts of ice so formed are carried out to the open sea by currents, carrying glacial till with them; when they eventually melt, they deposit this sediment.

It follows from their mode of transportation that although glacial marine sediments are composed of the same sort of material as glacial sediment on land, this material will be arranged quite differently: on land, we find the coarser material arranged in [moraines](#), while the finer material is carried off by meltwater and spread across [outwash plains](#). At sea, by contrast, the coarser sediment will be deposited at random among the finer sediment.

2.20.2 Glacial marine sedimentary rocks: how do we know?

We can observe present-day glacial marine sediment in the Arctic and Antarctic. The areas marked in cyan on the map to the right show where glacial marine sediments are presently being deposited. These sediments have sufficiently distinctive characteristics that when we find rocks



Marine sediments: cyan areas represent glacial marine sediment.

with the same characteristics in the geological record, we are entitled to deduce that they are [lithified](#) glacial marine sediments.

The marine fossils in such rocks identify them as being marine. What identifies them as glacial is their unique structure. This consists principally of finely ground rock flour, often lacking layering, scattered throughout with **dropstones**: boulders and cobbles which have been rafted out to sea on glacial ice and then sank when the ice melted.

Being glacial material, dropstones are often unrounded or poorly rounded. Furthermore, some of them will betray clear signs of their glacial nature by being polished and striated as a result of being dragged along at the base of the glacier.

These features allow us to identify glacial marine sedimentary rocks. Our ability to do so helps us to find out about climatic conditions in the past; we shall discuss this further when we address [paleoclimatology](#).

2.21 Saline giants



Salt mine, Wieliczka, Poland.

Saline giants are vast deposits of soluble minerals. What do we mean by “vast”? Well, to take one example, the Louann Salt covers 800,000 square kilometers and is four

kilometers deep, amounting to some seven quadrillion tonnes of salt. Even a small saline giant, such as the one found in the Michigan Basin, covers an area of 100,000 square kilometers and has a depth of 250-350 meters.

In this article we shall review what is known, and what may plausibly be conjectured, about the formation of saline giants.

2.21.1 Difficulties

Although precipitation of salts from seawater can be observed today in an ordinary bucket, the formation of saline giants cannot be observed anywhere. This is not really surprising: the geological record shows that the formation of saline giants has only happened at certain times and in certain places, and it is not unexpected that we should happen to be living in one of the times when no saline giants are forming.

While it is not unexpected, it is annoying. Our understanding of sedimentary rocks is in other cases greatly enhanced by the fact that we can see the sediments being deposited in the present day. In the case of saline giants, we lack this information and must do the best we can.

2.21.2 Evaporation of seawater

Seawater contains a variety of dissolved ions, such as (in descending order of abundance by mass) Cl^- , Na^+ , SO_4^{2-} , Mg^{2+} , Ca^{2+} , K^+ , HCO_3^- and Br^- ; these eight ions alone make up more than 99% of the dissolved ions in seawater, and other ions can be neglected for the purposes of this article.

When seawater evaporates, these precipitate out as minerals such as halite (NaCl) and gypsum ($\text{CaSO}_4 \cdot 2\text{H}_2\text{O}$). The evaporation of 1 liter of seawater will produce around 35 grams of evaporites.

The degree of evaporation required for precipitation varies from mineral to mineral: so gypsum will begin to precipitate out when seawater has been reduced to about 30% of its original volume, but it needs to be reduced to 10% of its volume for the precipitation of halite.

The relevant facts are summarized in the table below. The first column specifies the mineral, the second gives its abundance as a percentage of all the minerals precipitated, and the table as a whole has been ordered roughly according to the ease with which the various minerals precipitate out, from those that precipitate out most readily down to the most soluble. Minerals which occur only in the tiniest traces have been omitted.

The figures given here are based on the pioneering work of Usiglio, still thought to be reasonably accurate: more information will be found [here](#).

The upshot of all this is that as sea water evaporates, a small quantity of calcite will be deposited first. As evap-

oration continues, gypsum will be deposited: as there is much more gypsum than calcite, and as most of the calcite will have been deposited already, this means that the gypsum will swamp the calcite being deposited, and so the resulting rock will be almost exclusively gypsum. Similarly, when the halite starts being deposited, more halite will be deposited than gypsum or calcite, and the result will essentially be halite. A really intensely concentrated brine, reduced to a few percent of its original volume, will precipitate other salts, but the greater abundance of halite will ensure that it predominates.

We should note also that further dehydration of gypsum ($\text{CaSO}_4 \cdot 2\text{H}_2\text{O}$) will remove the water molecules associated with the calcium sulphate, converting it to anhydrite (CaSO_4).

It follows that saline giants will, broadly speaking, consist either of halite, gypsum, or anhydrite.

2.21.3 Models for the formation of saline giants

Saline giants are invariably covered over by a blanket of more conventional sediment, otherwise they would long ago have been washed away by rain (if on land), or dissolved in the sea (if underwater). This observation leads us to a question which is initially perplexing: how in the world can they have formed in the first place? The sea dissolves soluble minerals, and is nowhere near the saturation point at which they must start precipitating out. How, then, is it even possible for these saline giants to form within a marine basin?

Your first guess might be that they are the result of a bit of sea becoming closed off from the main body of the sea and simply drying up. However, there is too much salt in saline giants to be accounted for by a single such event, since the drying up of a kilometer's depth of seawater would result in the deposition of only 14 meters of salt. What we need is some model in which the basin keeps being filled, either continuously or periodically, with new supplies of salt water.

We shall describe four such models. Note that although they cannot all apply to the same saline giant at the same time, it is perfectly possible for them to apply to different saline giants, or, conceivably, to the same saline giant at different times. In this sense, it is possible for all the models described to be correct.

- Model 1: A barrier with a small gap in it.

Our first model is this: suppose we have a sedimentary basin which is connected to the sea only by a very narrow channel. Combine this with an arid climate and little input of fresh water into the basin from rivers and streams or from rain. So long as the rate of evaporation is greater than the input of fresh water, the physical necessity that

the surface of the sea inside the basin must always be at the same level as the surface of the sea outside the basin ensures that salt water will always be flowing into the basin; and as water will be continually evaporating from it, leaving the dissolved minerals behind, this will increase the salinity of the water in the basin until it reaches the saturation point and precipitation occurs.

Something of the sort can be seen today: it was first calculated by Edmond Halley (of Halley's Comet fame) that the Mediterranean loses more water to evaporation than the input of fresh water; this, he realized, explained why there is always a current flowing into the Mediterranean through the Straits of Gibraltar.

We may note that the Mediterranean is indeed somewhat more salty than the Atlantic. However, there is no saline giant forming on the floor of the Mediterranean today. The reason is (so it has been calculated) that in order for this proposed mechanism to work, the cross-sectional area of the channel must be many orders of magnitude smaller than the surface area of the water in the sedimentary basin; otherwise the tendency of water to mix will prevent the water in the basin from ever becoming saturated enough for precipitation to take place. The Straits of Gibraltar today are simply not narrow enough for the formation of saline giants.

Now a channel of just the right size would be unstable: we might expect it within a short space of time to get blocked up or, alternatively, broadened out, either of which would spoil the proposed mechanism. The deposition of the saline giants under the Mediterranean took 300,000 years — a blink of an eye by the usual standards of geologists, but a long time for such a narrow channel to stay just the right width. For this reason, although we must regard this proposed mechanism as possible in theory, we should require very definite evidence to endorse it in any particular case.

- Model 2: A barrier overtopped at high tide.

Another proposal is of a barrier (a “sill”) between the main body of the sea and a sedimentary basin such that water will surmount it only at high tide. Again, this is possible, but as with the mechanism previously described, it needs to be just right, and, to stay just right for hundreds of thousands of years. This is possible but implausible. A further problem with such a model is this: the influx of water at high tide needs to just balance out the water lost by evaporation. For if the water input was less than the water lost, then the water in the basin would be reduced to an intermittent puddle which would not account for basin-wide sedimentation; and if the water input was greater than the water lost, then eventually the basin would fill up until, at high tide, it was at the same level as the main body of the sea, allowing mixing to take place. It is hard to see what effects could keep the situation in equilibrium, so that the basin is always reasonably full but never quite fills up.

- Model 3: A barrier overtopped by rises in sea level.

A third, similar model again requires a barrier completely blocking off the sedimentary basin, which is periodically overtopped, not as a result of the tide, but as a result of an increase in the global sea level caused by changes in the Earth's climate. Such a model would predict that layers of evaporites should alternate with layers of more conventional marine sediments deposited during periods when the basin is full. It seems that this is what we see in the case of the Mediterranean deposits (see [here](#) for further information) but it is by no means true of all such evaporites.

This requires a less precise set of circumstances than the previous model, in that fluctuations in global sea level caused by climate change might be expected to be greater in magnitude than local fluctuations caused by the tide.

This model might be combined with the first or second models: variations in sea level might alternately allow and prevent the mechanisms described in model one or two: again, we should then expect to see an alternation of evaporites with more conventional sediments.

- Model 4: A permeable barrier.

A fourth model is as follows: the basin is completely cut off from the main body of the sea, but by porous sediment or sedimentary rock, so that sea water can seep through the barrier. As with our other models, we require that the output of fresh water through evaporation should be greater than the input: however in this case, unlike the “narrow channel model”, the water in the basin is free to drop below sea level when it evaporates: this produces a pressure differential between the two sides of the barrier and ensures that water flows in just one direction, from the main body of the sea into the basin. The nearest analog to this model in the modern world would be the deposition of salt in lagoons.

You will notice that this model does not require anything to be just right: neither the height nor the width of the barrier are crucial: so long as the barrier is above sea level, the system described will work.

Such a model is immune to the problem of equilibrium that we raised with respect to the second model. For the lower the water level sinks in the basin by evaporation, the greater the pressure differential on the two sides of the barrier, and the greater the influx of water; and conversely, the higher the level in the basin, the less water will seep through the barrier. So we might well expect such a system to be in equilibrium, with the basin never either drying out completely or filling up so as to overflow the barrier.

2.21.4 Saline giants: what do we know and how do we know it?

All the models described above require two things: that the basin in which the saline giant is deposited should be nearly or totally isolated from the main body of the sea; and that the climate should be such that more fresh water is lost through evaporation than is input by rivers and rain. We can test whether these conditions were in place, and show that these are the conditions under which saline giants form. For example, the Mediterranean is in the present day *nearly* cut off from the sea, and it does not strain the imagination to suppose that 5.9 million years ago it was more isolated still. To take another example, it would be strange to see evaporites forming in the Gulf of Mexico today; but conditions were just right at the time when they formed, when it was almost, or entirely, blocked off from the main body of the sea by what is now West Africa. To take a third example, the Castile formation in the Delaware Basin (which, despite its name, is located in Texas and New Mexico) was ringed around by a *reef* complex during the time of its formation. Independent evidence from *paleoclimatology* also confirms that the climatic conditions were right for evaporite formation in each case.

So although in many cases there is controversy over which of the models we have described best explains the existence of a particular saline giant (a controversy which perhaps in some cases will never be fully resolved) this is really a dispute about details: geologists are agreed, and can confirm, that what is required is a basin which is almost, or entirely, cut off from the main body of the sea, plus an arid climate.

This explains why such deposits are rare in the geological record. There is no particular reason why this set of circumstances should be common: they occur by happenstance and not by any sort of geological inevitability.

2.22 Banded iron formations



Rock from a banded iron formation.

In this article we shall look at what banded iron formations are and how they might have formed. The reader may find it useful to look back at [the main article](#) on siliceous ooze and chert.

2.22.1 BIFs defined

Banded iron formations, or **BIFs** are sedimentary rocks consisting of alternating bands iron-rich sediment (typically **hematite**, Fe_2O_3 , and **magnetite**, Fe_3O_4) and iron-poor sediment, typically **chert**; the size of the bands ranges from less than a millimeter to more than a meter in thickness. The image to the right shows a fairly typical banded iron formation: the red bands are the **iron oxides**.

While BIFs have a wide geographical distribution, they are localized in time. They start to become common about 3.5 billion years ago), peak around 2.5 billion years ago, vanish about 1.8 billion years ago, make a small comeback around 1 billion years ago, and then essentially vanish from the geological record; none are being produced today.

2.22.2 BIFs and the rise of oxygen

With the exception of **saline giants** (as explained in [the previous article](#)) it is usually very easy to explain the origin of **sedimentary rocks**, because we can see identical sediments being deposited in the present: **chalk** looks just like **lithified calcareous ooze**, **tillite** looks just like **lithified glacial till**, **aeolian sandstone** looks just like **lithified desert sand**, and so on.

In the case of BIFs, however, no BIFs are being formed in the present, nor even recently. It seems, then, as though in searching for a cause for BIFs we must be looking for an event which could only have happened at in the past.

Fortunately, one comes to mind. According to biologists, the first living organisms neither produced nor consumed oxygen. Indeed, they would not have been able even to tolerate oxygen: oxygen is a very reactive gas, and is toxic to organisms which are not adapted to its presence (for example the modern bacterium *Clostridium botulinum*, which can only survive in the near-total absence of oxygen).

Biologists are also agreed that in the absence of oxygen-producing organisms, the atmosphere would have been very poor in free oxygen (i.e. the molecule O_2).

What does all this have to do with BIFs? Well, one of the interesting things about iron is that elemental iron (**Fe**) dissolves in water, whereas the various oxides of iron (as found in banded iron formations) precipitate out. The waters of the early Earth would certainly have had sources of iron, such as emissions from submarine volcanoes, and iron liberated from rocks by chemical weathering. It follows that when organisms arose that produced oxygen,

iron dissolved in the oceans would combine with dissolved oxygen to form iron oxides which would then have precipitated out, producing the iron oxides that characterize BIFs.

The iron would, indeed, form an “oxygen sink”; only after the iron had been used up in this way would O_2 have begun to constitute a large proportion of the atmosphere. The accumulation of oxygen in the atmosphere, which according to geological dating methods started about 2.4 billion years ago, is variously known as the **Great Oxygenation Event (GOE)**, the **oxygen catastrophe**, and the **oxygen crisis**.

2.22.3 How do we know?

The scenario given above is plausible according to biologists; indeed, if they're right about the history of early life, we should *expect* to see this kind of geological evidence of a rise in oxygen as a result of the rise of oxygen-producing organisms. And biology aside, it is certainly chemically very plausible: iron is soluble in water and **iron oxides** are not, and this at least is something we can check by direct observation. It is also plausible in that it explains the localization in time of BIFs in terms of something that we would expect to happen only once.

But was there in fact a change from an oxygen-poor to an oxygen-rich atmosphere? Studies of **minerals** before, during, and after the GOE answer this question in the affirmative.

The fact that the production of free oxygen is indeed a plausible explanation for BIFs is a point in favor of this scenario; what is more, this explanation is borne out by the nature of the **iron oxides** in BIFs: they tend to be iron oxides with a low oxygen to iron ratio such as hematite and magnetite, rather than, for example, goethite ($FeO(OH)$); which is what we would expect if they formed in conditions in which oxygen was still scarce.

However, we would be verging on circular reasoning if we explained BIFs by the advent of free oxygen, *and* if our evidence for this event consisted solely of BIFs. Fortunately, this is far from being the case: there are other indications of the GOE in the mineralogy of the early Earth.

For example, before, but not after, the date assigned to the GOE, the **minerals uraninite** (UO_2) and **siderite** ($FeCO_3$) can be found in river sediments (for information on how to identify such sediments, see the main article on **rivers**). The significance of this is that these minerals would not survive in waters containing dissolved O_2 , as all rivers do today; so the rivers that deposited them must have co-existed with an oxygen-poor atmosphere.

After the date assigned to the GOE, on the other hand, we see a great diversification of mineral types in the geological record, as after the GOE new minerals could then be produced from old ones by oxidization; and it is just

such minerals that we find after the GOE. (For more details, see Sverjensky and Lee (2010) **The Great Oxidation Event and Mineral Diversification**, *Elements*, 6(1).)

So we have abundant data pointing to the rise of free oxygen in the atmosphere; and such a rise would explain, indeed necessitate, the extensive formation of iron oxide deposits such as are found in BIFs.

2.22.4 BIFs: some questions

It seems, then, as though we have a good explanation for BIFs. However, the reader should bear in mind that what I have sketched here is only a broad outline of a broad consensus. There is still controversy over details.

The reader may well already be puzzling over a couple of these details. Firstly, why is the sediment in the dark bands of BIFs so frequently **chert**? And why are BIFs banded at all — why were the chert and iron oxides not deposited simultaneously as a mixture?

The chert may have been deposited by silica-forming organisms. It is true that we don't find in it the **tests** of diatoms and radiolarians, so it probably didn't have its origin as the sort of **siliceous ooze** produced by these organisms today; but one cannot rule out the possibility of other silica-producing organisms active in the pre-Cambrian and now extinct.

Alternatively, if there were no silica-producing organisms in the pre-Cambrian, then **silica** could just have built up in the seas until it reached saturation point and precipitated out by itself.

The rhythmic nature of deposition seems to suggest a cyclic variation in conditions. One candidate cause is a repeated cycle of ecological boom and bust. On biological grounds, it would be reasonable to suspect that the earliest oxygen-producing organisms could not tolerate high levels of oxygen. (This is not as paradoxical as it sounds: most organisms can't live in an environment full of their own waste products. Animals, for example, produce CO_2 and would suffocate in an atmosphere dominated by it.)

This suggests the following scenario: oxygen producing organisms would grow and flourish until they had produced toxic levels of oxygen; the population would then collapse almost to nothing, surviving only in low-oxygen refuges; the oxygen would be removed from the atmosphere by combining with the iron in the water to produce the iron oxide bands of BIFs; in these low-oxygen conditions the oxygen-producing organisms could once more increase in number, and the cycle would begin again. This would plausibly account for the periodic precipitation of iron oxides. It is even conceivable that the oxygen-producing organisms were identical with the (hypothetical) **silica**-secreting organisms mentioned above.

But here we have gone into the realms of speculation and controversy. It is quite possible that these questions

will remain controversial: since geologists labor under the handicap of not being able to watch BIFs form in the present, BIFs will never be understood with quite the same certainty as other [sedimentary rocks](#).

Chapter 3

Plate tectonics

3.1 Physical properties of rocks

In order to fully understand plate tectonics and the evidence for it, it is necessary for the reader to know a little about the physical properties of rocks. In this article we provide a brief introduction to the concepts involved.

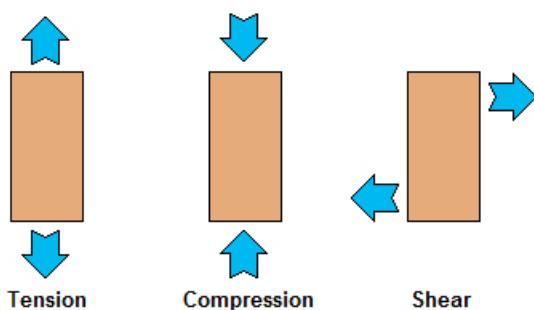
3.1.1 Stress and strain

In ordinary English, stress and strain are more or less synonymous. In physics, they refer to different though related quantities.

Stress is a measure of the force per unit area exerted on a surface of a deformable body. That is, roughly speaking, stress is to solids what pressure is to gasses, and like pressure it is measured in **pascals (Pa)**; that is, in newtons per meter squared.

Strain is the deformation of a body as a result of stress. In geology, strain is given by the length by which a rock expands or contracts divided by the length it was originally: because this is the ratio of a length to a length no units are associated with it.

3.1.2 Tension, compression, and shear



Tension, compression, and shear.

The stress on a rock (or any other material, for that matter) can be classified as tension, compression, or shear, as illustrated in the diagram to the right.

Rock is strong under compression but relatively weak under tension and shear. This is a result of the microscopic structure of rock: it contains microscopic cracks which are forced open and enlarged by tension and shear but which are forced closed by compression.

This is why a small overhang on a cliff will easily break under its own weight (being subjected to shear) whereas the rock at the foot of the same cliff will bear the much greater weight of all the rock above it, as in that case it is being subjected to compression.

3.1.3 Elastic and plastic behavior

A material is said to be **elastic** if it recovers from stress — that is, if, having been bent or extended or compressed under shear stress or tension or compression, it snaps back into its original configuration when the stress is released.

A material is said to be **plastic** if, on the contrary, once stress has squeezed it into a certain shape, it retains that shape; plasticine, for example, is plastic at room temperature and surface pressure.

When a solid is placed under stress, its behavior is at first elastic; then (with increasing stress) plastic; then with the addition of enough stress it fractures.

A solid which undergoes very little plastic deformation between elastic behavior and fracturing is said to be **brittle**. In colloquial English we usually reserve this word for things which are both brittle in the technical sense and also require little stress to break, such as egg-shells; in its technical use in physics, however, a substance such as diamond is also brittle in the technical sense: diamond may not break easily, but it will break before it undergoes any significant plastic deformation.

The opposite of brittle is **ductile**.

A material will have greater resistance to fracture if it is under a high surrounding pressure; and it will be more ductile at higher temperatures.

The reader should also bear in mind that the rate at which stress is applied may be significant: a force rapidly applied may produce fracture which, if more slowly applied, may produce deformation. The material known as Silly Putty is famous for clearly demonstrating this property:

it deforms under gentle pressure from one's fingers but shatters if hit with a hammer.

3.1.4 Application to rocks

We should now explain how all this applies to rocks in particular.

Rocks on the surface will exhibit elastic and brittle behavior, since they are cold and at low pressure. At depth, the pressure will be greater, increasing their brittle strength (that is, their resistance to fracture) and the temperatures will be higher, decreasing their ductile strength (that is, their resistance to plastic deformation).

Below the depth at which the ductile strength is less than the brittle strength, the rocks will be fully ductile and plastic. Some people describe the rocks below this depth as molten, but this is not accurate: they are not a liquid, but rather a ductile solid, like plasticine.

As a result, rocks near the surface tend to fracture under stress creating geological faults; more deeply buried rocks tend to fold.

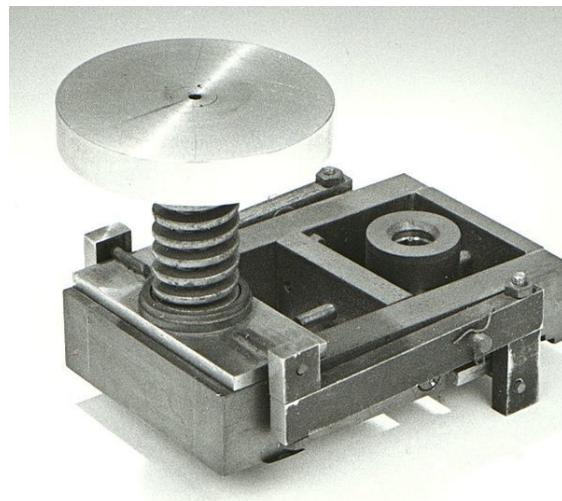
Earthquakes are also a phenomenon of the upper, brittle, elastic part of the rock. When two pieces of the Earth's crust try to move past one another, their mutual friction impedes them and they bend very slightly. Earthquakes are caused when the potential energy of the bent rocks is sufficient to overcome the resisting friction and they snap back, releasing the stored energy in the form of kinetic energy. This is only possible if the behavior of the rocks is elastic rather than plastic. Consequently we do not expect (and do not find) deep earthquakes except when they are associated with **subduction** (as will be discussed in a subsequent article).

3.1.5 How do we know?

The behavior of rocks at surface temperatures and pressures are easy to verify. To find out how they would behave at greater temperatures and pressures requires special equipment.

In a simple experiment often repeated with different kinds of rocks, scientists can take a cylinder of rock and compress it with a piston with varying degrees of confining pressure. With little or no confining pressure, the rock will fracture, as you would expect. However, under greater degrees of confining pressure (such as would be experienced by a rock buried at a depth of one or two kilometers) the rock does not fracture, but rather it deforms like plasticine: or, to put it another way, it is ductile rather than brittle.

More recent experiments have reached greater levels of sophistication. By using lasers to heat rock samples, and a device known as a **diamond anvil cell** to exert pressure on them, it is possible to simulate temperatures and



The very first diamond anvil cell, now on display in the Gaithersberg Museum, Maryland.

pressures such as are found deep within the Earth.

Such methods do not tell us everything we would like to know. Reproducing conditions in the very core of the Earth would require some sort of breakthrough in materials technology. Another thing that is hard to simulate is the effect of *time*. We know that materials are more likely to deform and less likely to shatter when stress is applied gradually: so what happens if you apply a gentle stress to a rock over a period of millions of years?

Such questions can to some extent be answered with reference to established notions in physics; but clearly if all such questions could be answered with total accuracy with reference to purely theoretical considerations, then geologists wouldn't spend so much money on diamond anvil cells and lasers.

That being said, what we do know is sufficient for us to understand plate tectonics; certainly it is quite enough for an introductory course such as this one.

3.2 Seismic waves

In this article we shall review some important facts about the physics of **seismic waves** (that is, waves generated by earthquakes).

This article is background reading for the following article on the structure of the Earth, which is itself background reading for subsequent articles on plate tectonics.

3.2.1 Surface waves and body waves

When an earthquake occurs, it is the cause of seismic waves, including both waves that travel along the surface (**Love waves** and **Raleigh waves**) and waves that travel



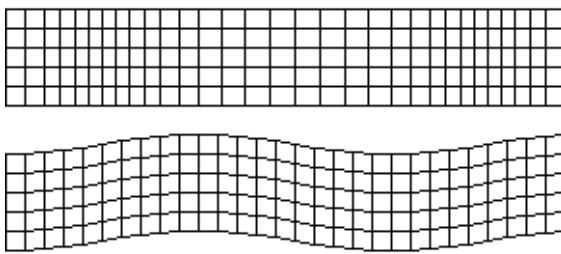
Earthquake damage, Tubigon, Bohol, Philippines

through the body of the Earth (**P-waves** and **S-waves**, known collectively as **body waves**).

In this and succeeding articles we shall be interested only in body waves, since by traveling through the Earth they give us clues about the Earth's interior.

3.2.2 P-waves and S-waves

P-waves are waves of compression and tension, like sound waves; indeed, they *are* sound waves, traveling through rock rather than through air. S-waves are waves of shear: that is, of displacement at right angles to the direction of travel of the wave, resembling the waves produced by shaking the end of a rope.



P-waves (top) and S-waves (bottom).

The image to the right shows the motion of P-waves (top) and S-waves (bottom). Look carefully at the picture of P-waves. If you focus on any particular vertical line, you will see that it is merely oscillating from side to side: it is the regions of compression that are moving from left to right, while the medium itself has no net motion. Similarly in S-waves, it is the displacement that moves from left to right, while the medium through which the waves move exhibits no such motion.

If we use κ to represent the incompressibility of a medium, μ to represent the rigidity, and ρ to represent the density, then the velocity of P-waves through that medium will be given by

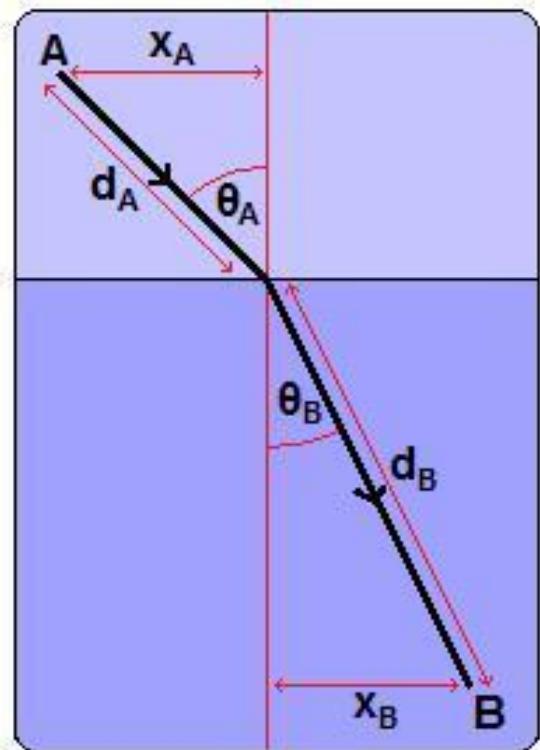
$$v_P = \sqrt{\frac{\kappa + \frac{4}{3}\mu}{\rho}}$$

and the velocity of S-waves by

$$v_S = \sqrt{\frac{\mu}{\rho}}$$

From these formulas we can immediately see that a P-wave will always travel faster than an S-wave through the same medium. We can also see that S-waves will not travel through liquid at all, since liquid has no rigidity and so in the case of liquids $\mu = 0$.

3.2.3 Refraction and least time



Refraction.

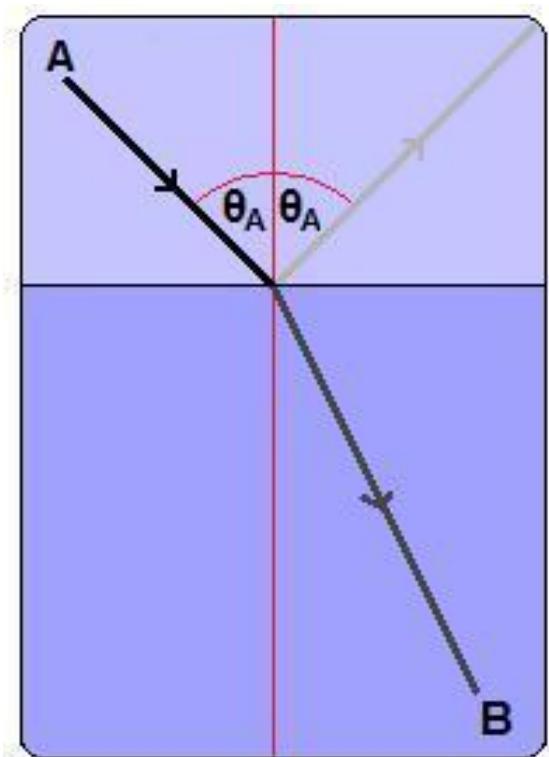
The **principle of least time** says that a wave traveling through a medium will take, not the shortest route as measured by distance, but rather the quickest route between two points.

The word **refraction** means the change of direction undergone by a wave when it passes from a material which permits travel at one speed to a material which permits travel at another speed. The existence of refraction is a direct consequence of the principle of least time: in the generalized diagram of refraction shown to the right, the path of the ray is the quickest route from point A to point B.

Using the principle of least time, we can say exactly how refraction should take place: a little simple mathematics tells us that the ratio of the velocity in medium A to the velocity in medium B should be equal to the ratio of $\sin \theta_A$ to $\sin \theta_B$; or to put it another way, equal to the ratio of x_A/d_A to x_B/d_B . This is known as **Snell's Law**. Where the velocity changes smoothly and gradually through the medium, this will result in the ray taking a curved path.

The upshot of all this is that if we know the wave velocities associated with each point in an object then we know exactly how a wave will travel through it, since its motion is determined by the principle of least time.

3.2.4 Partial reflection



Refraction and partial reflection.

When a wave encounters a sudden transition between media with different associated velocities, so that the wave is refracted at an angle, some of the energy will also be reflected back in a direction determined by the well-known law that the angle of incidence is equal to the angle of reflection (see the diagram to the right). This is why one can see faint reflections in windows and in water.

P-waves and S-waves behave in the same way as more familiar waves such as light, but with one difference: light is exclusively a transverse wave, and reflects back as one. When either a P-wave or an S-wave is reflected, however, the reflection will be composed of both P and S-waves.

When such reflections of P and S-waves are detected, this

tells us that they are being reflected off some sort of sharp boundary between rocks having different physical properties and hence different associated velocities.

3.2.5 How do we know?

The properties of body waves can be studied in the laboratory. They can also be derived from more basic principles, simply working from the fact that they are waves and must therefore obey the physics of waves. (Note that in physics a "wave" is not just something that behaves in a kind of wave-y sort of way, but something the dynamics of which can be described by the wave equation. The mere fact that body waves *are* waves therefore tells us quite a lot about them.)

3.2.6 Seismic tomography

As discussed, the routes taken by P- and S-waves through the interior of the Earth are determined completely by the velocities at which these waves travel at each point in the interior.

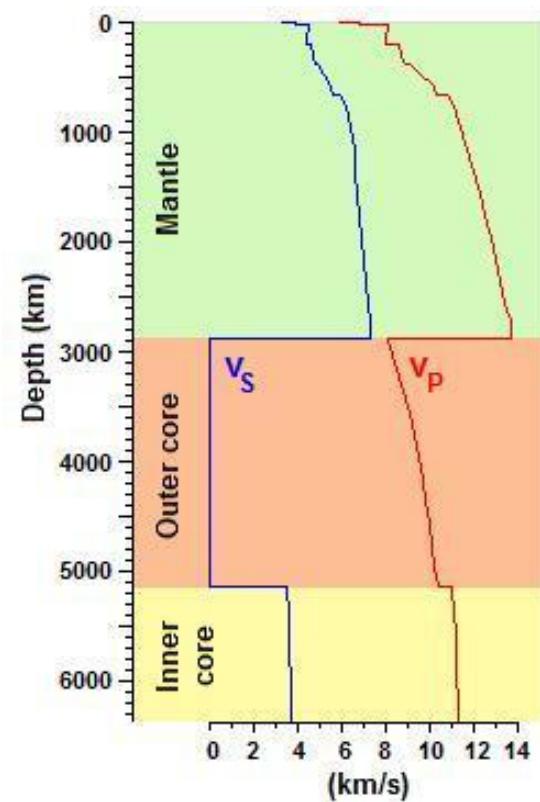
If these velocities were the same for every point in the Earth, then the time it took for a wave to travel from the **focus** of an earthquake (the focus being the point where the earthquake occurs, not to be confused with the **epicenter**, which is the point on the surface directly above the focus) to an earthquake detector (a **seismometer**) would be proportional to the distance of a straight line drawn from the one to the other through the body of the Earth.

But this is not the case. By studying the data showing how long the waves do take to travel through the Earth, it is possible to determine the velocities of P-waves and of S-waves at each point in the Earth.

Doing so is what mathematicians know as an "inverse problem": it might be compared to trying to reconstruct the shape of an object by observing its shadow. It would obviously be much easier to work the other way round and deduce the shadow from the object; and similarly the *general* problem of **seismic tomography** — that is, of discovering the inner structure of an object by studying the passage of body waves through it — would be very difficult to solve.

But fortunately we do not have to solve this problem in the general case, but only for one particular object: the Earth. This problem has a property that makes it particularly easy to analyze.

Consider the fact that the time it takes for a P-wave or an S-wave to get from the focus to a seismometer depends to a good degree of approximation only on the angle of separation between the focus and the seismometer. This tells us that (again to a good degree of approximation) the values of v_P and v_S at any particular point within the Earth must depend only on the depth below the surface of



Velocities of P- and S-waves in the Earth's interior.

the Earth and not on the longitude and latitude (in technical terms, we may say that the values of vP and vS are **spherically symmetric**).

Consequently, what looked like a three-dimensional problem involving finding values of vP and vS for every depth, longitude, and latitude can be reduced to the one-dimensional problem of finding values of vP and vS at every depth.

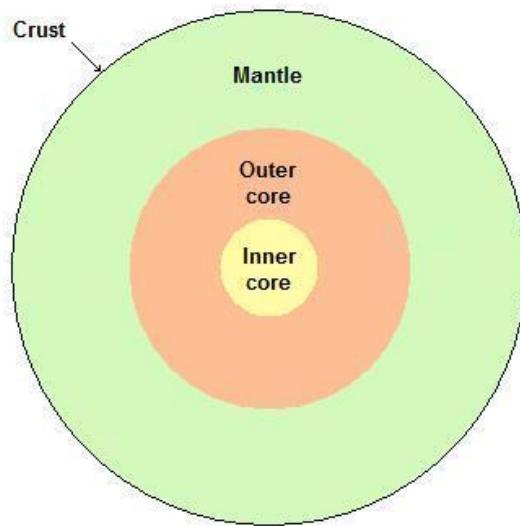
This fact makes it possible to produce a graph such as that shown to the right, relating vP and vS to depth; the figures are taken from the Preliminary Reference Earth Model (here). One fact that you should notice immediately is that the speed of vS drops to 0 in the outer core, showing that it is liquid.

I have said that the travel time of body waves depends on the angular separation between the earthquake focus and the seismometer *to a good degree of approximation*; but this is not exactly and perfectly the case. By studying the small variations from this rule, it is possible to detect low and high velocity **anomalies** within the Earth: volumes where the waves travel slower or faster than what we should expect if the Earth's interior was perfectly spherically symmetric. Work on this is ongoing.

3.3 Structure of the Earth

In this article we shall review some key facts on the structure of the Earth, and discuss how they are known. We shall refer back to facts already discussed in the articles on igneous rocks, seismic waves, and physical properties of rocks; readers may wish to refresh their memories on these subjects before reading further.

3.3.1 Structure of the Earth



Cross-section of the Earth.

The Earth can be divided by composition into the **crust**, **mantle**, and **core**, as shown to scale in the diagram to the right.

The crust

The crust consist mainly of igneous and metamorphic rocks with a relatively thin layer of sediment and sedimentary rocks on top.

Crust comes in two varieties: continental and oceanic. Continental crust consists mainly of **felsic rocks** such as granite, and is about 30-50 km thick, varying from place to place; oceanic crust consists mainly of somewhat more **mafic rocks** such as basalt, and is about 5-10 km thick.

The nature of these rocks suggests the reason why there is a crust as such. The reader may at this point find it useful to review the main article on igneous rocks. To summarize the most important points, felsic rocks (i.e. those that have a high **silica** content) also have a lower density and a lower melting point than more mafic rocks.

The **minerals** known as **feldspar**, for example, form 60% of the Earth's crust; their melting points range from 600°C - 1000°C, depending on their exact chemical composition, and a density between 2.55 and 2.76. Compare

this to the **ultramafic olivine** of the upper mantle, with a melting point ranging between 1200°C - 1900°C and a density of 3.27–3.37.

This immediately suggests why the Earth should have a crust as such: that is, a region chemically distinct from the mantle. The more **felsic** minerals in the mantle would melt, because they have a lower melting point than the major constituents of the mantle; they would rise, being less dense by virtue both of their composition and their molten state; once they had erupted on the surface and cooled, they would “float” on the mantle as a result of their lower density. So even if the earth started off in a relatively homogeneous state (as geologists think it did) this process, known as **differentiation**, would ensure that the Earth ended up with a crust of rocks composed of minerals more **felsic** than those in the mantle.

The mantle

The mantle consists of a further 2890 kilometers of denser **ultramafic** rocks.

It is sometimes wrongly stated that the mantle consists of molten rock. We know that this cannot be true because **S-waves** pass through the mantle, which would not be possible if it was a fluid (see the main article on seismic waves for more details). One reason for the popular belief in a liquid mantle is that, after all, **lava**, a liquid, erupts from out of the mantle. However, this lava is produced by *partial* melting of the solid mantle. This, incidentally, is why it has a different composition from the mantle, being more **felsic**. Although the mantle is not liquid, it does flow: technically, it is a ductile solid, as explained in the article on physical properties of rocks.

The core

The core is the innermost part of the Earth, having a radius of 3,400 km. It can be divided into the outer core, which is molten, and the inner core, with a radius of 1,220 km, which is solid. (This arrangement may seem strange at first, but recall that the inner core is at greater pressure and so will have a higher melting point.)

The core is made mainly of iron. As with the existence of the crust, this can be explained on the hypothesis of differentiation: just as the light substances rose to the top, so the denser substances would sink to the bottom.

The lithosphere and athenosphere

The division into crust, mantle, and core partitions the Earth according to the composition of the rocks. Sometime, however, it is more useful to group the crust and the uppermost layer of the mantle together as the **lithosphere**. What these have in common is that they are brittle and **elastic**, as opposed to the **plastic** and **ductile**

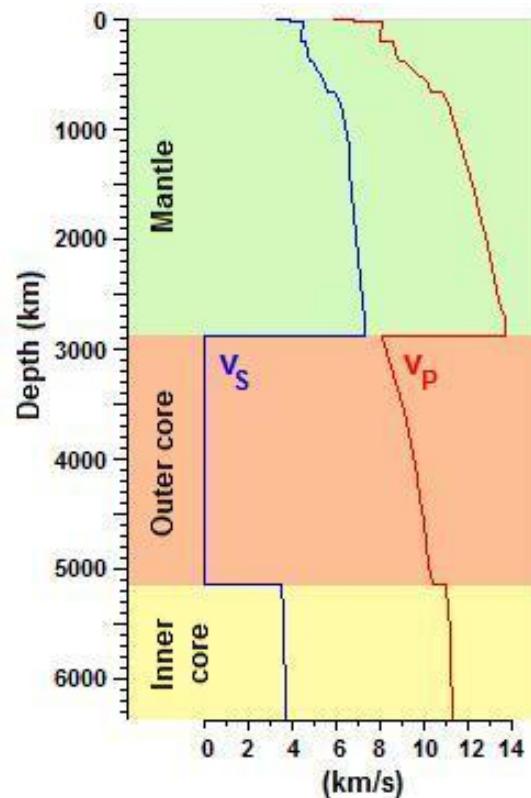
rocks in the rest of the mantle. The lithosphere ranges in thickness from 40-200 km, varying from place to place; it is thicker under continental crust. The concept of the lithosphere is especially important in the context of plate tectonics, as the plates in plate tectonics are not (as is sometimes stated) plates of the Earth’s crust: they are plates of the lithosphere.

The portion of the mantle immediately below the lithosphere is called the **athenosphere**. This is the weakest part of the mantle, because although it is at a lower temperature than the deeper rocks, it is also at a lower pressure.

3.3.2 How do we know?

We shall now, as usual in these articles, sketch out how the knowledge described in the preceding sections was obtained. Let us first of all consider the facts that constrain any attempts to make a model of the physical properties of the Earth’s interior.

vP and vS



Velocities of P- and S-waves in the Earth’s interior.

As was explained in the article on seismic waves, it is possible by studying earthquakes to discover the velocities of P- and S-waves at various points within the Earth. The results are summarized in the chart to the right.

Spherical symmetry

As explained in the previous article, we call a property spherically symmetric if it varies only with the distance from the center of the Earth and not with latitude and longitude. Probably no geological property (except distance from the center itself) is exactly spherically symmetric; but many of them can be demonstrated to be very nearly so: in what follows we shall use “spherically symmetric” to mean “spherically symmetric to a good degree of approximation”.

As discussed in our article on seismic waves, the velocities of P- and S-waves (v_P and v_S) are spherically symmetric properties. It would be very remarkable if this was the case, and yet the properties of the Earth on which these velocities depend was not.

In the case of density, there is good evidence that it is spherically symmetric: for if it was then the force of gravity at the surface of the Earth would be (to a good degree of approximation) the same at any point on the surface; which it is. So the evidence is that density (which we shall denote by the Greek letter ρ) must be spherically symmetric. It immediately follows that the same must be true for pressure, since this can be calculated from density.

Now consider the fact that for any point in the Earth the velocity of S-waves (v_S) is given by the formula $v_S = \sqrt{\mu/\rho}$, where μ is the rigidity and ρ is the density. So given that v_S and ρ are both spherically symmetric, it follows that μ must be also. Furthermore, the velocity of P-waves (v_P) is given by $v_P = \sqrt{(\kappa + 4\mu/3)/\rho}$. So given that v_P , μ , and ρ are spherically symmetric, it follows that so is κ .

By reasoning of this sort, exploiting the inter-relatedness of the properties that interest us, we can show that they are all spherically symmetric. So to construct a first approximate model of the Earth, we only need to associate each *depth* within the Earth with a value for gravity, density, incompressibility, and so forth.

Boundary conditions

We know the strength of gravity at the surface of the Earth, because we can measure it directly; we also know the gravity at the center of the Earth, since in any spherically symmetric body this must be precisely 0. We know the temperature at the surface and the rate of heat flow. We know the pressure at the surface: 1 atmospheric pressure. We know the mass of the Earth, which can be easily deduced from experiments measuring Newton's constant G . Since we know this and the volume of the Earth, we also know its average density.

These all serve as constraints on any Earth model. For example, if we think we know a function relating density to depth, we can easily calculate what the surface gravity of the Earth should be if this function was correct.

Interrelatedness of properties

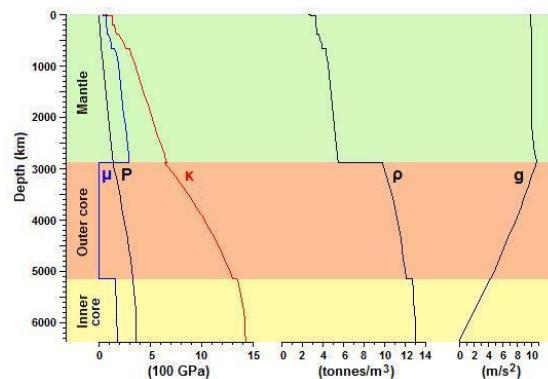
We have already used the formulas relating the properties to deduce the spherical symmetry of some properties from the spherical symmetry of other properties. But the relationships between them allow us to be much more precise than that. For example, if we know that $v_S = \sqrt{\mu/\rho}$, then knowing the relationship between depth and v_S , and the relationship between depth and ρ , we automatically know the relationship between depth and μ . And knowing this, and knowing the relationship between depth and v_P , we can exploit the formula $v_P = \sqrt{(\kappa + 4\mu/3)/\rho}$ to tell us the relationship between depth and κ — and so forth.

This means that it is not necessary or even possible for us to form separate hypotheses as to the values of the various physical properties that interest us. The values that we do know (v_P and v_S) place constraints on the values that we would like to know.

Results

To summarize: any model of the values of physical variables within the Earth must be constrained by:

- The known values of v_P and v_S
- Boundary conditions
- Spherical symmetry
- Self-consistency



Graphs showing the relationship of rigidity, incompressibility, pressure, density, and gravity to depth. Figures are taken from the Preliminary Reference Earth Model.

These constraints are sufficient for geologists to work out figures for pressure, density, gravity, incompressibility, and so forth. Some results are shown in the graphs to the right.

Here μ represents rigidity; κ incompressibility; P pressure; ρ density; and g the force of gravity.

Mineral composition

Hypotheses about the mineral composition of the Earth must of course be constrained by our model of its physical properties: the minerals must have the right density, rigidity, etc to account for these properties.

To date, it has been possible to drill a little over 12 km into the crust and take samples and make temperature measurements. The results are that continental crust, beneath any layers of sediment that have been deposited, is indeed composed of **felsic granites**, **gneisses** and so forth; and the oceanic crust of more **mafic rocks** such as **basalt** and **gabbro**. We can also study sections of ocean crust that have been thrust up onto the land — **ophiolites**. These will be the subject of a subsequent article.

There are a number of clues to the composition of the mantle.

- Volcanic eruptions sometimes bear up to the surface fragments of **peridotite**; their broken jagged shapes indicate that they must have been torn from the parent rock by the force of the eruption. Unfortunately, such eruptions originate from a maximum depth of about 180 km, so they only give us a sampling of the upper mantle.
- Except at subduction zones (where volcanoes recycle the material of the crust) volcanoes and rifts generally emit **basaltic lava**; it can be experimentally shown that this is just what would be produced by partial melting of **peridotite**.
- The base of ophiolites is **serpentinite**, a rock produced from **peridotite** in the presence of heat and water; that is, under the conditions present in the upper mantle.
- **Peridotite** has the right density to account for the values of ρ inferred from the seismological data.
- Minerals change their **phase** with pressure: the same elements in the same proportion adopt a more compact configuration. For example, at the pressures found at a depth of about 400 km, the mineral **olivine** (the main mineral constituent of **peridotite**) changes to **wadsleyite**. The seismological data indicate just the change of density at this depth that we would expect if such a change of phase took place. The other discontinuities in the density of the upper mantle are explicable in a similar manner.
- We do not see any discontinuities that would definitely indicate the substitution below some depth in the mantle of some different material altogether.

Although a cautious geologist would not claim absolute certainty as to the composition of the lower mantle, there

is general agreement that the mantle is made of more or less the same stuff from the top of the mantle down to the top of the core.

Finally, the core. The differentiation of the core into an outer liquid core and an inner solid core is based on the study of **S**- and **P**-waves. **S-waves** don't travel through the outer core, proving that it is liquid; **P-waves** travel faster through the inner core, showing an abrupt transition that corresponds well to a change of phase from liquid to solid.

Given the density of the core, it must be composed mainly of iron. To be sure, it could in principle be composed of a mixture of something much heavier than iron, such as gold, mixed with something much lighter. However, iron is the only element in the Earth, in the Solar system, or in the Universe generally that is both dense enough and common enough to account for the mass.

The iron probably contains an admixture of about 8% nickel, since these elements are usually found in association in these proportions. However, this iron-nickel mix would actually be somewhat too heavy to account properly for the mass of the Earth; it would follow that there must also be a proportion of lighter elements. The abundant light elements silicon and oxygen are favorite candidates for this role.

3.4 Geomagnetic reversals

In this article we shall consider the evidence for reversals of the Earth's magnetic field.

3.4.1 Paleomagnetism

When **igneous rocks** are formed, as the temperature of the rock falls beneath what is known as the **Curie temperature** (roughly speaking, the temperature above which a material cannot be magnetized and below which it can) iron-based **minerals** such as **magnetite** and **hematite** are magnetized by the Earth's magnetic field, indicating the directions of the Earth's north and south magnetic poles at the time that the rocks are formed.

Sedimentary rocks can also indicate these directions: when sediment is deposited gently in a low-energy environment such as the deep ocean floor, magnetized grains of **magnetite** and **hematite** will orient themselves to the Earth's magnetic field like so many tiny compass needles, indicating the directions of the north and south magnetic poles at the time the sediment was deposited.

When geologists realized that this was the case, it was immediately obvious that studying the magnetism of ancient rocks would tell them about the Earth's magnetic field as it existed in the deep past: **paleomagnetism**.

3.4.2 Geomagnetic reversals

The study of paleomagnetism led to the discovery of magnetic field reversals. At present the Earth's magnetic field exhibits what is known as **normal polarity**: that is, it has the magnetic north pole near the geographic north pole, and the magnetic south pole near the geographic south pole. But studies of the paleomagnetism of ancient rocks showed evidence that in the past the Earth has sometimes had **reversed polarity**, with the magnetic north pole in the southern hemisphere and vice versa.

Apparently, then, the Earth periodically undergoes **geomagnetic reversals**, in which the north and south magnetic poles switch ends.

3.4.3 Geomagnetic reversals; how do we know?

We can see today that magnetic **minerals** in sedimentary and **igneous rocks** align themselves with the present direction of the magnetic field; and the physics of this is well-understood — this is just what they ought to do.

We can also see that some rocks laid down in the past are magnetically aligned in the opposite direction. Moreover, all the rocks dated to a particular time in the past will have the same alignment: rocks that date to 60 million years ago will all have reversed polarity no matter where you look.

This really leaves us only with one plausible explanation. We can, to be sure, think of alternative *implausible* explanations. Perhaps the laws of physics themselves keep changing back and forth; or perhaps every now and then all the continents rapidly rotate 180 degrees in perfect synchrony. But it requires less of a stretch of the imagination to suppose that the poles themselves are moving.

This is supported by the fact that the poles are detectably moving. No-one has ever witnessed a geomagnetic reversal, which is hardly surprising if, as the evidence shows, they only take place about once every 100,000 years. but the poles certainly move: the difference between true north and magnetic north has been the subject of measurement and inquiry since the sixteenth century. At the present time of writing, the north magnetic pole is moving at around 40km/year, an unusually fast rate for anything to happen in geology. While the observation of this **secular variation**, as it is called, does not absolutely confirm the occurrence of geomagnetic reversals in the past, it certainly gives credence to the possibility of their occurrence.

Finally, in recent decades it has been possible to perform numerical simulations of the behavior of the Earth's core, starting with the work of Glatzmaier and Roberts in 1995 (which you can read [here](#)). Briefly, they put into a large and powerful computer the relevant laws of physics and the best estimates of the major forces and energies in the

Earth's core. We may omit the details, but the important thing to note is that their model contained nothing explicitly relating to the motion of the poles, so that if they are found to move in the model, this must be implicit in the physics of the situation.

And their models did indeed exhibit polar reversals. The models also exhibited certain other features which must give us confidence in their broad accuracy: first, the fact that they exhibited a self-sustaining electromagnetic field at all; secondly, that they exhibit the sort of secular variation that we can observe today; third, that they exhibited one phenomenon which up until that time had never even been guessed at: that the solid inner core is rotating slightly faster than the rest of the planet: that is, it revolves slightly more than once a day. This really remarkable prediction has since been supported by seismological data (which you can read about [here](#)).

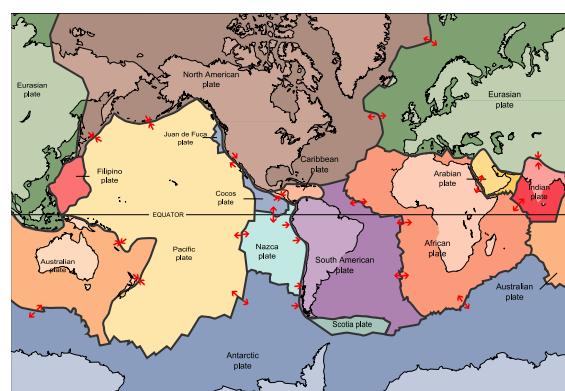
We may then regard the existence of geomagnetic reversals in the past as being on a very firm footing, since they *should* happen in principle and all the evidence shows that they *did* happen in practice.

3.5 Plate tectonics: overview

3.5.1 Introduction

This article consists of a short introduction to the underlying concepts of plate tectonics. Detailed discussion will be deferred to subsequent articles; but it is a good idea for the reader to have a grasp of the big picture in order to understand the relevance of the details.

3.5.2 Plate tectonics: the mechanism



The essential idea of plate tectonics is very simple. The **lithosphere** of the Earth is divided into **plates**, as shown in the map to the right: and these plates move. **Plate tectonics** is the study of the motion of the plates in the present and in the past.

Observing that the plates shown on the map interlock with

no gaps between them, you might at first guess either that they couldn't move at all, or that they would have to do so as a body, with no movement of the plates relative to one another.

However, there are two mechanisms which allow this relative motion. The first is **subduction**: at one edge, a plate can be forced underneath another and into the mantle. The second is **sea-floor spreading**: plates can move apart at a rift which is continuously filled up by intruding magma produced by partial melting of the underlying mantle, creating a continuous supply of fresh oceanic crust.

This means that a plate can move along like a conveyor belt: it can move along from east to west (for example) if it is being subducted into the mantle at the west end and spreading at the east end.

3.5.3 A deeper theory?

There is a great deal of disagreement among geophysicists as to the exact nature and the relative importance of the mechanisms driving spreading and subduction.

The essential mechanism is agreed to be convection. The **mantle** is hotter at the bottom, near the **core**, than it is at the top, near the **crust**. What's more, the **mantle** below the **lithosphere** is ductile. Given the temperature difference and the viscosity of the **mantle**, it is possible to prove that the **mantle** must convect: that is, hot material must rise to the top, cool there, and sink to the bottom again. Since it cannot rise and sink in the same place, the **mantle** must flow laterally below the **lithosphere**.

However, there are conflicting opinions and indeed conflicting evidence as to the exact nature of the convection, and of its interaction with the **lithosphere**. So we still lack what one might call the theory of the theory of plate tectonics. What we would like is a model sufficiently good to predict such things as the average speed of a plate and the average size of a plate; but this has so far eluded geologists.

The emphasis of this textbook is however on historical geology: if we can establish that the plates do move in the present and have moved in the past, this is quite sufficient for our purposes.

3.5.4 Plate tectonics: how do we know?

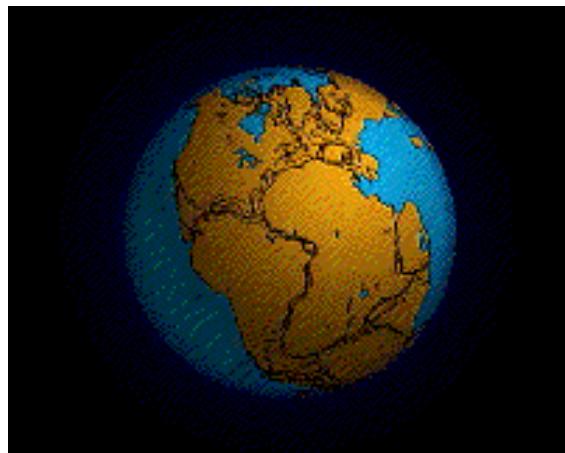
The fact that plates are moving now can be confirmed by direct measurement, as will be discussed in [the next article](#). To demonstrate their past motion, and the mechanisms operating now and in the past, it is necessary to study many diverse categories of evidence: too many, indeed, to compress into the space of a single article. In the following articles we shall look at the evidence piece by piece.

3.6 Continental drift

Continental drift is the theory that the continents have shifted their positions over time. Today, the subject of continental drift has been subsumed into **plate tectonics**, the science which deals with every aspect of the motion of **lithospheric plates**.

To write about “continental drift” rather than “plate tectonics” is therefore something of an anachronism. However, there is such a lot to say about plate tectonics that it seems necessary for us to break down the subject into smaller pieces that are easier to swallow. In this article, therefore, we shall examine the notion that the continents have shifted so far as we can do so without discussing the mechanisms of plate tectonics.

3.6.1 The drift theory



The rifting of Pangaea.

As first formulated, the drift theory was that the continents had once been joined together in a single large continent (**Pangaea**) which then split into the modern continents, which then drifted into their current positions. We now know that this was only one of a series of similar events by which continents have alternately been pushed together and pulled apart. The animation to the right shows the splitting apart of Pangaea.

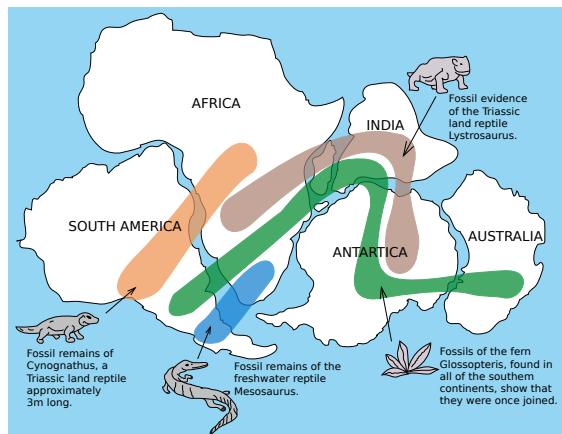
When the drift theory was first proposed, the “**drifters**”, as they were called, had no idea *how* the continents could have moved, and such proposals as they came up with were contrary to reason and evidence (as was pointed out by their opponents, the “**fixists**”). But, as we now know, the evidence they produced *that* the continents had moved was correct, and should have convinced more geologists than it did at the time.

3.6.2 How did they know? The initial evidence for drift

As explained in the introduction to this article, what we will present here is the most superficial evidence that continents have shifted in their relative positions; specifically, that the Americas were once attached to the Old World along the eastern edge of the former and the western edge of the latter. What is presented below is therefore by no means the whole or even the majority of the evidence for continental drift; it is merely all the evidence that we can present without looking at the mechanisms of plate tectonics.

When continental drift was first proposed, the arguments rested on four key pieces of evidence.

- A glance at a world map shows that it *looks like* the Americas and the Old World once fitted together. Today, by measuring the submerged borders of the continental shelf rather than the coastline, and by using computers and the result in spherical geometry known as Euler's theorem, we can see just how good this fit is. Yet even without these techniques, a glance at a world map shows that there is something to be explained.
- There are geological features that seem to be interrupted by the Atlantic: for example, rock formations and mountains which appear to start in North America and finish in Scotland. We could perhaps try to ascribe this to coincidence or to forces as yet unknown; but when we take this in conjunction with the rest of the evidence, it is more parsimonious to conclude that the Americas and the Old World were once contiguous, which entirely explains why the east coast of the Americas and the west coast of the Old World underwent the same geological processes.



Distribution of fossil flora and fauna.

- Similar remarks might be made about the ranges of certain flora and fauna as indicated by their distribution in the fossil record. For some examples, see the map to the right, which shows the distribution of various fossil species across the southern continents, which are represented in the relative positions inferred by the "drifters". Such a distribution would be very difficult to explain if the continents had been at their present positions when these species lived. "Fixists" used to postulate "land bridges" rising and falling between continents to explain these **biogeographical** features, but this was very much an *ad hoc* explanation, for which they could provide no mechanism; and it would not explain other features that are explained by drift.

Another piece of evidence that is inexplicable without drift is provided by past glaciations. For example, studying the **striations**, **moraines**, and other evidence for glaciation in South America, we find evidence of **glaciers** traveling westwards from the east coast of South America — as though glaciers had somehow climbed out of the Atlantic to roll across South America! But this is impossible and absurd: glaciers can't climb out of the sea. However, when we observe that there is also evidence of glaciers moving across Africa towards the west coast of that continent, and that dating methods put the South American and African glaciers at the same date, we can see that the absurdity is avoided by the hypothesis that the two coastlines were once contiguous.

- ### 3.6.3 How do we know? More modern evidence for drift

To the observations mentioned in the previous subsection we may now add further evidence that seems to belong in this article, since like the evidence in the previous section it does not require us to know *how* the continents move.

First, there is the fact that the continents *are* moving right now, as can be measured by **GPS** (the **Global Positioning System**), by **SLR** (Satellite Laser Ranging), and by **VLBI** (Very Long Baseline Interferometry).

Moreover, the continents are moving at the right rate. That is, if we apply geological dating methods to see how long ago the Americas parted from the Old World, if we measure the distance across the Atlantic, and if we measure the rate of drift using GPS, SLR, and VLBI, we find the numbers to be consistent.

If this in itself is not conclusive evidence that the continents have been moving for millions of years, it is at least highly consistent with that proposition.

We might also mention in this subsection the phenomenon of **apparent polar wander**, something which was only noticed in the mid-1950s, just before the plate tectonics revolution really got underway. You will recall from the article on the Earth's magnetic field that the magnetic poles observably drift around somewhat; also that

information about the Earth's magnetic field is preserved in the rocks.

It follows that we should be able to find out how the positions of the magnetic poles have changed in the past. However, when we try to do this, we find that the rocks on different continents give discrepant answers: all the rocks in South America (for example) will tell us that 60 million years ago (again, just as an example chosen at random) the north magnetic pole was in such-and-such a place, whereas the African rocks will all tell us that no, it wasn't, it was in another place altogether. It looks, in fact, as though there were two north poles, one controlling the deposition of rocks in Africa but having no effect on the rocks in South America, while the other affected South American **paleomagnetic** data while leaving Africa untouched — and the situation only gets worse when we start taking into account the data from Australia and India and so forth, and apparently discover that there were *lots* of north poles 60 million years ago, one for each continent.

But we can reconcile all these contradictory data with the hypothesis of continental drift. All we need to do is suppose that when the **paleomagnetic** data were laid down, they were all perfectly consistent and showed the north pole as being in exactly the same place, but that since then the continents have shifted their positions relative to one another and to the poles, leaving the **paleomagnetism** in the rocks pointing in different directions for each continent.

This is more than an *ad hoc* hypothesis to account for the **paleomagnetic** data. If we want to reconcile the data, we have to do more than wave our hands in the air and say: "what if the continents were in different positions once?" Instead, we have to propose specific positions that they were in: and we find that this exercise leads us to the same conclusions as we arrive at when studying the other data, such as **biogeography**, **glacial striations**, **geological formations**, and other evidence which we shall mention in subsequent articles.

3.6.4 The drift controversy: a historical footnote

When the drift hypothesis was first proposed, it met with stiff opposition; opposition which, with the 20/20 view of hindsight, seems unjustified. Drift could, after all, explain a number of anomalous features in geology; whereas the "fixists" had only an *ad hoc* explanation for the facts of biogeography and no explanation for anything else explained by drift.

The drifters were handicapped by having no good explanation for how drift could occur; indeed, they were handicapped by having a *bad* explanation: that continents plowed through oceanic **crust** like icebergs through water; an explanation which was demonstrably false.

Also, they did not then have the technology to prove that the continents are still in motion: they had no **GPS**, no **SLR**, and no **VLBI** to appeal to. If they could have shown that the continents were indeed moving, then the problem of explaining how could have been left to future geologists.

Nonetheless, in retrospect the drifters should have been granted more acceptance than they were. It is, after all, usually the case that scientists accept the existence of a phenomenon before finding a mechanism for it; science could hardly have progressed otherwise.

However, some popular authors have exaggerated the opposition to continental drift, portraying the leader of the "drifters", Wegener, as a modern-day Galileo persecuted by a scientific rather than a religious establishment. This is far from the truth. Wegener was well-respected as a scientist: his textbook on meteorology became a standard work; his book on continental drift was translated into several languages; he had a professorship specially created for him at an Austrian university; he won the support of some other geologists, mainly in Britain and continental Europe; and he was able to get funding for his research. In fact, he died of overexertion on an expedition to Greenland costing \$1.5 million in today's money, and if no-one had taken him seriously as a scientist he might have lived beyond 50.

Whatever the rights and wrongs of the drift controversy, it is a fact that drift was not widely accepted until the revolution in geology attendant on the birth of plate tectonics. We shall discuss this newer and more conclusive evidence in the next few articles of this textbook.

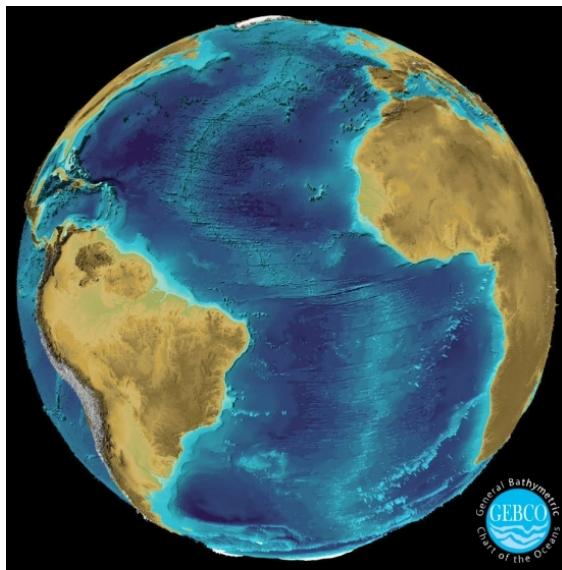
3.7 Sea floor spreading

In this article we shall explain what sea-floor spreading is, and the role it plays in plate tectonics; we shall conclude, as usual, with an explanation of how we know that sea-floor spreading is taking and has taken place. The reader will find it useful to be familiar with the article on **geomagnetic reversals** and the article on **marine sediments** before reading further.

3.7.1 The nature and role of sea-floor spreading

The sea floor is divided by a system of mountain ranges (**mid-ocean ridges**) each with a deep valley running down the center (**mid-ocean rifts**); on the bathymetric map to the right you can clearly see the mid-Atlantic ridge.

According to the theory of plate tectonics, **plates** move apart at the rifts. As the **lithospheric plates** move apart, this makes a gap into which magma intrudes; it also reduces the pressure on the **asthenosphere** below, causing



Bathymetric map showing the mid-Atlantic ridge.

partial melting of the **mantle** material. The intrusion of this material ensures that the rift is always being filled up by a fresh supply of oceanic **crust**. This whole process is known as **sea-floor spreading**.

One common misconception is that the intrusion of the **magma** at the rifts causes the motion of the **lithospheric plates**. In fact, geologists are well-agreed that this does little or nothing to cause the motion, rather, as explained in the previous paragraph, it is actually the parting of the plates at the ridges which causes the intrusion of the **magma**.

Nonetheless, sea-floor spreading plays a crucial role in plate tectonics: if the plates were unable to move apart at rifts, they would be unable to move at all.

In the remainder of this article we shall survey the evidence for sea-floor spreading.

3.7.2 Sea floor spreading: how do we know?

The proposition that the sea floor spreads out from the mid-ocean rifts, and has been doing so for millions of years, implies a diverse assortment of testable predictions, all of which turn out to be true.

Paleomagnetism

As we discussed in a previous article, the Earth's magnetic poles keep swapping their positions. This leads us to a prediction. If **igneous rocks** have been formed at and spreading out from the mid-ocean rifts, then when we look at the **paleomagnetic record** in the **igneous rocks** that form the oceanic crust, what we ought to see is a pattern of stripes of alternating normal and reverse magnetism

parallel to the mid-ocean rift and symmetrical around it: and this is in fact what we see. It was this discovery that almost overnight turned the concept of **continental drift** from a minority view among geologists to a widely-accepted idea..

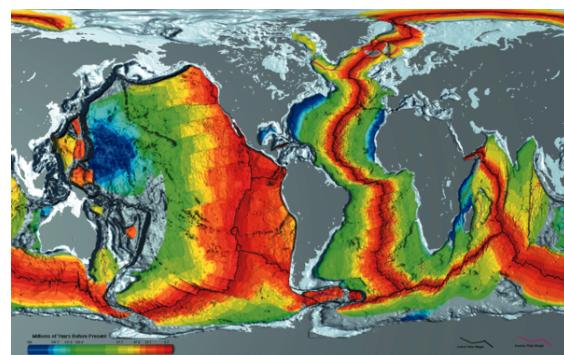
Heat flow

Heat is only conducted very slowly through large bodies of rock. Consequently, if hot rock is produced at the ridges and spreads out from them (cooling, of course, as it does so) we expect the flow of heat from the sea floor to be greatest at the ridges and to gradually decline as we look at the sea floor further away from them. And this is in fact the case (see Pollack et al, 1993, **Heat flow from the earth's interior: analysis of the global data set**, *Reviews of Geophysics* 31(3), 267-280, 1993).

It is this cooling process that explains why the rifts are flanked by ridges on either side, which gradually slope down as we get further from the rift: the newly produced rock is hotter, and therefore has greater volume than the older rock; the older rock, having moved further from the rift, has had more time to cool down and so to contract.

This, by the way, explains why islands capped with coral so often subside into the sea, as mentioned in the article on reefs: as the **basalt** of the islands is carried further geographically from the rift, and further in time from the heat in which it was formed, it and the rest of the oceanic **crust** below it will cool, contract, and therefore subside.

Dating



Map showing dates obtained for oceanic crust. Rocks are progressively old at greater distances from the rifts.

For the same reason, if the sea floor is spreading out from the rifts, another obvious prediction of the theory is that if geologists apply their dating methods to the **basalt** sea-floor on either side of a rift, the rocks will be found to be older the further out they are from the ridge system; as is the case, as shown in the map to the right.

Accumulation of sediment

Similarly, **sediment** will have been accumulating on the older parts of the sea-floor for longer than on the newer parts around the mid-ocean rifts, resulting in a deeper sedimentary layer further out from the rifts: this is also the case.

Fossils

Likewise, if the theory of sea-floor spreading is correct, then at any point on the sea floor the **fossils** found by drilling down to the bottom of the sea-floor sediment will be those deposited when that bit of the sea-floor was freshly produced at the rift.

This means that if we look at these deepest-buried fossils, we will see older and older fossils as we look further and further from the ridge; as a result we will see a greater proportion of extinct species. And this is in fact what we see.

The layer-cake effect

Different sediments tend to accumulate on different parts of the ocean floor. If the ocean floor stayed still, then, other things being equal, we would expect a sample of the sediment from any particular place on the sea floor to be pretty much the same all the way down.

But according to the theory of sea-floor spreading, the sea-floor has been continuously moving outward from the ridge systems like a conveyor belt, which implies that different sediments will have settled over the same portion of sea-floor as it moved.

So, for example, marine **carbonates** settle on the mid-Atlantic rift and rise, because in those shallow waters are above the **carbonate compensation depth**. Further out, where the waters are deeper, only **pelagic clay** will settle. So if the sea-floor really has been acting like a conveyor-belt, then when we take a sediment sample from areas of the Atlantic where pelagic mud settles, we should find this clay overlies a layer of **limestone** that settled when that portion of sea-floor was nearer the ridge; which is what we find.

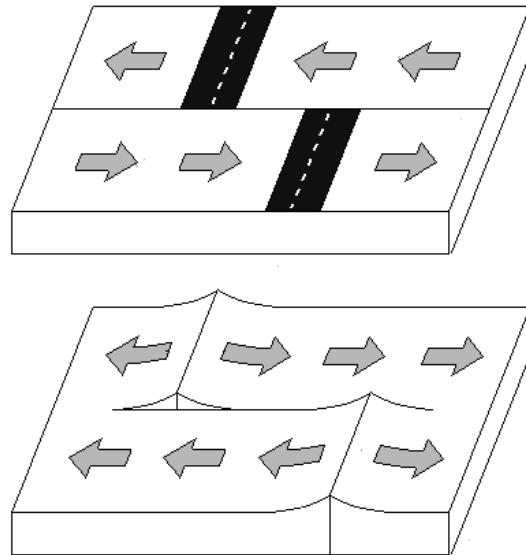
And in general we can state the rule that for any particular spot on the ocean bed, the layers of sediment from bottom to top should be consistent with the journey of the sea-floor from the ridge outwards; which is what we find.

So, for example, the conveyor belt moving northwest from the east Pacific Rise west of South America towards Japan crosses the equator and the region where **siliceous ooze** is deposited. So near Japan we should and do find (from bottom to top) **carbonate** sediments deposited in the shallow waters at the Pacific Rise; **pelagic clay** from deeper and non-equatorial waters; **calcareous/siliceous ooze** as the conveyor belt crosses the equator; and more

pelagic clay that accumulates north of the equator.

Transform faults

The mid-ocean ridges do not run in a continuous line on the ocean floor: rather, they are discontinuous, being displaced laterally along their length at faults, as can be seen in the map near the top of this article. This leads to an interesting prediction.



Top, an ordinary left strike-slip fault; bottom, a transform fault at a mid-ocean ridge.

The top picture in the diagram to the right shows an ordinary strike-slip **fault** such as the San Andreas fault, with a road cutting across it displaced by motion along the fault. From either side of the fault, one sees the road as being displaced to the left, making it a *left* fault. Clearly, if you were standing on one side of the fault during an earthquake, you would see the land on the other side of the fault moving to the left relative to you.

Now consider the lower picture in the diagram. Standing on either side of the fault and looking at the ocean floor on the other side of the fault, you would see the mid-ocean ridge as having been displaced to the left on the far side of the fault.

But if geologists are right about the sea floor spreading out from the mid-ocean ridges, then if you stood on one side of the fault and looked across it during an earthquake, you would see the sea floor on the opposite side moving to the *right* relative to you. And this is what we do in fact observe.

Rate of continental drift

The measured rate of sea-floor spreading at the ridges agrees with the measured rate of **continental drift**, and

the inferred rate at which geologists calculate it must have taken place in the past.

Structure of oceanic crust

From drilling through the oceanic **crust**, and by looking at the objects known as **ophiolites**, we can find out the structure of the crust, which is consistent with the theory of sea floor spreading and completely inexplicable without it. I shall not go into details here, as this topic will be covered in the main article on ophiolites.

Conclusion

All these disparate lines of evidence add up to a convincing demonstration that the sea floor is currently spreading from the mid-ocean rifts, and has done so in the past.

3.8 Subduction



A volcano in the Andes, South America, a symptom of subduction.

Subduction is the process whereby the edge of one **plate** slides under another and into the **mantle**.

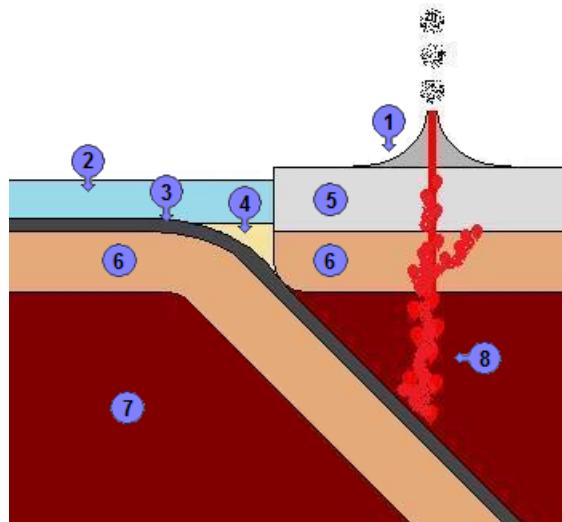
In this article we shall examine the evidence for subduction and the role it plays in plate tectonics. The reader should already be familiar with the previous article on sea floor spreading.

3.8.1 Subduction: an overview

As we have discussed in the article on sea floor spreading, rock is generated at mid-ocean rifts; when it is produced in this way it is hot and therefore buoyant. As rock spreads out from the rifts, it cools down, and also thickens as material from the upper **mantle** accumulates on the underside of the cooling **crust**, both of which processes make it denser. Eventually, as a result of this, it becomes gravitationally unstable and liable to plunge into the **atenosphere**.

It cannot simply sink straight down like a foundering ship, for two reasons. First, it is still attached to the more buoyant portion of the plate. Second, in order to do so it would have to displace the material of the **atenosphere** in a way that is physically implausible. To grasp this point, consider a sheet of paper floating on the surface of a bowl of water. The paper is more dense than the water and should sink, but it cannot sink straight down in its horizontal attitude without pushing up the water surrounding it. However, if one edge of the paper is forced beneath the water, then the sheet of paper can descend down and sideways, sliding edge-on through the water.

And this is what happens to oceanic **crust** when it subducts; forced beneath another more buoyant plate, it slides edge-first into the **atenosphere**. The portion of the **plate** being thrust down in this way is called the **slab**. A typical angle of descent is 45 degrees, but the slab can descend at any angle from near-horizontal to near-vertical.



Subduction. Key: (1) a volcano (2) the sea (3) oceanic crust (4) a trench containing an accretionary wedge (5) continental crust (6) lithospheric mantle (7) the atenosphere (8) partial melting of the slab.

The diagram to the right shows a cross-section of the process. (The reader should note that the diagram is somewhat schematic in nature and not entirely to scale.)

In the diagram, oceanic **crust** (on the left) is being subducted under continental **crust** (on the right). The triangular gap formed where it bends to go under the continental **crust** is known as a **trench**. For reasons that we shall discuss below, the subducted slab undergoes partial melting at a depth of about 120 km, producing volcanoes above the melting.

In the diagram I have chosen to show oceanic crust being subducted beneath continental crust. An oceanic plate can equally well subduct beneath another oceanic plate with much the same results, only forming a chain of volcanic islands and **seamounts** instead of a volcanic moun-

tain range.

In the diagram I have shown the trench filled up with sediment, forming what is known as an **accretionary wedge** or **accretionary prism**. However, not all trenches have such a feature: this depends on the rate at which sediment is transported into the trench and the rate at which it is subducted along with the plate. As a general rule, an accretionary wedge will be present when the trench is close to a continent, as this ensures a greater supply of sediment than can be found further out to sea.

3.8.2 Evidence for subduction

In this section, we shall look at the evidence for subduction.

The existence of sea-floor spreading

As we have seen in the article on sea floor spreading, new crust is generated at the mid-ocean rifts and spreads outward from them, and has been doing so for millions of years.

This raises the obvious question of where it goes to. The earth's surface cannot simply inflate like a balloon; and looking at the sea floor, we can see that neither does it ruck up like a carpet too big for the room it's laid in. The only possible explanation is that crust is being destroyed, and the only reasonable way that this can happen is for the crust to be recycled back into the **mantle**. So the evidence for sea floor spreading is on the face of it evidence that subduction must happen: or at the very least that *something* must be destroying oceanic **crust**.

Age of the sea floor

One consequence of the production of the sea floor at rifts and destruction of sea-floor at subduction zones is that the sea floor should consist of rocks younger than the Earth itself. And this is what we find: the oldest rocks on the ocean floor date to about 200 million years old, as compared to 4 billion or so years for the oldest continental rocks.

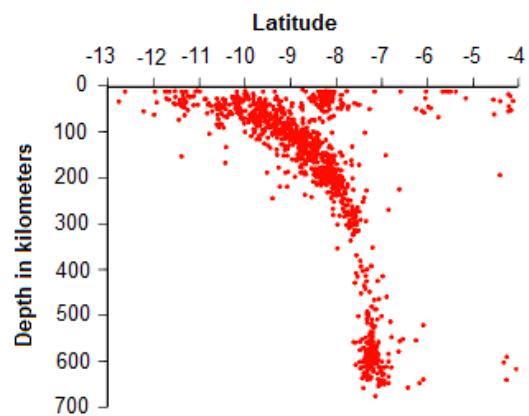
For the same reason, the layers of **sediment** on the ocean floor are thinner than those on the continents, and the oldest **fossils** are more recent.

As with the previous point, this is good evidence that the oceanic **crust** is being destroyed by *some* mechanism. We can now turn to the evidence that the mechanism is indeed subduction.

Earthquakes

If a slab of cold, brittle, elastic rock is really being subducted into the **athenosphere**, then we should expect it to

undergo earthquakes as large bodies of rock in this state always do when placed under stress.



Plot of recorded earthquakes where the Indo-Australian plate is subducted under the Lesser Sunda Islands.

Because it is possible to measure the depth of an earthquake (see [here](#) for details) as well as its latitude and longitude, observations of earthquakes allow us to form a good picture of what happens when a slab descends into the **athenosphere**. The diagram to the right, for example, shows the locations of earthquakes of magnitude 4.5 or greater occurring between 1973 and 2004 where the Indo-Australian plate is subducted under the Lesser Sunda Islands. The profile of the slab is clearly traced out by the incidence of earthquakes. (The US Geological Survey's [Earthquake Hazards Program](#) has many such maps of subduction zones.)

We should note that without subduction there shouldn't be earthquakes below the **lithosphere** at all. In rocks below the **lithosphere** the temperature and pressure are such that rocks under stress at that depth should bend, stretch and flow, producing no earthquakes. The only plausible explanation for quakes at these depths is that colder more brittle rock from the surface has been thrust down into the **athenosphere**. (In considering this, the reader should bear in mind that the slab is several kilometers thick and that heat flows very slowly through rocks, so there is no difficulty in a subducted slab remaining **brittle** and relatively cold for millions of years after it enters the **athenosphere**.)

Volcanoes and subduction melting

Water lowers the melting point of rock in a manner similar to how salt lowers the melting point of ice; the result is that when the wet rocks of the **crust** reach the right zone of temperature and pressure, about 120 km below the surface, the rocks of the **crust** undergo partial melting. (The fact that this should occur at 120 km depths can be verified in the laboratory by simulating the appropriate temperature and pressure.)

Naturally the hot and relatively less dense molten rock

rises through the athenosphere and escapes to the surface as **lava**, forming volcanic mountain chains if the slab is subducting under continental crust, or chains of volcanic islands and **seamounts** if the slab is being subducted under a continental crust. It is estimated that up to 20% of subducted rocks are returned to the crust by this mechanism.

Because we can use seismic readings to measure the angle of the slab, and because we know that partial melting occurs at a depth of about 120 km, we can use simple geometry to calculate where, relative to the subduction zone, the volcanic arcs should occur. The fact that they always do occur in the right place supports the idea that geologists really do know what's going on in the athenosphere.

Trenches

The existence of trenches were once a mystery to geologists. But the mechanism of subduction makes everything clear: there must be a trench, roughly triangular in cross-section, between a descending slab and the plate under which it is subducted.

If this is the true explanation, then we should expect to find these trenches (and under the right circumstances **accretionary wedges**) in geographical association with the earthquakes and volcanism which we also attribute to subduction; and this is in fact what we find.

Conclusion

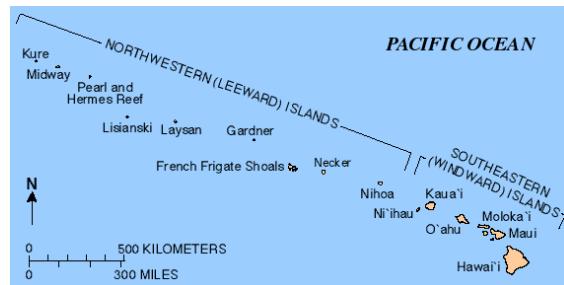
The Earth looks as it should if subduction really is taking place; what is more, there are phenomena such as earthquakes in the athenosphere, or the presence of chains of volcanoes at just the right distance from subduction zones, which no-one can explain on any other hypothesis. It is therefore reasonable to conclude that subduction really is taking place.

3.9 Hotspots

In this article we shall discuss the evidence for hotspots and what, if anything, they can tell us about plate tectonics.

3.9.1 Aseismic ridges

In several places the sea floor is scarred by what are known as **aseismic ridges**. These consist of a linear trail of volcanic islands or **seamounts**. When dating methods are applied to these, it is invariably found that the volcanic rocks along the ridge are arranged in order of age, with the youngest rocks at one end and successively older rocks



The Hawaiian islands. Image from USGS.

as we look further along the ridge. Any current volcanic activity will be found at the young end of the ridge.

The canonical example of an aseismic ridge is the Hawaiian islands, shown in the map to the right.

Dating methods reveal that the islands are arranged in order of age from east to west, with Hawaii being the youngest, Maui being older than Hawaii, Oahu being older than Maui, and so on along the island chain. Hawaii itself (the “Big Island”) is volcanically active; Maui, next in line, has not erupted since the eighteenth century; Molokai and all the islands further west are considered to be extinct. To the east of Hawaii is an area of seafloor volcanism known as Lohi (not marked on the map); if this activity continues for ten or twenty thousand more years it will build a new Hawaiian island younger still than Hawaii.

3.9.2 Hotspots

With the discovery of **plate tectonics**, geologists were able to explain aseismic ridges. They said: suppose there are places in the **mantle** where, for some reason, hot rock tends to rise in **mantle plumes**. The point on the surface above such a plume would be a **hotspot**, a place of high volcanic activity.

Now consider the fact that the **lithospheric plates** are *moving*, as explained in the previous articles. This would perfectly explain such things as the Hawaiian island chain: currently, the mantle plume is under Hawaii, but before that it was under Maui, and before that it was under Molokai, and so on. If you picture a conveyor belt slowly moving over a blowtorch, then you have the theory of hotspots in a nutshell: the mantle plume is the blowtorch, the plates in motion are the conveyor belt, and the trail of burned and molten material is the volcanoes.

The reader may be wondering why mantle plumes exist, and why they apparently persist in the same place for millions of years. The geological community replies that this is “not fully understood”, which is how geologists say: “We have no idea. Can you get us some research funding and come back in ten years?”

3.9.3 Hotspots: how do we know, and what do they tell us?

Hotspots are sometimes adduced as evidence for the motion of plates. But we should be extremely cautious in saying so. Hotspots, we are told, stand still, so the trail of islands and **seamounts** they leave behind them indicate the motion of plates. But how do we know that hotspots stay still? Geologists do not as yet have such a good grasp on the dynamics of the **mantle** that they can declare on theoretical grounds that hotspots *should* stay still. Rather, the fact that they stay in one spot is itself an inference from the fact that the plates are moving: if we calculate the motion of the plates using dating, **paleomagnetism** and so forth, we see that the trails made by the aseismic ridges are explicable on the hypothesis that the hotspots are stationary. Direct measurement also supports this hypothesis: for example, the Hawaiian islands are measurably moving westwards; if the hotspot stayed still, this would account for the observations.

So the idea that hotspots are more or less stationary is well-supported by the evidence, but all the evidence for this comes down the evidence that the **plates** are moving. If we were then to take it as a premise that hotspots are stationary, and use this as evidence that the **plates** are in motion, we would be guilty of the grossest form of circular reasoning. Some textbook accounts of plate tectonics fall into this trap.

There is, however, a way in which the existence of hotspots does suggest that the plates are moving. For it is a fact that all the aseismic ridges of a given **plate** will trend in the same direction: if one aseismic ridge runs east-west and is oldest at the east end, then that will be true of every aseismic ridge on that **plate**: for example, the Hawaiian islands display the same trend as the Louisville Seamounts. Now it is more parsimonious to believe that all the hotspots are staying still and that the common trend of the aseismic ridges is produced by the motion of the **plate**, than it is to believe that the **plate** is standing still but that all the hotspots on any given **plate** just happen to be moving in the same direction.

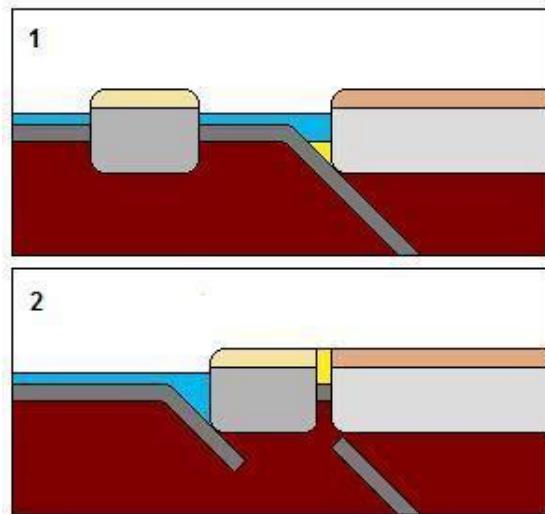
For this reason, the evidence from hotspots is indeed evidence for the motion of plates.

3.10 Terranes

In this article we shall discuss what a terrane is and how the existence of terranes can be explained by the mechanisms of **plate** tectonics already discussed.

3.10.1 What is a terrane?

A **terrane** is a part of a landmass, bounded by tectonic faults, which is different in many ways from the main landmass to which it is attached.



Schematic diagram of terrane docking.

Geologists ascribe this to the terranes having originally been separate from the main landmass, and being borne up against the edge of the landmass by the motion of plates. As we have discussed, oceanic crust is usually simply subducted beneath continental crust. But if some material (for example an island) can not or does not subduct (because of its greater thickness or lower density than oceanic crust, or for some other reason) then instead it will **dock** against the edge of the landmass.

One example of terrane formation is shown in highly schematic form in the diagram to the right. In picture 1 we see a microcontinent traveling from left to right along with the oceanic **crust** in which it's embedded; at the right of the picture we see the oceanic **crust** subducting beneath a continent.

In picture 2 the microcontinent has collided with the continent, incidentally squeezing up the **sediment** in the **accretionary wedge** between them. In this particular example, this has produced not one but two terranes: the former island and the former sedimentary wedge. The former microcontinent is now the left end of the continent, and the oceanic **crust** is now **subducting** beneath that.

The result is that after the docking has taken place we have three regions sitting side by side which are bounded by **faults** and which have no common geological history. Of course, *after* the point at which they dock, they can and will share a history; a **lava flow**, or an **fall** of **volcanic ash**, or deposition of **sediment**, can then cover them all in common; or a river could cut its way through all three. But *until* they join, there is no reason why they should have anything in common.

The map to the right (from the United States Geological Service) shows some of the terranes on the western coast of North America, where they have been particularly well studied. The terranes are color-coded by type.



Terranes on the west coast of North America.

Note that not all terranes are formed in exactly the manner of our example above; however, they all do appear to have been formed in the same *general* manner, i.e. by material being transported by the motion of plates to the western edge of the continent and sticking there.

3.10.2 Terranes: how do we know?

We have said that a terrane will have a different geological history from the surrounding areas on the same landmass. For example:

- The terrane will have different types of sedimentary rocks from the main landmass.
- The sedimentary rocks will typically contain different fossils from the main landmass.
- Also, the fossils and sediments often indicate a different environment from the main landmass; they may, for example, be indicative of a marine environment.
- Not only this, but the fossils in a terrane may indicate a different climatic zone from the landmass to

which it is attached: for example, they may be tropical fossils, such as coral, found in an environment which is far from being tropical.

- The terrane will often look like it was once something *in particular* other than just another part of the continent; it will look like it has formerly been oceanic sediment, or a volcanic island arc, or a piece of oceanic crust.
- The basement rock of the terrane is often different from that of the main landmass in obvious ways: we may, for example, see basalt in the terrane adjoining granite in the main landmass.
- The basement rock will frequently yield different dates on different sides of the fault when absolute dating methods are applied.
- Extensive geological features such as lava flows or layers of volcanic ash may terminate abruptly at the edges of the terrane.
- The terrane will have a different paleomagnetic history from the landmass to which it is adjoined.

What is more, these features will go together. That is, the fossil evidence, the paleomagnetic evidence, the evidence from dating, the different sediments, the different basement rocks, and so forth, will all identify exactly the same fault-bounded area as being a terrane. If you really wanted to, then perhaps you could try to explain each of these features away one by one (or at least try to imagine that there's some explanation which you can't think of yet) but how would you go about explaining why these features of a terrane are coterminous?

Terranes look, in short, as though they came from somewhere else and were somehow tacked on to the main body of the landmasses to which they are now attached. And so it is reasonable to conclude that this is in fact what happened.

Furthermore, we can understand why they should have been transported and attached in this way. What would otherwise be a puzzle becomes an *expectation* in the light of plate tectonics. What would be more natural than to find seamounts, volcanic islands, and marine sediments scraped off the edge of a subducting plate?

This view of terranes is strengthened by the observation that the evidence shows that terranes became joined to landmasses at times and in places where on other grounds we think that subduction was taking place. For example, since the Atlantic ocean is being formed by rifting, we should not expect to find any terranes on the Atlantic side of America which date to younger than the Atlantic; and we don't. Meanwhile we find plenty of more recent terranes on the west side of America, where the Pacific oceanic crust is subducting beneath the continental crust.

For these reasons, it makes sense to explain the terranes as being caused by the motion of **plates**. Indeed, this successful explanation of the existence of terranes must be taken as evidence for the **theory of plate tectonics**, and for the **subduction** of oceanic **crust** in particular, insofar as the existence of terranes makes no sense at all without that theory and make perfect sense in light of it.

3.10.3 Note on vocabulary

The word “terrane” originally just meant “an area having a preponderance of a particular rock or rock groups”. The terranes which we've talked about in this article were called “exotic terranes” or “allochthonous terranes”, “tectonostratigraphic terranes”, “accreted terranes” and so forth. However, the word “terrane” is now pretty much synonymous with an area distinct from the surrounding landmass because it has been transported by the motion of plates and has then docked against another landmass.

The reader should bear this in mind when looking at older texts: the words “exotic terrane”, for example, do not imply anything *more* exotic than what in this text we shall simply call a “terrane”.

3.11 Ophiolites



Pillow basalt from an ophiolite, Italy.

In this article we shall discuss ophiolites, how to recognize them, how they originate, and how we know this.

The reader may find it helpful to re-read the article on **igneous rocks** before going any further.

3.11.1 What is an ophiolite?

Here we shall describe a classic “Penrose type” ophiolite; geologists recognize variations on this type for reasons that we will discuss later in this article.

An ophiolite, then, is a sequence of rocks exposed above

sea-level and usually emplaced on or in continental **crust**. This sequence consists of (reading from the top down):

- Marine sedimentary rocks such as chert, limestone, and rock formed from **pelagic clay**. This top layer is often not counted as being part of the ophiolite, since this term is usually reserved just for the **igneous rocks** involved.
- **Basalt**, which is by definition an **extrusive** rock; often in the form of **pillow basalt**, which the reader should remember is only formed underwater.
- A **sheeted dike** complex: **dikes** which do not (as they usually do) intrude into some other rock, but which rather stand side by side rather like books on a shelf.
- **Gabbro**, which you should remember is the **intrusive** counterpart of basalt.
- Ultramafic rocks.

This sequence is known as the **pseudostratigraphy** of an ophiolite.

The diagram to the right shows an cross-section of such an ophiolite. To be precise, it shows a cross-section of a *vertical* ophiolite, in which the layers are horizontal and the sheets are vertical. We may perfectly well find an ophiolite lying on its side, as a result of the tectonic events that emplace it; but the relationships between the rocks remain the same.

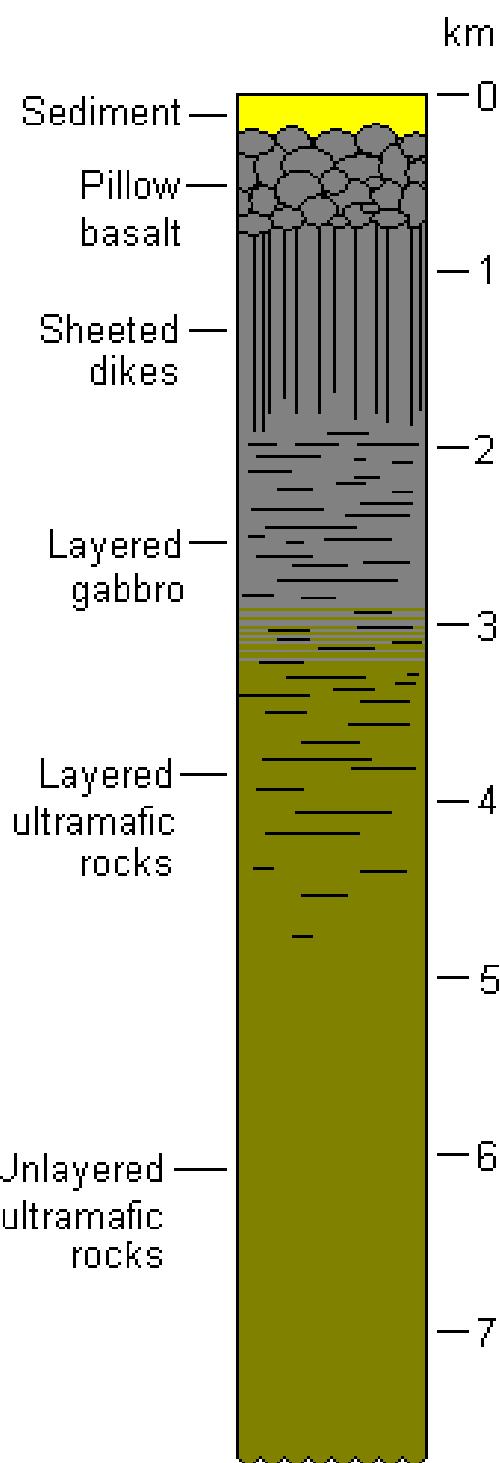
So, an ophiolite is recognized by being something which looks like that — but what *is* an ophiolite? According to geologists, an ophiolite is a section of oceanic **lithosphere** from the upper mantle through the **igneous crust** and up to the **sediment** on top. (You may note that this means that an ophiolite is a kind of terrane; the subject of ophiolites is of sufficient complexity that I have thought it best to give them their own article.)

To assert that ophiolites are slices of oceanic **lithosphere** is a bold claim, since it is not immediately apparent what such a thing would be doing emplaced on or in continental **crust**. It is therefore time to ask ...

3.11.2 Ophiolites: how do we know?

The reason that geologists think that ophiolites are pieces of oceanic **lithosphere** is that this is exactly what they look like.

The marine sediments on the top of an ophiolite obviously speak of a marine environment. The **basalt** is the characteristic **igneous rock** of the sea-floor, and the fact that it is usually **pillow basalt** agrees with this, since **pillow basalt** only forms underwater.



Cross-section of an ophiolite.

The sheeted **dikes** became explicable with the discovery of **sea floor spreading**. Since the mid-ocean rifts are linear, the route between the **magma** and the surface would naturally be shaped like a sheet, resulting in the formation of a **dike**. As the two sides of the rift move apart, this would lead to more dikes intruding between the previous dikes.

The **gabbro** in ophiolites would represent material which was not extruded, moving out laterally from the **magma** chamber beneath the rift.

The **ultramafic rocks** would represent a portion of the upper **mantle**: as we have discussed in the article on the structure of the Earth, this is what we expect the mantle to be made of.

After the discovery of **sea floor spreading** and the significance of sheeted dikes, geologists were quick to realize that if ophiolites represented a cross-section of the oceanic **crust**, then this would account extremely well for the results obtained by seismic studies of the oceanic crust.

It has been possible to drill and sample the oceanic crust through the sediment, through the **pillow basalt**, through the sheeted dikes, and down into the region where the dikes give way to **gabbro**; this was achieved at Hole 1256D of the Ocean Drilling Project. At the time of writing, no-one has managed to drill all the way from the sea-floor sediment to the upper **mantle**, but as far as anyone *has* drilled the oceanic **crust** does indeed look just like an ophiolite — and not just in the large-scale **pseudostratigraphy**, but in fine detail (see [here](#), for example, for a discussion of how closely **gabbro** recovered from the hole resembles gabbro from an ophiolite.)

Where drilling cannot take us, nature has obligingly offered us another way to look at oceanic **crust**: so-called **tectonic windows** such as Hess Deep. This is a rift in the ocean floor comparable in size to the Grand Canyon: submarine exploration of the Deep has shown that the sides of the rift display the same **pseudostratigraphy** as an ophiolite, with **ultramafic rocks** at the bottom, then **gabbros**, then sheeted dikes, and so forth.

For all these reasons, geologists identify ophiolites as portions of oceanic **lithosphere**. Now when oceanic **lithosphere** moves towards continental **lithosphere** it is usually **subducted** beneath it, as discussed in the article on **subduction**; but if for some reason some of it was **obducted** (thrust up onto a continent) perhaps because the zone of subduction was so close to the ridge where the crust was produced that the crust was still elevated and low in density, then what we would get would be an ophiolite.

3.11.3 Note on varieties of ophiolite

I would be painting an inaccurate picture of ophiolites if I did not point out that they come in several varieties. The particular type of ophiolite I have been using as an example does not even constitute a majority of those known to geologists. I have used this type as an example because they shed an interesting light on **sea floor spreading**, but I should now mention some of the variations on the ophiolite theme.

In this article I have taken as paradigmatic those ophiolites which display a complete sequence of **ultramafic**

rocks, gabbro, sheeted dikes, basalt, and marine sediment. But not all ophiolites have sheeted dike complexes, and some also lack the gabbro.

Likewise I have concentrated on one particular mode of production of ocean crust, at mid-ocean ridges. I have been describing MORB (mid-oceanic ridge basalt) ophiolites; but geologists also recognize other varieties such as LIP (large igneous province) ophiolites and SSZ (supra-subduction zone) ophiolites. In [this article](#) the reader can find out more about SSZ ophiolites, with special reference to why they don't have sheeted dike complexes.

Finally, I have mentioned only one way that ophiolites get attached to a continent, but geologists know of others. We should note that all such mechanisms must involve extremely rare events, since the world's total stock of ophiolites can only be a fraction of one percent of the total oceanic crust produced during the Earth's history. This explains why there is frequently controversy among geologists as to the mechanism by which particular ophiolites are emplaced; it is much easier to understand the general rule (i.e. subduction) than it is to understand what must necessarily be rare freak events.

If, then, I was to attempt to describe all the varieties of ophiolites in terms of both their igneous origin and their mode of emplacement, to explain how geologists tell one from the other, and to say why in certain cases there is debate, I should run the risk of unbalancing the textbook, which would have to be retitled *Historical Geology and Ophiolites*. Instead, let us try to say what ophiolites have in common.

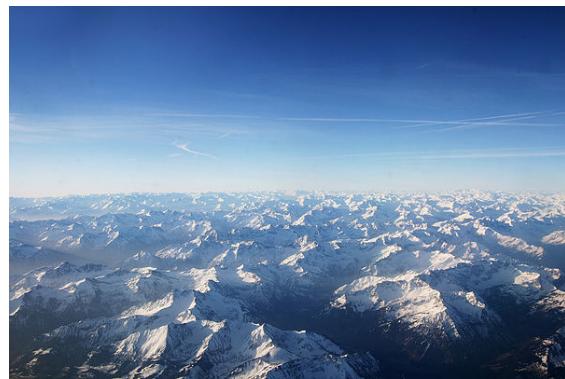
In terms of their composition, what they have in common is that they contain *igneous* marine rocks, and so clearly originated as oceanic lithosphere.

In terms of their history, what they have in common is that they cry out for an explanation in terms of *plate tectonics*.

There are other reasons behind *plate tectonics* why marine *sedimentary* rocks are found in continents on top of *continental* crust, as we shall discuss in later articles, but when we find actual *igneous* oceanic *crust* emplaced in a continent, then we have to seek an explanation in terms of the forces we know of that move bits of the Earth's crust around.

3.12 Orogeny

In this article we shall discuss the concept of orogeny, and investigate how *plate tectonics* causes the formation of mountains. It will be helpful to the reader to recall the facts discussed in the articles on *subduction*, *terranes*, and *ophiolites*.



The Alps.

3.12.1 Note on the word “orogeny”

The term **orogeny** means, literally, “the formation of mountains”. It is sometimes used to mean just that; so, for example, we may use the term “subduction orogeny” to refer to the volcano- and island-arc-forming process described in [the article on subduction](#).

However, “orogeny” has come to mean a period of intense **folding** and **faulting** of large areas of crust as a result of lateral pressure.

The two meanings may be, and often are, combined, so that it means **folding** and **faulting** of this kind *which causes mountains*.

When I use the term “orogeny” without any further qualification (i.e. unless I specifically say that I am talking about *subduction orogeny*) I shall be using the word in this last sense.

3.12.2 Orogenic mechanisms

In orogeny, mountains are formed when a continent meets another continent, a micro-continent or an island arc, and instead of one **subducting** beneath the other, they push together and buckle, forming mountains. Hence if we wish to distinguish this mechanism from subduction orogeny, we may call it “*collisional orogeny*”.

Now, this mechanism for mountain-forming is at least plausible, for the following reasons:

(1) When (for example) two continents collide, each will be too thick, and too low in density compared to the *athenosphere*, for one to simply subduct under the other.

(2) There is no reason why, when the two continents meet, the forces that brought them into collision should then regard their job as done and knock off: whatever propelled one continent into another will, it seems safe to suppose, continue to impel the continent.

(The reader should note, however, that the continents do not continue to move because of their “momentum”, as some people carelessly say. The momentum even of

something with the mass of a continent is negligible when it's only moving at a speed that can be measured in centimeters per year. Rather, as we have said, they continue to move because the forces that brought them together continue to push them together.)

(3) We know that rocks can deform under stress; in fact, geologists perform experiments with plasticine to gain insights into orogeny, on the grounds that two slabs of plasticine in collision should behave on a small scale over a small period of time much as two continents in collision will behave on a larger scale over a longer period of time.

But it is one thing to be plausible, another thing to be right. In order to show that the concept of collisional orogeny is more than just the thought of an idle hour, it is necessary to look at the details of mountain ranges.

3.12.3 How do we know?

We are used to the fact that mountains come in belts (the Urals, the Himalayas, the Appalachians, and so forth) but when you think about it the fact is rather striking. Why do mountains come in long and relatively thin strips rather than being arranged in big circles (for example) or being dotted randomly here and there about the landscape?

Well, a subduction orogeny will produce belts (such as the Andes) because the boundary between two plates is necessarily a line; and collisional orogeny will produce belts (for example the Himalayas) for exactly the same reason: the boundary (the **suture zone**) between two adjoining pieces of continental crust must in the nature of boundaries be linear.

When we look in detail at orogenic belts, we see that the nature of the folds they contain is consistent with the theory of their origin. Consider a piece of deformable material such as a floor rug. If you push it from one end so as to shorten the area it occupies, it must necessarily ruck up in a series of parallel folds; if this process continues some of the folds will become **recumbent**: they will fall over sideways.



A recumbent fold.

This is just what we see in orogenic belts. The photo-

graph to the right shows a fine example of a recumbent fold belonging to the Caledonian orogeny.

Then there is the matter of sedimentology. Before the discovery of plate tectonics, geologists recognized that non-volcanic mountain ranges tended to consist largely of prisms of marine sediment (often metamorphosed); they just didn't know why. It was simply the fact that these great accumulations of sediment would form in the ocean and then climb out of the sea and become mountains, and the most that geologists could say about it was that this was just the sort of thing that tended to happen from time to time.

But this can be explained in terms of collisional orogeny: before a continent meets another continent (or micro-continent, or island arc) there must be sea floor between them; this will contain the sediment of the continental shelf and slope. Naturally when the two landmasses meet this sediment will be squeezed between them.

Because in order for the two landmasses to meet the oceanic crust between them must be **subducted**, we might well expect to find signs of subduction orogeny accompanying collisional orogeny: we would expect to see volcanoes on at least one landmass or the other, or volcanic island arcs between them which would then be trapped when they meet to form a terrane. And this is in fact what we find.

The idea that orogenies are collisional also explains why **ophiolites** are found in mountain belts. Recall that these are sections of oceanic crust: what, then, are they doing in the middle of the Alps, the Appalachians, the Himalayas, the Urals, the Apennines, the Klamath Mountains, the Cascade Mountains ... ?

But this is all explained on the collisional theory of orogeny. These ophiolites are there because that's where the ocean was, before two landmasses collided; they remain as clear evidence of the suture zone.

3.12.4 Case study: the Himalayas



Folded rocks in the Himalayas, Kali Gandaki, Nepal.

As an example of all that we have been discussing, con-

sider the Himalayas.

Note first of all that they do in fact form a mountain belt; they have a long thin topography. Along the north edge, we have the Transhimalayas, which look just like they were produced by **subduction** volcanism; further south we have 10 - 17 kilometers of marine sedimentary rock and metamorphosed sediments; at the Indus Suture Zone where the two plates meet we have **ophiolites**.

The whole Himalayan range has very evidently been smashed, mashed, folded and faulted, and contains many recumbent folds.

To the south of the Himalayas is of course India, which may be considered one vast **terrane**, having its own distinct geological, paleomagnetic, and **fossil** history up until the point where it crashed into south Asia; if, as this evidence suggests, they were once separate, they must at some point have collided.

Finally, we may note that in this case the collision is still going on: India is measurably still moving north by about 5cm/year, the Himalayas are still measurably undergoing uplift at about 1cm/year, and Asia north of the Himalayas is still measurably stretching and deforming as a result of the collision. So when we find something that looks similar to the Himalayas but where the motion has ceased, we are entitled to infer motion in the past.

Chapter 4

Stratigraphy

4.1 Actualism



Explanations involving magic are usually frowned upon in the natural sciences.

In this article we shall discuss the concept of actualism and some common misconceptions which surround it.

4.1.1 What is actualism and why?

Actualism in geology is the idea that the facts of geology can and should be explained by in terms of the sort of physical processes that actually happen.

As such, it can be considered both as a scientific theory (that the facts *can* be explained by real processes) and as a methodological principle (that they *should* be so explained).

You might wonder why there is a need to have a name for this. You might also wonder why it is necessary to men-

tion it particularly in a textbook on geology, since after all this is a universal scientific principle. Chemists (to take one example) are just as much actualists as geologists, because all scientists are; and yet they do not include a section on actualism in their textbooks.

The reason why it is mentioned particularly in geological textbooks is a historical one. For many centuries people have been trying to explain geology in terms of non-actual, magical processes: explaining, for example, that the Earth's strata were produced by God turning off the force of gravity and then turning it back on; or that God created fossils when he made the Earth so that coal-miners even when underground would have visible signs of his presence. Various religious sects still promote non-actual concepts of geology to this present day.

4.1.2 Actualism as a theory

As we have said, actualism (considered as the assertion that the geological record *can* be explained in terms of real processes) should be regarded as a scientific theory. Why? — because it is testable. We can look at the rocks, and we would recognize if there was something in the geological record which could not be explained in terms of real processes.

In fact, as you will recall from previous articles, what we find is that we can explain what we see: we can explain glacial till in terms of glaciers, marine limestone in terms of the deposition of calcareous ooze, chemical weathering in terms of chemistry, paleomagnetism in terms of continental drift, saline giants in terms of the evaporation of seawater, and so forth.

If there are still things that are not yet perfectly explained, such as the question of how *exactly* glaciers make drumlins, then we can hardly regard that as a falsification of the theory, but merely an area in which more work needs to be done, for it is not plain that drumlins *cannot* be explained by a better understanding of actual glacial processes and would instead require the invocation of a non-actualistic being such as the Drumlin Fairy to fill this minor gap in our knowledge.

If, on the other hand (for example) we split open two

leaves of slate and found therein the first chapter of Genesis written in quartz, then this would falsify actualism; we could not even imagine that one day we would find any ordinary physical process that would explain the phenomenon: we know too much about the way in which the world works to consider that even for a moment.

4.1.3 Actualism as a methodological principle

Considered as a methodological principle, actualism may be stated in the phrase: “If it looks like a duck and it quacks like a duck, it’s a duck”.

Take aeolian sandstone, for example. It looks exactly like lithified aeolian sand; we can understand it perfectly well in those terms. Therefore, this is the most parsimonious way to understand it: it is simply unnecessary to imagine an unknown unobserved process to explain what can be explained by a known observable process.

Now, there are some people who (for religious reasons) dislike this: they wish, for example, that **aeolian sandstone** was something else altogether. These people, let us hasten to say, are perfectly entitled to their own beliefs. But they are not entitled to pass such beliefs off as scientific: when they daydream about alternative magical processes that might have formed something that *looks exactly like* lithified aeolian sand formed by actual processes, they have abandoned the scientific method in favor of wishful thinking.

For the proposition: “If it looks like a duck and it quacks like a duck, it’s a duck” stands at the heart of all scientific thought. We may imagine that it is not a duck; we may imagine that it is in fact a magical fairy disguised as a duck. It is in a sense as easy to imagine this as that it is a duck, for the human imagination is not constrained by actualism. But the scientific method *is* constrained by actualism: within that method we cannot put the fairy hypothesis ahead of the duck theory, we cannot even place them on the same level. The idea that it is a duck is to be preferred for all scientific purposes unless and until we find evidence that it’s a magic fairy. Those who prefer to think otherwise have not merely stepped outside the edifice of the scientific method, they are throwing bricks through its windows.

4.1.4 Naturalism

We should note that the adoption of actualism is not the same as the adoption of **philosophical naturalism** (the rejection of the existence of processes other than the physical). Supernatural beings and processes may perfectly well exist; it is no function of this textbook to pronounce on such a question. It is simply that we can see no evidence in the geological record that such processes have ever been involved in geology; hence actualism succeeds

as a theory, as I have explained.

And, that being so, we are obliged to uphold it as a methodological principle. In the words of William of Conches: “God can make a cow out of a tree, but has He ever done so? Therefore show some reason why a thing is so, or cease to hold that it is so.” Let us concede that God can make a cow out of a tree; but unless we have a reason to think that he has done so, we must explain the historical origins of any particular cow as involving a mommy cow and a daddy cow; there is nothing to justify the idea that God made the cow out of a tree even once we have admitted his power to do so if he really wanted.

4.1.5 Uniformitarianism

The view which we have called “actualism” is sometimes (perhaps more commonly) known as **uniformitarianism**.

However, this word is often misleading. As a term in the history of science, it often refers to ideas some of which no living geologist considers to be true. And as a term in religious apologetics, it often refers to ideas which no geologist in the entire history of geology has *ever* considered to be true.

For that reason, in this text I have thought it best to retire the old word and go with the more modern term “actualism” instead.

The term “uniformitarianism” is misleading in itself: for when modern geologists call themselves uniformitarians, what are they claiming to be *uniform*? No less than the laws of nature themselves — but not necessarily anything else. Every geologist will insist that many things have *not* been uniform over the course of the Earth’s history: its flora and fauna, for example, have not stayed the same; its temperature has not stayed the same; the composition of its atmosphere has not stayed the same; the arrangement of continents has not stayed the same; the global climate has not stayed the same.

What has apparently stayed the same is that throughout all this change the laws of nature have been uniformly unbroken, and only actual processes have taken place. In modern parlance, a “uniformitarian” geologist asserts no more than that; he or she has no general belief in *uniformity*, merely in actualism.

4.2 Steno’s principles

Steno’s principles of stratigraphy are among the oldest of geological principles; they explain how we can look at **sedimentary rocks** today and figure out facts about the sediment at the time of deposition.

They are named for Nicholas Steno (1638 - 1686), a Catholic bishop and polymath scientist who set forth his

principles in his *Dissertationis prodromus* of 1669. He was beatified in 1988, making him officially just one step away from being a saint; the scientific world has shown its admiration for him by naming craters on Mars and the Moon in his honor.

In this article we shall explain, discuss, and illustrate his principles.

4.2.1 The principle of superposition



Beds of sandstone and siltstone: the lower strata were deposited first.

The **principle of superposition** simply says that when **sediments** are deposited, those which are deposited first will be at the bottom, and so the lower sediments will be the older. This is because sediment is deposited from above, because gravity operates in a downward direction, and because sediment does not readily pass through other sediment.

Note that this only applies to sediment; it does not necessarily apply to **igneous rock**, because whereas sediment is deposited from above, **magma** oozes up from below. So, for example, an igneous **sill** intruding into sedimentary rock will be *younger* than the rocks immediately above it.

One snag may occur to you. It is all very well to say that the sediment, when originally deposited, was laid down from the bottom up. But does it necessarily follow from this that when we look at sedimentary rocks the lowest is the oldest? For tectonic processes such as folding can completely overturn a section of rock, or, just as bad, turn it on its side so that we can't tell which was originally up and which was down.

Fortunately, there are many clues within the rocks which allow us to discover which way up they were originally; these are known as **way-up structures** and will be the subject of **the next article**. Once we have used these indications to discover which way up the rocks were when the sediment was deposited, we can then apply Steno's principle to sort out their relative ages.

4.2.2 The principle of original horizontality

The **principle of original horizontality** states that sediment is originally laid down flat.

This needs some qualification. After all, some sediment is not originally deposited in flat **beds**, as you will know from the articles on sedimentology: for example aeolian sand is laid down in **cross-beds**, and the **angle of repose** of a sand dune can be as much as 34° from the horizontal.



Cross-bedding, Dry Fork Dome, Utah

Yet you will also recall that the cross-beds are laid down in sets formed as the dune rolls across the desert landscape, and these are much more nearly horizontal. Consider the photograph to the right of aeolian sandstone in Utah.

The cross-beds themselves reflect the angle of repose of the original sand-dunes; but the *sets* of cross-beds are horizontal.

To take another example, you may remember that the **progradation** of a delta forms **foreset beds**, which show the slope of the **delta** into the **lake** or sea into which it builds. And yet these foreset beds considered as a whole will form a roughly horizontal layer sandwiched between the **topset** and **bottomset** beds.

So even apparent exceptions can, looked at the right way, be seen as instances of Steno's rule; and having taken them into account, we find that we can look at rock formations and identify the plane of horizontality at the time that the sediment was originally deposited.

4.2.3 The principle of original continuity

The **principle of original continuity** states that sediment is originally deposited in a continuous horizontal sheet until it meets some obstacle, such as the base of a cliff, or it tapers off because of its distance from its source, or **grades** laterally into another sort of sediment.

The qualifications in that sentence may seem at first to make the principle futile. What is it saying but "sediment will go on until it stops"?

An example will make the concept clearer. Consider the photograph to the right of **hoodoos** at Drumheller, Canada.



Hoodoos at Drumheller, Canada.

You should easily be able to make out three kinds of sediment: there is hard dark mudstone at the bottom, followed by lighter-colored softer sandstone, and then the “caps” visible on the taller hoodoos are formed from sandstone which is harder and darker than that which underlies it.

Now, it is inconceivable that geological processes originally deposited these sediments in the forms of hoodoos: for one thing, there are no processes that would do that; and for another thing, the sediment, when unlithified, wouldn't have stayed standing if it *had* been deposited in such a way.

What must have happened is that each of the three layers was originally deposited as a continuous sheet, which subsequently underwent lithification and then erosion, producing the forms visible today.

4.2.4 Steno's principles and actualism

As we have just explained the principle of actualism [in the previous article](#), it is worth taking a while to examine how Steno's principles are just special cases of this more general principle.

For example, when discussing the processes that formed the hoodoos above, we appealed twice to actualism; we wrote: “It is inconceivable that geological processes originally deposited these sediments in the forms of hoodoos: for one thing, *there are no processes that would do that*, and for another thing, *the sediment, when unlithified, wouldn't have stayed standing*”. We appealed to what we know of geological processes to say that the hoodoos couldn't have been deposited as hoodoos; we appealed to known laws of physics to point out that even if there was such a process, the unlithified sediment would have slumped and collapsed.

Again, consider the principle of superposition as it applies to the hoodoos. It seems plain that mud must have been deposited first, then the paler sand, and then the darker sand. In saying so, we are tacitly taking it as axiomatic that science-fictional processes such as levitation

or teleportation did not come into play. If they did, then we have no way of knowing which rock layer came first: maybe the dark sand was deposited first but hung unsupported meters above the sea floor until the gap was filled in by the lower sediments which somehow managed to pass through the existing layer.

We can imagine such a thing happening, just as we can imagine a dragon or a unicorn, but the only basis for practicing geology is to ignore this hypothetical magical event and to work on the basis that this did not in fact happen.

4.3 Way-up structures

As was noted in the previous article, it is perfectly possible for rocks to be overturned by tectonic processes. Hence, before we can apply the principle of superposition to discover the relative ages of the strata, we must first identify which way up the rocks were when they were formed.

Fortunately there are many indications we can use to find this out, known as way-up structures. In this article I shall list some of them

Note that I do not claim that this list is complete; these are simply some of the most commonly cited way-up structures.

4.3.1 Mud cracks

Mud cracks (also known as **desiccation cracks**) are formed in mud when it dries, and examples can be found preserved in the geological record. These form a distinctive structure, with their polygonal forms and the roughly V-shaped cross-section of the cracks; not only is there nothing else like them, but also there is nothing that looks exactly like mud cracks apart from going up where mud cracks go down. Hence mud cracks can be used as way-up structures.

4.3.2 Ripple marks and cross-bedding

Ripples have curved troughs and sharp crests, and a convex shape as seen from above; as with mud cracks, we may note that there is nothing that looks exactly like a ripple only upside-down. Hence they form way-up structures.

The **cross-bedding** in **aeolian sand** is also convex on the upper side, as sand dunes are steeper at the top and have a shallower curve near the base (see, for example, the picture in the previous article).



Flame structures.



Potholes, Blyde River Canyon, South Africa

4.3.3 Flame structures

Flame structures are formed when a denser sediment, (typically sand) is deposited on top of a less dense sediment (typically mud). The difference in density forces the mud to flow upward in what are known as **diapirs**, producing a distinctive flame-like structure in which the “flames” are always at the top. The photograph to the right shows an unlithified example.

(We may note in passing that diapirs are technically a counterexample to the principle of superposition: some of the earlier sediment has managed to rise above some of the later sediment. However, as diapirs are fairly easy to identify, they cause little confusion in practice.)

4.3.4 Graded beds

Some forms of deposition produce graded beds: for example, turbidity currents produce beds which grade upwards from coarse to fine material. Although occasionally reverse grading can be seen in turbidites, one typically finds turbidites in large stacks, so there is no difficulty in discerning the general trend and identifying the occasional example of reverse grading as being the odd man out.

Again, in meandering rivers we find fining-up sequences at point bars; rivers do not produce fining-down sequences.

4.3.5 Erosion

Currents will often incise structures into the sediment over which they flow: flute marks, scour marks, sole marks, etc, producing distinctive impressions in the underlying beds which are then filled in with other sediments (we have discussed this particularly in our article on turbidites, but such structures are produced in other environments such as rivers).

In general, it is easy to look at a surface where two sed-

iments meet and determine which of the two sedimentary rocks was eroded, and which was laid down over the eroded surface. If, for example, potholes formed by a river or glacier are subsequently filled up with mud, it is not difficult to conclude that the mud conformed itself to the potholes rather than the potholes to the mud, and that consequently the potholes indicate the lower surface.

4.3.6 Fossils attached to the surface

Some fossils form attached to the ground, and display a distinct difference between up and down: so, for example, branching corals will branch upwards, not downwards; a tree-stump found still with its roots intact and embedded in seat-earth shows that the roots were down and the stump was up; stromatolites will have a flat base and a convex top.

4.3.7 Fossils not attached to the surface

Even when fossils are movable and can be overturned, we can sometimes learn from them. For example, trilobites, being broad, flat creatures, are not readily overturned, and are usually found belly-side down as they were in life. Although one or two might get flipped over, if we have a fair number of trilobites we can distinguish the statistical trend, and say which direction was most likely up.

Bowl-shaped shells such as individual valves of **bivalves** will, in a current (such as that produced on beaches by the tide) tend to come to rest convex-side up, as the reader may easily observe by taking a stroll on a beach. Again, this will not be true of every single shell, but so long as we have a fair number of shells, we can gauge the general trend and use it to figure out which way was up.

4.3.8 Trace fossils

The casts thrown up by invertebrates as they burrow are naturally found on the surface.



Lithified dinosaur tracks.

Some burrowing invertebrates make burrows which serve as way-up structures. For example, some make distinctive U-shaped burrows: naturally the openings are at the surface, so the prongs of the U point up and its bowl points down.

The indentations made by footprints are necessarily convex, and any sediment which fills them in will be concave, forming a way-up structure.

The photograph to the right shows the footprints of some large theropod dinosaur. Note also the mud cracks. Clearly this particular piece of rock is the right way up.

4.3.9 Geopetal structures

Geopetal structures are formed when a hollow object (such as a shell) becomes partly filled with sediment (such as mud). This provides us with a naturally occurring spirit level allowing us to tell up from down, and indeed the plane of horizontality, at the time when the sediment was deposited.

4.3.10 Bubbles in igneous rock

When igneous rock is formed, bubbles of trapped volcanic gas will, of course, rise through the still-molten lava because of their lower density; for this reason, if we find bubbles in a solidified lava flow, they will tend to be at the top rather than the bottom.

4.3.11 Structures in lava flows



Pahoehoe.

While the base of a lava flow will conform itself to the ground over which it flows, the top of the flow will usually not be flat, but may take on a number of distinctive forms: pillows if it was formed underwater, and **aa** or **pahoehoe** if it forms on land (the unusual terms are Hawaiian in origin).

The photograph to the right shows the distinctive “ropy” structure characteristic of the surface of pahoehoe: clearly this rock is the right way up.

4.4 Fossils



Close-up of petrified wood, Canyonlands National Park, Utah.

In this article we shall discuss the processes by which fossils are formed, and the circumstances under which this

occurs.

4.4.1 Formation of fossils

In the process of **replacement**, the original material of bone, shell, or other tissue is replaced by minerals from the surrounding sediment.

In **permineralization** minerals (typically **silica**) fill in the spaces and voids within a fossil, including the interiors of cells.

Permineralization and replacement often go hand in hand (in which case the organism is said to have undergone **petrification**) but it is possible to have one without the other. The photograph to the right shows a piece of petrified wood.

A fossil **mold** is produced when sediment is packed around a organic remains which are then destroyed by decay, leaving a void in the shape of the organism.

A **cast** is produced from a mold when minerals fine enough to percolate through the gaps in the sediment then fill in the void left where the organism used to be.

There are also unaltered fossils; although in popular usage to say that something is a fossil is often to imply that it has been mineralized, in scientific usage “fossil” can refer to any remains which are **prehistoric**: that is, which are older than any written human history. So, for example, people speak of “fossil mammoths” even though their bones are unaltered by mineralization.



A gnat in amber.

Small organisms such as insects and spiders can become trapped in the sticky resin of trees, which hardens to become **amber**; the magnified image to the right of a tiny gnat trapped in amber shows the exquisite level of preservation that this process achieves.

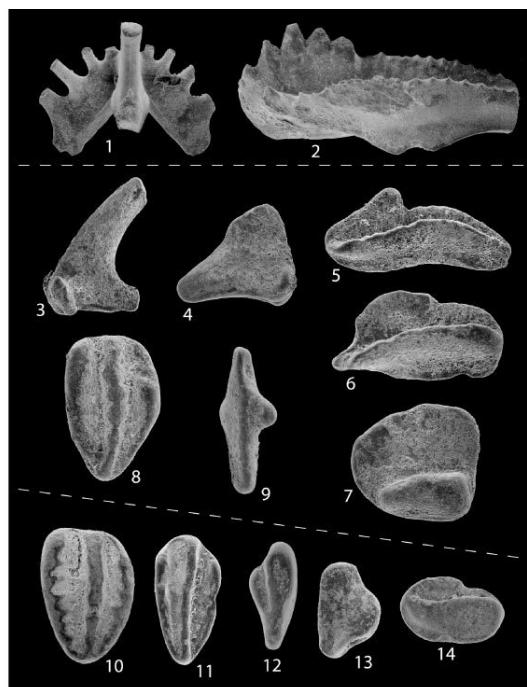
Finally, we may note the existence of **trace fossils** such as footprints or worm casts: these are preserved like other sedimentary structures by the lithification of the sediment in which they are formed.

4.4.2 Conditions for fossilization

Following the death of an organism, several forces contribute to the dissolution of its remains. Decay and predators and/or scavengers will typically rapidly remove the flesh; the hard parts, if they are separable at all (i.e. if the organism doesn't just have one big hard part such as the shell of an **ammonite**) can be dispersed by predators, scavengers or currents; the individual hard parts are subject to chemical weathering and erosion, as well as to splintering by predators and/or scavengers, which will crunch up bones for marrow and shells to extract the flesh inside. Also, an animal swallowed whole by a predator, such as a mouse swallowed by a snake, will have not just its flesh but some and perhaps all its bones destroyed by the gastric juices of the predator.

It would not be an exaggeration to say that the typical vertebrate fossil consists of a single bone, or tooth, or fish scale. The preservation of an intact skeleton with the bones in the relative positions they had in life requires a remarkable and fortuitous circumstance: burial in **volcanic ash**; burial in **aeolian sand** due to the sudden slumping of a sand dune; burial in a mudslide; burial by a **turbidity current**; and so forth.

The mineralization of soft parts is even less common, and is seen only in exceptionally rare chemical and/or biological conditions. How rare? Well, consider the story of the conodont animals.



Conodont structures.

From 1856 onwards paleontologists noticed that the fossils record from the Cambrian period right through to the end of the Triassic was littered with microfossils known as **conodont structures**. They guessed, correctly as it hap-

pens, that these were the scattered teeth of some unknown type of otherwise soft-bodied organism. The photomicrograph to the right shows some conodont structures.

In 1934 Schmidt and Scott discovered conodont structures grouped together on the same bedding plane, arranged in symmetric pairs; that is, in this case the conodont structures had presumably not been scattered but lay in the relative positions they would have had in life. But since the soft tissues were not preserved the nature of the **condont animals** remained a mystery.

Not until 1983 did paleontologists discover a specimen in which the soft tissues of a conodont animal had been preserved — one hundred and twenty-seven years after the very abundant conodont structures had first been described.

4.4.3 Mineralized fossils: how do we know?

We take it for granted today that mold, cast and mineralized fossils are the relics of organic life. It may surprise the reader to learn that this was once a minority view, verging literally on heresy. Instead, it was widely believed that they were not: the most common view being that the fossils grew in the rocks as the result of a mysterious force known as *vis plastica*.

This explanation fitted nicely with the religious views of the time. Many fossils, if interpreted as the relics of once-living organisms, would have to represent species that had gone extinct, since no-one could find their modern equivalents. Now theologians argued that God, being perfect, would not have made any species so badly that it would go extinct; dissenting scientists such as Robert Hooke were obliged to guard themselves carefully against accusations of impiety.

It is then at least possible to suppose that mineralized fossils are not in fact mineralized remains of organisms. How would we argue against someone who was inclined to doubt it?

We might point out the existence of partially mineralized fossils: in fact, this was pointed out to supporters of *vis plastica*, who replied that the sequence was in fact the other way round, from rock to mineralized fossil to unmineralized fossil.

We might point out, as was pointed out at the time, the similarity between some fossils and living organisms. It would be remarkable if a process occurring in the rocks should produce just the same sorts of forms as are also produced by organic processes, giving rise to things that look exactly like sharks' teeth or sea-urchins.

We might also look at the consistency within rock formations. For example, one sort of sandstone will contain terrestrial plants and animals, and the large cross-beds found in **aeolian sand**; another will contain seashells and

the same sort of symmetrical ripples found on a beach. On the basis of the mineralization theory, this is explicable and indeed expected: but how strange it would be if the *vis plastica* somehow managed to bring forth just those tableaux that would *look like* (but not be) the relics of former ecosystems.

From our perspective in the twenty-first century we might appeal to actualism. We know the composition of rocks down to the very arrangement of **atoms** in their constituent minerals, and there seems to be no room for a mechanism for mineralization: the process of mineralization only requires the sediment and organism to become more chemically homogeneous, which is much more chemically plausible than the reverse.

This seems a fair reply to the proponents of *vis plastica*. Another view proposed at about the same time was that the fossils were created by God when he created the Earth. This is certainly conceivable (an omnipotent God can do what he wants) but it might be answered in a similar way. From an actualistic view, we would point out that this invokes a miracle where none is apparently necessary, violating the scientific method. And from consideration of the nature of the fossil record we would have to say that in that case God has gone to extraordinary lengths to deceive us by making it look exactly like we are looking at the lithified relics of times gone past; we may imagine this, but it is grossly inconsistent with the traditional view of God, which supposes honesty to be among his virtues.

Since the hypotheses of initial creation or of *vis plastica* as the origin of fossils are, so far as I know, currently held by no-one whatsoever, the foregoing discussion may seem somewhat in the nature of a needless digression. However, it does emphasize the point, which I feel is worth making, that any statement in a geology textbook, no matter how much we take it for granted today, had to be discovered by someone; evidence had to be produced, and arguments had to be made — often against determined and dogmatic opposition.

So if we *now* take it for granted that mold, cast, and mineralized fossils are what we think they are, then this is not an unfounded assumption: we can afford to take it for granted because the case has been so well-made that perhaps no textbook except this one takes the trouble to review the question.

4.5 Principle of faunal succession

In this article we shall introduce the principle of faunal succession, and discuss how we know it is valid and why it should be so. In subsequent articles we shall discuss further how it can be applied to stratigraphy. The reader will find it useful to have read previous articles on Steno's principles, way-up structures, and **fossils**.



Heteromorph ammonites.

4.5.1 The principle of faunal succession

By using the principle of superposition and by using way-up structures to determine the up and down directions at the time of deposition, we can find the order of deposition in an assemblage of strata. This means of course that we can also find the order of deposition of the fossils within the strata.

Now, looking at some particular assemblage of strata, we may find that the deposition of species A stopped before the deposition of species B started. (We can write this for short with the notation $A < B$.)

Then the **principle of faunal succession** says that if we look at a different location and find species A and B, then we will also find that $A < B$. This is not to say that we will always find A in the same assemblage of strata as B, or vice versa, but it says that *if we do*, then we will find that A was deposited before B; that the order is the same in different locations. While, as discussed later in this article, exceptions to this are certainly possible (which is why we do not call the principle a **law**) the principle is very generally applicable, and so it does deserve to be called a principle.

Note that the relation $<$ is **transitive**: that is to say if $A < B$ and $B < C$, then we will also find that $A < C$. (Again, this is not to say that we always will find A and C in the same location, it just tells us what order they will be in if we do.) This allows us to establish a linear order on the fossil record, as we shall discuss further in our article on the geological column.

4.5.2 How do we know, and why is it so?

How do we know? Because we looked. The principle is a simple one which tells us directly what we should see if we look, so it is easy to verify that it is generally valid.

The question of *why* it is so is more interesting. It is also in a sense irrelevant, since it is possible to verify the principle and apply it without having the faintest idea why it should be true; and in fact this is what geologists did when the principle was first discovered.

However, we do now understand the reason why the principle holds good; it is an elementary consequence of the **theory of evolution**. It would take a textbook equal in length to this one to explain the theory of evolution and to sketch out the nature of the evidence for the fact of evolution. However, for the purposes of understanding the principle of faunal succession, the reader really needs to understand only one corollary of the theory: that any particular species will only evolve *once*.

So it is not possible for species A to evolve before species B in North America and for species B to evolve before species A in South America; each species must arise at one time in one place.

Nor is it possible for species A to arise and go extinct, for species B to arise and go extinct, and then for species A to arise a second time; again, species A can only evolve once, and extinction is final.

(Those readers, if any, who deny evolution will just have to look on the validity of the principle of faunal succession as one of life's little mysteries, or to be more accurate as one of life's huge, gigantic, preposterously enormous mysteries; but the practical validity of the principle is beyond question.)

Now, the underlying mechanism of the principle does allow for the principle to be violated in particular instances. It would for example be possible to have $A < B$ in one location and $B < A$ in another, through some such scenario such as the following.

- Species A arises at location X.
- Species A spreads to location Y.
- Species A goes extinct at location X.
- Species B arises at location X.
- Species B spreads to location Z.
- Species B goes extinct at location Z.
- Species A spreads to location Z.

It would then be the case that at location X we have $A < B$, but at location Z we have $B < A$. This sort of elaborate dance is quite unlikely to happen, and if it ever does, it can't happen very often, since the principle of faunal succession does in fact generally hold good.

In summary, the principle of faunal succession *should* work in theory and *does* work in practice.

4.5.3 Note on vocabulary

The principle of faunal succession does not just apply to **fauna** (i.e. animals), so strictly speaking the principle should be called: “The principle of faunal, floral, fungal, and everything-else-al succession”, but that would be inconvenient; and in any case, the name of the principle is so well established that it is too late to do anything about it now.

4.6 Index fossils



In this article we shall discuss what an index fossil is, why they are useful, and what qualifies a fossil to be an index fossil.

4.6.1 Why do we need index fossils?

Consider the **fauna** living on the continental shelves of the east and west Atlantic ocean (the **littoral** fauna). Obviously these faunas will be different — they have had 130 million years to follow separate evolutionary trajectories, and since they require shallow water they aren't going to cross from one side of the ocean to the other.

So when we look at the successions in the fossil record of the littoral faunas on each side of the Atlantic, we will see *two* faunal successions, an eastern and a western. What's more, the sediments in which they are deposited will not particularly correlate, since they will have different points of origin.

However, mixed in with the littoral fauna there will also be fossils of free-floating or free-swimming surface organisms: **pelagic** fauna. Now such species will spread throughout the ocean, since there's nothing to stop them from doing so.

So we can and do find the same pelagic species on both sides of the Atlantic. We can use such species to correlate the other littoral species and the sediments in which they are deposited. For if we find a pelagic species (call it S) somewhere in the fossil record on both coasts, then this gives us a way to divide up both successions into species that were deposited before S, along with S, or after S.

Given enough such species, we can find many such correlations, and this would allow us to represent the eastern and western deposition on the same timeline.

Such species would be examples of **index species**, and their fossils would be examples of **index fossils**, which we may define generally as follows: an index fossil is a fossil of a species that was sufficiently widely distributed that its fossils can be used to correlate the deposition of fossils and sediments in widely separated locations.

4.6.2 What makes a good index species?

Obviously, the first requirement for an index species is that it should indeed be widely distributed. Pollen makes a good index fossil, being wind-borne; so do foraminifera, which are pelagic species as in our example.

An index species should also be readily preserved in the fossil record. Birds, for example, would make bad index fossils, because although there are many species which have wide (indeed intercontinental) ranges, they fossilize very poorly: their skeletons come apart easily, and then their delicate honeycombed bones are highly susceptible to decay. Typically, only fragments will survive, and even if these were common, which they are not, it would still be hard to tell one species from another by studying them.

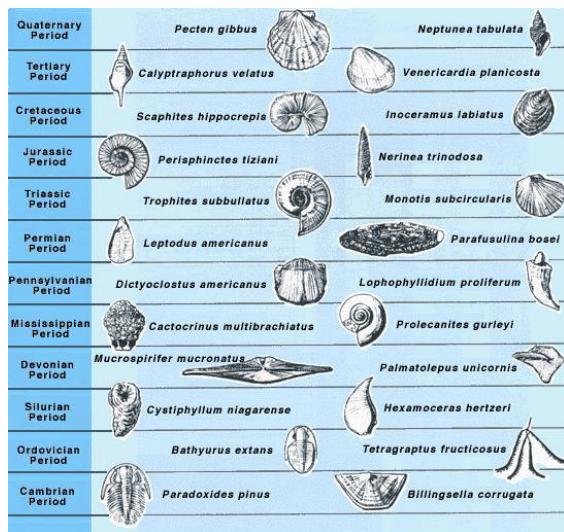
Finally, we would ideally like an index fossil to have a short time of deposition as a proportion of the fossil record, since we want to use it to identify a particular chapter in the history of deposition: it should represent a geographically broad but temporally narrow slice of the record.

Some examples of marine index fossils are shown in the table below. Note that they are not claimed to form an evolutionary sequence, any more than a list of Presidents does: they are a *temporal* sequence, and should be understood as such.

4.7 Geological column

In this article, I shall explain what the geological column is, how it is constructed, and what relationship it bears to the geological record. I shall also provide a rough description of the geological column summarizing some of the major trends observed in it.

The reader should already have read the the article on Steno's principles, the article on the principle of faunal



Index fossils.



Spinosaurus, a dinosaur of the Cretaceous.

succession, and the article on index fossils before reading further.

4.7.1 Construction of the geological column

By using the principles explained in the previous articles ([superposition](#), [faunal succession](#), the use of [index fossils](#)) it is possible to produce an account of the order of deposition of the organisms found in the fossil record, noting that one was deposited before the other, that the deposition of such-and-such a group starts after the deposition of some other group ceases, and so forth.

This means that fossils and the [sedimentary rocks](#) that contain them can be placed in order of their deposition.

The resulting table is known as the **geological column**.

4.7.2 Prolog to a sketch of the geological column

Below, I sketch out the major geological systems from the Vendian onwards. Note that it is written from the bottom upwards, so that the earliest-deposited fossils are at the bottom; the reader may therefore find it tells a more coherent story if read from the bottom upwards.

It is no more than a sketch: it records the appearance and disappearance of major groups, rather than individual species; and it has been divided into the large stratigraphic units known as **systems**, which geologists would divide into **series**, which they would further divide into **stages**, which they would then further subdivide into **zones**. I am, then, only giving the broadest outline of the geological column; those who require the finer details must look elsewhere.

I have not attached any dates to the geological systems discussed here, because as we have not yet reached our discussion of [absolute dating](#), it would be premature to do so. All that our study of [fossils](#) and their [faunal succession](#) tells us is the *order* of deposition. (It is for this reason that I have used the increasingly obsolete term “geological column” rather than “geological timeline”; it is not a timeline until we get round to attaching dates to it.)

I have also avoided using terms such as “evolution” and “extinction”. From a biological standpoint, it is obvious that these are the underlying cause of the patterns in the fossil record; but as with the evolutionary explanation of the [principle of faunal succession](#) this biological explanation is irrelevant to the practice of geology. For the purposes of doing stratigraphy it doesn’t really matter if dinosaurs appear in the geological column because they [evolved](#) from more basal archosaurs or because they parachuted out of the sky from an alien spaceship, and it doesn’t matter if they disappear from the geological column because they went extinct or because they all went to live in cities on the Moon; what matters is that we can find out where their [fossils](#) come in the sequence of deposition.

4.7.3 A sketch of the geological column

Quaternary

Marked by the existence and spread of modern humans and the decline and disappearance of many groups of large fauna extant in the Neogene.

Neogene

Contains recognizable horses, canids, beaver, deer, and other modern mammal groups. The Neogene also contains many large mammalian fauna no longer extant: glyptodonts, ground sloths, saber-toothed tigers, chali-



Skull of a Smilodon (commonly known as a “saber toothed tiger”).

cotheres, etc. First hominds found in Africa.

Paleogene

Marked by the diversification of mammals and birds. Among the mammals we see the first that can be easily identified with modern mammalian orders: primates, bats, whales, et cetera. Similarly representatives of many modern bird types are identifiable in the Paleogene, including pigeons, hawks, owls, ducks, etc. Now-extinct groups of birds found in the Paleogene include the giant carnivorous birds known colloquially as “terror birds”.

Cretaceous

Here we see the diversification of angiosperms (flowering plants) from beginnings around the Jurassic-Cretaceous boundary; representatives of modern groups of trees such as plane trees, fig trees, and magnolias can be identified in the Cretaceous. Here also we see the first bees, ants, termites, grasshoppers, lepidopterans. Dinosaurs reach their maximum diversity; some of the best known dinosaurs such as *Triceratops* and *Tyrannosaurus* are found in the Cretaceous. Mosasaurs appear near the end of the Cretaceous, only to disappear at the Cretaceous-Paleogene boundary, which also sees the last of the dinosaurs (excluding birds, which biologists classify as dinosaurs) and the last pterosaurs, plesiosaurs, ichthyosaurs, ammonites, rudists, and a host of other groups.

Jurassic

This system is notable for the diversification of dinosaurs. It has the first short-necked plesiosaurs (pliosaurs); first birds; first rudists and belemnites. Mammals are certainly present, but tend to be small and insignificant by compar-



Craspedites, an ammonite from the Jurassic period.

ison with reptile groups. The first placental mammals are known from the Upper Jurassic.

Triassic

The Triassic contains the first crocodiles, pterosaurs, dinosaurs, lizards, frogs, snakes, plesiosaurs, ichthyosaurs, and primitive turtles. Whether or not there were mammals in the Upper Triassic depends on what exactly one classifies as a mammal. The Triassic-Jurassic boundary sees the loss of many groups, including the last of the conodonts, most of the large amphibians, and all the marine reptiles except plesiosaurs and ichthyosaurs.

Permian

This system is noted for the diversification of reptiles: the first therapsids (mammal-like reptiles) and the first archosaurs (the group including crocodiles and dinosaurs). It also has the first metamorphic insects, including the first beetles. It has the first trees identifiable with modern groups: conifers, ginkgos and cycads. Many species and larger groups come to an end at or shortly before the Permian-Triassic boundary, including blastoids, trilobites, eurpterids, hederellids, and acanthodian fish.

Carboniferous

This system contains the first winged insects. Amphibious vertebrates diversify and specialize. The Carboniferous has the first reptiles, including, in the Upper Carboniferous, the first sauropsid, diapsid, and synapsid reptiles. Foraminifera become common. All modern classes of fungi are present by the Upper Carboniferous.

Devonian

The Devonian has the first (wingless) insects; the first ammonites; the first ray-finned and lobe-finned fish; the first amphibious vertebrates; the first forests. Terrestrial fungi become common. The first seed-bearing plants appear in the Upper Devonian. The last placoderms are



Crinoids of the Carboniferous.

found at the Devonian-Carboniferous boundary. Almost all groups of trilobite have disappeared by the Devonian-Carboniferous boundary, but one group (Proetida) survives until the Permian-Triassic boundary.

Silurian

In the Silurian, coral reefs are widespread; fish with jaws are common; it has the first freshwater fish; first placoderms (armour-plated fish); the first hederellids; the first known leeches. Diversification of land plants is seen.

Ordovician



The Ordovician trilobite Flexicalymene meeki, curled up in a ball presumably for defensive purposes.

In the Ordovician system we see the first primitive vascular plants on land; jawless fishes; some fragmentary evidence of early jawed fishes. Graptolites are common, and the first planctonic graptolites appear. Bivalves become common. The first corals appear. Nautiloids diversify and become the top marine predators. Trilobites diversify in form and habitat. The first eurypterids ("sea

scorpions" appear in the Upper Ordovician. Trilobite forms such as Trinucleoidea and Agnostoidea disappear at the Ordovician-Silurian boundary, as do many groups of graptolites.

Cambrian

This system sees the first animals with hard parts (shells, armor, teeth, etc). Trace fossils reveal the origin of the first burrowing animals. Trilobites are common; chordates exist but are primitive. Archaeocyathids are common reef-forming organisms in the Lower Cambrian and then almost completely vanish by the Middle Cambrian. Condonts are first found in the Upper Cambrian. Many groups of nautiloids and trilobites disappear at the top of the Cambrian, but some groups survive to diversify again in the Ordovician.

Vendian

This system contains the first complex life, including sponges, cnidarians, and bilaterians.

4.7.4 The geological column and the geological record

We should distinguish between the geological *record* and the geological *column*. The geological record is a *thing*: it is the actual rocks. The geological column is not a thing, it is a *table* of the sort given above. To ask questions such as "where can I go to see the geological column?" or "how thick is it?" is therefore a category error along the lines of asking how many people can be seated around the Periodic Table.

The relationship between the geological column and the geological record is this: when we look at a series of strata in the geological record and use the principle of superposition and way-up structures to discover the order of deposition of the fossils in it, then if we find that A < B in the strata, this will correspond to B being shown above A in the geological column. The geological column is therefore a particularly simple and neat way of recording what relationships we do and don't find in the geological record.

However, the geological column is not a *picture* of what we find in the geological record. There are three reasons for this.

First, as we know, the geological record is folded and faulted in some places. Recall that when we write A < B we are talking about the original order of deposition of fossils, as reconstructed by using the principle of superposition and way-up structures: it does not necessarily mean that A is actually below B; whereas the geological column is always depicted as a vertical column with A below B when A < B.

Second, by using index fossils geologists produce a single time-line for the entire planet; but clearly any particular location will only have local fossils: the column, if writ-

ten out in full, would show exclusively South American Cretaceous dinosaurs above exclusively North American Jurassic dinosaurs, but these will not in fact be found in the same assemblage of strata.

The third reason is that deposition will typically not happen continuously in one place: **sediment** is deposited in low-lying areas; it would not be deposited on top of a mountain. What's more, an elevated area will typically undergo **erosion**: not only will fresh sediment not be deposited, but existing **sedimentary rocks** and their **fossils** will be destroyed. Also, marine sediment will be destroyed by **subduction**, so the sediment of the oceanic crust will be no older than the ocean that it's in, and even then only at the edges — it will be considerably younger near the **mid-ocean rifts**.

Consequently, the meaning of the geological column is not that any location in the geological record will look like the geological column: the column is merely an elegant way of representing the facts about faunal succession.

4.7.5 The geological column: how do we know?

As was explained at the start of this article, the geological column is constructed using ideas introduced in previous articles: the **principle of superposition**, the **principle of faunal succession**, and the use of **index fossils**.

Note that the geological column does not except in the weakest sense constitute a **scientific theory**. It does resemble one, because there is a sense in which it suggests what we are likely to observe, which is the role of a theory; but it is essentially descriptive in nature. That is, it does not really predict the sequence of fossils that we *will* find, it is determined by, and summarizes, the sequences of fossils that we *have* found. Since it is likely that what we will find tomorrow will be similar to what we have been finding for the past couple of hundred years, the column is in that sense predictive, but its predictive power goes no further than that.

So if tomorrow we found that some trilobites were deposited above the Permian system, we should simply amend the geological column to reflect this, and it would be surprising not because it contradicted the geological column as such, but because in centuries of paleontology no-one has yet made such a discovery.

Compare this with how we would feel if we consistently found violations of the **principle of faunal succession**. This would present a difficulty in theory, and would require us to give up on the **principle of faunal succession** (and to give up on using it to construct a geological column, something that would then become impossible). But finding something that contradicts the geological column as it stands is merely unlikely in practice, not in theory, and would only require us to revise the geological column in one particular detail (i.e. to take the new dis-

covery into account) without requiring us to rethink any fundamental ideas.

So the geological column is trustworthy simply because it is no less, but *no more*, than an up-to-date summary of our knowledge, and so it can be taken as such. To which we might add that after all these years of looking at the fossil record it is extremely unlikely that we'll find anything so unusual as to require any major revision of the column.

4.8 Unconformities

In this article we shall look at the various types of unconformity in the geological record, and discuss why they exist and how they can be recognized.

4.8.1 What are unconformities, and why do they exist?

As we mentioned in the article on the geological column, we do not expect the fossil sequence to be complete in any given location, because there will be times when no sediment is deposited in that location, and/or times when it is destroyed by erosion after deposition.

Such an episode will show up in the geological record as an **unconformity**: a surface between successive strata representing a period of erosion or of no deposition. These come in several varieties, listed below.

4.8.2 What do unconformities look like?

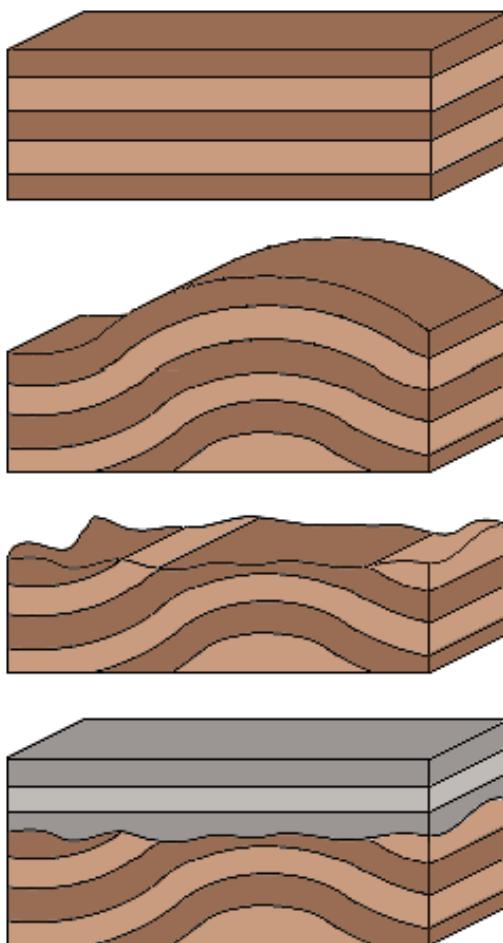
In an **angular unconformity** the underlying beds meet the overlying beds at an angle. The diagram to the right shows the stages of the process.

First, **sediment** is laid down. Then tectonic events destroy its original horizontality, in this case by creating an upward fold. Then an **erosional** surface is produced, truncating the **beds**. Then more sediment is laid down in horizontal beds on top of the erosional surface. In the example in the diagram this has produced two sets of angular unconformities, one at each side of the diagram.

In a **disconformity**, the same thing happens, except that the underlying beds are not distorted by tectonic events, so that the first and second collection of beds lie parallel to one another rather than meeting at an angle, but they are still separated by an erosional surface.

A **nonconformity** is like an **disconformity**, except that the underlying rock is **igneous** or **metamorphic** rather than **sedimentary**. The presence of an erosional surface indicates a time when this basement rock was exposed to **weathering** and **erosion**, and so cannot have been protected by a blanket of sediment.

Finally, consider what would happen in a location where sediment is deposited, then it ceases, and then erosion



Origin of an angular unconformity.

does *not* take place, and then deposition starts again. This would produce what is known as a **paraconformity**. What would that look like? There would be no meeting of beds at an angle, there would be no erosional surface, there would just be beds of sediment lying on top of more beds of sediment.

Except that if the period of non-deposition lasted for any significant amount of time we would see a sudden jump in the faunal succession: where the geological column shows fossils in order (for example) P, Q, R, S, T, U, (counting from the bottom upwards) then at the location of the paraconformity we would see P, Q, T, U, where P and Q correspond to the first episode of deposition, T and U correspond to the second episode of deposition, and the missing fossils R and S correspond to the time at which no deposition was taking place. Note that if this jump forward in the faunal succession was just because there was no fauna around to be deposited, then we would see non-fossil-bearing sediment between Q and T; but instead we see nothing at all.

Obviously we will see a similar jump in the faunal succession in the case of angular unconformities and disconfor-

mities, but in the case of a paraconformity this is all that there is to see. This makes the identification of paraconformities less blatantly obvious than the identification of other unconformities, but since they look just like what we should see in locations where deposition stopped for a while and then restarted, it is reasonable to conclude that that's what they are.

Note that this only serves to identify sufficiently long periods of non-deposition. In cases in which the interruption of deposition only lasted for a small period of time then all we would see would at most be a **bedding plane**, and perhaps not even that.

4.9 Faults

In this article we shall look at what faults are, what causes them, and how we can recognize them. It will be helpful if the reader is familiar with the articles earlier in this textbook on the physical properties of rocks, on terranes, and on the principle of faunal succession.

4.9.1 Causes and appearance of faults

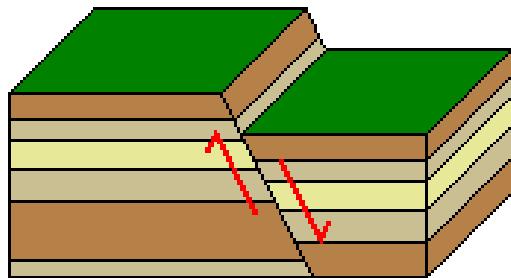
A geological **fault** is a planar fracture in a volume of rock caused by motion of one side with respect to the other. Motion along faults, and indeed the faults themselves, are caused by tectonic events; by the stretching or compression of the **crust**. This causes the rock to fracture when it is near enough to the surface to be **brittle** rather than **ductile** (as explained in the article on the physical properties of rocks).

Faults are classified according to the nature of the motion producing them. **Dip-slip faults** are those which involve vertical as well as horizontal motion; these can be classified as **normal faults**, where the landscape is being pulled apart, and **reverse faults**, where one part of it is being pushed over another. These are more easily illustrated than described, and are depicted in the block diagram to the right. The reader should note that there is nothing particularly normal or common about "normal" faults: the name is just a name. In dip-slip faults the rock lying above the fault is known as the **hanging wall** and that below the fault as the **foot wall**.

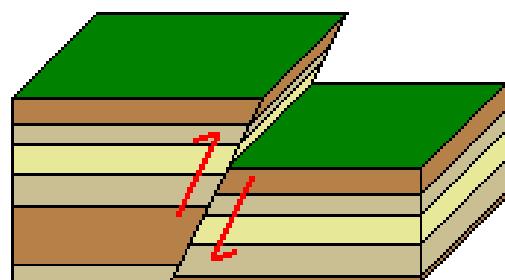
A **thrust fault** can be defined as a reverse fault in which the angle of the fault is more than 45° from the vertical. An interesting point to notice is that in a reverse fault, and especially in a thrust fault, older rocks end up directly above younger rocks, presenting geologists with an interesting stratigraphic puzzle; we shall return to this point later.

In a **strike-slip fault**, the blocks on either side of the fault move laterally but not vertically with respect to one another in a direction parallel to the fault: the San Andreas Fault is the most notorious example of this variety

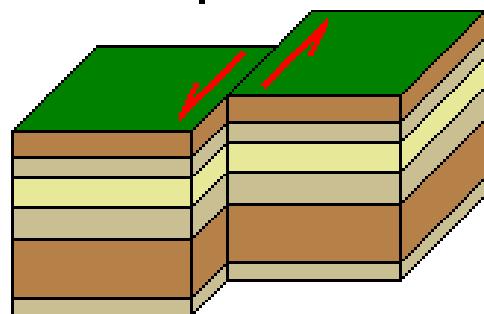
A normal fault



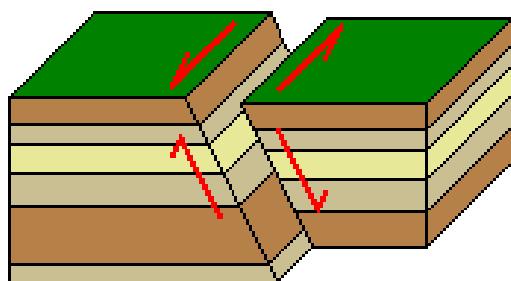
A reverse fault



A strike-slip fault



An oblique fault



Types of faults.

of fault. The transform faults discussed in the article on sea-floor spreading are a special case of strike-slip faults.

Strike-slip faults can be categorized as right (or **dextral**) and left (or **sinistral**) according to their direction of motion: in the case of a left fault, for example, anyone stand-

ing on one side and looking at the other when it was moving would see the other side moving to the left. The strike-slip fault depicted in the block diagram is a left fault.

An **oblique fault** combines elements of a dip-slip fault and a strike-slip fault.

In these block diagrams I have portrayed faults in which an originally continuous piece of landscape has been disturbed by a fault. However, as discussed in the article on terranes, faults will also arise when two previously disjoint land masses are forced together by plate motions. For more information about such faults, see the article on terranes.

We shall now turn to the question of how we would go about identifying inactive faults: that is, faults where the two sides of the fault are no longer moving relative to each other.

4.9.2 Fault rocks

When a fault is still active then it is easy to spot it. The San Andreas Fault, for example, is hardly inconspicuous, and the motion along it is measurable.

In inactive faults, where motion has ceased, we will still, perhaps, be able to discern a crack in the rocks, perhaps accompanied by a sudden discontinuity of rock types: but perhaps the crack is just a crack, and the discontinuity is just an **unconformity**. What we would like is evidence that there was once motion along the suspected fault plane. There are a number of clues that point us in this direction.



Fault gouge.

In the block diagrams above, the fault itself is just shown as a straight line with no width. In reality, this is not the case. The two sides of the fault do not fit neatly together and slide smoothly past one another; instead they grind off fragments, often large ones, and crush and mill them, producing **fault rocks** such as **fault breccia**; the milling process will continue until the breccia no longer impedes motion along the fault, producing finer material filling the gaps between the coarser **clasts**. **Fault gouge** is a similar rock but with finer clasts. In the photograph to the right, courtesy of the U.S. Forest Service, fault gouge can

clearly be seen interrupting the horizontal strata on either side of it.

Fault breccia and fault gouge can be found in active faults. For example, when engineers constructed the aqueduct between Owens Lake and Los Angeles, they were forced to tunnel through the San Andreas Fault, and found themselves tunneling through a thick sheet of fault breccia and gouge. So when we find something that looks like this in the geological record, but not associated with any present motion, then it is reasonable to conclude that this too was produced by motion along a fault; not just because of its similarity to rocks associated with modern faults, but because the nature of the fault rocks themselves unmistakably indicate processes of fracturing, crushing, and grinding. When we also note that these rocks form a narrow sheet sandwiched between other rocks, and that the **clasts** of the **breccia** match in composition the rocks that they're sandwiched between, the conclusion becomes irresistible.



Mylonite.

Another rock characteristic of faults is **mylonite**. This is formed at depths where deformation of rocks is more ductile. The effect of this on the fabric of the rock is to crush its component **minerals** and draw them out in streaks, producing a wood-like grain as shown in the photograph to the right, with the grain being parallel to the direction of motion of the fault.

Although mylonite is formed at depth, we can still see it in active fault zones because it is **exhumed**: that is, it is brought out of the Earth by the rising side of the fault if its motion has a dip-slip component. For example, the Alpine Fault in New Zealand is a dextral-reverse fault with the hanging wall rising at a rate of 6-9 mm/yr: the grain of the exhumed mylonite on the hanging wall is consistent with exhumation by the motion of the fault.

So when we find mylonite in the geological record in the shape of a fault and sandwiched between stratigraphic evidence for faulting (as will be discussed below) we can take this as an indication that we're looking at a fault, even if it is no longer active.

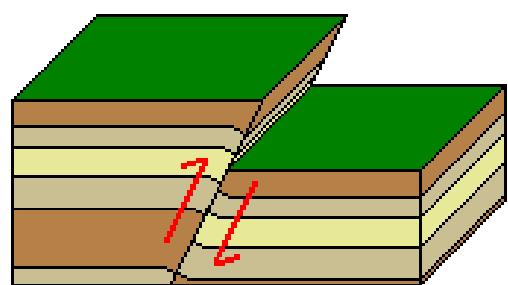
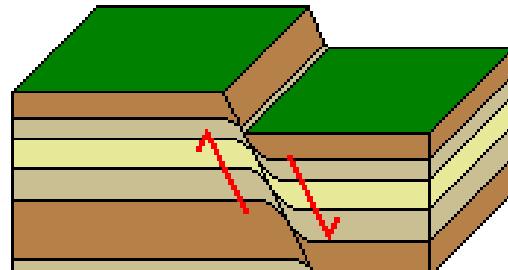
Slickensides are smoothed and striated surfaces produced by the friction between the two sides of the fault, or

rather between the two sides of the fault and the **breccia** between them. These are not dissimilar to the smoothed and striated surfaces left by **glaciers**. However, slickensides often take on a much higher polish than rocks smoothed by glaciers, which is why they are sometimes called **fault mirrors**.

4.9.3 Faults and stratigraphy

In the case of a dip-slip or oblique fault, if the rocks are stratified and if we are able to look at the fault side-on, we are able to see discontinuities in the originally continuous layers at the fault plane, as illustrated in the block diagrams above. Sedimentary **strata**, lava flows, and volcanic **sills** will not join up across the fault. If we are lucky, we will be able to see how the strata originally joined up, and so figure out the extent and direction of motion; but if the motion has been sufficiently great and **erosion** has been sufficiently severe, or if the rocks do not have distinctive **strata**, or if we are not privileged to have a side view of the fault, this will not be possible.

In the case of strike-slip faults there is no vertical movement, and so not such a pronounced disturbance of lateral continuity. However, strike-slip faults disrupt linear features such as **dikes** and **riverbeds**.



Faults with drag folds.

When two large pieces of landscape are pulled past one another, there will inevitably be frictional resistance, and this can pull the material of the rocks backwards (i.e. in the opposite direction to their directions of motion) distorting the structure of the rock and producing **drag folds**. The phenomenon is illustrated in the block diagrams to the right.

In a reverse fault, particularly a thrust fault, there will

be places where older fossils (according to their usual arrangement in the fossil record) appear above younger fossils, in apparent, though not real, violation of the principle of faunal succession. If this was the only sign indicating a reverse fault, then perhaps we might suspect that we weren't looking at a reverse fault, but at an actual violation of the principle. But of course we can look for the other signs of a thrust fault, as listed above: slickensides, drag folds, fault breccia, etc; and we can also check that when we subject the rocks to absolute dating, the sequence of ages suggested by the fossils is confirmed by the dates, and that we really are looking at older fossils pushed by a thrust fault over younger fossils.

4.10 Folds



Folds in rock.

In this article we shall look at folds and how to recognize them in the geological record. The reader will find it useful to have read the articles on [orogeny](#), [Steno's principles](#), and [way-up structures](#) before continuing.

4.10.1 Origin and appearance of folds

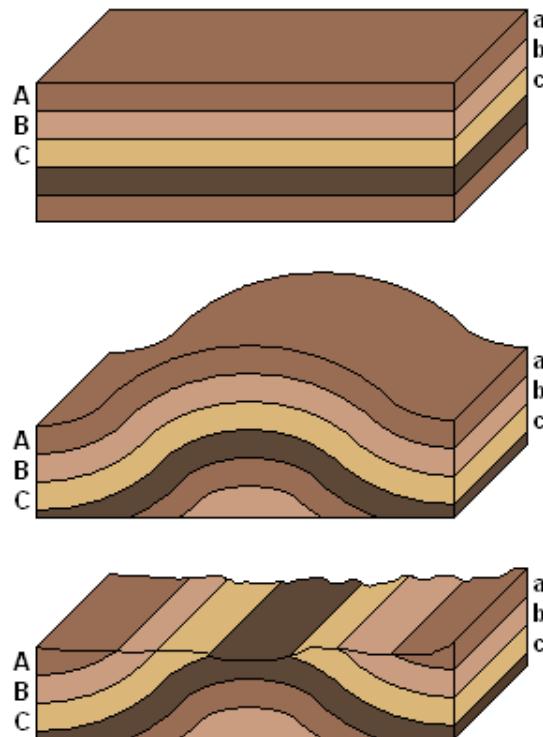
The appearance of folds shows that they are produced by the landscape being pushed sideways, as described in the article on [orogeny](#), rather than being pushed up or pulled down from below.

This explains the normal appearance of folds (prior to erosion) as a set of parallel ridges where the landscape has been folded up (**anticlines**) alternating with troughs where it has been folded down (**synclines**). The photograph to the right shows an anticline on the left and a syncline on the right, seen from a side-on perspective.

Sufficient lateral motion will push the folds themselves over sideways, resulting in a **recumbent fold**, as shown in the article on [orogeny](#).

4.10.2 Folds: how do we identify them?

When, as in the photograph above, a fold is intact and we can see it from the side, it is obviously a fold.



Origin and subsequent erosion of a fold.

But now consider a sequence of events such as that shown in the block diagrams to the right: sediment is laid down in flat layers according to the principle of original horizontality and the principle of original continuity, then lithified, folded, and finally eroded.

As a result of the erosion, the shape of the fold will be destroyed. Even then, if we could see the rocks from the side it would not be hard to deduce that this was an eroded fold. But geologists are not usually so lucky, and instead they have to view the landscape from the top.

Nonetheless, it would not be hard to deduce what happened. Looking at the rocks, a geologist could see that the formation labeled A is the same kind of rock as a, B is the same as b, and so on. Applying the principles of original horizontality and original continuity, the only possible explanation would be that the rocks were folded and then eroded. The symmetric pattern of bands of different rock types across the landscape are by themselves an excellent indication of an eroded fold.

Even if the types of sediment didn't vary during deposition, there would be plenty of other evidence. Looking at the individual beds in the rocks, we would find them sloping up to the right on one side of the anticline, and up to the left on the other. Again, the principles of original

horizontality and **original continuity** would tell us that at one point the layers would have been flat and met up in the middle.

Similarly we could look at the **way-up structures** in the rocks: on one side of the anticline they would indicate that the original up direction had been tilted to the left, and on the other side to the right.

One interesting sign of folding is that **clasts** and **fossils** within the rocks can be stretched and deformed as the rock gradually folds. Amusingly, before this process was well-understood, early paleontologists would identify the same species as two different species according to how the specimens were stretched: a fossil fish stretched one way would appear long and thin, whereas another originally identical fish stretched the other way would look short and wide.

Indications such as these allow us to identify folds even if they have been **eroded**, and even if we can't see them from the side.

4.11 Walther's principle

In this article we shall look at what Walther's principle is and why it works.

4.11.1 The principle

Walther's principle can be stated as follows: If sediment A is succeeded vertically by sediment B without an **unconformity** between them, then sediment A will also be succeeded *horizontally* by sediment B in some direction.

The somewhat abstract statement of the principle will be clarified by a few examples.

4.11.2 Examples

Walther himself was led to formulate his principle by looking at **fluvial** rocks. As a **meandering river** shifts its course, the riverbed (consisting, for example, of **gravel**) is overlaid by the sand of a **point bar** and then the mud of a **flood plain**. But this is exactly the same sequence as one would see moving laterally outward from the center of the river out to the bank: gravel, then sand, then mud. It is this lateral succession of sediments, plus the fact that the river shifts, which *causes* the vertical succession in a **fining-up sequence**, so naturally they are going to be the same.

Similar effects are caused by shifts in sea level. Consider the various depositional environment and their associated types of sediments (**facies**) that we would see on the sea floor as we move further from the shore and out to sea: a typical progression would be sand grading into mud which grades into **calcareous ooze**.

Now consider what happens if the sea level rises and so the shoreline moves inland (a marine **transgression**). The sandy facies would move landwards; so would the muddy facies, and similarly with the calcareous ooze. Now this means that in some locations we will see a vertical succession of sand giving way to mud; and further seaward a succession of mud giving way to limestone.

In a marine **regression**, when the sea level falls, the facies would of course move in the opposite direction, reversing the vertical succession.

The reader should note that this is why finding a physically continuous rock formation (of sandstone, for example) does not indicate that it was all laid down at the same time. The seaward end of it could have been laid down at the beginning of a marine transgression, and the landward end of it at the end of the transgression. So long as the transgression was sufficiently gradual, this would produce a continuous layer of sandstone (perhaps sloping slightly upwards from the seaward to the landward direction) and yet one end of it would have been laid down at a different time from the other. Such a transgression can be detected in the geological record because although the sandstone will remain sandstone throughout its length, the **fossils** at the landward end will be younger than those at the seaward end.

As a final example, consider the deposition of marine sediment on a moving oceanic **plate**. A particular section of the plate moves (let us say north, for example) from a location where **pelagic clay** is deposited to a location where **calcareous ooze** is deposited. Then it follows that when mud was being deposited on that part of the plate, **calcareous ooze** was being deposited to the north of it, and when **calcareous ooze** is being deposited on it, **pelagic clay** will be deposited to the south of it. The horizontal succession from south to north is **pelagic clay** followed by **calcareous ooze**, which is just the same as the vertical succession from bottom to top.

4.11.3 Walther's principle: how do we know?

The principle is a simple one: all it does is tell us what we should see if we look; and we do in fact see what the principle tells us we ought to see.

What's more this is what we ought to see in theory, because as you can see in our examples it is the horizontal succession of sediments that *causes* the vertical succession of sediments, so naturally the two successions are going to be the same.

The principle can have exceptions, but they will necessarily be rare. If there is sudden shift in the mode of deposition (such as a **tsunami** rapidly laying down a bed of sediment known as a **tsunamite**) then because of the underlying mechanism the succession of sediment will not obey Walther's principle. However, such an event will

tend to be so violent in its nature as to be erosional and to produce an **unconformity**; and so it would not constitute a counterexample to Walther's principle, which by definition only applies to beds of sediment without an unconformity between them.

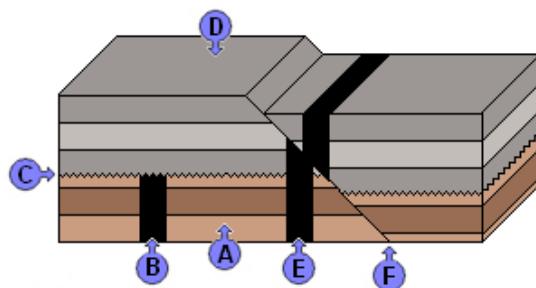
It is usually the case that a change in the mode of deposition at a given place will correspond to a gradual geographical shift in the places in which different types of sediment are deposited (or, as in our final example above, a shift of the geography itself beneath the different areas of deposition). For this reason Walther's principle will be generally if perhaps not universally applicable.

4.12 Cross-cutting relationships

Besides the principles already explained, we can determine the relative ages of various geological features by studying the way they intersect with one another: their **cross-cutting relationships**. In this article I shall explain how this can be done.

4.12.1 The principle of cross-cutting relationships

The **principle of cross-cutting relationships** may be stated as follows: when one geological feature cuts through another, the former is the younger and the latter is the older of the two features.



Cross-cutting relationships.

For example, consider the diagram to the right.

The brown **sedimentary rocks** (A) must be older than the **dike** (B) that cuts through the **strata**; the **dike** must be older than the **erosional surface** which truncates it (C) which is older (by the principle of superposition) than the gray **sedimentary rocks** (D) which overlie it; this rock must be older than the **dike** (E) cutting through the gray strata; and then this **dike** must be older than the **fault** which cuts through it (F).

One exception might occur to you: what if we have an outcrop of rock which is then covered over by sediment? The outcrop will cut through the resulting sedimentary strata, but be the older of the two. Whether or not this is

really an exception to the principle depends on how you look at it; but if we consider it to be an exception, it is a *recognizable* exception, for the outcrop would be weathered and eroded where it formerly projected above the surface, and from this erosional surface we would therefore be able to see that it was the older component of an **unconformity**.

4.12.2 How do we know?

The reasoning behind these conclusions is **actualism** of the most straightforward sort. Is it really necessary to argue that an erosional surface is younger than the eroded rock, or that a **fault** is younger than the faulted rock? This is a matter as much of logic as of physics, for it is simply to assert that the surface of a thing cannot precede the thing.

Why must a **dike** or similar intrusion be younger than the rock into which it intrudes? Because a sheet of molten rock would not stand up by itself: where the **magma** forming a dike pierced through to the surface, then it would not go on up into the air building a **dike**, but rather it would ooze along the ground creating a **lava flow**.

Hence we can indeed use these cross-cutting relations to establish the relative ages of geological features.

4.13 Igneous rocks and stratigraphy

In this article we shall briefly recap the facts about **igneous rocks** as they relate to **stratigraphy**. The reader may find it helpful to go back and re-read the main article on igneous rocks.

4.13.1 Igneous rocks and stratigraphy

From the facts about **stratigraphy** and **igneous rocks** that we have covered in previous articles, we may make the following statements:

When they are first formed, extrusive **igneous rocks** (**lava flows** and **volcanic ash**) will be younger than the **sedimentary rocks** below them, and will be older than the **sedimentary rocks** which subsequently form above them.

Why so? Well, by definition, extrusive igneous rocks are **igneous rocks** which are deposited on the surface, and so the principle of superposition applies to them.

Note that as usual the original position of the rocks can be altered by tectonic events, and if this has happened the original order must be recovered by studying **way-up** structures and observation of the **faults** and **folds** that have been formed. Using our $<$ notation, we may write that if rock A $<$ rock B $<$ rock C, where rocks A and C are



Intrusive dikes, Madeira.

sedimentary and rock B is extrusive, then rock A was deposited before rock B, and rock C was deposited after it.

Now consider **intrusive igneous rocks**. If we have an intrusive sheet-like structure (a **sill**) then this is necessarily younger than the **sedimentary rocks** below it. But it is also younger than the **sedimentary rocks** that were above it when it formed. (Note that something must have been above it when it formed, otherwise it would not be **intrusive**.)

But it is only older than the **sedimentary rocks** above it *when it formed*. Suppose that igneous rock B intrudes between **sedimentary rocks** A and C. Then further sediment is deposited on top of C, forming **sedimentary rock** D. Then B is younger than C but older than D.

When considering **intrusive igneous structures** such as **dikes** and **plutons**, we may appeal to the principle of **cross-cutting relationships** to say that they are younger than the **sedimentary rocks** through which they penetrate.

The reader should bear in mind that we can tell the difference between **intrusive** and **extrusive igneous rocks** just by looking at them: **intrusive igneous rocks** are **coarse-grained**, and **extrusive igneous rocks** are **fine-grained**. So even though a **sill** and a **lava flow** are the same shape and may have just the same chemical composition, we can still tell the difference between them by their **texture**.

4.13.2 Why is this important?

So far, we have only been able to consider the relative ages of features in the geological record; we can say that this **stratum** was deposited before that, and that the deposition of this **fossil** preceded the deposition of the other. But this only provides us with relative ages: we can say that A is older than B, but how much older? Older by a minute? A day? A million years? We have relative ages, but so far we have no actual *dates*.

Now the crucial thing about **igneous rocks** is that we can assign actual dates to them by analysis of the rocks, something we can't usually do directly with **sedimentary rocks** or **fossils**.

A geologist, if handed (for example) a fossil trilobite, cannot perform a physical or chemical analysis of the fossil and tell you how old it is. But if that same geologist finds the trilobite in the field lying in **sedimentary rocks** between two **lava flows**, then s/he can find the actual dates of the lava flows, and can then tell you that the trilobite is younger than the oldest lava flow and older than the youngest lava flow, and by assigning dates to the lava flows can tell you that the date of the **fossil** lies between them.

This brings us on to the topic of **absolute dating**, how it is done, and why it works.

Chapter 5

Absolute dating

5.1 Concepts in absolute dating



A hourglass, one of the earliest methods of measuring the passage of time.

In this article I shall define absolute dating, and shall discuss the conditions that we would require to use a geological process as the basis for absolute dating.

5.1.1 Absolute dating defined

In the articles on stratigraphy we looked at what is called **relative dating**, where we could say that one geological feature was older or younger than another but without actually putting dates on them. By contrast, **absolute dating** allows us to assign dates to geological features.

To avoid confusion later on, let us say at once that the “absolute” in “absolute dating” is not short for “absolutely correct”. In this context, the word “absolute” is not the opposite of “approximate” but just of the word “relative”. A method of dating which was accurate to within a billion years either way would still technically speaking be a method of absolute dating, it would just be a very bad one.

5.1.2 Conditions for absolute dating

Suppose that we wanted to find out how long it has been since an hourglass was set running by measuring the amount of sand in the lower bulb. To do so successfully, we would need to assure ourselves of the following conditions:

- That we can in fact accurately measure the amount of sand in the lower bulb.
- That we know the rate (or rates, if it may have varied) at which the sand has trickled from the upper bulb to the lower.
- That there is no hole in the lower bulb from which sand may have leaked out.
- That there was no sand in the lower bulb when the process started.
- That the process has been uninterrupted, and that since the hourglass was set running it hasn't spent time lying on its side, or standing the other way up.

Given these conditions, we can find out how long the hourglass has been running. If we wish to use a geological

process as the geological equivalent of an hourglass, we would want to have similar conditions: we would like to find some quantity which we can measure reliably (corresponding to the condition that we can measure the amount of sand in the lower bulb of the hourglass); which increases or decreases from a known quantity (corresponding to the lower bulb of the hourglass being empty when it starts running) at a known rate (corresponding to knowing the rate of flow of sand); and so forth.

Note that the conditions we have given for the hourglass are ideal conditions which we would require to know *exactly* how long it is since the hourglass started running. If conditions are less than ideal, we may still be in a position to come up with an approximate figure which is better than nothing. If, for example, we cannot measure precisely the amount of sand in the lower bulb, but we have good reason to think that our estimate of it must be within 10% of the true value, then we also have a good reason to think that we can give the time it's been running to within 10%. Or again, if there may have been sand in the lower bulb when it started running, but we have a good reason to think that there can't have been very much, then we also have a good reason to think that the figure which we get for the time can't be very wrong; and so on.

The geological processes that we'll be discussing work rather like that. We have no reason to believe that nature will be so obliging as to always provide us with the ideal conditions that would provide us with ideal clocks; but by analysis of the geological conditions as they actually are, we can often set limits to how imperfect the geological clocks can be.

For educational purposes, we shall start with a method which barely works at all under any circumstances. This will be the subject of the next article.

5.2 Erosion, deposition, and time



Arches formed by erosion, Arches National Park, Utah.

One obvious way you might think of to estimate times is to look at a feature of erosion or deposition, measure how much erosion or deposition has taken place, measure

the rate of erosion or deposition, divide the former measurement by the latter, and derive a length of time as our answer.

In this article we shall look at some of the problems involved in using such procedures as dating methods.

5.2.1 Rates of erosion and time

As an example, we might look at a wave-cut platform, see how far the waves cut away the cliff per year or per decade, see how extensive the platform is, and so find out how long the waves have been cutting the platform.

Or again, we might look at the depth of a canyon, measure the rate at which it is being cut, and then see how long it would have taken to cut it at that rate.

Of course this sort of reasoning could only work if as in these examples the erosional process only cuts away *part* of the landscape. We need to measure the depth of the canyon by comparison to the remaining rock; or the erosion caused by the waves by measuring the length of the remaining platform. If we were just looking at a horizontal erosional surface, then even if we knew the rate at which it was being eroded, we could not just by looking at the erosional surface figure out how much material was there originally and has been removed.

5.2.2 Rates of deposition and time

We might also look at the rates of deposition of sedimentary or igneous rock.

We could for example look at the sediment deposited at a mouth of a river, figure out how much sediment it transports each year, and see how long the river mouth has been in that location.

Or again, we might look at a volcano, see from historical records how often it erupts, see how large the typical lava flow is, and figure out how long it took to build the volcanic cone.

5.2.3 Some problems

There are, however, problems with these approaches.

One problem is this: geological events vary in intensity, and the larger they are, the rarer they are. So, for example, geologists will talk of a “ten-year storm”, one of an intensity that only occurs one year in ten; a “hundred-year storm”, of a magnitude that only happens one year in a hundred; and so on. Similar things may be said of volcanic eruptions, of rivers flooding, of earthquakes, and of pretty much anything else.

Now this presents us with a difficulty. If we look at the rates of erosion and deposition as they are happening now, we may be discounting the largest events. In

principle it might be possible (for example) that much or even most of the erosion forming a wave-cut platform is performed by thousand-year storms of a magnitude that we have never actually observed on that stretch of coast. Similarly, much or most of a volcanic cone might have been formed by eruptions of a magnitude that that particular volcano only ever undergoes every hundred thousand years.

Then again, when we look at erosion or sedimentary deposition, we must also consider long-term alterations in **climate**. A river, for example, eroding its banks and depositing sediment at its mouth may be a mere trickle in a cold dry climate when compared to its rate of flow in a hot moist climate; and as we shall see in the articles on **paleoclimatology**, climates have indeed undergone long-term variations such as these.

In some cases, deposition may have stopped altogether for a time, forming a **paraconformity**; and if the paraconformity is brief enough not to cause a significant discontinuity in the faunal succession, we might never know about it; and if there are enough such paraconformities, then we could be missing sizable chunks of time when we measure the thickness of sediment and try to estimate its age.

(Note that the problems we have been discussing do not all operate in the same direction: some would cause us to overestimate durations, and some to underestimate them.)

And one more problem: suppose we wanted to use these techniques to find the age of a **fossil** in (for example) the Tonto Group, which lies near the base of the Grand Canyon. Naively, we might try looking at the layers of sediment lying above it, estimating how long it took the limestone, sandstone, shale, etc to be deposited, add it all up, and arrive at a figure.

The trouble is that the rocks contain a number of **unconformities** between the bottom and the top, and the top itself, the Colorado Plateau, is also an eroded surface. Each of these surfaces represents vanished sediment which took a certain amount of time to be deposited which we cannot even estimate, because we don't know how much sediment there was; and which took a certain amount of time to be eroded which we also can't estimate for exactly the same reason.

So even if we managed to overcome all the other problems we have mentioned and produce good ball-park estimates for the length of time it took to produce each rock formation, we would be unable to put a date on the lowest rocks and the fossils they contained. The best this would do for us, even if we overcame all the other problems and got our figures exactly right, is supply us with a minimum date which could be too low by any quantity at all.

For these reasons, nineteenth-century geologists barely attempted to put dates on rocks. The best they could do was say that the Earth was old. How old? Very old. And a **geological period** (the unit of time corresponding to a

system) was long. How long? Very long. That millions of years were involved rather than hundreds or thousands was very obvious to them; but so also was the fact that considerations of **erosion** and **deposition** would not permit them to perform absolute dating.

5.3 Dendrochronology



Tree rings, Hillborough forest, UK.

Dendrochronology is the technique by which we can identify the age of a piece of wood by studying the growth rings it contains.

In this article we shall examine how it works, how we know it works, and the limitations of the technique.

5.3.1 How dendrochronology works

It is a well-known fact that many tree genera will produce one new **growth ring** each year. This means that, as every schoolchild knows, you can find out how old such a tree is by chopping it down and counting the rings.

This in itself would not be particularly useful. However, it is also the case that the rings produced are of different thicknesses according to the weather in each particular year, with a good year corresponding to a thicker growth ring. This is of some interest to paleoclimatologists, but what is important from the point of view of **absolute dating** is that this produces a sequence of growth rings of different thicknesses which is almost as distinctive as a fingerprint. Imagine for the sake of simplicity that there are only two thicknesses of rings: large ones and small ones. Then over a mere twenty years, over a million different sequences of large and small rings could potentially form, and which one actually does will depend on the weather over those twenty years.

This still leaves one question. If we were to dig up a piece of wood with the distinctive dendrochronological “fingerprint” of (let us say) 10,000 years ago, how would we know that that was in fact the time that it was the fingerprint of? To recognize the fingerprints of a criminal at a crime-scene, we need to have his fingerprints

in our files. In the same way, to recognize a 10,000 year dendrochronological “fingerprint” we would have to know what a 10,000 year fingerprint looks like. To know that, we would have to find a piece of wood which we knew to be 10,000 years old to take the fingerprint of. It seems, then, that we can't do dendrochronology unless we already have a way to determine the age of a piece of wood. If this was the case, it would be one of the more useless scientific techniques.

However, there is a way out. Suppose we take a core sample from a tree which grew between 1500 AD and the present; that gives us fingerprints for the past 500 years or so. Now suppose we find dead wood which, unknown to us, represents growth from 1100 AD to 1600 AD. This will have a fingerprint, and the last 100 years of its fingerprint will match the first 100 years of the tree we sampled. Observing the identities between these fingerprints, we can now put a date on each of the rings of the dead wood, which allows us to extend our knowledge of what the fingerprints look like back to 1100 AD, four hundred years before the living tree took us. Now if we find another dead sample which runs from 800 AD to 1250 AD, its tree-rings have a 150-year overlap with the known sequence, we can use this to date it, and then we can extend the sequence still further. By continuing this process with older and older samples of wood, we can build up data stretching back tens of thousands of years. This technique is known as **crossdating**; similar principles can be employed in other absolute dating methods.

5.3.2 Limitations of the technique

From the point of view of a geologist, tens of thousands of years is not very much. It is useful to an archeologist, but to a geologist that's just the recent past. And it seems very unlikely that the technique will ever take us much further.

The problem is that wood is not readily preserved; for it to last a long time, it must have been preserved under fairly unusual conditions; perhaps in an anoxic peat swamp, or buried under volcanic tuff. What's more, not all kinds of wood are suitable for the task. Some trees are **complacent**: that is, they produce growth rings of about the same thickness whatever the weather is like; whereas other kinds of trees don't produce exactly one growth ring per year, which also makes them unsuitable for dendrochronology.

So while dendrochronology may be an excellent technique so far as it goes, its scope is limited by the ability of archeologists to locate the right pieces of old wood; and these are scarce and become progressively scarcer as we go back through the geological record.

5.3.3 Dendrochronology: how do we know?

We can check that trees, or at least the kinds of trees we use for dendrochronology, do in fact add one ring per year. We can also check that different trees do produce the same pattern of thick and thin rings. Such observations tell us that dendrochronology should work in principle.

And in practice, when we cut down a tree with a known date of planting and count its rings, we can verify that they do in fact give its age.

We can also look, for example, at the timbers in an old building of known date. If dendrochronology works, then we would predict that the dates it gives for the timbers should not be later than the date of construction.

Or we can look at the charred timbers from cities destroyed by a volcano with a known date of eruption. For example, we can look at Herculaneum, which was destroyed by the eruption of Pompeii in 79 AD, and we would predict that dendrochronological dates for the charred timbers would not post-date 79 AD. The success of such predictions confirms the accuracy of dendrochronology.

Finally, we can note that dendrochronology is in close agreement with other techniques described in this textbook; techniques which are based on completely different principles. Even if we can imagine some unusual conditions in the past that might have messed up dendrochronology in some undetectable way, we should also have to suppose that other unusual conditions messed up other dating methods *in such a way that they would still concur with dendrochronology*. This is an extravagant conjecture: it is more parsimonious to conclude that the reason that all the methods concur is that they all actually work.

We shall have more to say on this subject in later articles.

5.4 Varves



Proglacial lake, Glacier National Park, Montana.

In this article we shall examine what a varve is, how they can be used for **absolute dating**, and when they cannot.

The reader may find it useful to go back and re-read the article on **glaciers** before continuing with this article.

5.4.1 What is a varve?

In its original definition, a varve was a sedimentary feature in a **proglacial lake**, consisting of a couplet of coarse and fine sediment. Such varves are deposited in proglacial lakes annually because of the seasonal changes in the **ablation** of the glacier and the amount of **meltwater** feeding the lake.

Since then the definition of a varve has been extended so that it can be used to describe any layer which is deposited annually, the varves in proglacial lakes being only one example.

In this article we shall discuss varves in the wider sense, since they are equally good for **absolute dating** whatever the origin of the **sediment**.

5.4.2 Varves and absolute dating

Given such a situation, there is no difficulty in principle in finding the age of any varve; we just start from the one that was deposited this year and count backwards. (In practice there may be technical difficulties, but the principle is straightforward enough.)

Of course, this only works if there is still a source of sediment, so that we can identify this year's varve and know which year we're counting from. Once the source of sediment is cut off, the link with the present is severed, and unless we could find some other method to place an absolute date on one of the varves, the only thing we can tell from them is the difference in age between two varves, but not how old either of them is.

This limits the use of varves for absolute dating. We can count back thousands or tens of thousands of years, but as we shall see in later articles this is only a short span of time in proportion to the much longer history of the Earth.

5.4.3 Varves and cross-dating

We introduced the idea of **cross-dating** in the article on **dendrochronology**. The same principle can sometimes be applied to varves. In varves in **proglacial lakes**, for example, the thickness of the layers will vary yearly, as a hotter year will result in more melting, more **outwash**, and more deposition in the **coarse-grained** part of the varve.

However, it is difficult to use this principle as it is used in **dendrochronology**, to link together varve sequences in the present and in the past, and to associate varves in a lake

currently being fed by **glacial** meltwater with a different lake in a different location where the **glacier** has melted entirely and the source of sediment has been removed.

Rather, the technique is more usually used on core samples taken from different locations in the same lake, in order to reduce error: if a minor disturbance has taken place in one location within the lake, removing one or more varves, then it is possible to use cross-dating with a sample taken from a different point in the lake to determine that this has taken place and to correct the chronology.

5.4.4 Varves: how do we know?

We can look in a lake or other environment of deposition, and see that varves are deposited on an annual basis; then we can take a core sample and see that the layer deposited in the past look just like those that are being deposited on an annual basis in modern times. It is an obvious conclusion that layers that look just like annually deposited layers were in fact annually deposited.

But what if the deposition has long ceased, and a succession of varves has been buried and **lithified**? How can we recognize them then? This question has various answers, depending on what sort of varves we're looking at. In classic varves, i.e. those found in a proglacial lake, the appearance of the sedimentary couplets is quite distinctive, and we may be fortunate enough to find **striated dropstones**, unambiguous evidence of a **glacial** depositional environment. In other cases it might not be so easy.

However, as we have pointed out, such ancient varves would usually not allow us to carry out absolute dating in any case, because they would have lost their link with the present: by counting varves we could determine the time interval between any two of them, but not the age of either of them.

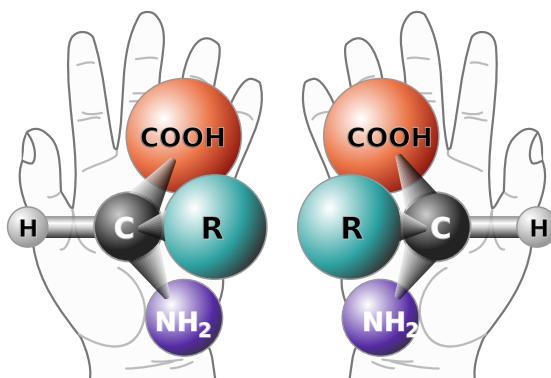
5.5 Amino acid dating

In this article we shall discuss the principles behind **amino acid dating** (also known as **racemization dating**); we shall discuss how it ought to work, and why it often doesn't.

5.5.1 Chirality

An object is said to have **chirality** if it is not possible to make it into a mirror-image of itself by turning it round. For example, a shoe is chiral: you cannot turn a left-foot shoe into a right-foot shoe by turning it round or flipping it over. On the other hand, an object such as a table-knife is not chiral: if you have it lying on the table so that the blunt edge is on the right and the serrated edge is on the left,

then you can produce the mirror-image of this situation by rotating the knife around its long axis.



Enantiomers.

Some molecules are chiral. For example, consider the two molecules in the picture to the right. They both have exactly the same chemical formula, but one is left-handed, and the other is right-handed. They are said to be **enantiomers** of one another.

When we make chiral molecules using ordinary chemical processes, we usually produce equal quantities of both enantiomers. Such a mixture is said to be **racemic**.

However, biological processes produce molecules with a distinct chirality: all the amino acids are “left-handed” (with the exception of glycine, which is not chiral) and all the sugars are “right-handed”.

5.5.2 Racemization

So when an organism dies, its amino acids are left-handed. But after its death, the amino acids can spontaneously change their chirality, flipping from being left-handed to right-handed, and indeed back again.

The result of this process is that eventually the amino acids will collectively become racemic: each particular amino acid will have one chirality or another, but after a sufficient amount of time, collectively the amino acids won't favor one enantiomer over another. This process is known as **racemization**.

We should note that although the underlying basis for this process is random, and that *in principle* the amino acids could by some statistical fluctuation become less racemic and more chiral, the laws of statistics ensure that in practice if we are looking at a large enough sample of amino acids, the chances are astronomically remote that such a thing will occur.

So the process of racemization looks like a good candidate for one of nature's clocks. We know that when an organism dies, its amino acids will all be left-handed; and we know that as time progresses the amino acids will become continually more and more racemic.

So it would seem that if we want to know how long it was since an organism died, all we have to do is see how racemic its amino acids are. And this would work, on one proviso. The process of racemization would have to go at a constant rate, and we'd have to know what it was.

And this is where the whole idea breaks down.

5.5.3 How do we know it works?

It doesn't.

The problem with racemization is that it depends on chemical processes that are affected by temperature, humidity, and the nature of the original material undergoing racemization. As a result, it isn't possible to say that racemization happens at such-and-such a rate.

However, it does have some applications. Suppose we examine a *particular* material (let us say **tests** of the foraminiferan *Neogloboquadrina pachyderma*) in a particular environment (let us say in mud in Arctic waters) and by comparing it with a dating method we know we can rely on, we establish that under these conditions racemization *does* happen at a reasonably steady rate.

In that case we could use the foraminiferans to date sediment in places where we aren't able to use **radiometric dating**. (For it would be strange and anti-scientific to conjecture that the rate of racemization of the shells in the Arctic mud is constant whenever we can check it, but variable when we can't.)

Just this was established by Kaufman et. al. (2008) in their paper *Dating late Quaternary planktonic foraminifer *Neogloboquadrina pachyderma* from the Arctic Ocean by using amino acid racemization*, *Paleoceanography*, 23(3). It gives the reader some idea of the difficulties of the method that they were obliged to use the single common **foram** species *N. pachyderma*, having found that racemization rates differed even between different species of forams.

So dating by racemization can have a few applications, but the conditions under which it can confidently be applied are rather rare. What's more, racemization happens quite fast by geological standards, so, like the other methods of **absolute dating** we have discussed so far, dating by racemization cannot take us far back in geological time.

All this is not to say that the reader should dismiss out of hand results obtained by amino acid dating; but it can be trusted only when the people applying it have taken care to ensure that they are using it in a context in which it is known to work. In early papers, before geologists and archaeologists had learned the pitfalls associated with amino acid dating, inaccurate dates were presented with much more confidence than they deserved, and such papers should not be relied on.

5.6 Radioactive decay

As a prelude to the articles on radiometric dating, it is desirable that the reader should know something about the mechanisms of radioactive decay. This article provides a fairly non-technical explanation of what it is and how it works.

5.6.1 Isotopes

The reader should recall from high school that the **nucleus** of an atom consists of **protons** (positively charged particles) and **neutrons** (uncharged particles which have almost exactly the same mass as protons). The nucleus is surrounded by a cloud of **electrons**, negatively charged particles having negligible weight. The number of electrons is equal to the number of protons.

The number of protons in an atom is its **atomic number**, and the sum of the protons and the neutrons gives its **atomic weight**.

The chemical properties of an atom are determined by the behavior of its electrons, and so are in effect determined by its atomic number. Hence in chemistry atoms are classified into **elements** according to their atomic number: an element such as carbon, for example, is defined by having an atomic number of 6. However, two atoms can have the same atomic number and different atomic weights. So, for example, ^{12}C (carbon-12) has six protons and six neutrons, whereas ^{14}C (carbon-14) has six protons and eight neutrons. They are both carbon, and they both behave chemically as though they are carbon, but they have a different atomic weight. So they are said to be the same element, namely carbon, but to be different isotopes of carbon. An **isotope** is defined by its atomic number *and* its atomic weight.

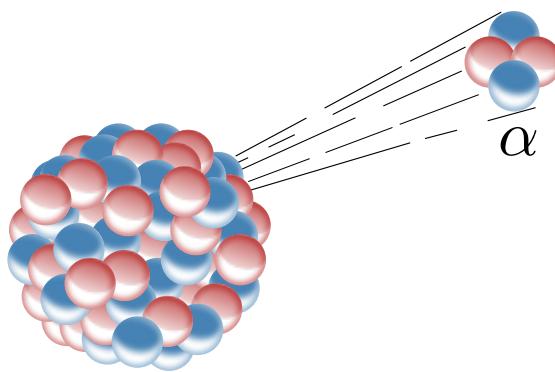
In the example in the previous paragraph, we have shown the notation used for isotopes. An atom with six protons and eight neutrons is written as ^{14}C . The fact that it has six protons is revealed by the “C”, which is the chemical symbol for carbon; by definition, all carbon atoms have six protons. The fact that it has eight neutrons is revealed by the little “14” written above and to the left of the “C”: this is the atomic weight of the isotope, and so since the atomic weight is the number of protons plus the number of neutrons, and since all carbon atoms have six protons, this tells us that this isotope of carbon must have eight neutrons.

5.6.2 Radioactive decay

Radioactive decay may be defined as any spontaneous event which changes the state of the nucleus, emitting energy from the nucleus in the process. With the exception of gamma decay, which need not concern us here, this will involve changing the number of protons, or neu-

trons, or both, and so also changing the atomic number, the atomic weight, or both.

There are a number of mechanisms by which decay may take place. For our purposes, the most important are:



Alpha decay: a nucleus emits two protons and two neutrons.

- **Alpha decay.** In this form of decay, the nucleus ejects an **alpha particle** consisting of two neutrons and two protons, reducing the atomic number by two and the atomic weight by four.
- **Beta minus decay.** In this form of decay, one of the neutrons in the atom is converted to a proton by the atom emitting an electron. Hence, the atomic number goes up by one, while the atomic weight stays the same.
- **Beta plus decay.** In this form of decay, a proton is converted into a neutron by the emission of a positron (a particle like an electron only positively charged) the result being that the atomic number goes down by one while the atomic weight stays the same.
- **Electron capture.** In this form of decay, one of the atom's own electrons combines with one of its protons, converting the proton into a neutron. This reduces the atomic number by one while leaving the atomic weight the same.

When decay takes place, the original atom is called the **parent atom**, and the new atom produced by decay is called the **daughter atom**. Those isotopes which are produced by radioactive decay are said to be **radiogenic**. Not all isotopes undergo decay: those that do are called **unstable** isotopes (or **radioactive** isotopes) and conversely those that don't are called **stable**. So for example ^{12}C is stable and will go on being ^{12}C forever; by contrast ^{14}C is unstable and has a tendency to decay into ^{14}N (nitrogen-14). As we can see from this example, it is perfectly possible for different isotopes of the same element to differ in their stability.

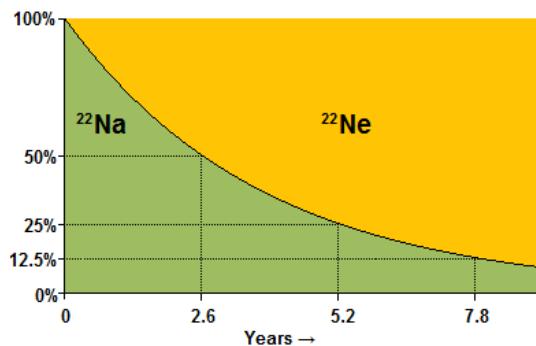
The reader should note that when a parent atom decays to a daughter atom, the daughter is not necessarily stable; sometimes the daughter will undergo further decay. Such a situation is described as a **decay chain**.

5.6.3 Statistics of radioactive decay

It is important to understand how and why radioactive decay takes place. According to physicists, radioactive decay occurs at random: an atom of (for example) ^{22}Na (sodium-22) will undergo beta decay and produce an atom of ^{22}Ne (neon-22) just because its number has come up.

The age of the atom has nothing to do with it. Consider, by analogy, a man playing Russian Roulette with a six-shooter. Every single time he plays, he has a one-in-six chance of dying, and this is true no matter how long he's been playing. The same is true of radioactive decay. If we have an atom of ^{22}Na , then no matter how old it is, it has a 50% chance of decaying in the next 2.6 years; and if it survives that period of time, then it has a 50% chance of decaying in the *next* 2.6 years; and so on.

Consider what this means if we have a large sample of ^{22}Na . Because the sample is large, its behavior will closely approximate our statistical expectations for it: so after 2.6 years 50% of the atoms will have decayed to ^{22}Ne ; after a further 2.6 years 50% of the *remaining* ^{22}Na will have decayed (leaving only 25% of the original ^{22}Na); after a further 2.6 years 50% of *those* ^{22}Na atoms will have decayed, leaving 12.5% of the original sample; and so forth.



Exponential decay.

This figure of 2.6 years is known as the **half-life** of ^{22}Na : that is, the time over which an atom of ^{22}Na has a 50% chance of decaying; or, equivalently the time over which our statistical expectation is that 50% of the sample will have decayed.

Such situations are mathematically well-understood, and can be represented by the equation for **exponential decay**:

$$N(t) = N_0 \times 2^{-t/2.6}$$

where N_0 is the original quantity of ^{22}Na , 2.6 is the half-life in years, t is the time elapsed as measured in years, and $N(t)$ is the quantity of ^{22}Na left in the sample at time t .

The situation can be represented by the graph to the right.

For ease of exposition, we have used ^{22}Na consistently as an example, but the same rules apply to all radioactive isotopes, the only difference is that the half-lives of different isotopes will be different: for example ^{107}Pd (palladium-107) has a half-life of 6.5 million years.

People are sometimes startled by such a statement: how, they ask, is it possible to say this when we haven't been watching a sample of ^{107}Pd for 6.5 million years to check that this is how long it takes for half of it to decay?

However, this is not really a problem. After all, by analogy, it is not necessary for a police officer to observe your car for an hour to report that you were traveling at 72 kilometers per hour. It is sufficient to observe you traveling at 20 meters in a single second, and then do the math.

In the same way, if we spend just a single year observing a sample of ^{107}Pd and see that over that time only 0.0000106638% of the sample decays, then we can write:

$$100 - 0.0000106638 = 100 \times 2^{-1/h}$$

and then solving the equation for h will give us the half-life in years.

It is obvious from this example that the exactness of our knowledge of the half-life will depend on the exactness with which we can measure the initial size of the sample and the rate at which it decays.

5.6.4 Invariance of the half-life

Each unstable isotope, then, has its own characteristic half-life. What is more, for each isotope, this half-life is constant: it is a property of the isotope, and virtually unaffected by external circumstances.

How do we know this? In the first place, it should be true in principle: it can be deduced from the underlying laws of quantum mechanics. In the second place it is confirmed by actual observation: shortly after the discovery of radioactive decay, scientists began trying to change the decay rate by subjecting unstable isotopes to heat, pressure, magnetism, and so forth, with negative results.

Some small variations have been observed in certain isotopes depending on which other chemicals they form molecular bonds with. An example is ^7Be (beryllium-7) in which the decay rate varies by about 1% according to the chemical environment it's in. There are theoretical reasons why ^7Be ought to be particularly susceptible to such effects: it is a very small atom, and it decays by electron capture. Even so, the variation is fairly small.

For the same reasons, even smaller variations have been achieved by putting ^7Be under intense pressure.

We should also mention the peculiar isotope ^{187}Re (rhenium-187). Because the energy emitted from the nucleus when it decays is so very small, it is possible to change its half-life by ionizing it. If it is completely stripped of all its electrons, its half-life falls from 43 billion years to 33 years! However, as this will happen nowhere on Earth outside a physics laboratory, we can take its half-life to be 43 billion years for all geological purposes.

So to really affect the half-life of an isotope one needs to resort to highly artificial methods — such as dropping it into the core of a nuclear reactor. In nature, and in particular in rocks, there are sound theoretical and observational reasons to conclude that unstable isotopes will have constant or very nearly constant half-lives and so will undergo decay, and will have undergone decay, in a regular and predictable manner.

5.6.5 Radiometric dating

The astute reader will probably have figured out where this is going. In the decay of unstable isotopes, we have a set of natural processes each of which goes at an utterly predictable rate.

Naively speaking, the idea behind radiometric dating goes something like this. If we have a rock and we know the original quantity of some radioactive isotope that the rock contained when it was first formed, and if we know its half-life, and if we can accurately measure the quantity of the isotope in the rock at the present time, then we can figure out how old the rock is. Or alternatively if we knew the original quantity and the present quantity of the daughter isotope of the radioactive isotope, then again we could figure out the age of the rock. Such a technique of absolute dating is known as **radiometric dating**.

Only how can we know the original composition of the rock? The answer to this question varies depending on which radioactive isotope we're talking about; so at this point it is time to stop talking in generalities and instead look at particular methods of radiometric dating in detail. These will be the subject of the next few articles.

5.7 K-Ar dating

In this article we shall examine the basis of the K-Ar dating method, how it works, and what can go wrong with it.

5.7.1 Decay of ^{40}K

^{40}K (potassium-40) is rather a peculiar isotope, in that it can undergo decay in three different ways: by beta minus

decay into ^{40}Ca (calcium-40); by electron capture into ^{40}Ar (argon-40); and by beta plus decay into ^{40}Ar again. It is possible to measure the proportion in which ^{40}K decays, and to say that about 89.1% of the time it decays to ^{40}Ca and about 10.9% of the time to ^{40}Ar . ^{40}K has a half-life of 1.248 billion years, which makes it eminently suitable for dating rocks.

5.7.2 K-Ar dating

Potassium is chemically incorporated into common minerals, notably hornblende, biotite and potassium feldspar, which are component minerals of igneous rocks.

Argon, on the other hand, is an **inert** gas; it cannot combine chemically with anything. As a result under most circumstances we don't expect to find much argon in igneous rocks just after they've formed. (However, see the section below on the limitations of the method.)

This suggests an obvious method of dating igneous rocks. If we are right in thinking that there was no argon in the rock originally, then all the argon in it now must have been produced by the decay of ^{40}K . So all we'd have to do is measure the amount of ^{40}K and ^{40}Ar in the rock, and since we know the decay rate of ^{40}K , we can calculate how long ago the rock was formed. From the equation describing radioactive decay, we can derive the following equation:

$$t = h \times \log_2(1 + R/c)$$

where

- t is the age of the rock in years;
- h is the half-life of ^{40}K in years;
- c is the proportion of ^{40}K which decays to ^{40}Ar rather than to ^{40}Ca (about 10.9%);
- R is the measured ratio of ^{40}Ar to ^{40}K .

5.7.3 Limitations of K-Ar dating

There are a number of problems with the method. One is that if the rocks are recent, the amount of ^{40}Ar in them will be so small that it is below the ability of our instruments to measure, and a rock formed yesterday will look no different from a rock formed fifty thousand years ago. The severity of this problem decreases as the accuracy of our instruments increases. Still, as a general rule, the proportional error in K-Ar dating will be greatest in the youngest rocks.

A second problem is that for technical reasons, the measurement of argon and the measurement of potassium have to be made on two different samples, because each measurement requires the destruction of the sample. If

the mineral composition of the two sample is different, so that the sample for measuring the potassium is richer or poorer in potassium than the sample used for measuring the argon, then this will be a source of error.

Another concern with K-Ar dating is that it relies on there being no ^{40}Ar in the rock when it was originally formed, or added to it between its formation and our application of the K-Ar method. Because argon is inert, it cannot be chemically incorporated in the minerals when they are formed, but it can be physically trapped in the rocks either during or after formation. Such argon is known as **excess argon**.

If the source of this argon is atmospheric contamination, then we can correct for this. The reasoning is as follows: the atmosphere does not only contain ^{40}Ar , but also ^{36}Ar . There is 295 times as much ^{40}Ar as ^{39}Ar in the atmosphere, and there is no reason why an atom of ^{40}Ar should be preferentially incorporated into rocks rather than an atom of ^{36}Ar , or vice versa. So this means that for every atom of ^{36}Ar we find in our sample, we can discount 295 atoms of ^{40}Ar as being atmospheric argon.

However, this only works if all the excess argon did indeed come from the atmosphere. But consider what happens if the argon came from deep within the Earth, where it was formed by ^{40}K decay, and was then trapped in **magma** or transported into the rock by hydrothermal fluid. Then the excess argon will not have the same $^{40}\text{Ar}/^{39}\text{Ar}$ ratio as is found in the atmosphere, and the formula that corrects for atmospheric carbon will not correct for this.

Finally, we must consider the possibility of argon loss. When a rock undergoes **metamorphism**, some or all of its argon can be outgassed. If all the argon was lost, this would reset the K-Ar clock to zero, and dating the rock would give us the time of **metamorphism**; and if we recognized the rock as **metamorphic** this would actually be quite useful. However, we cannot rely on all the argon being lost, and if it is not then when we apply K-Ar dating this will give us an essentially arbitrary date somewhere between the formation of the rock and the **metamorphosis** event.

For these reasons K-Ar dating has largely been superseded by Ar-Ar dating, which will be the subject of the next article.



Ultrapure argon glowing in a plasma lamp.

5.8.1 The isotopes

In the previous article I introduced you to ^{40}K , an unstable isotope of potassium which produces the **daughter isotope** ^{40}Ar by electron capture or beta plus decay.

The Ar-Ar dating method relies crucially on the existence of two other **isotopes**. ^{39}K is a stable isotope of potassium, which by definition means that it will not spontaneously undergo decay into another isotope. However, if you put it near the core of a nuclear reactor, so that it is bombarded by **neutrons**, then this will convert it into ^{39}Ar . This isotope of argon is quite unstable, having a **half-life** of only 269 years. Consequently, the amount of it found in rocks is negligible — *unless* you subject them to an artificial **neutron source**.

A crucial point to note is that because ^{39}K and ^{40}K are isotopes of the same **element**, they have the same **chemical properties**. Therefore when the rock first forms, some of the minerals in it will have more potassium in and some less, but all the **minerals** will have the same initial *ratio* of ^{39}K to ^{40}K , because since they have identical **chemical properties**, there is no way that the ^{40}K could preferentially end up in the **hornblende** and the ^{39}K in the **biotite**.

5.8.2 The method

First, you take your rock sample and place it near the core of a nuclear reactor. As a result, some of the ^{39}K is converted to ^{39}Ar as a consequence of the **neutron bombardment**.

Then you heat the rock sample to release the ^{39}Ar and the ^{40}Ar . The first of these, you will recall, is produced by our artificial neutron bombardment of the stable ^{39}K isotope; the second is produced by the natural decay of the unstable ^{40}K isotope in the rock.

5.8 Ar-Ar dating

Some of the problems of K-Ar dating can be avoided by the use of the related **Ar-Ar dating** method. In this article we shall explain how this method works and why it is superior to the K-Ar method. The reader should be thoroughly familiar with the K-Ar method, as explained in the previous article, before reading any further.

So if all has gone well, and if there were no problems with argon loss or excess argon, then the age of the sample would be given by the following formula:

$$t = h \times \log_2(1 + J \times R)$$

where

- t is the age of the rock in years;
- h is the half-life of ^{40}K in years;
- R is the measured ratio of ^{40}Ar to ^{39}Ar .

But what is J ? J is a factor which depends on the nature of the neutron bombardment. J is not calculated on theoretical grounds, but is found experimentally; alongside the sample we're interested in, we irradiate and then heat a sample of known age (a **standard**).

Measuring the ^{39}Ar and ^{40}Ar emitted from the standard, and knowing the time t that it was formed, we can put these figures into the equation above and solve it for J .

So now we know J , and we have measured the R -value of the sample we're actually interested in dating, so we can use these data to solve the equation for t , giving us the age we're looking for.

You will note that this means that we have to be able to date some rocks accurately using some method other than Ar-Ar, so that we can find a standard to use for the determination of J ; fortunately we can do this, and geologists have put a lot of effort into identifying rocks which can be accurately dated and used as standards.

5.8.3 Advantages of the Ar-Ar method

So far, all we seem to have done is taken the K-Ar dating method and made it much more complicated for no apparent reason. However, there are advantages to this more complex method.

In the first place, recall that one of the potential problems with the K-Ar method is that it requires two different samples, one to measure the potassium and the other to measure the argon; if the two samples had different chemical compositions when they first formed then this will introduce an error. However, in Ar-Ar dating the two isotopes of argon are both measured from the same sample, and so at least one potential source of error is eliminated.

The other important advantage of Ar-Ar dating is the extra data gained from **step heating**: instead of heating the irradiated sample to the highest possible temperature all at once, and so releasing all the argon all at once, we can increase the temperature in steps starting at a low temperature.

What's the point of this? Well, different minerals within the rock will give up their argon at different temperatures,

so each step will give us a ratio of ^{40}Ar to ^{39}Ar which we can use in the equation to calculate a date. Now, recall that we said that when the rock was first formed, the ^{39}K and ^{40}K from which these are derived must have appeared in the same ratio in each mineral, because both isotopes of potassium have the same chemical properties.

This means that if the rock cooled rapidly enough that all the minerals in it have the same date, and if there has been no argon loss, and if there is no excess argon added to the system, then the dates we calculate at each step of the heating will be the same date.

If we don't get the same date at each step, then we may be able to work out what's going on.

For example, if the date increases at each step, then we are quite possibly looking at a slow-cooling igneous rock in which different minerals crystallized out of the magma at different times, a possibility we can investigate further.

Or if we consistently get one date for the steps below (for example) 400°C , and consistently get another date in the steps above 400°C , then it seems as though argon loss occurred as a result of **metamorphism** at a temperature of about 400°C , with the younger date representing the date of the **metamorphism**, and the older date representing the formation of the rock; and we can investigate this clue further by looking for other evidence of the metamorphic event.

And if the dates we get are all over the place, then we are probably looking at excess argon. Now the bad news is that there is no way we can somehow manipulate this data to give us a correct date for the sample. But the good news is that we do *know* that there's a problem; whereas if we'd analyzed the same rock using the K-Ar method, then it would have supplied us with a date and there'd have been no sign in the K-Ar data of anything wrong with it.

For these reasons Ar-Ar dating has largely superseded **K-Ar dating**, although the simpler method is still employed in some cases where it is known to be unproblematic or where Ar-Ar is unsuitable for some technical reason.

5.9 Rb-Sr dating

In this article I shall introduce the Rb-Sr dating method, and explain how it works; in the process the reader should learn to appreciate the general reasoning behind the isochron method.

5.9.1 The isotopes

There are three isotopes used in Rb-Sr dating. ^{87}Rb (rubidium-87) is an unstable isotope with a half-life of about 49 billion years. It produces the stable daughter isotope ^{87}Sr (strontium-87) by beta minus decay. The third isotope we need to consider is ^{86}Sr , which is stable

and is not **radiogenic**, meaning that in any closed system the quantity of ^{86}Sr will remain the same.

As rubidium easily substitutes chemically for potassium, it can be found doing so in small quantities in potassium-containing minerals such as **biotite**, potassium **feldspar**, and **hornblende**. (The quantity will be small because there is much more potassium than rubidium in the Universe.)

5.9.2 Strontium in rocks

This means that if we wanted to date a rock, and if there was no ^{87}Sr present initially, and if we could measure the $^{87}\text{Sr}/^{87}\text{Rb}$ ratio present today, then it would be easy to derive a formula giving the age of the rock: it would be like that we used for **K-Ar**, except that as ^{87}Rb has only one decay mode, we could drop the term c .

But there is no reason at all to suppose that there was no ^{87}Sr present initially. When we produced the formula for **K-Ar** dating, it was reasonable enough to think that there was little to no argon present in the original state of the rock, because argon is an inert gas, does not take part in chemical processes, and so in particular does not take part in mineral formation.

Strontium, on the other hand, does take part in chemical reactions, and can substitute chemically for such elements as calcium, which is commonly found in **igneous rocks**. So we have every reason to think that rocks when they form *do* incorporate strontium, and ^{87}Sr in particular.

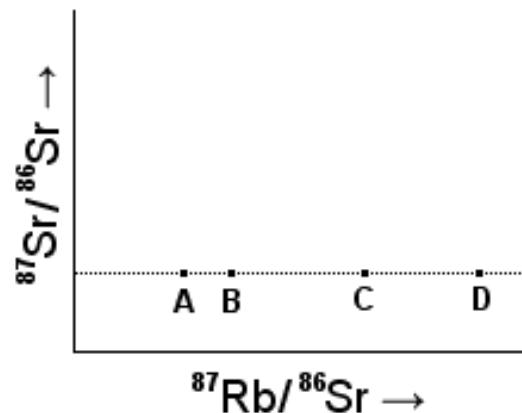
5.9.3 The isochron diagram

However, there is still a way to extract a date from the rock. In the reasoning that follows, the reader may recognize a sort of family resemblance to the reasoning behind **step heating** in the **Ar-Ar** method, although the two are not exactly alike.

The reasoning, then, goes like this. When an igneous rock is first formed, its minerals will contain varying concentrations of rubidium and strontium, with some minerals being high in rubidium and low in strontium, others being high in strontium and low in rubidium. We can expect these differences to be quite pronounced, because rubidium and strontium have different chemical affinities: as we have noted, rubidium substitutes for potassium, and strontium for calcium.

Now consider the distribution of the two strontium isotopes ^{87}Sr and ^{86}Sr . Because they are *chemically indistinguishable*, they will appear in the same ratio in every mineral at the time when it is formed: some minerals will have more strontium, some will have less, but all must necessarily have the same $^{87}\text{Sr}/^{86}\text{Sr}$ ratio.

The initial state of the rock may therefore be schematically represented by the graph to the right, which shows the initial states of four minerals imaginatively named A,

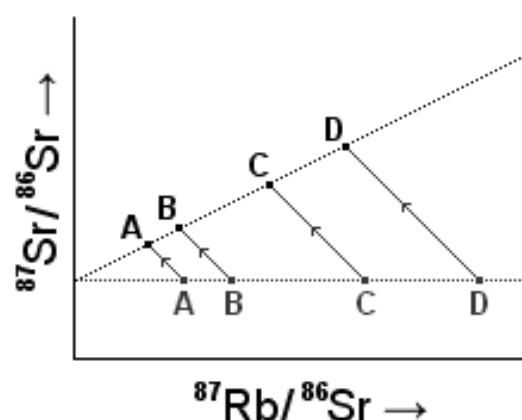


Starting position of an isochron diagram.

B, C, and D. They have different chemical compositions, and therefore have different $^{87}\text{Rb}/^{86}\text{Sr}$ ratios, but they all have exactly the same $^{87}\text{Sr}/^{86}\text{Sr}$ ratio, for reasons explained in the previous paragraph. Hence the dotted line connecting the four minerals and extended beyond them must be straight and horizontal, and the point at which it intersects the vertical axis is the initial value of the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio.

Now consider what will happen to this system over time, as the ^{87}Rb decays to ^{87}Sr . For each mineral, this will decrease the $^{87}\text{Rb}/^{86}\text{Sr}$ ratio and increase the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio.

This will have the biggest impact on the ratios of minerals such as D which have high initial $^{87}\text{Rb}/^{86}\text{Sr}$ ratios, and the smallest impact on the ratios of minerals such as A, which have low initial $^{87}\text{Rb}/^{86}\text{Sr}$ ratios. (If you have difficulty seeing this, try considering the extremal case of a mineral which contains no rubidium at all. Its $^{87}\text{Rb}/^{86}\text{Sr}$ ratio will be initially zero and will stay that way.)



The isochron diagram changes over time.

The effect of the decay process on the isotope ratios can again be plotted on a graph, as shown to the right.

We shall omit the math, but it happens to work out so that after any given period of time, the minerals will *still* lie on a straight line on the graph, as the diagram shows, and, crucially, the point at which this line intersects the vertical axis is *still* the initial value of $^{87}\text{Sr}/^{86}\text{Sr}$.

So now we can find a date for the rock. What we have to do is take samples from the rock consisting of different minerals, or at least of different mineral composition, so that our samples will all have different $^{87}\text{Rb}/^{86}\text{Sr}$ ratios.

For each sample we then measure its $^{87}\text{Rb}/^{86}\text{Sr}$ ratio and its $^{87}\text{Sr}/^{86}\text{Sr}$ ratio.

We then use the isochron diagram to find the initial value of the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio. This one additional piece of information about the *initial* state of the rock allows us to calculate its age.

5.9.4 Confounding factors

As with the other methods we've discussed so far, the Rb-Sr method will only work if nothing but the passage of time has affected the distribution of the key isotopes within the rock. And of course this is not necessarily the case. **Hydrothermal** or **metasomatic** events may have added or subtracted rubidium and strontium to or from the rocks since their formation; or a **metamorphic** event may have redistributed the rubidium or strontium among its constituent **minerals**, which would also interfere with the method.

However, barring an extraordinary coincidence, the result of such events will be that when we draw the isochron diagram, the minerals will no longer lie on a straight line. A small deviation from a straight line tells us that there is some uncertainty about the date, and this degree of uncertainty can be calculated; and if we get something which is nothing like a straight line, then the method simply doesn't supply us with a date. So just as step heating in Ar-Ar dating protects us from error, so too does the isochron method in Rb-Sr dating: it may not always lead us to the right date, but it is a good safeguard against our accepting one that is wrong.

5.9.5 Mixing

There is, however, one potential source of error which will not show up on the isochron diagram, since it is expected to produce a straight line. Suppose that the original source of the rock was two different **magmas** (call them X and Y) imperfectly mixed together so that some parts of the rock will be all X, some all Y, some part X and part Y in varying proportions. Then these different parts of the rock, when analyzed for their isotopic composition, will plot in a straight line on the isochron diagram; and the slope of this line, and the point at which it intercepts the vertical axis, will have nothing to do with the age of the rock, and everything to do with the compositions

of X and Y.

About half the time this will produce a straight line with negative slope: that is, it will slope down from left to right instead of up. Such a line must necessarily be produced by mixing, since a real isochron will always have positive slope: the rarity of such an occurrence tells us that mixing of this type must itself be rare.

We can also test for mixing using what is known as a **mixing plot**: if we draw up a graph of the composition of our sample in which the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio is the vertical axis (as in the isochron diagram) but the horizontal axis represents $1/\text{Sr}$ (the reciprocal of the quantity of both isotopes of strontium taken together) then if the rock was produced by this mixing process, then the points on this graph will lie along a straight line.

It can happen that if we produce a mixing plot for a perfectly good isochron, it will by some statistical fluke produce a straight line on the mixing plot; we would then be throwing out a perfectly good date. However, this is worth it: it would, as I say, require a fluke for this to happen, so if we reject dates based on the mixing plot, then we will be throwing out a hundred bad dates for every good one.

5.10 Other isochron methods

In this article I shall point out some other dating methods which work the same way as the **Rb-Sr method**. The reader who has not read [the article on the Rb-Sr method](#) will find this present article almost completely incomprehensible, and should go back and read it.

5.10.1 The isochron method generalized

I have introduced the **isochron** method in the context of rubidium and strontium. But is there anything particularly special about those two elements? Not really. For the isochron method to work, what we need are three isotopes with the following properties.

- (1) An **unstable isotope**. This should have a fairly long half-life if it is to be of any use in dating rocks, but not too long, or it will hardly undergo any decay at all. A figure expressible in billions of years is ideal. In the Rb-Sr method, we used ^{87}Rb .
- (2) A **stable daughter isotope of isotope (1)**. In the Rb-Sr method, we used ^{87}Sr .
- (3) An **isotope** which is the same **element** as isotope (2) and which is neither unstable nor radiogenic, so that in a closed system it remains constant in quantity. In the Rb-Sr method, we used ^{86}Sr .

Given a sets of three such **isotopes**, we can apply exactly the same reasoning as we did for ^{87}Rb , ^{87}Sr and ^{86}Sr , and it will be equally valid.

5.10.2 The isotopes

The table below shows some sets of three isotopes which can be treated like rubidium and strontium for the purposes of dating; the table also shows the half-life of the parent and its decay mode. The numbers (1) (2) and (3) are as in the section above.

5.10.3 Notes

- I said that isotope (3) should be stable. ^{186}Os is not in fact stable, but as it has a half-life of two quadrillion years, it might as well be.
- In the Lu-Hf method, again, ^{144}Nd is unstable, but has a half-life even longer than that of ^{186}Os .
- Similarly in the La-Ce method, *neither* isotope of cerium used is strictly speaking stable, but their half-lives are so enormously long that for all practical purposes they may be treated as stable.
- In using the K-Ca method, we have to make a slight mathematical adjustment to take into account the fact, mentioned in our article on the K-Ar method, that ^{40}K decays to ^{40}Ar as well as to ^{40}Ca .
- Similarly, ^{138}La can decay two ways, to ^{138}Ce or ^{138}Ba . As you can see from the table, both are susceptible to the isochron method.
- We have noted the peculiarities of the half-life of ^{187}Re in the article on radioactive decay. As we only have to consider how it behaves in rocks, and not in elaborate equipment in physics laboratories, we may take its half-life to be 43 billion years as given in the table.
- There are two entries for U-Pb because there are two parent isotopes we can use, ^{238}U and ^{235}U . Each decays to a (different) final stable element of lead by a complex decay chain.
- In practice, the U-Pb decay chain is usually exploited by methods other than isochron dating, for reasons that will be explained in the next article.

5.11 U-Pb, Pb-Pb, and fission track dating

In this article we shall discuss the basis of the U-Pb and Pb-Pb methods, and also fission track dating. The reader will find this article much easier to grasp if s/he has already mastered the material in the articles on K-Ar dating, Ar-Ar dating, and Rb-Sr dating.

5.11.1 The isotopes

There are a number of isotopes of interest in U-Pb dating. ^{238}U (uranium-238) decays to ^{206}Pb (lead-206) by a complex decay chain. It has a half-life of 4.5 billion years. ^{235}U (uranium-235) decays to ^{207}Pb (lead-207) by an equally complex decay chain, and has a half-life of 0.7 billion years.

It is also useful to know of the existence of ^{204}Pb (lead-204), which is neither unstable nor radiogenic.

5.11.2 Isochron dating and U-Pb

We can always try U-Pb dating using the isochron method, but this often doesn't work: the compositions of the minerals involved, when plotted on an isochron diagram, fail to lie on a straight line.

There seem to be two reasons for this. First of all, the straight-line property of the isochron diagram is destroyed when the isotopes involved get shuffled between minerals. Now lead and uranium are particularly susceptible to such shuffling in the event of even mild metamorphism. The other problem is that uranium is particularly susceptible to weathering. Now since all rocks are somewhat porous, and since we are pretty much obliged to date rocks from near the surface, it's hard to find instances in which uranium has not been lost.

5.11.3 Zircons



Zircon.

Zircon is the mineral ZrSiO_4 ; as you can see from its chemical formula, it is one of the silicate minerals. Although it is not abundant in igneous rocks, it is sufficiently common to be used for the purposes of radiometric dating.

It has two properties which make it useful for this purpose.

First of all, uranium will readily substitute for the zirconium (Zr) in the mineral, whereas lead is strongly rejected. For this reason we expect zircons, when formed, to contain some uranium, but virtually no lead.

Second, zircons are durable and chemically inert, able to resist chemical weathering and even high-grade metamorphism up to about 900°C.

Zircons, then, are relatively immune to the problems that make isochron U-Pb dating so difficult. But of course for isochron dating we need more than one mineral; zircons alone would not be enough.

However, these facts about zircons, combined with what we know about uranium, suggest an alternative method of dating.

5.11.4 The method

If there is no lead in the zircon originally, and if no lead or uranium has been added or subtracted to the zircon since its formation, then the following formula will hold:

$$t_{238} = h_{238} \times \log_2(1 + R_{238})$$

where t_{238} is the age of the zircon, h_{238} is the half-life of ^{238}U and R_{238} is the ratio of ^{206}Pb to ^{238}U .

Not only that, but since we have two uranium isotopes to work with, we will also have

$$t_{235} = h_{235} \times \log_2(1 + R_{235})$$

where t_{235} is the age of the zircon, h_{235} is the half-life of ^{235}U and R_{235} is the ratio of ^{207}Pb to ^{235}U .

Now because the zircon has only one age, it follows that if no lead or uranium has been added or subtracted from the zircon since its formation, we will have $t_{238} = t_{235}$, in which case the two t values are said to be **concordant**; whereas if lead and/or uranium has been added or subtracted, then it would require some sort of statistical fluke for the two t values to end up identical. So analysis of both the $^{206}\text{Pb}/^{238}\text{U}$ ratio and the $^{207}\text{Pb}/^{235}\text{U}$ ratio acts as a check on the correctness of the date we come up with in the same way that **step heating** does in the Ar-Ar method and the plotting of several minerals on an isochron diagram does for the Rb-Sr and related methods: it allows us to find out if the isotope ratios have been affected by something other than the passage of time, and to reject any “dates” calculated from the isotope ratios if this turns out to be the case.

It is possible to refine this date still further. If we suspect that the zircon, despite its chemical properties, still managed to incorporate a little lead at or after its formation, then since all lead isotopes are chemically the same, we can measure the amount of ^{204}Pb the zircon contains. Since we know the ratios in which the various lead isotopes are usually found, we can then apply the same sort

of correction we used to account for atmospheric argon in the K-Ar method.

While zircon has been the most popular mineral for U-Pb dating, other minerals have been employed, including apatite, monazite, titanite, allanite and, most interesting of all, xenotime.

5.11.5 Xenotime

There is a difficulty in using radiometric dating to put an age on **sedimentary rock**. The problem is that **sediment** is made up of **clasts** of some **parent rock**, and when we date these **clasts**, we are in effect dating the **parent rock** rather than the **the sediment** as such. If, for example, we apply U-Pb dating to a grain of zircon found in sandstone, we aren't dating the formation of the sandstone, we're dating the formation of the **granite** that the zircon came from; all we could say about the sandstone is that it must be younger than that.

However, it is possible to put a date on some sedimentary rocks using the mineral **xenotime** (YPO_4). Uranium can and often does substitute for the **element** yttrium, whereas lead cannot, making xenotime suitable for radiometric dating.

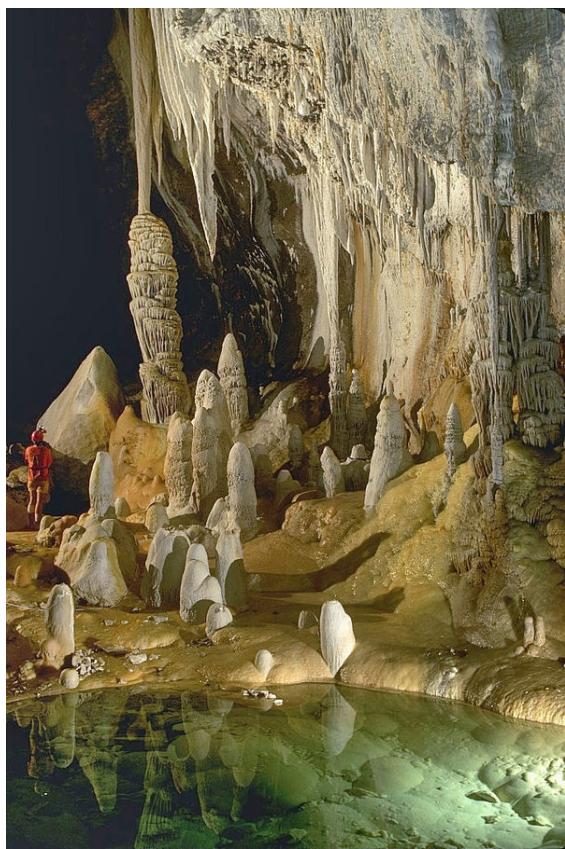
The key fact about xenotime is that since it has the same **crystal structure** as zircon, it can grow on zircon **crystals**, forming a crust; and this process, of course, cannot begin to take place while the zircon **crystal** is still locked inside its **parent rock**. The zircon will only start acquiring its xenotime crust after weathering and erosion have freed it from its **parent rock** and it becomes **sediment**.

So by dating the xenotime crust, we can find out when the zircon it's growing on became **sediment**; dating the zircon itself would tell you the age of the **parent rock**.

5.11.6 Speleothems

A **speleothem**, more colloquially known as a **cave formation**, is formed when minerals dissolved in water **precipitate** out of the water as it drips, seeps, or flows into a cave. The reader will probably be familiar with **stalagmites** and **stalactites**; more speleothems are shown in the photograph to the right.

Now, compounds of uranium are often highly soluble in water (this, indeed, is one of the major problems with U-Pb isochron dating) whereas compounds of lead are stubbornly **insoluble**. As a result, we expect speleothems when they are first formed to contain some uranium but little or no lead — just like zircons. So we can apply the same technique to speleothems as we do to zircons.



A cave with speleothems.

5.11.7 Pb-Pb dating

We can exploit our double system of ^{238}U - ^{206}Pb and ^{235}U - ^{207}Pb in another way. Suppose we separate the minerals in a rock, and plot a graph showing their $^{206}\text{Pb}/^{204}\text{Pb}$ ratios on one axis and their $^{207}\text{Pb}/^{204}\text{Pb}$ ratios on another (similar, though not identical, to what we did when constructing **isochron diagrams**). It can be shown mathematically that if the rock has been undisturbed, so that the **isotope** ratios reflect nothing but the passage of time, then just as with the **isochron diagrams** we've already discussed (though for a different reason) the minerals so plotted will lie on a straight line on the graph; and the age of the rock can be calculated from the slope of the line.

Unlike the ordinary **isochron methods** such as Rb-Sr, the Pb-Pb method does not allow us to deduce the original proportions of the various lead isotopes from the data acquired from the sample. Instead, we need to find this out some other way.

We can do this by finding minerals that contain lead but never contained any uranium, or only ever contained it in negligible quantities. Troilite (FeS) from iron-rich meteorites fits the bill: its present ratio of uranium to lead is so tiny that either the solar system and indeed the universe is many many times older than cosmologists think, or, given the long **half-life** (4.5 billion years) of ^{238}U , there

can hardly have been any uranium in the meteorites to start with, and so its decay can hardly have affected the lead isotope ratios of these meteorites.

You might perhaps doubt that meteorites would have the same initial lead isotope ratios as the Earth. Planetary scientists maintain that they should, for reasons which are somewhat beyond the scope of this textbook. Another reason for believing it is that if we calculate Pb-Pb dates on this basis, the dates we get are in agreement with dates produced by other methods where they can be applied: this would hardly be possible if we were using the wrong figures for the initial lead isotope ratios. So taking the figure derived from the troilite as an "anchor" for our calculations, we can then go ahead and apply the Pb-Pb method to rocks which do contain significant quantities of uranium.

Now, recalling that I began this article by explaining that the **isochron method** is no use for U-Pb, you may wonder why this Pb-Pb isochron should be any better. However, recall that one of the major problems with the U-Pb isochron is that uranium compounds are highly soluble and are easily removed from the rock by **weathering**. But when that occurs, the lead will still remain and can be used for Pb-Pb dating. What's more, even if some lead is also removed, then since all the lead isotopes are the same element, having the same merely chemical properties, there will be no tendency for one isotope to be lost in a greater proportion to the others, and so the isotope ratios will remain the same.

It is true, of course, that the removal of the uranium by **weathering** will slow down and even, if all the uranium is removed, completely stop the radiometric clock, so that we will not have an accurate measurement of the time after the **weathering** began, and Pb-Pb dating will therefore tell us that the rock is a little younger than it is. But only a *little* younger, because a typical chunk of **igneous** rock will only have spent a relatively short amount of time being exposed to **chemical weathering** compared to the time when it was not.

As with the isochron methods we've already met, the Pb-Pb isochron method carries its own built-in check on its correctness: if the rock has been seriously disturbed, so that the **isotope** ratios depend on something other than the passage of time, then when we plot the **minerals** on our graph, they will almost certainly not lie on a straight line, and we will not obtain a date.

5.11.8 Fission track dating

Finally, I should mention **fission track dating**. The decay chain by which uranium decays involves the emission of **alpha particles**, and as these particles travel through the rock they produce microscopic scars (**fission tracks**) in the **minerals** they pass through, which can be revealed by cutting and polishing the minerals and inspecting them through a microscope. A number of minerals are suitable

for this process, including apatite, zircon, and titanite.

The number of fission tracks in the minerals will depend on the quantity of uranium and the amount of time it's had to do damage. So, conversely, if we count the fission tracks and we measure the amount of uranium, then we can figure out how much time it must have taken to produce the fission tracks.

One weakness of this method is that the fission tracks will heal over if the rock is heated to about 200°C , so the fission track clock will be reset even by the mildest metamorphism.

5.12 Radiocarbon dating

In this article we shall discuss how radiocarbon dating works, the conditions under which it can be applied, and the limitations of the method.

5.12.1 The isotopes

There are three important isotopes underlying the process of radiocarbon dating.

^{14}N (nitrogen-14) is converted to ^{14}C (carbon-14) in the upper atmosphere as a result of bombardment by neutrons in so-called **cosmic rays**: high-energy particles bombarding the Earth's atmosphere from outer space. Such an isotope is said to be **cosmogenic**. On formation, the newly-born carbon atom quickly oxidizes to form a molecule of carbon dioxide (CO_2).

^{14}C is an **unstable isotope** of carbon, and so decays back to ^{14}N via beta decay with a **half-life** of about 5730 years. Because the quantity of ^{14}C being produced annually is more or less constant, whereas the quantity being destroyed is proportional to the quantity that exists, it can be shown that the quantity in the atmosphere at any given time will be more or less constant: the processes of production and decay of ^{14}C produces an equilibrium.

Also of importance is the **stable** carbon isotope ^{12}C ; this makes up 98.89% of atmospheric carbon, as opposed to only 0.000000001% ^{14}C . The balance is made up by the stable isotope ^{13}C , which need not concern us in this article.

5.12.2 The terrestrial carbon cycle

The terrestrial carbon cycle is fairly simple: plants get their carbon from the atmosphere via the process of photosynthesis; herbivores get their carbon from plants, and carnivores from the herbivores. After the death of the organism, processes of decay will return its carbon to the atmosphere, unless it is sequestered — for example in the form of **coal**.

This means that when an organism is alive, its ratio of $^{14}\text{C}/^{12}\text{C}$ will be the same as the ratio in the atmosphere. But of course when the organism dies it is cut off from the source of atmospheric carbon, the ^{14}C will start to decay to ^{14}N , and the ratio will begin to change.

5.12.3 The method

This immediately suggests a method of dating organic remains. If we measure the amount of ^{12}C and the amount of ^{14}C in an organic sample, then since we know the atmospheric ratio and the amount of ^{12}C present, we can deduce how much ^{14}C was present originally. And then since we know how much was present originally, since we can measure how much is present now, and since we know the decay rate of ^{14}C , it is trivial to compute the age of the sample.

This method is variously known as **radiocarbon dating**, **carbon dating**, **^{14}C dating**, or **C-C dating**.

One of the nice things about this method is that we don't have to worry about carbon being lost from the sample. Because we are measuring the abundance of two isotopes of carbon, and because isotopes of the same element will be chemically identical, no ordinary process can preferentially remove ^{12}C or ^{14}C , and so any process of carbon removal will leave the $^{12}\text{C}/^{14}\text{C}$ ratio the same, and the method will still work.

5.12.4 Limitations of the method

The method has various limitations. First of all, the quantity of ^{14}C is going to be small enough to begin with, being only 0.000000001% of atmospheric carbon, and then as the decay process progresses it's going to get smaller and smaller. After about 60,000 years the quantity will be too small for our instruments to measure accurately, and the best we'll be able to say is that the sample is about 60,000 years old or more. For this reason radiocarbon dating is of more interest to archaeologists than to geologists.

Two effects also interfere with the dating of very recent samples. The testing of thermonuclear weapons produced an increase in atmospheric ^{14}C , peaking in the mid-1960s; and the burning of fossil fuels has been causing an increase in atmospheric ^{12}C ; this has not been accompanied by a corresponding increase in ^{14}C because as the carbon in coal and oil is old, the amount of ^{14}C they contain is infinitesimal. Fortunately it is rarely necessary to use radiocarbon methods to date very recent samples.

Thirdly, it is in the nature of the method that it can only be applied to organic remains: it makes no sense to apply it to rocks or to mineralized fossils.

Fourthly, the carbon in the organic remains does have to originate with the terrestrial carbon cycle and with plants performing photosynthesis. If this is not the case, it is

sometimes possible to correct for the fact; in other cases it makes dating impossible.

For example, marine carbon behaves quite differently from carbon in the terrestrial cycle. The residence time of carbon in the ocean can be measured in hundreds of thousands of years (where the **residence time** of carbon is defined as the average time an atom of carbon will stay in the ocean). This increases the apparent age of the sample by about 400 years, depending on where in the ocean the organism lived and died. Given a latitude and longitude, an appropriate correction to the date is supplied by the **Marine Reservoir Database**.

Since humans eat seafood, this can also affect the carbon dating of humans, and what is worse it does so in an inconsistent manner, since human consumption of seafood varies with location and culture. However, the marine component of diet can be estimated by measuring the ratio of the stable isotopes $^{15}\text{N}/^{13}\text{C}$ in the sample: this will be higher the more seafood the individual consumed. This allows archaeologists to estimate the magnitude of this effect and correct for it.

Another source of carbon we have to take into account is the **weathering of limestone**. The result of this is to supply streams, rivers, and lakes with a source of dissolved **calcium carbonate**; if freshwater shellfish (for example) use this to construct their shells, then they are using a source of carbon which is millions of years old. Clearly applying radiometric dating in such a case is pointless. Another source of old carbon is the outgassing from volcanoes: in locations where this is a significant source of CO_2 , plants growing in the area will appear older than they actually are.

Even participation in the terrestrial carbon cycle does not *quite* guarantee the date: we could, for example, imagine termites eating their way through the wood of a 200 year old house; these termites would date to 200 years old or more (depending on the age of the tree). By and large, however, organisms tend to consume fresh vegetation or fresh meat, so this problem is unlikely to arise in practice.

5.12.5 Comparison with known dates

One way we can check the efficacy of radiocarbon dating is to compare the dates it produces with dates known on historical grounds, to ensure that it does indeed give us the right answer.

We can also compare radiocarbon dates with dates known on other grounds. For example, we have discussed the use of **varves** for dating; now since varves incorporate organic material as they are formed, we can check that when we radiocarbon date a varve, we get the same date for it as we obtain by counting the varves.

Also it is obviously possible to carbon-date one of the growth rings of a tree, and to compare the date produced by radiocarbon dating with the date produced by

dendrochronology. Such dates typically agree to within 1 or 2 per cent.

5.12.6 Calibrated dating

Although the radiocarbon dates agree closely with dendrochronology, they do not agree exactly. It is generally agreed that the dendrochronological dates should be considered the more accurate. The proportion of ^{14}C in the atmosphere is not absolutely constant; for example, it can be reduced by volcanic activity, since the carbon dioxide emitted by volcanoes is richer in ^{12}C than atmospheric carbon dioxide. By comparison the behavior of the genera of trees used in dendrochronology is more reliable and consistent.

It is therefore standard procedure to tweak the raw radiocarbon dates to bring them in line with dendrochronology, producing what are known as calibrated radiocarbon dates. This allows us to combine the greater accuracy of dendrochronology with the wider applicability of radiocarbon dating.

5.13 Cosmogenic surface dating



A surface exposed by the action of glaciers, Kosterhavet national park, Sweden.

In the article we shall discuss the techniques and applications of cosmogenic surface dating.

5.13.1 Applications

Unlike other dating methods, which tell us how long it is since a rock was formed, **cosmogenic surface dating** tells us how long a rock has been exposed on the surface.

In some cases, as when the rock is a lava flow, this amounts to the same thing. But there are other ways in which a rock can become exposed, as for example when a glacier erodes the sediment covering bedrock: when the glacier melts, the **bedrock** will be exposed.

5.13.2 Cosmogenic isotopes

In the article on radiocarbon dating we have already introduced one cosmogenic isotope, ^{14}C , which is produced by cosmic rays from ^{14}N .

For cosmogenic surface dating, the two most commonly used isotopes are the cosmogenic isotopes ^{10}Be , which is produced from ^{16}O and which has a **half-life** of 1.39 million years; and ^{26}Al , which is produced from ^{26}Si and which has a **half-life** of 717,000 years.

5.13.3 The method

Because the isotopes we're using have a short **half-life**, it follows that if a rock has been buried for a few million years the quantities of these isotopes will be negligible. But when the rock becomes exposed on the surface, and so exposed to cosmic rays, these cosmogenic isotopes will begin to accumulate in the rock.

The rate at which they do so will depend on a number of factors, including:

- The exposure of the rock. A nearby obstacle such as a mountain will shield the rock from cosmic rays coming from that direction, reducing the creation of cosmogenic isotopes.
- The elevation of the rock. If the rock is on top of a mountain, then the **cosmic rays** have less atmosphere to travel through to get to the rock, and so more of them will make the journey all the way to the rock without being absorbed in the atmosphere on the way.
- The depth from which we take the sample. **Cosmic rays** can penetrate a few meters through rock or soil, but the further they travel the more likely they are to be absorbed, so a rock sample will get more exposure to **cosmic rays** if it is taken from the surface than if it is taken from a meter down.

If we take all the relevant factors into account, and calculate, estimate, or simply measure the amount of **cosmic rays** a given rock is exposed to per year, and if we measure the quantities of the cosmogenic isotopes in a sample of the rock, then we can figure out how long the rock has been exposed.

5.13.4 Limitations of the method

The quantity of the relevant isotopes in the rock will not simply grow without limit with longer and longer exposure to **cosmic rays**; rather they will tend towards a maximum (a **secular equilibrium**): the point at which the cosmogenic cosmogenic production of **unstable isotopes**

is equaled by their destruction by **decay**. In practice, we are not going to be able to tell the difference between a rock which has reached 99.9% of this maximum and one which has reached 99.99%. Consequently, the practical limit for the use of cosmogenic surface dating seems to be about 10 million years; after that, one old rock looks much like another. The lower limit for application of the method seems to be about ten years, because of practical limits on the accuracy with which we can measure the quantities of the relevant isotopes.

5.14 U-Th, U-Pa, and Ra-Pb dating

In this article we shall discuss three similar methods that can be used to date marine and **lacustrine** sediments: the **U-Th**, **U-Pa**, and **Ra-Pb** methods.

5.14.1 The isotopes

The methods discussed in this article each require two isotopes: a **parent isotope** which is **soluble** (or the commonly occurring compounds of which are **soluble**) and a radioactive **daughter isotope** which is not **soluble**.

The table below shows three such systems together with the **half-life** of the **daughter isotope**, since this is the crucial figure.

5.14.2 The method

The **parent isotope** will be present dissolved in the ocean or in lakes, but when decay takes place the **insoluble daughter isotope** will precipitate out as **sediment** and will form part of the upper layer of marine or lacustrine sediment. It will subsequently be buried in its turn by further sediment, and being **radioactive** will undergo **decay**.

Now, if there was absolutely none of the **parent isotope** present in the sediment, then the calculation would be very simple: when we have dug down through the sediment up to the point where the **daughter isotope** is only half as abundant as it is on the surface, then we would have dug back through one **half-life**'s worth of time; and in general we could write:

$$t = h \times \log_2(N/N_s)$$

where

- t is the age of the sediment;
- h is the **half-life** of the **daughter isotope**;
- N_s is the quantity of the **daughter isotope** on the surface layer of sediment;

- N is the quantity of the daughter isotope at the depth we're trying to date.

That would be the simple case: however it will not necessarily be true that there will be none of the parent isotope in the sediment. There may well be some, but this is not a problem, since we can measure the quantity of the parent isotope present in the upper layers of sediment and take this into account in our calculations. The crucial point is that there will be more of the daughter isotope than could be accounted for by the decay of the parent within the sediment.

5.14.3 Note on the use of Ra-Pb

All the methods described in this article are somewhat limited in their usefulness by the short half-lives of the daughter isotopes. This is particularly true of ^{210}Pb ; since it has a half-life of only 22 years, this makes it useless for most geological purposes. However, it can be used to gauge the rates of deposition of marine sediment as an alternative to the use of sediment traps.

This method has a couple of advantages over sediment traps. First, it is quicker: it doesn't take long to obtain a sediment core sample, whereas a sediment trap has to be left in place for at least a year to produce useful results.

Second, use of Ra-Pb allows us to measure the sedimentation that has taken place over the course of a century or so and average it, reducing the effect of small-scale fluctuations on the figures we obtain.

5.14.4 Alternate use of U-Th

We can make an alternative use of the fact that ^{234}U is soluble and ^{230}Th is not.

First of all, this means that ^{234}U will be incorporated into the structure of marine organisms such as corals. Secondly, it means that ^{234}U will be incorporated into speleothems and ^{230}Th will not, just as with the U-Pb method discussed in the article on U-Pb and related methods.

There is, however, a difference between U-Pb and U-Th: ^{230}Th is radioactive. Whereas this was essential to its use in dating marine sediments, it is actually an inconvenience when dating organic remains or speleothems, since it means that the ^{230}Th will not only be produced by decay, but also destroyed by it.

As a consequence, what happens is that the quantity of ^{230}Th in the sample will tend towards secular equilibrium: the point at which the thorium is being produced at the same rate as it is being destroyed. This fact, combined with the practical difficulty of measuring whether the level of ^{230}Th has reached 99.9%, 99.99%, or 99.999% of secular equilibrium, limits the useful range of the method to about 500,000 years.

Because this method can be applied to organic materials, it can be correlated with the radiocarbon method, and the dates produced by both methods can be shown to be concordant.

5.15 Paleomagnetic dating

In this article we shall discuss how we can use the paleomagnetism in rocks to attach dates to them (**paleomagnetic dating**). The reader may find it useful to go back and read the main article on paleomagnetism before continuing.

5.15.1 Polar wander and dating

Once we have dated a sufficient number of rocks and measured the orientation of the magnetism they contain, we can build up a picture of how the position or apparent position of the poles over time.

So if we are then faced with a rock the date of which we do not know, then we do know (of course) the latitude and longitude at which we found it, and we can measure the orientation of its magnetism, and so we can look at the global picture we've built up of continental drift, and so figure out when the rock must have formed in order to have its magnetism oriented in just that direction.

5.15.2 Magnetic reversals and dating

Once we have dated a sufficient number of rocks and found out whether they have normal or reverse polarity, we can likewise build up a timeline for the occurrence of the reversals.

As noted in a previous article, magnetic reversals come at irregular intervals. This means that the pattern of normal and reverse polarity in an assemblage of rocks can be distinctive in the same way (though for a completely different reason) that growth rings in a tree can be distinctive. We might, for example, see a long period of reverse polarity, followed by six very quick switches of polarity, followed by a long period of normal polarity; and this might be the only time that such a thing occurs in our timeline.

So if we are presented with an undated rock, and we find a really distinctive pattern of paleomagnetic reversals within it, we may be able to identify the one time at which such a sequence of magnetic reversals took place.

5.15.3 Strengths and weaknesses of the method

The reader will observe that it is necessary to be able to date some rocks, in fact a lot of rocks, before paleomagnetic dating can be brought into play. You may there-

fore be wondering why, if we have perfectly good dating methods already, we don't just use them.

However, the advantage of paleomagnetic dating is that we can use it on different rocks from those susceptible to our ordinary methods of **absolute dating**: while most radiometric methods usually require igneous rocks, paleomagnetism can be measured in sedimentary rocks.

One problem which may arise is that the direction of the poles from a given location, or the pattern of magnetic reversals, may repeat over a long enough period of time, so that the **paleomagnetic** data we get when we measure these factors are not unique to a single time in the history of the Earth.

It is possible to get round this problem if we can find an approximate date of the rocks by other means. For example, if by considering their **stratigraphic** relationship to a datable **igneous rock** we can establish that they are (for example) less than 20 million years old, then it may turn out that the **paleomagnetic** data, though not unique over the whole history of the Earth, are unique over the course of the last 20 million years, and then we can go ahead and use paleomagnetic dating.

5.16 Sclerochronology



Arctica islandica, a favorite shell of sclerochronologists.

In this article we shall discuss the basis of **sclerochronology**, a method of dating shells and corals by analysis of their growth patterns.

5.16.1 Growth patterns in shells and corals

When shelly organisms grow, many types lay down bands of new growth in a way that regularly reflects the passage of time: for example, laying down one growth band per day (as many corals do), or one growth band every low tide, as mussels do.

Some corals lay down distinct bands of skeletal calcium carbonate on a daily basis and also display seasonal pat-

terns, so that they keep count both of days and of years. In the same way, mussels deposit their growth bands every low tide, but also show variations according to the phase of the moon, so that they keep count both of low tides and of lunar months. The photograph above right shows the clam *Arctica islandica*, a popular species with sclerochronologists: growth bands are visible in the shell.

It is possible to use these growth patterns to date recent shells (and so the **sediments** that contain them) in a manner analogous to **dendrochronology**. However, there is a more interesting way of using this data, which we shall discuss in the remainder of this article.

5.16.2 Tidal braking

The friction of the tides slows down the Earth's rotation: this is known as **tidal braking**. The effect, though small, is measurable by the high-precision clocks used by astronomers, and so can be established directly as well as on theoretical grounds: at present, the effect amounts to a day getting shorter by 2.3 milliseconds over the course of a century (see [here](#) for more details). This may not sound like much, but it adds up: over the course of 100 million years, that would add up to a change in the length of a day of 38 minutes.

This means that in the past days must have been shorter. As the length of a year is constant, this means that in the past there must have been more days per year: using the present rate of slowing as a basis, there would have been about ten more days per year a 100 million years ago.

We should note that in fact scientists do not simply extrapolate the present rate of slowing in a linear manner to calculate past rates of rotation, but rather calculate this from the physics of the Earth-Moon system. For the purposes of this article, it is not necessary to go into the details of the calculation.

5.16.3 Tidal braking and sclerochronology

This immediately suggests a way of dating corals and shellfish. Take corals as an example. As I have said, they lay down daily bands, and the way in which they do so displays seasonal fluctuations. This means that by counting the number of daily bands per year, we can find out how many days there were per year at the time when they were formed. In the same way, by looking at mussels we can find out how many low tides there were per lunar month when they grew.

So by calculating how tidal braking has changed the number of days in a year or a lunar month, we can put a date on the organisms: for example, a coral showing 375 daily growth bands per year must have grown around 100 million years ago.

5.16.4 Weaknesses of the method

The number of days per year or per lunar month changes so slowly over time that we cannot expect sclerochronology to be as precise as radiometric methods such as $U-Pb$. If a change of one day per year corresponds to the passage of 10 million years, then this limits the precision with which we can resolve the age of a shell or coral.

What is more, the change in day length is not as predictable as the decay of radioactive isotopes. Although there is a general tendency for the Earth to slow down, occasionally it has sped up slightly, as we know from direct measurement.

Over the longer term, the magnitude of tidal braking will depend on the exact interaction of the Earth and Moon. Such things as the position of the continents and of mid-ocean ridges will affect tidal patterns, and these change over time as we have seen in our discussion of plate tectonics. Then again, the formation of polar ice-caps, and the concomitant fall of sea-levels would speed up the Earth's rotation as a consequence of the law of conservation of angular momentum.

Because of these considerations, geologists prefer to use radiometric methods rather than sclerochronology where it is possible to do so, even though radiometric dating is rather more expensive. However this sclerochronological technique, even if it is rarely used in practice, has a distinct theoretical significance: it acts as an check on the validity of radiometric methods. When we find approximately 400 days per year in the Devonian period and about 390 in the Carboniferous (see J. Wells (1963) *Coral Growth and Geochronometry*, *Nature* 197(4871), 948-950) then this is in line with the dates put on these periods by radiometric methods. Since the mechanisms of coral and shell growth are completely unrelated to the process of radioactive decay, this provides a completely independent check on radiometric dating. The agreement between sclerochronology and radiometric dating is therefore a good reason to have confidence in both.

5.17 Tidal rhythmites and dating

In this article I shall explain what tidal rhythmites are, how they are formed, and their implications for dating.

As this article re-uses ideas introduced in the previous article on sclerochronology, the reader will need to have read that article first.

5.17.1 Rhythms and rhythmites

Rhythmites are sedimentary rocks which display a repetitive vertical succession of types of sediment. We have already discussed varves, which are a kind of rhythmite. In this article we shall be interested in rhythmites produced

by the action of the tide.

Tidal cycles include:

- A semidiurnal cycle. In most locations the moon produces two high tides and two low tides a day.
- A diurnal cycle. Some locations, such as the Gulf of Mexico, get only one high tide and one low tide per day.
- A mixed cycle. This can be seen in many locations on the west coast of America. In a mixed cycle, there are two high tides and two low tides per day, but one high tide is higher than the other, and one low tide is lower than the other.
- A fortnightly cycle. The highest high tides and lowest low tides (spring tides) occur at new moons and full moons, when the moon is either in line with or opposite the sun.
- Tidal cycles such as the apsidal cycle (presently lasting 8.85 years) and the nodal cycle (18.6 years) can also be distinguished; for the purposes of this article we may overlook them.

In addition to these tidal cycles, the rhythm of the seasons can also have their effect on sedimentary deposition. Varves are a special case of this, though typically varves are so thin that cycles of shorter duration are not discernible. Some sediments, however, will display a full range of cycles from semidiurnal to annual.

5.17.2 Rhythmites and dating

Any or all of the cycles mentioned above can be recorded in **nearshore sediments**. So it is possible to look at nearshore sedimentary rocks and, depending on which rhythms are recorded in the rock, to find out how many days there were in a month, or days in a year, or months in a year, or all of these facts, at the time when the rhythmite was deposited.

This allows us to subject these rhythmites to the same analysis as is used in sclerochronology. The fact that there is close agreement between the number of days in a year as calculated on the basis of rhythmites and by the use of sclerochronology is a reason to have confidence in both methods, since it is hard to see how both could be wrong and yet coincidentally in agreement.

The same caveats apply to the use of this type of rhythmite for dating as apply to sclerochronology, and for just the same reasons. Also as with sclerochronology, the agreement of data from rhythmites with dates produced by radiometric dating is a reason to have confidence in radiometric methods.

5.18 Fossils and absolute dating



Sandstone with fossil shells.

In this article we shall discuss how fossils can be used for the purposes of absolute dating.

5.18.1 Fossils and dating

We have already discussed the construction of the geological column. If our stratigraphic methods show that fossil A was always deposited below fossil B whenever we are in a position to compare their dates of deposition, then we can conclude that species A is older than species B. We can apply the same sort of reasoning to the stratigraphic relationships of fossils and datable rocks.

For example, suppose that using stratigraphic methods, we can show that a particular fossil is always older than rocks which are 14 million years old or less, and always younger than rocks which are 16 million years old or more, whenever we are in a position to make a comparison.

Now, it is a fundamental principle of science — arguably, the *only* fundamental principle of science — that a rule that works every time we can test it must be taken as true unless and until we find a counterexample. So in this case we would have to conclude that this fossil species is between 14 and 16 million years old wherever we find it, even in those cases where there are no datable rocks that we can compare it to.

But this means that we can now use the fossil species to date the sedimentary rocks in which it is found; and we can say that those fossils found in the same strata as this species must be the same age; those species which stratigraphy tells us are older than it is must be more than 16 million years old; and those species which stratigraphy tells us are younger than it is must be less than 14 million years old.

Hence we can use datable rocks to put dates on fossil species; and then we can use the fossil species to put dates on other rocks which would otherwise be difficult to date.

Those fossils we have described as "index fossils" are particularly suitable for this purpose, since they have a wide geographical distribution but only inhabit a thin slice of time.

5.18.2 Advantages of the method

There are three main advantages of using fossils for dating in this manner.

First of all, we may want to date a stratum which is a long way up or down from any rocks we can date using radiometric methods. In this case, the use of fossils will be absolutely the best method available.

Second, it is much faster than any more technical method. Why send a rock to a laboratory and wait for a reply when you can just glance at the fossils it contains and say: "Ah yes, Early Ordovician"?

Third, by the same token, it's much cheaper. Radiometric dating requires specialized equipment: lasers, spectrometers, or in the case of Ar-Ar dating a small nuclear reactor. Even the humblest items of equipment come at a price: laboratories that carry out U-Pb dating wash the bottles they use for two years continuously to eliminate contamination. Rather than employ the services of such a laboratory, it is so much cheaper for the geologist to recognize a well-known species of ammonite, trilobite, foraminiferan, or whatever, the age of which is already known.

5.19 Absolute dating: an overview



In this article, we shall take a look back at the methods of absolute dating, and see how we know that they can be relied on.

5.19.1 Basis of the methods

One argument in favor of the **absolute dating** methods presented in the preceding articles is that they *should* work in principle. If they don't, then it's not just a question of geologists being wrong about geology, but of physicists being wrong about physics and chemists being wrong about chemistry; if the geologists are wrong, entire laws of nature will have to be rewritten. Science, since it concerns just one universe with one set of laws, constitutes a seamless whole; we cannot unpick the single thread of **absolute dating** without the whole thing beginning to unravel.

Still, it has happened in the past that scientists have thought they'd got hold of a law of nature and then found out it was false. There is no particular reason to suspect that this will turn out to be the case when it comes to the laws underlying **absolute dating**; nonetheless, an argument from principle alone can never be entirely convincing. Let us therefore turn to the evidence.

5.19.2 Sea-floor spreading

You will recall from our discussion of sea floor spreading that the sea floor spreads out from **mid-ocean rifts**, and so ought to be younger nearer the rifts and progressively older further away from them.

What is more, we can measure the rate of spreading directly by **GPS**, **SLR**, and **VLBI**. This means that if we didn't have any other way of doing **absolute dating**, we would as a first approximation take the age of **basalt** on a spreading sea floor to be the distance from the rift divided by the rate of spreading.

Now if we estimate the age of the sea floor like that, then we get a good agreement with the dates produced by **radiometric methods**. It is hard to think that this is a coincidence; it is also hard to think of any mechanism that could produce this agreement other than that the rocks are as old as radiometric methods tell us.

5.19.3 Marine sediment

We began our discussion of absolute dating by saying that sedimentation rates could not be relied on for absolute dating. If there is one possible exception to this, it would be the deposition of marine sediment, since it is not subject to erosion, and since we would expect the rates of deposition of the various sediments to be, if not actually constant, then not subject to such a degree of variation as (for example) glacial till. Based on the known rates of deposition, we may therefore at least say that the depths of marine sediment found on the sea floor are *consistent* with the ages of the **igneous rocks** beneath them as produced by radiometric dating.

5.19.4 Radiometric dating and paleomagnetism

The polarity of the Earth's magnetic field is a global phenomenon: at any given time it will either be **normal** everywhere or **reversed** anywhere. So if our methods of **radiometric dating** are correct, then we would predict that rocks dated to the same age would have the same polarity, which they do.

If this does not completely prove that radiometric dating is correct, it does at least show that (barring a wildly improbable coincidence) there is at least a one-to-one relationship between the dates produced by radiometric methods and the true dates, and so it must be taken as an argument in favor of these methods.

5.19.5 Comparison with historical dates

It is possible to test **radiocarbon dating** by using it to put a date on historical artifacts of known date, and to show that it is usually very accurate.

It has also been possible to test **Ar-Ar dating** against the historical record, since it is sufficiently sensitive to date rocks formed since the inception of the historical record. For example, Ar-Ar dating has been used to give an accurate date for the eruption of Vesuvius in 79 A.D. as recorded by Roman historians at the time. (See Lanphere et al., ⁴⁰Ar/³⁹Ar ages of the AD 79 eruption of Vesuvius, Italy, *Bulletin of Volcanology*, 69, 259–263.)

5.19.6 Radiocarbon dating, varves, and dendrochronology

Because **varves** contain organic material, it is possible to compare the dates from varves with the dates produced by **radiocarbon dating**, and see that they are in good agreement. We also see close agreement between **dendrochronology** and uncalibrated radiocarbon dates. (I specify uncalibrated dates because as **radiocarbon dating** is calibrated against **dendrochronology**, the agreement of *calibrated* radiocarbon dates with dendrochronology is inevitable.)

Now, each of these three methods relies on a different underlying physical process: radioactive decay, outwash from **glaciers**, and the growth of trees. We can hardly suppose that there is some *single* mechanism which would interfere with all three of these very different processes in such a way as to leave the dates derived from them still **concordant**.

But it is equally far-fetched to imagine that three different mechanisms interfered with the three processes in such a way as to leave the dates **concordant**; that would require either a preposterous coincidence, or for natural processes to be actually conspiring to deceive us: an idea which is, if anything, even more preposterous.

Now, preposterous things do happen occasionally. But in this case there is a perfectly reasonable and straightforward explanation for why the dates are concordant, namely that they are correct.

5.19.7 Radiometric dating, sclerochronology and rhythmites

Similar remarks may be made about the agreement between radiometric dating of rocks, sclerochronology, and dating by rhythmites.

Are we to believe that one single mechanism interfered with the decay of radioactive isotopes, the secretion of calcium carbonate by molluscs, and the action of the tide? Absurd. But are we instead to believe that three separate mechanisms interfered with these processes in such a way as to leave all the dates **concordant**? That would be equally absurd. The straightforward explanation for the concordance of the dates is that they are in fact correct.

Consider the following analogy: a clockmaker sells us an electric clock, a pendulum clock, and a spring-driven clock, and guarantees that they are shockproof. Skeptical of the clockmaker's claim, we subject the clocks to shock: we shake them, drop them, hit them with hammers and shoot them out of a cannon. Throughout this process, they all go on showing exactly the same time. Is it plausible that we have damaged their very different internal mechanisms in such a way that they are all running fast or slow but still in perfect synchrony? Or is it more likely that they are synchronized because nothing that's happened to them has affected their working?

5.19.8 Agreement with relative dating

Relative dating by definition does not produce actual dates, but it does allow us to put an order on the rocks, and so if absolute dating is to be trusted, it should agree with this order, telling us, for example, that Ordovician rocks are older than Triassic rocks; and it does.

It is hard to see this as a coincidence; it is equally hard to think of some alternate explanation of why we can correlate isotope ratios or sclerochronological data with the relative order of rocks as deduced from stratigraphic methods — other than the straightforward explanation that absolute dating is producing the right dates.

5.19.9 Internal consistency of radiometric dates

In our discussion of radiometric dating, we have seen that many, indeed most, radiometric methods are self-checking.

So in the U-Pb method, we check that the two uranium isotopes produce concordant dates. In the Ar-Ar method,

we check that step heating yields the same date at every step. In Rb-Sr, Sm-Nd, Lu-Hf, Re-Os, La-Be, La-Ce and K-Ca dating, we check that the points we plot on the isochron diagram lie on a straight line.

These precautions allow us to throw out most data that have been produced by confounding factors such as atmospheric contamination, weathering, hydrothermal events, metamorphism, metasomatism, etc.

It is, as we have explained, possible for the occasional incorrect date to slip through this filter, since it is possible for some of these confounding factors to accidentally change the **isotope** ratios in such a way as to produce something that looks like a good date: apparently concordant dates for Ar-Ar or U-Pb, or a false isochron for the various **isochron** methods.

It would indeed be remarkable if this never happened, since one-in-a-thousand chances do in fact occur one time in a thousand. But by the same token, the other 999 times they don't, and so although any particular date produced by these methods might be called into question, it must be the case that the vast majority of dates that pass through these filters must be good; for we can hardly suppose that the confounding factors are actively conspiring to deceive us, and so these long-shot events must be as rare as statistical considerations would lead us to expect.

5.19.10 Mutual consistency of radiometric dates

You might perhaps suggest that if some unknown factor, contrary to our present understanding of physics existed that sped up or slowed down radioactive decay in the past, then we would expect the radiometric dates to be concordant whether they were right or wrong.

This is, as I say, contrary to our present understanding of physics, and so is mere unfounded speculation. What is more, the reader should recollect that "radioactive decay" is not the name of one process; it is the name of *any* process that rearranges the nucleus. So to leave dates produced by different radiometric methods still concordant, nature would somehow have to conspire to fool us by changing the rates of **alpha** decay, of **beta** decay, and of **electron capture**, in such a way that the different dating methods based on these different modes of decay come up with the same dates.

Another point to bear in mind is that a change in the rate of radioactive decay, even if it was carefully coordinated in this way, would still not change every radiometric date in the same direction: if, for example, radioactive decay sped up at some time in the past then this would make **U-Pb** or **Ar-Ar** dates older than they should be, but it would make the dates produced by cosmogenic surface dating **younger** than they should be.

5.19.11 Summary

It is possible to doubt any particular date obtained by **absolute dating** methods. But it would be bizarre to doubt the general picture they paint. For what we see is a massive agreement between the different **radiometric** methods, **varves**, **dendrochronology**, **sclerochronology**, **rhythmites**, **paleomagnetic** data, **deposition rates**, **sea floor spreading**, and **relative dating** methods.

For the dates obtained by **absolute dating** to be wrong in general and yet wrong in such a way as to be in agreement with one another and with other observations, we would have to suppose either that we are looking at an inconceivably massive coincidence, or that the whole Earth is a fraud designed to deceive us.

Ideas to the latter effect have actually been proposed from time to time; most notably by the nineteenth century religious zealot Philip Gosse, whose eccentric work *Omphalos* proposed that the Earth was a mere few thousand years old, but that God had created it to look much older. To this the Reverend Charles Kingsley memorably answered: “I cannot believe that God has written on the rocks one enormous and superfluous lie for all mankind”. That of course would be a theological rather than a geological question, and so is outside the scope of this textbook. What can be said is that geology is a science, and that in science it is necessary to proceed on the basis that the universe is not a lie; because if we believed that, we could believe that anything at all was the case and disregard all evidence to the contrary. The scientific method compels us, then, to disregard the possibility of divine malice; and mere natural processes, being mindless, cannot be actually malevolent.

What, then, of coincidence? Well, there are limits to the degree of coincidence we can believe in, otherwise again we could believe nearly anything. The scientific method requires us to discard such remote possibilities unless there is at least a hint of a shred of evidence for them.

We are left with the conclusion that the great majority of the dates produced by **absolute dating** methods must be reasonably accurate.

Chapter 6

Paleoclimatology

6.1 Paleoclimatology: introduction



Hurricane Frances, as seen from the International Space Station.

In this article we shall take a brief overview of what paleoclimatology is before getting down to specifics.

6.1.1 What is paleoclimatology?

As the name suggests, **paleoclimatology** is the study of ancient climates. **Climate** can be defined as broad trends in the weather: the fact that it rained in Seattle yesterday is merely weather; the fact that it often rains in Seattle is climate.

By definition, paleoclimatology must include the whole of climatology, *plus* some techniques for finding out about past conditions. Now, climatology is a complicated science, because the climate is composed of many parts which interact with one another in complex ways. A full introduction to the subject would have to deal with the influence and relations of such factors as the salinity of the ocean, the biological productivity of the land, the composition of the atmosphere, the effects of chemical weathering on climate, the effects of climate on chemical weathering, the deposition of coal, the short and long term effects of volcanoes, variations in the Earth's orbit, and dozens of other considerations. To explore all these factors in the detail they really deserve would so unbalance this textbook as to make it require a new title along the lines of *Historical Geology Plus Everything You Ever Wanted To Know About The Climate*.

Instead, we shall focus more on methods for finding out what the climate was like in the past than on understanding the underlying principles that drive the climate.

6.1.2 Methods: measurement

The gold standard of measurement would be **instrumental** measurement, such as is performed with a thermometer, a barometer, a rain gauge, etc. However, such measurements have only been made in the recent past, and obviously geologists have to look elsewhere for paleoclimatic data: they have to look at the sedimentary and fossil records.

In paleoclimatology, the relationship between the geological record and the facts is not so straightforward as in other areas of geology: there is nothing that stands in the same simple relationship to (for example) ancient temperatures as aeolian sand does to ancient ergs, or fossils do to ancient organisms. Instead, geologists have found it necessary to develop proxies. A **proxy** might be defined as something we can measure which is not the thing we'd actually like to measure, but which bears a known relationship to it. A proxy can be justified in various ways: by testing that it presently indicates what we think it indicates, by observation of the natural world or of laboratory experiments; on theoretical grounds that argue that the proxy ought to indicate what we think it indicates; by comparison to proxies in which we already feel confident; or any combination of these considerations. We shall look at a selection of these proxies in subsequent articles.

6.1.3 Methods: models

As we have noted above, the climate is complicated, climatic models are complicated, and to keep from doubling the length of this book we shall have to treat climate models more or less as black boxes containing the accumulated knowledge of climatologists.

That said, the construction of climate models is of interest to us: we can use models to figure out what the climate *should* have been like in the past, and then compare this with the evidence we have for what the climate actually

was like. If the two match up, then this increases our confidence in both our proxies and our models.

This also gives us a check on our reconstruction of past geological events. For example, if geologists tell us that there used to be a mountain range along the coast of a continent, and if climate models tell us that the mountain range should have created a **rain shadow desert** on its **leeward** side, then if we find evidence of a **desert** in the right place with the right date, this increases our confidence in the models *and* in the geologists' claim that there used to be a mountain range where they say it was. In this way paleoclimatology can not only draw evidence from the other fields in historical geology, but can also contribute evidence to them.

6.2 Sediments and climate



Clashach Cove, Scotland. Analysis of the exposed yellow sandstone reveals that this was once a hot arid sandy desert.

In this article we shall look at how various kinds of **sediment** are characteristic of the **climates** in which they are formed, so that the **lithified** counterparts of these sediments can be used as an indicator of climatic conditions in the past.

The reader may find it useful to look back at the main articles on **deserts**, **glaciers**, **paleosols**, **oids**, and **coal** to put the remarks below into their proper perspective.

6.2.1 Sediments and paleoclimates

There are a number of types of **sediment** which are symptomatic of the climatic conditions under which they formed.

- **Glacial** sedimentation and **erosion**. As was discussed in the article on **glaciers**, we can detect their former presence by such clues as **striations**, **glacial till**, and **dropstones**. Obviously the former presence of **glaciers** indicates a former cold climate.

- Similarly the presence of **aeolian sand**, with its distinctive large-scale **cross-bedding**, **pinstripe laminae**, etc, indicates an arid and usually a hot climate. The same may be said of features such as **playa lakes**, indicative of low rainfall and a high rate of evaporation.

- **Redbeds** are sedimentary rocks **cemented** together chiefly by **iron oxides**; these are characteristic of a dry climate.
- **Ooids** are formed only in warm, shallow, agitated water, and so are indicators of a warm climate.
- **Coal** requires a **peat swamp**, and therefore cannot form in an arid climate.
- The **soil** types known as **laterites** are produced in tropical conditions with seasonal alternation between a monsoon season and a dry season, and so the corresponding **paleosols** indicate such conditions where they are found in the geological record.

6.2.2 How do we know?

For information about how we know that glacial till is deposited by **glaciers**, or that coal is **lithified** peat, and so forth, the reader should refer to the main articles on those topics.

Given that knowledge, the inference from the sedimentary rocks to the **climate** in which they were formed is a fairly obvious one. It is difficult, after all, to suppose that in times gone past **glaciers** flowed in conditions of sweltering heat, and it is a downright contradiction in terms to think of a dry swamp. Similar remarks apply to the other sediment types listed above; unless the laws of physics or chemistry were significantly different in the past than they are in the present, it is hard to see how **sediments** in the past could be characteristic of different climates than the corresponding sediments in the present.

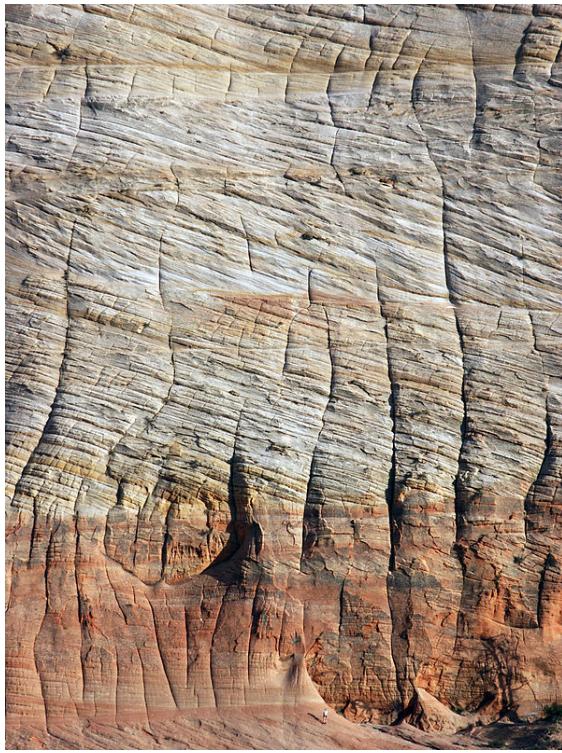
6.3 Paleocurrents

In this article we shall look at how we can use sedimentology to determine the direction of **paleocurrents**: former currents of wind or water.

The reader may find it helpful at this point to go back and look at the main article on **rivers** and on **deserts** to place the following article in its proper perspective.

6.3.1 Cross-bedding

It is possible to determine the past direction of currents of wind or water by examination of sedimentary structures



Cross-bedding in sandstone, Zion National Park, Utah.

in rocks.

In ripples and dunes formed by wind or water, the cross-beds consist of the remnants of beds formed by sediment cascading down the lee face of the dune. Where the *stoss* side of a ripple is preserved, it has a shallower slope than the *lee* side. These considerations allow us to deduce the former direction of currents by studying the arrangement of sediment laid down by them. For example in the photograph to the right we can see that the current (of wind, in this case) that formed the cross-beds was traveling roughly from left to right.

In the case of streams and rivers, the current direction is necessarily downstream. In the case of desert sand and desert winds, you might not suppose that there is a direction associated with the current: the wind is, after all, proverbially capricious; but it is not quite so capricious as a layperson might think. Pretty much any spot on the face of the earth has a *prevailing*, though not constant, direction in which the wind blows: for example in the Sahara the prevailing wind is north-easterly, and this is reflected by the motion of the dunes and the sets of cross-beds that they deposit.

6.3.2 Clast size

On a larger scale, the direction in which a river flows is indicated by the average *clast* size on its bed: there is a tendency for smaller *clasts* to be found further downstream. The reason for this is that if at any point in its course the

river slacks off due to becoming broader and/or deeper, this will diminish its capacity to carry the larger of the clasts that it has carried up to that point, and so they will be deposited. The river may well regain speed after that point, so that it could carry clasts of that size once more; but by that time they have already been deposited further upstream.

6.3.3 How do we know?

We can verify that these relationships between currents, sedimentary structures, and *clast* size hold good in the present.

Now in this case the principle of actualism is if anything even more compelling than usual, because we are dealing here with really basic laws of physics. Is it even conceivable that in the past the laws of fluid dynamics could have been so different that (for example) the *stoss* side of a ripple would be steeper than the *lee*?

The agreement between current direction as inferred from ripples and as inferred from *clast* size is another reason for confidence in these methods.

6.4 Biogeography and climate



Palm tree, Las Vegas, Nevada.

In this article we shall look at how *biogeography* relates to *climate*.

6.4.1 Species and their preferred climates

Pretty much every species, with the notable exception of our own, flourishes within a fairly narrow range of climatic conditions. This means that when we find representatives of that species preserved in the *fossil* record, we should conclude that those conditions were present at the date and location when and where they lived.

We can refine our knowledge of temperatures by looking at the overlap in the ranges of species. If one lives

in regions with an average annual temperature between 20°C and 24°C, and another inhabits regions with an average annual temperature between 22°C and 26°C, then when we find them together we can conclude that they were living in a time and place where the average annual temperature was between 22 and 24°C.

6.4.2 How do we know?

In order for this method to work, we need to know what climatic conditions suited the species we find in the fossil record.

If the species is still extant, we can of course just look at where it lives and what the extremes of temperature are in that location. You might speculate that perhaps without any visible morphological change the species could have evolved to adapt to different climatic conditions, so that the temperature range it lived in then is not the same as that which it inhabits now. While this is possible in principle, what we observe is that species do not evolve to adapt to shifting climatic conditions; rather, they shift their locations to follow the conditions that suit them. (Even plant species will migrate; although an individual plant cannot move, the spread of their seeds allows them to migrate as a species.)

Species, of course, come and go over time; but we can also look at larger groups. For example, since out of about 2,600 species of palm trees we know of none that will grow in a cold climate without human intervention, it is reasonable to conclude that this would be also be true of an extinct species of palm tree.

Still, extinction does present us with a problem. For example, with no modern representatives at all, how are we meant to know a warm-water trilobite from a cold-water trilobite? This difficulty is not quite insurmountable in principle. For example, if we frequently find some species of trilobite in association with *oolids*, then we would be inclined to think that it was a warm-water species, and then if we were to find it in the absence of other climatic indicators, we would take the trilobite species as an indication of a warm climate. That being said, the use of the biogeographic method of assessing paleoclimatic conditions does in practice get less and less useful the further back into the past we wish to look.

climatic indicator. We also noted that this is difficult when the species is extinct, since then we can no longer find by direct observation what climate is congenial to it.

But what if the morphology of a species was an indicator of the climate it prefers? In that case, we would be able to tell what climate it inhabited just by looking at its remains in the **fossil** record. One morphological feature commonly used for this purpose is the leaf shapes of plants.



Serrated



Entire

6.5 Leaf shape and temperature

In this article we shall discuss how the analysis of the shapes of leaves can be used to reconstruct past climates.

6.5.1 Leaf shape

In the previous article, we saw that if we know the environment preferred by a species, we can use it as a cli-

Plants in temperate climates tend to have leaves with **serrated margins**, i.e. they have jagged edges; plants in warmer and more humid climates tend to have what are known in botanical jargon as **entire margins**, that is, smooth and unserrated. The difference is illustrated in the photographs to the right.

Rather than there being a sharp cut-off between the temperate and tropical styles of leaves, there is a continuous relationship between the climate and the mix of leaf types found in it: that is, as the climate gets a little hotter and

wetter, the proportion of entire margins increases a little. This means that looking at a single leaf doesn't tell us that we are looking at a temperate or tropical climate; but looking at a whole lot of species will allow us to do something a whole lot better than simply dividing climates into tropical or temperate: we can actually estimate the average annual temperature.

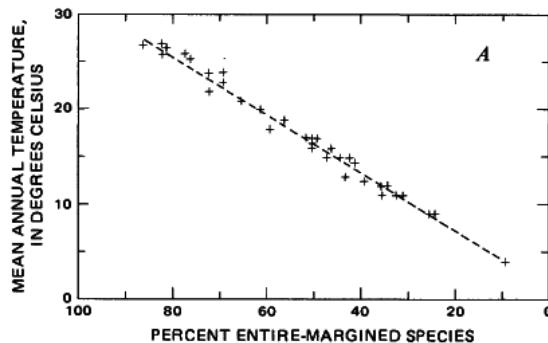
So once we have established the relationship between climate and the proportions of leaf shapes in the flora, we can use this relationship to find out about past climates where leaves have been preserved in the fossil record. Although leaves are not the most easily preserved material, there are some environments, such as lacustrine sediment, where they are preserved and can be used as a paleoclimatic indicator.

6.5.2 Other leaf characteristics

Using leaf shape as a climatic indicator began with counting the proportions of serrated to entire margins, but it didn't end there. Other characteristics such as leaf length, leaf width, leaf area, the presence or absence of a pointed tip also have a relationship to the climate.

The CLAMP (Climate Leaf Analysis Multivariate Program) group has measured many more characteristics of leaves and have used sophisticated methods to produce a formula relating these characteristics to climate; their website can be found [here](#).

6.5.3 How do we know?



Relationship between climate and leaf shape.

We can establish by observation that the ratio of temperate to tropical leaf styles is a surprisingly good indicator of average annual temperature, as illustrated by the graph to the right, showing the relationship between floras and temperature in the forests of East Asia (figure from J. A. Wolfe, 1979, Temperature parameters of humid to mesic forests of eastern Asia and relation to forests of other regions of the Northern Hemisphere and Australasia, USGS Professional Paper, 1106).

This means that if we look at the ratio of leaf types in the fossil record, we can use it to produce an estimate of the temperature at the time when the plants grew.

Now, you might wonder if this relationship between leaf types and temperature really held in the past. Well, the adaptation of fauna and flora to their environment does seem to be one of the great constants of nature, so in principle we would expect the relationship to have held in the past much as it does in the present. It is after all extremely unlikely on the face of it that serrated margins would suit temperate climates in the present and yet have been more suitable to tropical climates in the past.

However, it should be noted that the *exact* relationship between leaf types and climate varies from continent to continent: although more serrated leaves always correspond to cooler climates on every continent, the exact numerical relationship is not the same. So a graph similar to the one above but based on Australian data rather than Asian data would still be a good approximation to a straight line sloping downwards to the right, but the exact angle of slope would be different. Now, what can vary from place to place could also in principle vary from time to time. This means that although leaf shape is a useful indicator of past climate, it would be naïve to treat it as though the relationship between climate and leaf shape was quite so exact as the relationship between temperature and a thermometer.

6.6 Dendroclimatology

In this article we shall discuss the principles behind dendroclimatology. The reader may find it useful to re-read the article on dendrochronology before proceeding further.

6.6.1 How dendroclimatology works

You should recall from the article on dendrochronology that many species of trees produce annual growth rings. The dendrochronological method depends crucially on the fact that these rings will not be of constant width, but will be thicker when the tree in question has had a good year and thinner in bad years.

Now, from the point of view of a tree, the difference between a good year and a bad year is mainly going to be determined by the weather: by the amount of sun and the amount of rainfall in its growing season.

So by using dendrochronology we can put a date on a tree-ring, and by looking at the thickness of the ring we can find out about the climate in the year of its formation. This method is known as **dendroclimatology**.



Another bad year: old fire scars on a pine tree, Sweden.

6.6.2 Difficulties of the method

Some of the difficulties of dendroclimatology are, so to speak, inherited from dendrochronology: for example, a shortage of material limits the scope of dendroclimatology in time. Also, since annual rings don't form in the tropics, the use of dendroclimatology, like dendrochronology, is limited to temperate zones.

One difficulty with dendroclimatology is that although the weather is the most likely cause of a tree having a bad year, it is not the only possible cause: disease, fire, or other external causes can have the same effect. This difficulty is not insuperable: trees affected by such events will be the odd ones out, so analysis of a sufficiently large number of trees from different locations allows us to distinguish the climatic signal from the incidental noise.

Another difficulty is that we would expect both the temperature and the amount of rainfall to affect the growth of trees. We would ideally like a **proxy** for one or the other, whereas when we measure the width of a **growth ring** we are looking at the combined effects of both. One way around this is to study what are known as **limiting stands**. For example, if we look at trees near the snow-line of a mountain, then because storms tend to break over peaks, those trees will get all the water they need; the limiting factor on growth will be temperature, which will therefore determine the width of the growth rings.

6.6.3 How do we know?

We can compare the growth rings in trees with records of temperatures going back in some cases hundreds of years.

Mostly, there is agreement between direct measurement and climatological data. However, there is a discrepancy between the most recent records and recent tree ring growth. As the reader will doubtless be aware, global temperatures have been rising over the last few decades, something we know by consulting actual thermometers; but the dendroclimatological **proxy** has not kept pace with the change in climate. This leaves a question-mark hanging over dendroclimatology. If it doesn't tell us about events which we know have happened, then what isn't it telling us about events that we don't know about?

So the same kind of data that tells us that dendroclimatology is somewhat reliable also tells us that it isn't completely reliable. Together with the other difficulties with dendroclimatology, you might think that climatologists should abandon it and rely on proxies which are based on something less capricious than the growth of trees. However, it does have its advantages: it provides us with data specific to particular locations on land which are not covered by other proxies.

6.7 Scleroclimatology



Cypraecassis rufa. Seashells record climatic information.

In this article we shall look at how the composition of shells of marine organisms is affected by climate, and how we can therefore obtain clues about past climates by chemical and isotopic analysis of their shells. This collection of methods is known as **scleroclimatology**.

6.7.1 Oxygen isotopes

The ratio of the stable oxygen isotopes ^{16}O and ^{18}O in water (H_2O) can be shown to vary with temperature, with a higher $^{16}\text{O}/^{18}\text{O}$ ratio associated with lower temperatures.

Oxygen from the seawater is used by shell-forming marine organisms when they form their calcium carbonate shells, and as the two isotopes of oxygen are chemically identical, no isotope is preferentially involved in this process, with the result that the oxygen ratios in the shells reflect the oxygen ratios in the seawater.

This means, of course, that if we look at shells in the geological record (the tests of foraminiferans being a favorite choice for this) and if the calcium carbonate in the shells hasn't been replaced by some other mineral, and if we can put a date on the shells, then we can find the temperature of the seawater at the time and place where the shells were deposited: this proxy is known as $\delta^{18}\text{O}$. What's more, since it is possible to distinguish between the fossils of **planktonic** and **benthic** species, we get two figures for each location: the surface temperature and the temperature at the sea floor.

6.7.2 Mg/Ca and Sr/Ca

The elements magnesium (Mg) and strontium (Sr) lie in the same column of the periodic table as calcium (Ca) and so possess similar chemical properties. This means that magnesium and strontium can substitute for the calcium in calcium carbonate (CaCO_3). As these elements substitute more readily for calcium at higher temperatures, the proportions of magnesium and strontium substituting for calcium can be used as a temperature proxy.

6.7.3 Difficulties of the method

Obviously for this method to work the mineral composition of the shells must not be changed by fossilization. While this can present us with difficulties in obtaining appropriate material, it will not usually lead us to produce erroneous results; after all, if a shell has undergone mineral replacement so that it's no longer made of calcium carbonate then this is not something a geologist would easily be able to overlook.

One thing that interferes with the oxygen isotope method is that, paradoxically, actual glaciation has the opposite effect on global marine oxygen isotope ratios than mere low temperatures have on local isotope ratios. This is because when the glaciation of the Earth increases, water that evaporates from the seas is locked up in ice sheets; now ^{16}O , being lighter than ^{18}O , evaporates more readily, so that the result of increasing glaciation is a decrease in the $^{16}\text{O}/^{18}\text{O}$ ratio. To make sense of the data, it is necessary to disentangle these local and global effects by reference to other data.

When it comes to the Mg/Ca and Sr/Ca methods, the local mineral composition of the seawater can act as a confounding factor: for example, if the seawater in a particular location is particularly rich in magnesium for some reason, then this will also increase the Mg/Ca ratio.

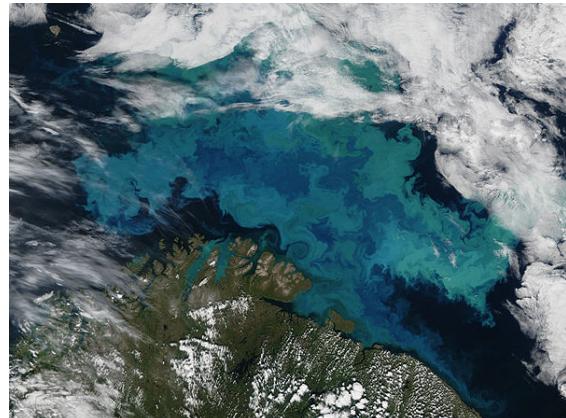
6.7.4 How do we know?

We can take samples of shell-forming organisms from locations with a known temperate and measure their isotopic and chemical ratios, and see how they relate to temperature. It is also possible to grow shellfish in tanks kept artificially at a controlled temperature and see what happens.

It is unlikely that the biochemistry of shell formation has changed significantly since organisms first started forming calcium carbonate shells; and it is almost unthinkable that the physics of the evaporation of water has changed at all. It is reasonable to conclude that we can take chemical and isotopic ratios in the past as proxies for past temperatures.

Finally, we can note that there is a good if not exact correlation between the paleoclimatic data obtained from shells and other paleoclimatic proxies.

6.8 $\text{Uk}'37$



Satellite image of a bloom of *Emiliania huxleyi* in the Barents Sea.

In this article we shall look at alkenones and how they can be used to construct the temperature proxy known by the somewhat cryptic name of $\text{Uk}'37$.

6.8.1 Alkenones

Alkenones are organic molecules having a chain-like structure, as organic molecules so often do. The length of the chain is given by the number of carbon atoms in it, so that we can speak of the alkenones C_{37} , C_{38} , etc. Note that these are not chemical formulas for the alkenones, which contain other atoms besides carbon; they merely record the length of the alkenone.

A further variation in the structure of alkenones is that they can be either doubly or triply unsaturated, a detail depending on the nature of their carbon bonds. So we can speak of the alkenones $\text{C}_{37:2}$ and $\text{C}_{37:3}$.

It is not really necessary for the reader to understand the organic chemistry of alkenones in any detail, or to know what “unsaturated” means; the important thing is that they exist and occur in different varieties.

6.8.2 Alkenones in nature

Only a small number of species are known to produce alkenones, all lying within the group of planktonic organisms known as haptophytes: the species *Gephyrocapsa oceanica*, the genus *Chrysotila*, and most importantly the incredibly numerous cocolithophore *Emiliania huxleyi*.

These alkenones have two useful properties. First of all, they survive conditions that would destroy most organic molecules. Consequently, alkenones have been found in marine sediment as much as 110 million years old. Second, the different varieties of alkenone are produced in different quantities at different temperatures. The relationship between temperature and the proportions of C₃₇ alkenones is given by:

$$T = 29.41 \times U^{k'}_{37} - 1.15$$

where U^{k'}₃₇ is the proportion of C_{37:2} and C_{37:3} which is C_{37:2}, and *T* is the water temperature in °C; specifically, since all the organisms that produce alkenones are planktonic, *T* gives us the *surface* temperature of the water.

The longevity of alkenones in the sediment means that we can use them as a paleoclimatic proxy.

6.8.3 Limitations of the method

There are two main limitations on the method. Firstly, the relevant organisms don't grow in polar waters, and so obviously can't be used to indicate their temperature. Secondly, the formula breaks down for very warm water. For the proportion of C₃₇ which is C_{37:2} can never be more than 100%, which means that the formula can never yield a *T* value greater than 28.3°C no matter what the actual temperature of the water is.

We should also mention a couple of potential confounding factors which have been discovered experimentally (see Prahl, Wolfe & Sparrow, *Physiological impacts on alkenone paleothermometry*, *Paleoceanography*, 18(2)). First, U₃₇^{k'} is reduced by conditions of low nutrition; second, it is increased by conditions of prolonged darkness.

6.8.4 How do we know?

We can measure the relationship between temperature and U^{k'}₃₇ in oceanic plankton; we can also grow plankton in the laboratory at controlled temperatures (see,

for example Prahl and Wakeham, *Calibration of unsaturation patterns in long-chain ketone compositions for palaeotemperature assessment*, *Nature*, 330, 367-369).

Can we be sure that this relationship held good in the past? It seems likely: if there is some reason why C_{37:2} is preferable in warm waters now, then why should it have been different in the past? But we could say this with more confidence if we knew *why* C_{37:2} is preferentially produced in warmer waters. As it is, biologists don't yet know why these haptophytes produce alkenones in the first place. Under these circumstances, perhaps we should be a little more skeptical of this biochemical proxy than of those based on known chemical and physical mechanisms.

6.8.5 Note on terminology

The subscript and superscript in U^{k'}₃₇ can also be written as U₃₇^{k'}, and when superscripts and subscripts are not available, people will write either U^{k'}37 or U37^{k'}. The reader wishing to research the subject further by performing an internet search should be aware of this variation.

6.9 *TEX*86

In this article we shall look at the *TEX*₈₆ temperature proxy, how it works, and how we know that it works.

6.9.1 GDGTs

The *TEX*₈₆ method is based on glycerol dibiphytanyl glycerol tetraethers (GDGTs). These come in various forms with more or fewer cyclopentane structures (the reader need neither know nor care what these actually are). The GDGTs of interest to us can be denoted as GDGT 1, GDGT 2, GDGT 3 and GDGT 4' (pronounced “four-prime”), where the numbers 1, 2, 3 and 4 correspond to the number of cyclopentanes.

(For the benefit of those readers who wish to research the *TEX*₈₆ method in the technical literature, I should point out that different papers use different numbering schemes; the one used here seems most suitable, because of the correspondence between the GDGT number and the number of cyclopentanes.)

6.9.2 Crenarchaeota and temperature

In nature these GDGTs are produced by the group of single-celled organisms known as the Crenarchaeota. As with the alkenones discussed in the previous article, the GDGTs resist processes that destroy most organic compounds, and so can be found in marine sediment; and

just as with the alkenones, the proportions of the different GDGTs produced by the Crenarchaeota varies with temperature, according to the formula:

$$T = 56.2 \times \text{TEX}_{86} - 10.78$$

where T is the temperature in $^{\circ}\text{C}$ and TEX_{86} (an abbreviation of “TetraEther indeX of tetraethers consisting of 86 carbon atoms”) is defined as the ratio of the sum of the quantities of GDGTs 2, 3 and 4' to the sum of the quantities of GDGTs 1, 2, 3 and 4'.

It should be noted that this relationship ceases to hold below about 5°C ; below this temperature the variation in TEX_{86} becomes negligible and so measurements of TEX_{86} can't distinguish between temperatures below that point.

6.9.3 How do we know?

We can measure TEX_{86} in living organisms and recent sediments, and measure the temperature of the water in which they are found; this is how the formula given above was derived.

However, unlike the U^{37}K method, the relationship is harder to demonstrate experimentally. Experiments *do* show that TEX_{86} increases with temperature in the lab; however for some unknown reason lab-grown cultures of Crenarchaeota produce less GDGT 4' than is found in Crenarchaeota in the wild, and so the exact relationship between temperature and TEX_{86} can't yet be replicated in the laboratory.

6.10 Ice cores



Annual layers in a Norwegian glacier.

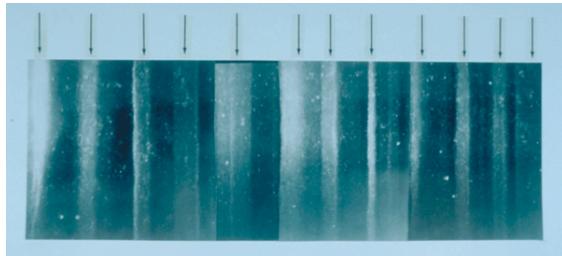
In this article we shall look at how core samples from ice can be used to give us information about paleoclimates. The reader may find it useful to look back at the main article on [glaciers](#) before reading further.

6.10.1 Ice layers

At any location where snow falls but does not melt (or at least does not completely melt) before the next year's snowfall, the snow will accumulate, and as each year's snow is buried by the further snowfall of succeeding years it compacts from loose snow to permeable **firn** to impermeable ice, at which point it is said to undergo **closure**. You should remember from the article on [glaciers](#) that any point at which snow accumulates like this must inevitably become the **accumulation zone** of a glacier.

Ice formed from summer snow is lighter and less dense than ice formed from winter snow; as a result, if the rate of accumulation is more than about 4 cm/year we get what in effect are **varves** formed from snow. At least in the upper part of the [glacier](#), these are plainly visible if we take a core sample, as you can see in the photograph below right.

6.10.2 Ice core data



A core sample from a glacier.

The fact that the ice cores contain visible annual layers means that just as with varves in lakes, we can count them down from the top, and assign a year to each one.

Looking at the thickness of each layer, we can quantify the amount of snowfall, or at least unmelted snowfall, in the summer and winter of each year. What is perhaps more interesting, we can measure the $^{16}\text{O}/^{18}\text{O}$ ratio of the water, which acts as a climatic indicator for reasons discussed in the article on [scleroclimatology](#); we can also measure the $^{1}\text{H}/^{2}\text{H}$ ratio of the water, which acts as a climatic indicator for the same reason: water **molecules** containing the ^{2}H **isotope** are slightly heavier, and so evaporate less readily.

Besides these data, when the permeable firn turns to impermeable ice, bubbles of the atmosphere become trapped within the ice. This allows us to analyze the past composition of the atmosphere, and quantify gasses which affect the **climate**, such as carbon dioxide (CO_2) methane (CH_4) and sulfur dioxide (SO_2).

Also, as with ordinary sediment, the ice cores will contain windborne particles including **volcanic ash**, **pollen**, and **loess** (dust produced by the action of [glaciers](#)).

6.10.3 How do we know?

We can see the annual formation of layers in ice, and we also understand the mechanism behind this in terms of seasonal variation. For the layers not to form at all in a given year, there would have to be a complete cessation of snowfall, which is very unlikely; for more than one layer to form per year we would need more than one warm period per year, which is almost impossible.

We can also verify that the data preserved in the layers reflects the climate by comparing the ice core record with direct measurements of climate made over the last few centuries, and with other climatic proxies.

To this we can add that the methods outlined in this article *should* work; they are based, after all, on very simple physical principles. Is it even conceivable that in times gone by water composed of heavy isotopes evaporated more easily than water composed of light isotopes? Or that once upon a time when firn underwent closure it trapped bubbles of something other than the atmosphere?

So the analysis of ice core data ought to work in principle, and comparison of the results with other data confirms that it does work in practice.

6.10.4 Difficulties with the method

Because firn doesn't turn to ice immediately, and because the air bubbles aren't trapped until it does, the year we calculate for the age of a layer will be different from the age of the atmosphere trapped in that layer. The difference (**closure time**) between the age of the gas and the age of the ice can be as much as 7,000 years, as is the case in ice cores from **Vostok**, or as little as 30 years at **Law Dome**.

Mathematical methods can be used to recover data on a finer scale than the closure time; however, such methods rely on knowing the closure time. Now, the climate has changed in the past (otherwise paleoclimatology would be a completely unnecessary science) and so it seems likely that closure time will have varied from time to time in the past just as it varies from place to place in the present. This introduces an element of uncertainty into the data.

Another problem is that at greater depths the annual layers become more and more indistinct, until in some cases they can't be made out at all. One can try in such cases to estimate the age of the buried ice by measuring its depth and the accumulation at the location in question, but again this would only work perfectly if the climate, or at least the snowfall, had remained constant each year. Where the ice cores contain **volcanic ash**, this can be dated by **radiometric methods**, allowing us to put correct dates on the layers in which they are found, but only up to the accuracy of the radiometric methods used.

There is one final weakness inherent in the method. By definition, any point where the annual rate of snowfall

exceeds the rate it melts must be the **accumulation zone** of a glacier; in the case of the Greenland and Antarctic ice sheets that most interest paleoclimatologists, the accumulation zones of continental glaciers. And a glacier must flow out from its zone or zones of accumulation to a zone or zones where it **ablates**. In short, the record in the ice is progressively being destroyed as well as created. This happens relatively fast by geological standards, providing us with a record that can be measured in hundreds of thousands of years; compare this with **proxies** based on **marine sediment**, which is only destroyed by the much slower process of **subduction**.

However, ice cores remain valuable because the ice does trap air in its pore spaces, providing us with a continuous record of the composition of the atmosphere.

6.11 Milankovitch cycles

In this article we shall discuss what Milankovitch cycles are, their effect on climate, and how we know that this effect exists.

6.11.1 The cycles

The reader should remember from high school that the reason why the Earth has seasons is that its axis of rotation is at an angle to the plane of its orbit. When the northern hemisphere is tilted towards the sun, then it is summer in the northern hemisphere and winter in the southern hemisphere; when the southern hemisphere is tilted towards the sun, it's the other way round.

The magnitude of this effect depends on how tilted the Earth's axis is, and this angle varies between 22.1° and 24.5° in a 41,000 year cycle.

The reader should also recall that the Earth's orbit is not perfectly circular: it is an ellipse with the Sun at one focus, meaning that the Earth is closer to the Sun during some months of the year than others. The effect of this is less than you might suppose: the Earth is five million kilometers closer to the Sun in January than in July, but this doesn't stop the Northern Hemisphere from undergoing winter.

The magnitude of this effect depends on how far the Earth's orbit deviates from being circular, and a number of factors affecting this figure add up to a cycle of about 100,000 years.

Finally, there is the precession of the Earth's axis. *At present*, as we have seen, the North Pole is tilted away from the Sun at the Earth's point of closest approach to the Sun. However, this too varies, in this case in a 21,000 year cycle.

These, then, are the three **Milankovitch cycles**; as they are of different lengths, their interaction will produce

rather a complex pattern as they go in and out of phase with one another. Together, they will affect both the total annual **insolation** (the amount of solar radiation that reaches the earth's surface) and also season variations in insolation.

Although the idea of Milankovitch cycles as a factor in the Earth's climate was initially greeted with some suspicion by climate scientists, it is now generally accepted that Milankovitch cycles account for about 60% of past variation in climate.

6.11.2 How do we know?

First of all, how do we know that Milankovitch cycles exist? A short answer is that the physics of the Solar System require that they should exist; a longer answer would require an introduction to celestial dynamics which would be excessive in length and out of place in what is supposed to be an introduction to historical geology.



Rhythmites caused by Milankovitch cycles.

From a geological perspective, we can look for the effects of the cycles on the sediments and **proxies** used in paleoclimatology. As Milankovitch cycles can't be the only thing affecting the climate, this is not so simple as demonstrating that the climate fluctuates perfectly in synchrony with the cycles; rather, statistical analysis is necessary to sort out the "signal" of the cycles from the "noise" produced by (for example) variations in volcanic activity. Such an approach confirms that the cycles have a real effect on the climate.

One interesting effect of the cycles is that in places where the nature of sedimentation is sensitive to the climate, we can see **rhythmites** with a period dictated by the lengths of Milankovitch cycles, as shown in the photograph to the right.

There remains one outstanding puzzle. In principle, the 100,000-year cycle should have less of an effect than the 41,000-year cycle. But for the last million years or so, the 100,000-year cycle has predominated; whereas prior to that the 41,000-year cycle was indeed more important,

in line with theory. This is known as the **100,000 year problem**, and serves as a useful reminder that our understanding of long-term climatic change is still imperfect.

6.12 Climate models

In this article we shall look at how **climate** models are constructed, their strengths and weaknesses, and how they can be applied to the study of plate tectonics.

6.12.1 Climate and climate models

First we should look at what factors should be represented in a model of the climate.

First of all, there is the **insolation**. This drives the climate, but it doesn't exclusively determine it, otherwise the weather at any point could be expressed as a function of time and latitude.

But in fact there is also the atmospheric circulation to be taken into account. This transports atmospheric heat and moisture from place to place. The atmospheric circulation is partly caused by the Coriolis Effect (fundamentally, because the Earth is rotating). However, the circulation is also caused by variations in the density of the atmosphere, which are caused by variations in the temperature and amount of moisture in the atmosphere; or, to put it another way, one of the main causes of the weather is the weather. It is this fact that makes modeling the weather or the climate particularly difficult; such systems are notoriously hard to model.

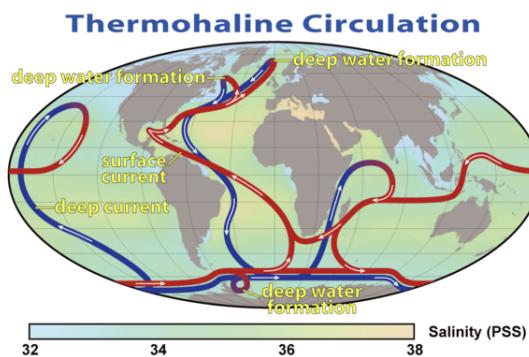
One important factor affecting the nature of the circulation is the location and nature of the landmasses. The ocean will absorb more heat than the land; and then the amount of heat absorbed by the land will depend on the nature of the ground cover, i.e. whether the land in question is desert, forest, covered by a sheet glacier, etc. Similar remarks can be made about moisture; obviously more water will evaporate from a sea or a lake than from a desert with the same **insolation**.

By modeling **insolation** and the circulation of the atmosphere, climatologists can produce what is known as an **atmospheric circulation model**, or **ACM**.

Such a model is the simplest that's any use at all, but it still doesn't tell the whole story. The ocean has its own circulation, and this also transports heat in tandem with the atmospheric circulation. A really good model should take this into account.

Surface ocean currents can be driven by the wind. Deeper in the oceans, we have the **thermohaline circulation**. As the name suggests, or would suggest if we were Greek, this circulation is driven by the temperature and salinity of the oceans, both of which affect the density of seawater. The differences in density drive vast currents, extending

thousands of miles and carrying 100 times more water than the Amazon River.



The thermohaline circulation.

The map to the right shows the thermohaline circulation; as you can see, it can be quite complex, with less dense currents actually flowing over denser currents flowing beneath them in a different direction.

The other thing that affects the ocean circulation is, of course, the positions of the continents, which constrain the flow.

By adding the ocean circulation to an ACM, climatologists can produce a **General Circulation Model**, or **GCM**.

Even then, there are things which we would like to add to a really good model of the climate. For example, we have noted above that ground cover affects the climate. But it is also the case that the climate will to a large extent determine the ground cover from location to location, causing forests here, and grasslands there; **deserts** in one place and **sheet glaciers** in another. Once more we have a case of the climate causing the things that cause the climate, and so we have another complicated set of interactions to model — if we can.

6.12.2 Accuracy of the models

The size and complexity of the climate are a challenge even to modern supercomputers, and limited processing power (in effect, how fast the computer can perform calculations) limits the accuracy and detail attainable in climate models. The most obvious problem is that of scale. No computer can be expected to simulate the behavior of each molecule of air in the atmosphere. Suppose, then, that as a coarser approximation we divided the Earth up into “cells” each one kilometer square, and used a model that assigned to each one at each step of the computer simulation a figure representing the average temperature and the average humidity of the atmosphere inside that “cell”. Then our model would be inaccurate by reason of this approximation; it would also be inaccurate by ignoring the oceanic circulation, and it would *still* involve simulating the interaction of half-a-billion cells; and so even such a degree of approximation might well leave

our model intolerably slow. Of course, scientists can always spend more time running slower models, but they rightly suppose that an imperfect model which returns results within the lifetimes of the researchers is, despite its acknowledged flaws, superior to one which delivers more accurate answers to their great-grandchildren.

Any model of the climate must therefore involve some degree of simplification and approximation; and any such model will therefore be wrong to a certain extent. It is for this reason that there are so many climate models: the researchers must choose in what way to simplify their models in order to reduce the degree of error, and it is not yet clear how best this may be achieved. As a consequence of this, the first report of the Paleoclimate Modeling Intercomparison Project (hereafter referred to as **PMIP1**) lists 22 climate models produced by 19 different research institutions, each in their own way striving after accuracy.

It is possible to find out how good or bad the various models are, and in which respects, by comparing their results to data. We can of course compare the climate now (as measured instrumentally) with the climate as the models say it should be; any model not producing reasonably good agreement has fallen at the first hurdle. But we can also get the models to simulate events in the past and compare them with proxy data. This was done by the PMIP, comparing models with proxies for the mid-Holocene (6,000 years ago) and the last glacial maximum (**LGM**, 21,000 years ago). At these dates, the climate was markedly different from the present; yet they are recent enough that the full array of **proxies** can be brought to bear on what the climate was actually like.

The results are such as to be encouraging to an optimist and disappointing to a pessimist. To briefly summarize the results, the models tend to be *qualitatively* correct: they correctly indicate the *nature* of the differences between the present climate and the climates of the mid-Holocene and the LGM; on the other hand, they are *quantitatively* inaccurate, tending to underestimate the *magnitude* of the differences between then and now.

It should be added that in the first phase of PMIP the research was confined to ACMs which did not take the oceanic circulation into account; most models also omitted the interaction between climate and ground cover. One would expect the results to be more accurate when more factors are taken into account; more complex models which do so will be reported on in the forthcoming second phase of PMIP.

6.12.3 Paleoclimatology in deep time

If we want to look back millions rather than thousands of years, then there is another factor we need to take into account: the fact that the continents have altered their positions. As their location affects both the atmospheric and oceanic circulation, modeling paleoclimates in deep time requires a reconstruction of their positions.

This leads to an interesting line of research. When we reconstruct the position of the continents at some past date, and use climate models to tell us what, in theory, the climate should have been like, does this agree with the sedimentary evidence of what the climate was like at that time?

The answer is “yes”. There are a number of results obtained from climate models which are consistent with **proxies** and **sedimentary** indicators of climatic conditions. For example:

- Continents at higher latitudes are, other things equal, colder than those nearer the equator. (This is rather obvious, but it is nice to see it confirmed by, for example, evidence of continental glaciers rolling over Africa at a time when paleomagnetic data tell us it was further south.)
- Changes in the average temperature of the Earth have more effect near the poles than the equator.
- Continental **glaciation** in Antarctica could only begin when the other southern continents were sufficiently separated from Antarctica to allow the existence of the Antarctic Circumpolar Current. (For further details see Katz et al. (2011) **Impact of Antarctic Circumpolar Current Development on Late Paleogene Ocean Structure**, *Science*, 332(6033); also Bijl et al (2009) **Early Palaeogene temperature evolution of the southwest Pacific Ocean**, *Nature*, 461.) These models are borne out by measurements of Mg/Ca, $\delta^{18}\text{O}$, TEX_{86} and U^{237} .
- Polar ice-caps will not form unless there are landmasses sufficiently near the poles.
- The interiors of **supercontinents** should be dry.
- The rise of the Himalayas is implicated in a significant strengthening of the Asian monsoons. See for example Valdiya (1999) **Rising Himalaya: Advent and intensification of monsoon**, *Current Science*, 76(4); also Clift et al (2008) **Correlation of Himalayan exhumation rates and Asian monsoon intensity**, *Nature Geoscience*, 1.

Now this sort of agreement should increase our confidence in our reconstruction of **continental drift**, in climate models, and in the use of sediments to reconstruct climates; for if any one of these three techniques was no good, then there would be no reason for the agreement between the models and the evidence.

6.13 Ice ages

In this article we shall look at what an ice age is, what causes ice ages, and how we can identify the evidence for



We are living in an ice age: Webb island, Antarctica.

ice ages in the geological record.

The reader will probably find it useful to look back at the main article on **glaciers** before reading further.

6.13.1 Definition of an ice age

An **ice age**, or to use the more technical term, a **glaciation**, may be defined as a time when **continental glaciers** are present. This, of course, means that we are currently experiencing an ice age, since there are continental glaciers on Antarctica and Greenland. To be more precise, we are living in an **interglacial**: a warmer interval within an ice age when the glaciers have retreated towards the poles.

6.13.2 Glaciations in the geological record

Geologists have identified and dated a number of ice ages in the geological record:

- The Huronian glaciation (or Makganyene glaciation) extended from 2400 to 2100 million years ago
- The Cryogenian glaciation lasted from 850 to 635 million years ago.
- The Andean-Saharan glaciation was from 460 to 430 million years ago.
- The Karoo Ice Age lasted from 360–260 million years ago.
- The Pliocene-Quaternary ice age is the one now in progress, and started about 2.6 million years ago.

The methods of dating events have already been explained in other articles; later in this article we shall discuss how we identify ice ages.

6.13.3 Causes of ice ages

We have quite a good idea of the causes of the present ice age, which can be explained by continental drift affecting the oceanic circulation. The causes of previous ice ages are more obscure and debatable.

This is because causality as such is not preserved in the geological record. We may be able to see that event A happened and was followed by event B, but the fact (if it is a fact) that event A caused event B must be established on theoretical grounds, by understanding the relationship between the events of the type in question. For example, it is not hard to deduce a causal relationship between the feet of dinosaurs and dinosaur footprints, because we have a good general understanding of the relationship between feet and footprints. But our understanding of climatic effects is hardly as secure.

A further point to consider is this: ice ages are rare and irregular in the geological record. This suggests that there may well not be a single cause for ice ages. If there was a *constant* reason why ice ages happen, then they would be a permanent feature of the geological record, or at least occur as cyclic events; the fact that they are sporadic suggests that they had singular causes. The best we can do in investigating the causes of an ice age is to look at the events leading up to it and see which of these events might, in principle, be the cause of that particular ice age.

Consider, for example, the Huronian ice age. Two explanations have been proposed for this.

(1) There is evidence (as was discussed in the article on banded iron formations) that oxygen levels rose dramatically in the time leading up to the Huronian. Abundant free oxygen would have combined with the **methane** (CH_4) in the atmosphere, converting it to carbon dioxide and water. Methane is a potent **greenhouse gas**; that is, it helps to keep the climate warm. Carbon dioxide is also a greenhouse gas, but much less potent than methane. The result of the rise of oxygen and the decline of methane should, therefore, have been a decline in global temperature.

(2) There was remarkably little volcanic activity between 2.45 and 2.2 billion years ago. This can be shown by analysis of **zircons**, which being resistant to weathering and erosion outlast the igneous rocks in which they are formed, and which can be accurately dated as explained in the article on uranium-based dating methods. The scarcity of zircons formed between 2.45 and 2.2 billion years ago as compared to their relative abundance both before and after strongly suggests a time of low volcanic activity.

Now volcanoes are prone to emit the greenhouse gas carbon dioxide; if volcanic activity virtually ceased, while the process of **chemical weathering** continued to remove carbon dioxide from the atmosphere, this would lead to a reduction of the greenhouse effect and a decline in global temperatures.

There is, of course, absolutely no reason why both these mechanisms shouldn't have worked together to produce the Huronian glaciation; but while this is possible, it is also possible that one or the other was of such a greater degree of significance that it would be reasonable to call it *the* cause of the Huronian glaciation.

Whichever of these mechanisms was the main cause, it would be an example of a one-off event. The transition from an atmosphere with little or no free oxygen to one with abundant free oxygen happened, and indeed could happen, only once in the history of the Earth. But a quarter-billion years of low volcanic activity is equally rare in the history of the Earth. It follows that studying the possible causes of this one ice age tells us nothing about the causes of the others, nor can studying the causes of the ice age that we're currently in tell us anything about the causes of the Huronian ice age.

While it can be difficult to say why a particular ice age occurred, it is relatively easy to determine *that* it occurred. We shall now turn our attention to the nature of the evidence.

6.13.4 Ice ages: how do we know?

As explained in the main article on **glaciers**, we can observe **sedimentary** and **erosional** features associated with **glaciers** existing today, or which have melted within the time that people have been keeping records of their location: features including **striations**, **moraines** of **glacial till**, **deposits** of **outwash**, **dropstones** in lakes, etc.

When we look at the evidence for recent **glacial** advances, we see the same sedimentary and erosional signs, but further south than the southern limit of the ice sheets in the present **interglacial**. By looking at the location of the lobe-shaped **terminal moraines**, we can find the southern extent of the ice sheets.

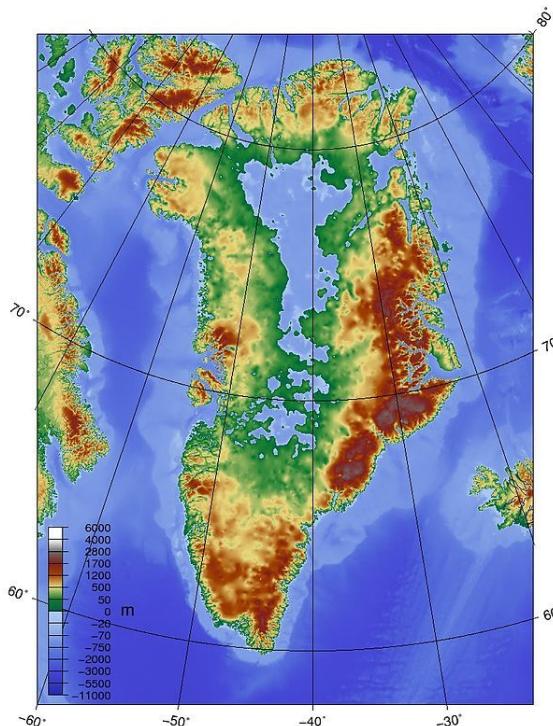
These observations would really be sufficient on their own; but in addition to this we can use **proxies** to investigate the advance and retreat of the **glaciers**. **Biogeographical proxies** such as **pollen** are particularly useful; as the events in question are so geologically recent, we can recognize the pollen of modern species and know exactly under which conditions they flourish.

Another source of information is **cosmogenic surface dating**. **Glaciers** are a powerful erosive force, and, as the **striations** show, **scrape** **sediment** off right down to the **bedrock**. When the **glacier** retreats in an **interglacial**, the **bedrock** is exposed to **cosmic rays**, and so we can use **cosmogenic surface dating** to date the retreat of the **glaciers**.

One interesting relic of the recent glacial retreat is the existence of so-called "**sky islands**". These are mountains found in the southern United States and Mexico which are home to a distinctive flora and fauna, and which are surrounded by dry grassland or scrubland which the sky

island fauna can't cross: hence the name *sky islands*. But if the species can't migrate from mountain to mountain, how did these species achieve their current range? In the light of paleoclimatology, the solution becomes obvious. When the climate of the southern U.S. was cooler than it is today, the *climate* of the plains would have been congenial to the species now found in sky islands. But when temperatures increased in the interglacial, they had nowhere to go but the cold moist refuge of the mountains.

Another indication of the recent *glacial* advance and retreat is **isostatic rebound**. Recall from our discussion of the structure of the Earth that the *lithosphere* is essentially floating on top of the denser *athenosphere*. When a region of the Earth is burdened by the weight of continental *glaciers*, this ought to push the *lithosphere* down into the *athenosphere*. When the *glaciers* retreat, the weight is lifted, and the *lithosphere* should very slowly bob back up. Geologists can measure just this happening, at a rate of about 1 cm/year, in regions which (according to sedimentary and other indications) were covered by *sheet glaciers* before the present *interglacial*.



Topographic map of Greenland.

As an example of what the burden of ice sheets can do to topography, consider the map to the right: it shows the present elevation of Greenland as it would be if the *ice sheets* weren't present. As you can see, Greenland is depressed below sea-level at the center where the *ice sheets* are thickest.

When we look further into the past, we should expect many of these indications to be absent: isostatic rebound, for example, wouldn't take a quarter of a billion years, and so would not remain as evidence of the Karoo ice age.

Similarly, as explained in the article on *biogeographical proxies*, these become less and less useful the further back into the past we wish to look.

However, we still have the *moraines*, the *striations*, the *dropstones*, etc. Are these sufficient to diagnose an *ice age*? It is reasonable to say yes: for we find these associated with present *glaciers*, and we also find them associated with the other indications listed above (*sky islands*, *isostatic rebound*, *proxies*, *cosmogenic surface dating*) when the deposition of the *till* is recent enough for these additional indications of the former presence of *glaciers* to have survived to the present. It would be really astonishing if in the cases where we lack this supplementary information, the *sedimentary* and *erosional* indications of *glaciation* were produced by some other process; it is reasonable to conclude that they are signs of an *ice age*, both because of their empirical association with the present *ice age*, and because (with the arguable exception of *drumlins*) we have a clear understanding of how and why *glaciers* produce these effects.

6.14 Sea level variations

Changes in global sea level are important both as a symptom and as a cause of *climate change*. In this article we shall look at what causes changes in global sea level, and how they can be detected in the geological record.

6.14.1 Causes of sea-level changes

One cause of sea level variations is the formation or melting of *continental glaciers*. It is estimated, for example, that if the *ice sheets* currently covering Greenland and Antarctica were to melt, the global sea-level would rise by approximately 70 meters.

Another factor that affects global sea level is the *mid-ocean ridges*. The more active they are, the larger they are, and the more water they displace. The ridges can also change in total length: when a *supercontinent* rifts apart, a *mid-ocean ridge* will be produced where it rifts; the production of the mid-Atlantic ridge by the rifting of *Pangaea* is the most recent example. As the ridge is in effect a mountain range 40,000 miles long, its formation clearly displaced a massive amount of water. *Hotspots* producing islands such as *Hawaii* have a similar though lesser effect.

The collision of continents can reduce sea-level by reducing the area of the continents as they compress into one another. For example, it has been estimated that the collision of India with south Asia and the compression resulting in the raising of the *Himalayas* would have lowered the sea-level by about 10 meters.

6.14.2 Local variations

The position of the shoreline can change as a result of the movement of the land, as a result of tectonic events or of isostatic rebound after depression caused by glacial cover. This means that there's no use in just looking at one or two places to estimate sea levels in the past, because what looks like a global fall in sea levels might in fact be a local rise in the level of the land.

Such local effects did in fact confuse the 18th century scientist Anders Celsius, whose measurements of the sea level along the Scandinavian coast convinced him that the seas must be shrinking. He was not to know that in fact Scandinavia was slowly rising out of the sea as a result of isostatic rebound.

To identify global changes in sea-level, it is necessary to look beyond these localized changes and see the bigger picture.

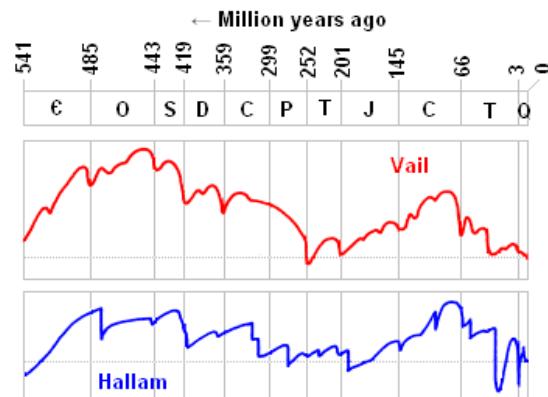
6.14.3 The Vail and Hallam curves

The **Vail curve** representing past global sea-levels is based on the work of Peter Vail and his associates. It is based on the study of unconformities in the geological record representing times of marine regression. An important part of Vail's work was establishing that these unconformities could be identified underground by the use of **reflection seismology**, a method which relies on studying the reflections of seismic waves artificially generated by producing small explosions.

The Vail curve has one major problem, from a methodological point of view: it is based on proprietary data of the Exxon-Mobil corporation which the general public is not allowed to see. This doesn't necessarily mean that the Vail curve is wrong, but it does mean that *if* some aspect of it is wrong, if the data on which it's based are incorrect in some way, or if Vail has somehow misinterpreted the data, then no-one would be able to tell. This makes it different from other results in geology, the quality of which is insured by the ability of scientists to cross-check each other's work.

The same problem does not arise with the **Hallam curve**, the brainchild of Anthony Hallam. This was constructed on the basis of a different methodology: Hallam calculated the area of the continents covered by the sea over the course of time by seeing where marine sediments were and were not deposited at various times, allowing him to sketch out a series of shorelines for the various continents over time; clearly when the land area shrinks globally, this corresponds to a global rise in sea level.

The Hallam and Vail curves are presented in the chart to the right. Note that these are "first order" curves which only show broad trends: they are not on a fine enough scale to show the fluctuations caused by recent advances and retreats of sheet glaciers.



The Vail and Hallam curves.

The reader will note the broad similarities between the two curves, which suggest that Hallam and Vail are basically on the right track, especially since this agreement is produced by two different methods. But the reader will also note the differences in detail, which suggest that one curve or the other, or probably both, are not completely accurate.

Their broad accuracy is confirmed by the fact that we see in them the features that we would expect to see; for example, we see a fall in sea level accompanying the assembly of Pangaea, and a rise in sea-level corresponding to the fracturing of Pangaea and the formation of the Mid-Atlantic Ridge. We can also see falls in sea-level corresponding to known times of glaciation.

Chapter 7

Appendices

7.1 Appendix A: Glossary and index

7.1.1 100,000 year problem

The question of why over the last million years, climatic variation has been driven by the 100,000 year Milankovitch cycle rather than the 41,000 year Milankovitch cycle. *Article:* Milankovitch cycles.

7.1.2 Aa

A type of lava flow, or the cooled and solidified rock produced by it, characterized by the rough jagged surface of the resulting rock. *Article:* Way-up structures.

7.1.3 Ablation zone

The end of a glacier; the point at which loss of ice by melting exceeds the supply of ice by the movement of the glacier. *Articles:* Glaciers, Glacial marine sediment, Ice ages.

7.1.4 Abrasion

Erosion of rocks caused by the sediments carried by wind or water. *Article:* Mechanical weathering and erosion.

7.1.5 Absolute dating

Dating methods which tell us how old a rock or fossil is, as opposed to relative dating. *Articles:* Concepts in absolute dating, Erosion, deposition, and time, Dendrochronology, Varves, Amino acid dating, Radioactive decay, K-Ar dating, Ar-Ar dating, Rb-Sr dating, Other isochron methods, U-Pb, Pb-Pb, and fission track dating, Radiocarbon dating, Cosmogenic surface dating, U-Th, U-Pa, and Ra-Pb dating, Paleomagnetic dating, Sclerochronology, Tidal rhythmites and dating, Fossils and absolute dating, Absolute dating: an overview

7.1.6 Abyssal plain

The flat terrain found at the bottom of the ocean beyond the continental margin. *Article:* Marine sediments.

7.1.7 Accretionary prism

An accumulation of sediment which forms in a trench. *Article:* Subduction.

7.1.8 Accretionary wedge

A synonym for accretionary prism. *Article:* Subduction.

7.1.9 Accumulation zone

The beginning of a glacier; the zone in which snowfall exceeds the loss of snow by melting or evaporation. *Article:* Glaciers.

7.1.10 Acidic rock

An outdated and inaccurate term for felsic rock. *Article:* Igneous rocks.

7.1.11 ACM

Abbreviation for atmospheric circulation model. *Article:* Climate models.

7.1.12 Actualism

The observation that the geological record can be explained in terms of the sort of geological processes that actually happen. *Articles:* Actualism, Steno's principles.

7.1.13 Aeolian

Having to do with the wind. *Article:* Deserts.

7.1.14 Aeolian sandstone

Sandstone formed from sand deposited by the wind, i.e. desert sand. *Article:* Deserts.

7.1.15 Al

Chemical symbol for the element aluminum. *Article:* Chemistry for geologists.

7.1.16 Alkenones

Organic molecules produced by certain planktonic organisms, used in the temperature proxy known as U^{37}_K . *Article:* U^{37}_K .

7.1.17 Alluvial fan

A fan-shaped deposit of sediment left where a mountain stream reaches a plain. *Articles:* Deserts, Rivers.

7.1.18 Alpha decay

Radioactive decay involving the emission of an alpha particle. *Article:* Radioactive decay.

7.1.19 Alpha particle

A particle consisting of two protons and two neutrons. *Article:* Radioactive decay.

7.1.20 Aluminosilicate

Any of an important class of silicate minerals in which the structure involves aluminum-based as well as silicon-based tetrahedra. *Article:* Minerals.

7.1.21 Alpine glacier

Synonym for valley glacier. *Article:* Glaciers.

7.1.22 Amber

A mineraloid formed from the solidified resin of trees. *Article:* Fossils.

7.1.23 Amino acid dating

A rather unreliable method of absolute dating based on measuring the racemization of organic remains. *Article:* Amino acid dating.

7.1.24 Amorphous

Lacking a crystal structure. *Article:* Minerals.

7.1.25 Ammonite

An extinct marine mollusc. *Articles:* Geological column, Index fossils.

7.1.26 Amphibole

A group of silicate minerals in which the SiO_4^{4-} units are bonded to form a double chain. *Article:* Silicate minerals, Igneous rocks.

7.1.27 Angle of repose

The maximum angle from the horizontal that a heap of a given type of sediment can assume without collapsing. *Article:* Steno's principles.

7.1.28 Angular unconformity

An unconformity in which the older strata meet the younger strata at an angle, the older strata being truncated by the erosional surface. *Article:* Unconformities.

7.1.29 Anomaly

In geology, the term “anomaly” means a measurement at some place of some quantity which is different from the average or background value for that quantity. This should not be confused with the usage of the term “anomaly” in the philosophy of science, where it means a measurement or observation which cannot be reconciled with current theory. In geology, the term has no such implication. *Article:* Sea floor spreading.

7.1.30 Antecedent river

A river which is present before the uplift of the hills through which it flows. *Article:* Rivers.

7.1.31 Anthracite

A very black, hard, and shiny form of coal produced by metamorphism. *Article:* Peat and coal.

7.1.32 Anticline

Structure formed when rocks are folded upwards. *Article:* Folds.

7.1.33 Antidune

A rounded dune-like structure found in rivers of the right velocity and having a sandy bottom. Because they erode by the transport of sand grains from the lee side of one antidune to the stoss side of the next, the net effect is that while the sand moves downstream, the antidunes move upstream. *Article: Rivers.*

7.1.34 Antisyncline

An upward fold in rocks. *Article: Folds.*

7.1.35 Aphanitic

An igneous rock is said to be aphanitic if the crystals in it are too small to be seen with the naked eye. In this textbook I have tended to use the more straightforward term “fine-grained”. *Article: Igneous rocks.*

7.1.36 Apparent polar wander

Apparent secular variation recorded in the paleomagnetic record which is actually caused by the motion of plates relative to the poles. *Article: Continental drift.*

7.1.37 Ar

Chemical symbol for the element argon. *Articles: Chemistry for geologists, K-Ar dating, Ar-Ar dating.*

7.1.38 Ar-Ar dating

Argon-argon dating, a form of radiometric dating. *Article: Ar-Ar dating.*

7.1.39 Aseismic ridge

A long linear trail of volcanic islands and seamounts caused by a plate passing over a hotspot. *Article: Hotspots.*

7.1.40 Archaeocyathids

Early reef-building organisms, shaped rather like goblets and secreting skeletons of calcium carbonate; they went extinct at the end of the Cambrian period. *Article: Reefs.*

7.1.41 Argillaceous

Having to do with mud. May be used to qualify the nature of a rock, e.g. argillaceous sandstone would be sandstone

with a significant amount of mud mixed in with the sand. *Article: Sedimentary rocks.*

7.1.42 Arenaceous

Having to do with sand. May be used to qualify the nature of a rock, e.g. arenaceous mudrock would be mudrock with a significant amount of sand mixed in with the mud. *Article: Sedimentary rocks.*

7.1.43 Arenite

Alternative term for sandstone. *Article: Sedimentary rocks.*

7.1.44 Arkose

Sandstone which contains an appreciable quantity of feldspar as well as the more usual quartz. The grains are often poorly sorted and not well rounded. *Article: Sedimentary rocks.*

7.1.45 Athenosphere

The portion of the mantle just below the lithosphere. *Article: Structure of the Earth.*

7.1.46 Atom

A nucleus of protons and neutrons orbited by electrons arranged in electron shells. *Articles: Chemistry for geologists, Radioactive decay.*

7.1.47 Atmospheric circulation model

A climate model which only takes into account the circulation of the atmosphere and not the oceanic circulation. *Article: Climate models.*

7.1.48 Atomic number

The number of protons in the nucleus of an atom. *Article: Chemistry for geologists.*

7.1.49 Atomic weight

The sum of the number of protons and neutrons in the nucleus of an atom. *Article: Chemistry for geologists.*

7.1.50 Attrition

Erosional processes whereby the **clasts** transported by wind or water are broken or worn down. *Article:* Mechanical weathering and erosion.

7.1.51 Aureole

A ring of metamorphic rock formed around an igneous intrusion by **contact metamorphism**. *Article:* Metamorphic rocks.

7.1.52 B

Chemical symbol for the element boron. *Article:* Chemistry for geologists.

7.1.53 Ba

Chemical symbol for the element barium. *Articles:* Chemistry for geologists, Other isochron methods.

7.1.54 Backshore

That part of a beach which is above the high-water line. *Article:* Nearshore sediments.

7.1.55 Bajada

The merging of two or more alluvial fans. *Article:* Deserts.

7.1.56 Banded iron formation

Sedimentary rock consisting of alternating bands of iron oxide and other sedimentary rock, typically **chert**. *Article:* Banded iron formations.

7.1.57 Bar

A local accumulation of sediment, usually sand, such as forms in between the channels of a braided stream or offshore from a beach. *Article:* Rivers.

7.1.58 Barrier island

An island formed at the mouth of a river running at right angles to the direction of the **distributary streams**. *Article:* Deltas.

7.1.59 Basalt

A mafic intrusive igneous rock, black in color and aphanitic. *Article:* Igneous rocks.

7.1.60 Basic rock

An obsolete and inaccurate term for **mafic rock**. *Article:* Igneous rocks.

7.1.61 Be

Chemical symbol for the element beryllium. *Article:* Chemistry for geologists.

7.1.62 Bed

A layer in a **sedimentary rock**. *Article:* Sedimentary rocks.

7.1.63 Bedding

A structure found in **sedimentary rocks** in which the rock is visibly composed of numerous layers (beds). *Article:* Sedimentary rocks.

7.1.64 Bedding planes

The planes dividing the **beds** in a bedded rock. *Article:* Sedimentary rocks.

7.1.65 Bedrock

The solid **rock** underlying unlithified sediment. *Articles:* Glaciers, Rivers.

7.1.66 Beta decay

A type of radioactive decay including **beta plus** and **beta minus** decay; the term is sometimes used to include **electron capture** as well. *Article:* Radioactive decay.

7.1.67 Beta minus decay

A form of radioactive decay in which one of the neutrons in the atom is converted to a proton by emitting an electron. *Article:* Radioactive decay.

7.1.68 Beta plus decay

A form of radioactive decay in which a proton is converted into a neutron by the emission of a positron. *Article:* Radioactive decay.

7.1.69 BIF

Abbreviation for banded iron formation. *Article:* Banded iron formations.

7.1.70 Biogeography

The study of the geographical distribution of living or extinct organisms. *Articles:* Continental drift, Biogeography and climate.

7.1.71 Bioturbation

Changes in the structure of sediment caused by the activity of living things. *Article:* Nearshore sediments.

7.1.72 Bituminous coal

The commonest form of coal: less peat-like than sub-bituminous coal, but not as hard, black, and pure as anthracite. *Article:* Peat and coal.

7.1.73 Bivalve

A member of a group of molluscs characterized by being enclosed in two shells (valves). Common examples are mussels, clams, and oysters. *Articles:* Reefs, Way-up structures.

7.1.74 Body waves

Seismic waves which pass through the body of the Earth rather than traveling on its surface; a collective term for S-waves and P-waves. *Articles:* Seismic waves, Structure of the Earth.

7.1.75 Borax

An evaporite mineral having the chemical formula $\text{Na}_2\text{B}_4\text{O}_7 \cdot 10\text{H}_2\text{O}$. *Article:* Deserts.

7.1.76 Bottom-set beds

Horizontal beds of sediment deposited on the sea or lake floor in front of a delta. *Article:* Deltas.

7.1.77 Bouma sequence

The characteristic pattern of sediment deposited by a turbidity current. *Article:* Turbidites.

7.1.78 Br

Chemical symbol for the element bromine. *Articles:* Chemistry for geologists, Other isochron methods.

7.1.79 Braided stream

A stream or river in which the current repeatedly splits into smaller streams which merge back together and then split again, and so forth. *Article:* Rivers.

7.1.80 Breccia

A rock consisting of large unrounded fragments cemented together. *Articles:* Sedimentary rocks, Faults.

7.1.81 Brittle

A material is said to be brittle if with increasing stress it undergoes very little plastic deformation between elastic deformation and shattering. *Article:* Physical properties of rocks.

7.1.82 Brown clay

Term occasionally used for pelagic clay. *Articles:* Pelagic clay, Marine sediments.

7.1.83 C

Chemical symbol for the element carbon. *Articles:* Chemistry for geologists, Radiocarbon dating.

7.1.84 Ca

Chemical symbol for the element calcium. *Article:* Chemistry for geologists.

7.1.85 Calcareous

Having to do with calcium carbonate. *Articles:* Calcareous ooze, Marine sediments.

7.1.86 Calcareous ooze

A calcareous sediment found over large areas of the ocean floor, consisting of the shells of small organisms. *Articles:* Calcareous ooze, Marine sediments.

7.1.87 Calcite

A mineral consisting of calcium carbonate in a trigonal crystal system.

7.1.88 Calcium carbonate

The chemical CaCO_3 . Most shells are formed of this, as are the rocks limestone and marble. *Articles:* Calcareous ooze, Marine sediments.

7.1.89 Carbonate

A molecule with the negative ion CO_3^{2-} ; also a rock consisting of carbonates, particularly limestone. *Articles:* Calcareous ooze, Marine sediments.

7.1.90 Carbonate compensation depth

The depth at which calcium carbonate will dissolve faster than it is deposited; hence, the depth below which calcareous ooze will not accumulate. *Article:* Calcareous ooze.

7.1.91 Carbon dating

Alternative term for radiocarbon dating. *Article:* Radiocarbon dating.

7.1.92 C-C dating

Alternative term for radiocarbon dating. *Article:* Radiocarbon dating.

7.1.93 Carbon dioxide

The molecule CO_2 . A gas at temperatures and pressures found on Earth, and forming 0.038% of the Earth's atmosphere. *Article:* Chemical weathering.

7.1.94 Carbonic acid

The acid H_2CO_3 . Although this is a very weak acid, it is extremely common, because it can be formed from the reaction between carbon dioxide and water. Because of

this, it plays an important role in chemical weathering. *Article:* Chemical weathering.

7.1.95 Cast

A fossil produced when a mold is filled with minerals. *Article:* Fossils.

7.1.96 Cave formation

Alternative term for a speleothem. *Articles:* U-Th, U-Pa, and Ra-Pb dating, U-Pb, Pb-Pb, and fission track dating.

7.1.97 CCD

Abbreviation for carbonate compensation depth. *Article:* Calcareous ooze.

7.1.98 ^{14}C dating

Alternative term for radiocarbon dating. *Article:* Radiocarbon dating.

7.1.99 Ce

Chemical symbol for the element cerium. *Articles:* Chemistry for geologists, Other isochron methods.

7.1.100 Cementation

The binding of **clasts** together by a finer material, typically silica, calcium carbonate, or iron oxide, to form a **clastic** rock. *Article:* Sedimentary rocks.

7.1.101 Cemented tuff

Volcanic ash which has lithified by the process of cementation, as opposed to welded tuff. *Article:* Volcanic ash.

7.1.102 Chain silicate

A **silicate mineral** in which the **silicate tetrahedra** are bonded together in the form of a chain, i.e. each tetrahedron is attached to just two other tetrahedra (except, of course, at each end of the chain). *Article:* Silicate minerals.

7.1.103 Chalk

Rock which, under a microscope, is clearly composed of the tests of calcium carbonate-secreting micro-organisms. *Article:* Calcareous ooze.

7.1.104 Chemical sediment

A chemical sediment is one deposited by precipitation rather than by mechanical processes such as wind or water; or by biological processes such as the growth of coral. Note however that some authors will include biological processes as a subcategory of chemical processes; our articles do not follow this usage. *Article:* Sedimentary rocks.

7.1.105 Chemical weathering

Weathering caused by chemical processes (most commonly by some or all of the constituent minerals of a rock being dissolved by carbonic acid); as opposed to mechanical weathering. *Article:* Chemical weathering.

7.1.106 Chert

A sedimentary rock composed of silica, having an amorphous or very fine-grained structure. *Article:* Siliceous ooze.

7.1.107 Chirality

The handedness of an organic molecule. *Article:* Amino acid dating.

7.1.108 Cirque

A large bowl-shaped depression formed at the accumulation point of a valley glacier, with the bowl lacking about a quarter of its rim to let the glacier flow out. *Article:* Glaciers.

7.1.109 Cl

Chemical symbol for the element chlorine. *Article:* Chemistry for geologists.

7.1.110 Clast

A piece of rock detached by erosion or weathering from a larger rock. *Article:* Sedimentary rocks.

7.1.111 Clastic

Composed of **clasts**. *Article:* Sedimentary rocks.

7.1.112 Clay

The term clay can either, depending on the context, refer to a class of sheet aluminosilicate minerals, or to clasts with a diameter of less than 1/256 mm. As clay in the second sense is usually also clay in the first sense, this causes less confusion than you might think. *Article:* Sedimentary rocks.

7.1.113 Claystone

Sedimentary rock composed of clay. *Article:* Sedimentary rocks.

7.1.114 Climate

Broad trends in the weather; i.e. the tendency of a location to be hot and humid, or dry and cold. *Article:* Paleoclimatology: introduction.

7.1.115 Closure

The point at which snow has been so far compacted into ice that the air trapped in it is completely sealed off from the atmosphere. *Article:* Ice cores.

7.1.116 Closure time

The time between snowfall and closure, varying from location to location. *Article:* Ice cores.

7.1.117 Coal

Coal is peat which has been lithified by compaction, heat, or both. *Article:* Peat and coal.

7.1.118 Coalification

The chemical processes by which peat is turned into coal. *Article:* Peat and coal.

7.1.119 Coarse-grained

Composed of crystals of large size; the opposite of fine-grained. *Article:* Igneous rocks.

7.1.120 Cobble

A **clast** between 64 and 256mm in diameter, especially one that exhibits **rounding**. *Articles:* Sedimentary rocks, Rivers.

7.1.121 Coccolith

A calcareous plate forming part of the shell of a **coccolithophore**; a common constituent of calcareous ooze. *Article:* Calcareous ooze.

7.1.122 Coccolithophores

A group of micro-organisms clad in **coccoliths**. *Article:* Calcareous ooze.

7.1.123 Compaction

Decrease in volume of **sediment**, caused by the pressure induced by being buried under yet more sediment. *Articles:* Sedimentary rocks, Ice cores.

7.1.124 Complacent

A tree is said to be **complacent** if the thickness of its growth rings is unaffected by annual variations in temperature. *Article:* Dendrochronology.

7.1.125 Compression

Stress that produces shortening of a solid along the direction in which force is applied. *Article:* Physical properties of rocks.

7.1.126 Concordant

Of dates, in agreement with one another. *Articles:* U-Pb, Pb-Pb, and fission track dating, Absolute dating: an overview.

7.1.127 Conglomerate

A **conglomerate** is a rock consisting of large **clasts** (pebble-sized or larger) cemented together; it is common usage (which we have followed in this text) to use the term to imply that the clasts are **rounded**, as distinct from a breccia. *Article:* Sedimentary rocks.

7.1.128 Conodont

Term used ambiguously to refer either to a **conodont animal** or a **conodont structure**; context usually makes it clear which. *Article:* Fossils.

7.1.129 Conodont animals

A group of extinct primitive chordates having no hard parts except for **conodont structures**. *Article:* Fossils.

7.1.130 Conodont structures

The hard parts of a conodont animal. *Article:* Fossils.

7.1.131 Contact metamorphism

Metamorphism caused by close proximity to a source of heat, such as an intrusion of **magma**; as opposed to regional metamorphism. *Article:* Metamorphic rocks.

7.1.132 Continental drift

The theory that continents have shifted their positions over time; now subsumed into the theory of plate tectonics. *Article:* Continental drift.

7.1.133 Continental glacier

A **glacier** covering a large area and flowing outwards from its accumulation zone under the pressure of its own weight, as distinct from a **valley glacier**. *Articles:* Glaciers, Ice ages.

7.1.134 Continental margin

The **continental shelf**, **slope**, and **rise**. *Article:* Marine sediments.

7.1.135 Continental rise

The shallowly sloping (approximately 1 degree from horizontal) terrain between the **continental slope** and the **abyssal plain**. *Article:* Marine sediments.

7.1.136 Continental shelf

That part of a continent which is underwater, lying between the unsubmerged portion of a continent and the **continental slope**. *Article:* Marine sediments.

7.1.137 Continental slope

A shallow slope, typically between 4 and 10 degrees from horizontal, found between the continental shelf and the continental rise. *Article:* Marine sediments.

7.1.138 Corals

A group of marine organisms. Hard corals secrete skeletons of calcium carbonate and so act as reef-forming organisms. *Article:* Reefs.

7.1.139 Core

The innermost 3,400 km of the Earth, composed mainly of iron. *Article:* Structure of the Earth.

or

A sample of ice or rock recovered from the Earth's crust by drilling. *Article:* Ice cores.

7.1.140 Cosmic dust

Dust fallen from outer space, i.e. micrometeorites. Although they can be found in pretty much all kinds of sediment, they are proportionally most abundant in pelagic clay due to its slow rate of deposition. *Article:* Pelagic clay.

7.1.141 Cosmic rays

Streams of high-energy particles which bombard the Earth from outer space. *Article:* Cosmogenic surface dating.

7.1.142 Cosmogenic

Of isotopes, produced by cosmic rays. *Articles:* Cosmogenic surface dating, Radiocarbon dating.

7.1.143 Cosmogenic surface dating

A method of absolute dating which gives the time since a rock became exposed on the surface. *Article:* Cosmogenic surface dating.

7.1.144 Country rock

The rock into which an igneous rock intrudes. *Article:* Igneous rocks.

7.1.145 Covalent bond

A bond between atoms in which they share electrons. *Article:* Chemistry for geologists.

7.1.146 Creep

Transport of clasts by wind or water by means of rolling them along the ground, river bed, sea bed, etc. *Articles:* Mechanical weathering and erosion, Rivers.

7.1.147 Cross-bedding

Bedding in which the beds, instead of being deposited horizontally, are deposited at an angle, as a result of deposition by a current of wind or water; in the simplest case, where the current has a continuous direction, the beds will have a downward slope in the direction of the current. *Articles:* Sedimentary rocks, Deserts, Rivers, Deltas, Nearshore sediments, Way-up structures, Paleocurrents.

7.1.148 Cross-cutting

An igneous rock such as a dike which cuts through the beds of country rock is said to be cross-cutting. *Article:* Cross-cutting relationships.

7.1.149 Crossdating

The correlation of dates from different sources. *Article:* Dendrochronology.

7.1.150 Crust

The upper layer of the Earth, varying from about 5 - 50 km thick, distinct from the mantle by having a different chemical composition, being composed of less dense and more felsic rocks. *Article:* Structure of the Earth.

7.1.151 Crystal

A large molecule composed of smaller chemical units chemically bonded together in a regular repetitive arrangement. *Article:* Minerals.

7.1.152 Crystal habit

The shape or shapes in which a mineral will typically grow. *Article:* Minerals.

7.1.153 Crystal system

One of the seven basic geometrical arrangements in which the atoms of a crystal can be arranged: triclinic, monoclinic, orthorhombic, tetragonal, trigonal, hexagonal or cubic. *Article:* Minerals.

7.1.154 Curie temperature

Very roughly speaking, the temperature above which a material cannot be magnetized and below which it can. *Article:* Geomagnetic reversals.

7.1.155 Cyclosilicate

A synonym for ring silicate. *Article:* Silicate minerals.

7.1.156 $\delta^{18}\text{O}$

A proxy for temperature based on oxygen isotope ratios. *Articles:* Scleroclimatology, Ice cores.

7.1.157 Decay

The destruction of organic remains by organic processes. *Articles:* Peat and coal, Soils and paleosols, Fossils.

or

Radioactive decay. *Articles:* Radioactive decay.

7.1.158 Decay chain

A sequence of events in which one isotope decays to another via an intermediate sequence of unstable isotopes. *Article:* Radioactive decay.

7.1.159 Deflation

The erosion of fine particles from dry soil by the wind. *Article:* Deserts.

7.1.160 Deflation lake

A lake caused when deflation has caused a hollow the bottom of which lies below the water table. *Article:* Deserts.

7.1.161 Delta

The body of sediment deposited when a river flows into a lake or the sea. *Article:* Deltas.

7.1.162 Dendrochronology

A method of dating wood by studying the annual growth rings produced by the tree. *Article:* Dendrochronology.

7.1.163 Deposition

All those processes which add sediment to a surface; the opposite of erosion. *Article:* Sedimentary rocks.

7.1.164 Desert

An area of exceptionally low rainfall. Note that although the stereotypical desert is hot and sandy, in geological terms a desert is defined solely by a shortage of rain or snow. *Article:* Deserts.

7.1.165 Desert pavement

A stony surface often found in deserts. *Article:* Deserts.

7.1.166 Dессification crack

Alternative term for a mud crack. *Article:* Way-up structures.

7.1.167 Detrital

Composed of clasts; synonymous with clastic. *Article:* Sedimentary rocks.

7.1.168 Dextral

A fault is said to be dextral if someone standing on one side of the fault and looking at the other when there is motion along the fault would see the other side moving to the right. *Article:* Faults.

7.1.169 Diamond anvil cell

A device used in experimental petrology to subject small samples of rock to large amounts of stress. *Article:* Physical properties of rocks.

7.1.170 Diapir

A sedimentary structure formed by one type of sediment flowing upwards through another as a result of pressure. *Article:* Way-up structures.

7.1.171 Diatomite

A very light and porous rock formed from diatom tests that have undergone little in the way of compaction and recrystallization. *Article:* Siliceous ooze.

7.1.172 Diatoms

A group of single-celled algae which produce siliceous tests; a major source of siliceous ooze. *Article:* Siliceous ooze.

7.1.173 Differentiation

The mechanism by which an originally homogeneous Earth separated into crust, mantle, and core. *Article:* Structure of the Earth.

7.1.174 Dike

A vertical or near-vertical sheet of igneous rock which intrudes into the country rock. *Articles:* Igneous rocks, Cross-cutting relationships, Igneous rocks and stratigraphy, Ophiolites.

7.1.175 Dip-slip fault

A **fault** in which much of the motion of the rocks on either side of the fault is vertical: hence either a **normal fault** or a **reverse fault**. *Article:* Faults.

7.1.176 Disconformity

An **unconformity** in which the underlying strata are parallel with the overlying strata. *Article:* Unconformities.

7.1.177 Dissolved

A substance (a **solute**) is said to be dissolved in another substance (a **solvent**) if it is mixed with it in such a way as to acquire the phase of the solvent. *Articles:* Chemistry for geologists, Chemical weathering.

7.1.178 Distributary

A smaller stream flowing out of a larger river, as opposed to a **tributary**, which flows in. *Article:* Rivers.

7.1.179 Diurnal

Occurring once daily. *Article:* Tidal rhythmites and dating.

7.1.180 Docking

The union of a **terrane** with the landmass to which it becomes attached. *Article:* Terranes.

7.1.181 Drag fold

A structure formed in a rock as friction drags the material in it backwards relative to its motion along a **fault**. *Article:* Folds.

7.1.182 Drift

Any **sediment** deposited by a glacier. *Article:* Glaciers.

7.1.183 Drifters

Term for the early supporters of **continental drift**; the opposite of “fixists”. *Article:* Continental drift.

7.1.184 Dropstone

A stone which has traveled out to sea on a “raft” of **ablated** glacial ice, and has been deposited when the ice melted. *Articles:* Glaciers, Glacial marine sediment.

7.1.185 Drumlin

A smallish hill shaped somewhat like the back of a spoon, deposited by **glaciers** in a manner not fully understood. *Article:* Glaciers.

7.1.186 Ductile

A material is said to be **ductile** if, under stress, it will undergo a great deal of **plastic** deformation before it breaks. The opposite of **brittle**. *Article:* Physical properties of rocks.

7.1.187 Dune

A mound of sand formed by the action of wind or water. *Articles:* Rivers, Deserts.

7.1.188 Dunnite

An ultramafic rock consisting entirely of olivine. *Article:* Igneous rocks.

7.1.189 Elastic

A material is said to be elastic if it recovers from stress: that is, if when the stress is removed it returns to its original conformation. The opposite of **plastic**. *Article*: Physical properties of rocks.

7.1.190 Electron

A particle with negative charge and negligible mass found orbiting the nucleus of an atom. *Articles*: Chemistry for geologists, Radioactive decay.

7.1.191 Electron capture

A form of radioactive decay in which the radioactive atom's own electrons combines with one of its protons, converting the proton into a neutron. *Article*: Radioactive decay.

7.1.192 Electron shell

An orbit followed by electrons about an atomic nucleus. *Articles*: Chemistry for geologists, K-Ar dating, Ar-Ar dating.

7.1.193 Element

Atoms are classified into elements according to their atomic numbers, which determine their chemical properties; this is a broader classification than the division into isotopes, which also takes into account their atomic weights. *Articles*: Chemistry for geologists, Radioactive decay.

7.1.194 Enantiomers

Molecules which are mirror images of one another. *Article*: Amino acid dating.

7.1.195 Entire margins

Leaf margins which are smooth rather than serrated, characteristic of warm humid climates. *Article*: Leaf shape and temperature.

7.1.196 Epicenter

The point on the Earth's surface directly above the focus of an earthquake. *Article*: Seismic waves.

7.1.197 Erg

A sandy desert. *Article*: Deserts.

7.1.198 Erosion

Any process capable of breaking up rocks or soils and transporting the resulting **clasts**. *Article*: Mechanical weathering and erosion.

7.1.199 Erratic boulder

A **boulder** which does not fit in with the geology of its surroundings, transported from its place of origin by a glacier. *Article*: Glaciers.

7.1.200 Evaporite

Any chemical **sedimentary rock** the precipitation of which was produced by the partial or complete evaporation of the water containing the **dissolved** minerals or which the rock is composed. *Articles*: Sedimentary rocks, Deserts, Saline giants.

7.1.201 Event horizon

An extensive geological feature all of which was laid down at exactly the same time; e.g. volcanic ash from a single volcanic eruption. *Article*: Volcanic ash.

7.1.202 Evolution

In biology, heritable change in a line of descent. Outside of biology, the term may be used colloquially to refer to any sort of change or development, as in (for example) "the evolution of jazz from ragtime". *Article*: Principle of faunal succession.

7.1.203 Excess argon

Argon which is not radiogenic; a potential source of error in Ar-Ar dating. *Articles*: K-Ar dating, Ar-Ar dating.

7.1.204 Exponential decay

A quantity is said to undergo exponential decay if its magnitude as a function of time t can be expressed in the form ab^{-ct} . *Article*: Radioactive decay.

7.1.205 Extrusive rock

Any igneous rock formed by lava pouring out on the surface (where the “surface” includes on the sea floor, under a glacier, or anywhere except under rock) as opposed to intrusive rock, which remains trapped within the country rock into which it intrudes. Extrusive rock can be distinguished from intrusive rock by its larger crystal size. *Article:* Igneous rocks.

7.1.206 Facies

A facies is a body of sediment or sedimentary rock characteristic of a particular depositional environment. *Article:* Walther’s principle.

7.1.207 Fault

A planar fracture or discontinuity in a volume of rock. *Articles:* Faults, Terranes.

7.1.208 Fault breccia

Breccia produced by the crushing action of motion along a fault. *Article:* Faults.

7.1.209 Fault gouge

Material similar to fault breccia but finer in texture. *Article:* Faults.

7.1.210 Fault mirror

Alternative term for slickenside. *Article:* Faults.

7.1.211 Fauna

Animals (in the broadest possible sense, including birds, fish, crustaceans, molluscs, etc). *Article:* Principle of faunal succession.

7.1.212 Fe

Chemical symbol for the element iron. *Article:* Chemistry for geologists.

7.1.213 Feldspar

A group of aluminosilicate minerals with a lattice structure. *Article:* Igneous rocks.

7.1.214 Felsic rocks

Rocks which are high in silica and feldspar and low in magnesium and iron. The opposite of mafic rocks. *Article:* Igneous rocks.

7.1.215 Fine-grained

Composed of crystals or clasts of small size; the opposite of coarse-grained. *Article:* Igneous rocks.

7.1.216 Fining-up sequence

A form of grading upwards from coarse to fine sediments associated with rivers. *Article:* Rivers.

7.1.217 Firn

Snow which has compacted, but not so far as to become ice. *Article:* Ice cores.

7.1.218 Fissile

Of a rock, having the property of splitting easily in a given direction (e.g. between bedding planes). *Article:* Sedimentary rocks.

7.1.219 Fission track dating

A form of absolute dating which involves counting the fission tracks in a rock. *Article:* U-Pb, Pb-Pb, and fission track dating.

7.1.220 Fission tracks

Microscopic scars left in minerals by alpha particles. *Article:* U-Pb, Pb-Pb, and fission track dating.

7.1.221 Fixist

Term for the early opponents of continental drift; the opposite of “drifter”. *Article:* Continental drift.

7.1.222 Flame structure

A sedimentary structure formed when a denser sediment (typically sand) is deposited on top of a less dense sediment (typically mud) which then penetrates it by seeping upwards; hence, a kind of small diapir. *Article:* Way-up structures.

7.1.223 Flaser deposits

Deposits in which light and heavy sediments alternate, characteristic of nearshore environments. *Article:* Nearshore sediments.

7.1.224 Flood plain

The flattened, sediment-rich area formed by the action of rivers on a landscape. *Article:* Rivers.

7.1.225 Flume

An artificial channel used by geologists to study the transport of sediment by water in the laboratory. *Articles:* Rivers.

7.1.226 Fluvial

Having to do with rivers. *Article:* Rivers.

7.1.227 Focus

The point in the Earth at which an earthquake originates. *Article:* Seismic waves.

7.1.228 Foliation

The arrangement of sheet silicates in parallel planes in some metamorphic rocks, due to pressure causing realignment of the sheets in planes at right angles to the direction of pressure. *Article:* Metamorphic rocks.

7.1.229 Foot wall

Rocks lying above a dip-slip or oblique fault. *Article:* Faults.

7.1.230 Foram

Short form of foraminiferan. *Article:* Calcareous ooze.

7.1.231 Foraminiferans

A group of micro-organisms which secrete calcareous tests; one of the most common constituents of calcareous ooze. *Article:* Calcareous ooze.

7.1.232 Foreset beds

Beds of sediment sloping down at the front of a delta into the sea or lake into which it discharges. *Article:* Deltas.

7.1.233 Foreshore

That part of the nearshore which is uncovered at high tide. *Article:* Nearshore sediments.

7.1.234 Fossil

Organic remains, or traces of organic activity such as footprints, preserved in the geological record. *Article:* Fossils.

7.1.235 Gabbro

Mafic intrusive igneous rock; the intrusive equivalent of basalt. *Article:* Igneous rocks.

7.1.236 GCM

Abbreviation for general circulation model. *Article:* Climate models.

7.1.237 GDGTs

Glycerol dibiphytanyl glycerol tetraethers, organic molecules produced by the single-celled organisms known as Crenarchaeota, used in the TEX_{86} temperature proxy. *Article:* TEX_{86} .

7.1.238 General circulation model

A climate model which takes into account both the atmospheric and the oceanic circulation. *Article:* Climate models.

7.1.239 Geological column

A table showing the order of the faunal succession in the fossil record. *Article:* Geological column.

7.1.240 Geomagnetic reversal

A change in state from normal polarity to reversed polarity, or vice versa. *Articles:* Geomagnetic reversals, Paleomagnetic dating.

7.1.241 Geopetal structure

A structure formed when a hollow object is partially filled with sediment, allowing us to use it as a way-up structure. *Article:* Way-up structures.

7.1.242 Glacial

Having to do with glaciers. *Articles:* Glaciers, Ice ages, Glacial marine sediment, Ice cores.

7.1.243 Glacial outwash

Sediment carried out of a glacier by meltwater. *Article:* Glaciers.

7.1.244 Glacial polish

The smooth (but striated) surface produced on a rock by the polishing action of a glacier passing over it. *Article:* Glaciers.

7.1.245 Glaciation

Synonym for ice age. *Article:* Ice ages.

7.1.246 Glacier

A moving mass of ice. *Articles:* Glaciers, Ice ages, Glacial marine sediment, Ice cores.

7.1.247 Glass

Any igneous rock with an amorphous structure, produced by lava cooling too fast to allow the formation of crystals. *Articles:* Rocks, Volcanic ash.

7.1.248 Global Positioning System

A method for finding one's location on the surface of the Earth; used in geology to measure the motion of plates. *Article:* Continental drift.

7.1.249 Gneiss

A metamorphic rock of high grade with a distinctive streaky appearance produced by the separation out of chain silicates into streaks. *Article:* Metamorphic rocks.

7.1.250 GOE

Abbreviation for Great Oxygen Event. *Article:* Banded iron formations.

7.1.251 Goethite

The iron oxide mineral FeO(OH) . *Articles:* Deserts, Banded iron formations.

7.1.252 GPS

Abbreviation for Global Positioning System. *Article:* Continental drift.

7.1.253 Grade

The degree to which a rock has undergone metamorphism, depending on the amount of heat to which it has been exposed. *Article:* Metamorphic rocks.

7.1.254 Grading

Change in size of clasts between two points; most typically from large clasts at the bottom of a layer to small clasts at the top. *Articles:* Turbidites, Way-up structures.

7.1.255 Grainflow lamina

A lamina formed in sand dunes when sand at the crest of the dune avalanches down the lee face of the dune. *Article:* Deserts.

7.1.256 Granite

A felsic intrusive igneous rock; the intrusive counterpart of rhyolite. *Article:* Igneous rocks.

7.1.257 Gravel

Sediment consisting of clasts 2mm in diameter and upwards. *Article:* Sedimentary rocks.

7.1.258 Greenhouse gas

A gas such as carbon dioxide or methane which helps keep the Earth warm by trapping heat in the atmosphere. *Article:* Ice ages.

7.1.259 Greywacke

Sandstone consisting of quartz, feldspar, and small rock fragments embedded in a clay matrix. *Article:* Sedimentary rocks.

7.1.260 Ground mass

Finer material in which larger clasts (in the case of sedimentary rock) or crystals (in the case of igneous rock) are embedded. A synonym for matrix. *Article:* Igneous rocks.

7.1.261 Group

In chemistry, elements which lie in the same column of the periodic table, with similar chemical properties as a result of having similar situations in their outer electron shells. *Article:* Chemistry for geologists.

7.1.262 Growth ring

A layer of wood produced by a tree on an annual basis, used in dendrochronology. *Articles:* Dendrochronology, Dendroclimatology.

7.1.263 Gypsum

A mineral consisting of hydrated calcium sulfate ($\text{CaSO}_4 \cdot \text{H}_2\text{O}$) with a monoclinic crystal system. *Article:* Saline giants.

7.1.264 H

Chemical symbol for the element hydrogen. *Article:* Chemistry for geologists.

7.1.265 Half-life

The half-life of an isotope is the length of time in which an atom of that isotope has a 50% chance of undergoing radioactive decay. *Article:* Radioactive decay.

7.1.266 Halite

Rock salt (NaCl). *Article:* Saline giants.

7.1.267 Hallam curve

A reconstruction of past variations of sea level based on sedimentary evidence of transgressions and regressions. *Article:* Sea level variations.

7.1.268 Hanging wall

Rocks lying above a dip-slip or oblique fault. *Article:* Faults.

7.1.269 Hematite

The iron oxide mineral Fe_2O_3 . *Articles:* Deserts, Banded iron formations.

7.1.270 Herringbone crossbedding

A form of cross-bedding in which the direction of slope alternates, as a result of oscillatory flow. *Article:* Nearshore sediments.

7.1.271 Hf

Chemical symbol for the element hafnium. *Articles:* Chemistry for geologists, Other isochron methods.

7.1.272 Hoodoo

A pillar of rock produced by erosion. *Article:* Steno's principles.

7.1.273 Horizon

A distinct layer in a soil, formed by pedogenetic processes. *Article:* Soils and paleosols.

7.1.274 Hornfels

A large group of metamorphic rocks produced from sedimentary rocks by contact metamorphism. *Article:* Metamorphic rocks.

7.1.275 Hotspot

A stationary point of high volcanic activity above which plates pass, creating an aseismic ridge. *Article:* Hotspots.

7.1.276 Humus

Decaying organic matter in soil. *Article:* Soils and paleosols.

7.1.277 Humic coal

Coal produced by the deposition of land plants in swamps, as opposed to sapropelic coal. *Article:* Peat and coal.

7.1.278 Hydrothermal

Having to do with hot water. *Article:* Metamorphic rocks.

7.1.279 Ice age

A time at which sheet glaciers are present on some regions of the Earth's surface. *Article:* Ice ages.

7.1.280 Ice sheet

Synonym for continental glacier. *Articles:* Glaciers, Ice ages.

7.1.281 Igneous rock

Rock formed by the cooling of lava (in which case the rock is said to be **extrusive**) or magma (in which case the rock is said to be **intrusive**). Igneous rocks can also be classified by their mineral composition from **felsic** to **ultramafic**. *Article:* Igneous rocks.

7.1.282 Index fossil

A fossil of a species that was sufficiently widely distributed that its fossils can be used to correlate the deposition of **fossils** and **sediments** in widely separated locations. *Articles:* Index fossils, Fossils and absolute dating.

7.1.283 Index mineral

Any **mineral** which forms only at certain temperatures and pressures, and which can therefore be used as an index to the conditions under which certain **metamorphic rocks** were formed. *Article:* Metamorphic rocks.

7.1.284 Index species

A species suitable for the production of **index fossils**. *Article:* Index fossils.

7.1.285 Inert

Of an element, unable to participate in chemical reactions. *Article:* K-Ar dating.

7.1.286 Inosilicate

A synonym for chain silicate. *Article:* Silicate minerals.

7.1.287 Insolation

The quantity per area of solar radiation reaching a given location. *Articles:* Milankovitch cycles, Climate models.

7.1.288 Insoluble

Incapable of becoming **dissolved**. *Article:* Chemistry for geologists.

7.1.289 Interference ripples

Ripples caused by two currents flowing (one after the other) at or near right-angles to one another. *Article:* Nearshore sediments.

7.1.290 Interglacial

A time of glacial retreat during an **ice age**. *Article:* Ice ages.

7.1.291 Interfingering

A complex pattern of **sediments** in which different sedimentary types (e.g. **sand** and **mud**) interpenetrate in interlocking wedges broadly similar to the pattern made by the fingers of two hands laced together. *Article:* Deltas.

7.1.292 Internal drainage

A drainage pattern typical of **deserts**, in which rivers flow into the **desert** and evaporate. *Article:* Deserts.

7.1.293 Intrusive rock

Rock formed by **magma** penetrating **country rock** but not reaching the surface as **lava**. As the **magma** will cool slowly, intrusive rock can be distinguished from **extrusive** rock by the relatively large size of the **crystals** of which the former is composed. Such rock is said to **intrude** into the **country rock**. *Article:* Igneous rocks.

7.1.294 Inverse grading

Grading where the size of clasts varies from small clasts at the bottom of a layer to large clasts at the top. *Article:* Turbidites.

7.1.295 Ion

An atom which has gained or lost electrons, giving it a negative or positive charge respectively. *Article:* Chemistry for geologists.

7.1.296 Ionic bond

A bond between atoms in which one atom donates electrons to another. *Article:* Chemistry for geologists.

7.1.297 Iron oxides

Minerals containing both iron and oxygen, as the name suggests: examples include hematite and goethite. *Articles:* Deserts, Banded iron formations.

7.1.298 Isochron dating

A form of radiometric dating involving the construction of an isochron diagram. *Articles:* Rb-Sr dating, Other isochron methods.

7.1.299 Issochron diagram

A graph showing the isotope ratios of various minerals found in the same rock, used in isochron dating. *Articles:* Rb-Sr dating, Other isochron methods.

7.1.300 Isostatic rebound

The process whereby land which has formerly been depressed by overlying weight (for example of an ice sheet) rises when the weight is removed. *Articles:* Glaciers, Ice ages.

7.1.301 Isotope

Atoms are classified into isotopes according to their atomic number and their atomic weight. *Articles:* Chemistry for geologists, Radioactive decay.

7.1.302 K

Chemical symbol for the element potassium. *Articles:* Chemistry for geologists, K-Ar dating.

7.1.303 K-Ar dating

Potassium-argon dating, a form of radiometric dating. *Article:* K-Ar dating.

7.1.304 Karst topography

The distinctive landscape produced by the chemical weathering of limestone. *Article:* Chemical weathering.

7.1.305 K-Ca dating

Potassium-argon dating, a form of radiometric dating. *Article:* Other isochron methods.

7.1.306 Kettle

A small lake formed by glacial outwash being deposited around a largish chunk of ice left behind by a retreating glacier; when the residual chunk of ice melts, this leaves a depression which will typically fill with water, producing a kettle. *Article:* Glaciers.

7.1.307 La

Chemical symbol for the element lanthanum. *Articles:* Chemistry for geologists, Other isochron methods.

7.1.308 La-Ba dating

Lanthanum-barium dating, a form of radiometric dating. *Article:* Other isochron methods.

7.1.309 La-Ce dating

Lanthanum-cerium dating, a form of radiometric dating. *Article:* Other isochron methods.

7.1.310 Lacolith

An intrusion between two strata, similar to a sill but thicker and lens-shaped. *Article:* Igneous rocks.

7.1.311 Lacustrine

Having to do with lakes. *Article:* Lakes.

7.1.312 Lake

An inland body of water fed by rivers, streams, or sometimes by seepage of groundwater. *Article:* Lakes.

7.1.313 Lamina

A very thin bed, no more than a few millimeters thick. *Article:* Deserts.

7.1.314 Lateral moraine

Sediment which accumulates along the sides of a valley glacier, having fallen or been scraped off the walls of the valley. *Article:* Glaciers.

7.1.315 Laterite

A soil type characteristic of a tropical climate alternating between a monsoon season and a dry season. *Article:* Sediments and climate.

7.1.316 Lattice silicates

Silicate minerals in which the silicate tetrahedra are bonded together to form a three-dimensional lattice. *Article:* Silicate minerals.

7.1.317 Lava

Molten rock which has reached the surface, as opposed to magma, which is still trapped beneath it. *Article:* Igneous rocks.

7.1.318 Lava flow

Igneous rock formed by lava flowing on the surface.

7.1.319 Leached ions

Ions dissolved in water as a result of chemical weathering; as opposed to residual minerals. *Article:* Chemical weathering.

7.1.320 Lee

The side of a mountain, dune, antidune, or generally any hill-shaped geological feature, which is on the downstream side of a current of wind or water. The opposite of stoss. *Article:* Rivers.

7.1.321 LGM

Abbreviation for last glacial maximum. *Article:* Climate models.

7.1.322 Lignite

The softest form of coal; the next stage in the formation of coal from peat after peat itself. *Article:* Peat and coal.

7.1.323 Limestone

Rock formed from calcium carbonate, usually in the form of calcite. *Article:* Calcareous ooze.

7.1.324 Limiting stand

Trees the growth of which we would expect to be limited by a single factor (such as temperature) because they grow in an environment with an abundant supply of other factors necessary for growth (such as rainfall). *Article:* Dendroclimatology.

7.1.325 Lineation

The arrangement of chain silicates in parallel lines in certain metamorphic rocks formed under pressure: the pressure forces these silicates to orientate themselves at right-angles to the direction of pressure. *Article:* Metamorphic rocks.

7.1.326 Lithification

The conversion of sediment into a sedimentary rock by such processes as compaction and cementation. *Article:* Sedimentary rocks.

7.1.327 Lithosphere

The Earth's crust together with that portion of the mantle which, like the crust, is brittle and elastic rather than plastic and ductile. *Article:* Structure of the Earth.

7.1.328 Littoral

Having to do with the coast. *Article:* Index fossils.

7.1.329 Loess

Fine wind-borne sediment produced by the action of glaciers. *Article:* Ice cores.

7.1.330 Longshore bars

Bars of sediment running parallel to a beach. *Article:* Nearshore sediments.

7.1.331 Longshore current

The component of a nearshore current that flows parallel to the shoreline. *Article:* Nearshore sediments.

7.1.332 Longshore drift

The motion of sediment along the shore as a result of the fact that waves that approach the shore obliquely will

recede from it at right-angles to the shoreline. *Article:* Nearshore sediments.

7.1.333 Love waves

A type of seismic wave which travels along the surface of the Earth rather than through it. *Article:* Seismic waves.

7.1.334 Lu

Chemical symbol for the element lutetium. *Articles:* Chemistry for geologists, Other isochron methods.

7.1.335 Lu-Hf dating

Lutetium-hafnium dating, a form of radiometric dating. *Article:* Other isochron methods.

7.1.336 Mafic rock

A rock rich in magnesium and iron, and poor in quartz and feldspar. *Article:* Igneous rocks.

7.1.337 Magma

Molten rock which has not reached the surface, as opposed to lava. Some authors will define magma as any molten rock, in which case it would be proper to say that "lava is magma on the surface". However, in this text I have preferred the usage which makes magma and lava two distinct non-overlapping categories of molten rock. *Article:* Igneous rocks.

7.1.338 Mantle

A zone of ultramafic rock lying below the Earth's crust and above its core. *Article:* Structure of the Earth.

7.1.339 Mantle plume

A column of hot rock rising in the mantle below a hotspot. *Article:* Hotspots.

7.1.340 Marble

A metamorphic rock formed by the metamorphism of limestone. *Article:* Metamorphic rocks.

7.1.341 Massive rock

Rock which does not display bedding (in the case of sedimentary rocks) or foliation (in the case of metamorphic rocks), giving the rock a uniform and homogeneous appearance. This term is not used in our articles, so as to avoid confusion with the common use of "massive" to mean "very big"; we have instead used more transparent terms such as "unbedded". *Article:* Sedimentary rocks.

7.1.342 Mass wasting

Erosion caused by gravity. *Article:* Mechanical weathering and erosion.

7.1.343 Matrix

Finer material in which larger clasts (in the case of sedimentary rock) or crystals (in the case of igneous rock) are embedded. A synonym for groundmass. *Article:* Igneous rocks.

7.1.344 Meander

A broad loop in a stream or river. *Article:* Rivers.

7.1.345 Meandering stream

A stream which flows in a series of meanders. *Article:* Rivers.

7.1.346 Mechanical weathering

Weathering caused by mechanical processes that break up a rock, as opposed to chemical weathering. *Article:* Mechanical weathering and erosion.

7.1.347 Medial moraine

A moraine formed by the union of two lateral moraines when two valley glaciers flow together to form a single glacier. *Article:* Glaciers.

7.1.348 Metamorphic grade

The degree of metamorphism undergone by a metamorphic rock. *Article:* Metamorphic rocks.

7.1.349 Metamorphic rock

A rock which has had its texture or composition changed by heat and/or pressure. *Article:* Metamorphic rocks.

7.1.350 Metamorphism

Changes in the texture or composition of a rock brought about by heat and/or pressure. *Article:* Metamorphic rocks.

7.1.351 Metasomatism

Changes associated with contact metamorphism in which the parent rock mixes and/or reacts with the intrusive igneous rock and the hot fluids associated with its eruption. *Article:* Metamorphic rocks.

7.1.352 Methane

The gas CH_4 , a potent greenhouse gas. *Article:* Ice ages.

7.1.353 Mg

Chemical symbol for the element magnesium. *Article:* Chemistry for geologists.

7.1.354 Mica

A group of sheet aluminosilicate minerals. *Articles:* Igneous rocks, Metamorphic rocks.

7.1.355 Mid-ocean ridge

Elevated sea-floor on either side of a mid-ocean rift. *Articles:* Sea floor spreading, Sea level variations.

7.1.356 Mid-ocean rift

The rift between two plates at which sea floor spreading occurs. *Article:* Sea floor spreading.

7.1.357 Milankovitch cycles

Periodic changes in the inclination of the Earth's axis and the shape of its orbit. *Article:* Milankovitch cycles.

7.1.358 Mineral

A solid with a particular chemical composition and structure. *Articles:* Minerals, Silicate minerals.

7.1.359 Mineraloid

Anything which is like a mineral in some respects but doesn't quite fit the definition. *Article:* Minerals.

7.1.360 Mixing plot

A graph showing the composition of a rock on which the plotted points will fall in a straight line if the rock was produced by the mixing of different sources of magma. *Article:* Rb-Sr dating.

7.1.361 Mold

A fossil formed when sediment is packed around organic remains, which are then destroyed, leaving a void in the sediment in the shape of the remains. *Article:* Fossils.

7.1.362 Molecule

A collection of atoms bonded together. *Article:* Chemistry for geologists.

7.1.363 Moraine

A deposit of till. *Article:* Glaciers.

7.1.364 Mud crack

A small-scale geological structures produced in mud as it dries. *Article:* Way-up structures.

7.1.365 Mudstone

Rock formed from clay or silt which is not bedded: lithified mud which is bedded is known as shale. *Article:* Sedimentary rocks.

7.1.366 Mylonite

A rock produced at depth by the action of two sides of a fault rubbing against one another, having a distinctive "grain" indicating the direction of motion of the fault. *Article:* Faults.

7.1.367 Na

Chemical symbol for the element sodium. *Article:* Chemistry for geologists.

7.1.368 Nd

Chemical symbol for the element neodymium. *Articles:* Chemistry for geologists, Other isochron methods.

7.1.369 Ne

Chemical symbol for the element neon. *Articles:* Chemistry for geologists, Radioactive decay.

7.1.370 Nearshore

The zone in which the sea bed is affected by waves. *Article:* Nearshore sediments.

7.1.371 Neosilicate

A silicate mineral in which the silicate tetrahedra are isolated from one another. *Article:* Silicate minerals.

7.1.372 Neutron

A particle with no charge and approximately the same mass as a proton; together with protons, neutrons make up the nucleus of an atom. *Articles:* Chemistry for geologists, Radioactive decay.

7.1.373 Nonconformity

An unconformity in which the older rocks are igneous or metamorphic. *Article:* Unconformities.

7.1.374 Normal fault

A dip-slip fault in which the hanging wall moves downwards relative to the foot wall. *Article:* Faults.

7.1.375 Normal polarity

The polarity of the Earth's magnetic field as it is at present. (Note that there is nothing particularly normal about this state of affairs.) The opposite of reversed polarity. *Article:* Geomagnetic reversals.

7.1.376 Nucleus

A fragment of shell or stone around which an ooid forms. *Article:* Ooids and oolite.

or

The core of an atom, consisting of protons and neutrons. *Articles:* Chemistry for geologists, Radioactive decay.

7.1.377 O

Chemical symbol for the element oxygen. *Articles:* Chemistry for geologists, Banded iron formations.

7.1.378 Oasis

A small lake found in a desert. *Article:* Deserts.

7.1.379 Obduction

A process in which one plate colliding with another is thrust over it instead of beneath it; the opposite of subduction. *Article:* Ophiolites.

7.1.380 Oblique fault

A fault which combines elements of a dip-slip fault and a strike-slip fault. *Article:* Faults.

7.1.381 Obsidian

Felsic volcanic glass. *Article:* Igneous rocks.

7.1.382 Olivine

A silicate mineral with the chemical formula $(\text{Mg},\text{Fe})_2\text{SiO}_4$ and an orthorhombic crystal system. *Articles:* Minerals, Igneous rocks, Structure of the Earth.

7.1.383 Ooid

A small roughly spherical particle consisting of calcium carbonate layers formed around a nucleus of sand or shell. Required the action of waves for formation, and is therefore formed in shallow seas. *Article:* Ooids and oolite.

7.1.384 Oolite

Limestone composed of ooids cemented together. *Article:* Ooids and oolite.

7.1.385 Oolith

Either a rock formed from ooids, or a single ooid, depending on context. *Article:* Ooids and oolite.

7.1.386 Oolitic limestone

Synonym for oolite. *Article:* Ooids and oolite.

7.1.387 Opal

Amorphous hydrated **silica**, of which the precious stone known as opal is only one particularly pretty example. *Article:* Siliceous ooze.

7.1.388 Opal compensation depth

The depth at which **siliceous** material will dissolve faster than it is deposited; hence, the depth below which marine **chert** will not form. *Article:* Siliceous ooze.

7.1.389 Ophiolite

A section of oceanic crust which has been thrust up above sea-level. *Article:* Ophiolites.

7.1.390 Orthosilicate

A synonym for **neosilicate**. *Article:* Silicate minerals.

7.1.391 Orogeny

The formation of mountains; or the faulting and folding of a large area by lateral pressure; or the formation of mountains by this process. *Article:* Orogeny.

7.1.392 Os

Chemical symbol for the element **osmium**. *Articles:* Chemistry for geologists, Other isochron methods.

7.1.393 Oscillatory flow

The washing back and forth of water on the foreshore as a result of the action of the tide. *Article:* Nearshore sediments.

7.1.394 Outwash

Light **sediment** carried by **meltwater** from the ablation zone of a **glacier**. *Article:* Glaciers.

7.1.395 Outwash plain

A flat area of outwash sediment in front of a **glacier**. *Article:* Glaciers.

7.1.396 Oxbow lake

A crescent-shaped **lake** formed when a **meandering** stream changes its course, leaving one of its **meanders** cut off from the stream. *Article:* Rivers.

7.1.397 Oxygen catastrophe

Synonym for the **Great Oxygenation Event** *Article:* Banded iron formations.

7.1.398 Oxygen crisis

Synonym for the **Great Oxygenation Event** *Article:* Banded iron formations.

7.1.399 Oyster

A **reef-forming** bivalve. *Article:* Reefs.

7.1.400 Pa

Abbreviation for **pascals**.

or

The chemical symbol for the element **protactinium**. *Articles:* Chemistry for geologists, U-Th, U-Pa, and Ra-Pb dating.

7.1.401 Pahoehoe

A type of **lava** flow, or the cooled and solidified **rock** produced by it, characterized by a ropey and billowy surface texture. *Article:* Way-up structures.

7.1.402 Paleoclimatology

The study of ancient climates. *Article:* Paleoclimatology: introduction.

7.1.403 Paleocurrents

Ancient currents of wind and water the direction of which can be deduced from the analysis of **sedimentary rocks**. *Article:* Paleocurrents.

7.1.404 Paleomagnetic dating

A form of **absolute dating** based on analysis of the paleomagnetic data in the rocks. *Article:* Paleomagnetic dating.

7.1.405 Paleomagnetism

The geological record of the past history of the Earth's magnetic field. *Articles:* Geomagnetic reversals, Sea floor spreading, Paleomagnetic dating.

7.1.406 Paleosol

Fossilized soil. *Article:* Soils and paleosols.

7.1.407 Pangaea

The last supercontinent to exist, prior to its rifting and the formation of the Atlantic Ocean. *Article:* Continental drift.

7.1.408 Pannotia

A supercontinent that existed before Pangea. *Article:* Continental drift.

7.1.409 Paraconformity

An unconformity without an erosional surface. *Article:* Unconformities.

7.1.410 Parent isotope

A radioactive isotope which undergoes radioactive decay to produce a daughter isotope. *Article:* Radioactive decay.

7.1.411 Parent rock

The original rock from which a metamorphic rock is formed by metamorphism. *Article:* Metamorphic rocks.

7.1.412 Pascal

Unit of stress: 1 pascal = 1 newton/square meter. *Article:* Physical properties of rocks.

7.1.413 Pb

Chemical symbol for the element lead. *Articles:* Chemistry for geologists, U-Pb, Pb-Pb, and fission track dating, U-Th, U-Pa, and Ra-Pb dating.

7.1.414 Pb-Pb dating

Lead-lead dating, a form of radiometric dating. *Article:* U-Pb, Pb-Pb, and fission track dating.

7.1.415 Peat

Waterlogged and partially decomposed vegetable matter. Note that in geological usage peat does not just refer to gardeners' peat (formed from sphagnum moss) but to any vegetable matter that has undergone peatification. Peat is the sediment from which coal is formed. *Article:* Peat and coal.

7.1.416 Peatification

The partial decomposition of waterlogged vegetable matter, turning it into peat. *Article:* Peat and coal.

7.1.417 Pedogenesis

The process of turning sediment into soil by chemical weathering and the activity of organisms (plants growing in it, burrowing animals such as worms, the addition of humus etc). *Article:* Soils and paleosols.

7.1.418 Pelagic

Having to do with the open sea. *Articles:* Marine sediments, Index fossils.

7.1.419 Pelagic clay

Fine-textured sediment deposited on the abyssal plain. *Articles:* Pelagic clay, Marine sediments.

7.1.420 Peridotite

An ultramafic igneous rock consisting mainly of olivine with a little pyroxene and amphibole. *Article:* Igneous rocks.

7.1.421 Periodic table

A tabular arrangement of the elements which gives insight into their chemical properties. *Article:* Chemistry for geologists.

7.1.422 Permineralization

A process forming mineralized fossils in which the voids in the original material are filled by minerals. *Article:* Fossils.

7.1.423 Petrification

A process forming mineralized fossils in which they undergo both **replacement** and **permineralization**. *Article:* Fossils.

7.1.424 Phaneritic

An igneous rock is said to be **phaneritic** if the **crystals** in it are large enough to be seen with the naked eye. The opposite of **aphenetic**. *Article:* Igneous rocks.

7.1.425 Phase

Whether a substance is solid, liquid, or gas; if solid, its **crystal structure** or lack thereof. *Article:* Chemistry for geologists.

7.1.426 Phenocryst

A large **crystal** embedded in the more **finely-grained** ground mass of a **porphyritic** rock. *Article:* Igneous rocks.

7.1.427 Philosophical naturalism

The rejection *a priori* of the existence of the supernatural; a position completely unnecessary to the practice of geology. *Article:* Actualism.

7.1.428 Phylosilicate

A synonym for **sheet silicate**. *Article:* Silicate minerals.

7.1.429 Pillow basalt

Basalt with a distinctive shape consisting of a set of “pillows”; formed underwater as a result of the more rapid cooling of lava on contact with water. *Articles:* Igneous rocks, Ophiolites.

7.1.430 Pinstripe lamina

A very thin **lamina** of very fine **clasts**, formed in and characteristic of aeolian sand dunes. *Article:* Deserts.

7.1.431 Plastic

A material is said to be **plastic** if it does not recover from **stress**; that is, having been squeezed by **stress** into a given form, it retains that form when the **stress** is removed. The opposite of **elastic**. *Article:* Physical properties of rocks.

7.1.432 Playa

A flat-bottomed basin in a **desert** which periodically fills with water to form a shallow temporary **lake**. *Article:* Deserts.

7.1.433 Plutonic rock

Alternative term for **intrusive rock**. *Article:* Igneous rocks.

7.1.434 Plate

A piece of the **lithosphere** bounded by **faults**. *Articles:* Plate tectonics: overview, Sea floor spreading, Subduction, Hotspots, Terranes.

7.1.435 Plate tectonics

The study of the motion of the Earth’s **plates**. *Articles:* Plate tectonics: overview, Sea floor spreading, Subduction, Hotspots, Terranes, Ophiolites, Orogeny.

7.1.436 PMIP

The Paleoclimate Modeling Intercomparison Project, a project comparing results from **climate models** with the evidence from **paleoclimatic proxies**. *Article:* Climate models.

7.1.437 Point bar

A **bar** of **sediment** formed on the inner bank of a **meander**. *Article:* Rivers.

7.1.438 Polymorph

Two **minerals** having the same **chemical formula** but a different **crystal structure** are said to be **polymorphs**. *Article:* Minerals.

7.1.439 Pond

A small **lake**. *Article:* Lakes.

7.1.440 Porphyritic

Of a **rock**, containing some large **crystals** embedded in a more **finely-grained** ground mass. *Article:* Igneous rocks.

7.1.441 Precipitation

When a chemical formerly dissolved in water settles out of it as a solid sediment, this is called precipitation. *Article:* Saline giants.

7.1.442 Prehistoric

Preceding written human history; the fairly arbitrary line before which organic remains are considered to be fossils. *Article:* Fossils.

7.1.443 Primary rock

Term sometimes used for igneous rock. *Article:* Igneous rocks.

7.1.444 Principle of cross-cutting relationships

The principle that when one geological feature cuts through another, the former is the younger and the latter is the older of the two features. *Article:* Cross-cutting relationships.

7.1.445 Principle of faunal succession

Roughly speaking, the principle that if the fauna and flora in one location are found in one stratigraphic order, the same species will not be found in a different order in another location. *Article:* Principle of faunal succession.

7.1.446 Principle of least time

The principle in physics that a wave traveling through a medium will take the quickest route between two points. *Article:* Seismic waves.

7.1.447 Principle of original continuity

The principle that when sediment is laid down, it will extend continuously until either it meets an obstacle or tapers off with increasing distance from the source of the sediment. *Article:* Steno's principles.

7.1.448 Principle of original horizontality

The principle that when sediment is laid down, it is usually laid down more or less flat. *Article:* Steno's principles.

7.1.449 Principle of superposition

The principle that when sediment is laid down, the sediment most recently deposited will be on the top. *Article:* Steno's principles.

7.1.450 Proglacial lake

A lake fed by meltwater from a glacier. *Article:* Glaciers.

7.1.451 Progradation

The building out of a delta into the sea by deposition of sediment. *Article:* Deltas.

7.1.452 Proton

A positively charged particle of about the same mass as the neutron; together with neutrons, protons form the nuclei of atoms. *Articles:* Chemistry for geologists, Radioactive decay.

7.1.453 Proxy

A quantity which we can measure which bears a known relationship to a quantity that we can't measure but would like to; for example measuring past oxygen isotope ratios in shellfish as a substitute for measuring past temperatures. *Articles:* Paleoclimatology: introduction, Leaf shape and temperature, Scleroclimatology, U^{37}_{K} , TEX_{86} .

7.1.454 Pseudostratigraphy

A term used to describe a situation where rock is layered, but the layers do not represent successive deposition; for example, the layers found in an ophiolite. *Article:* Ophiolites.

7.1.455 Pumice

A form of volcanic glass filled with air bubbles. *Article:* Igneous rocks.

7.1.456 Pyroclastic flow

A current of air laden with volcanic ash, which resists dispersion into the surrounding air because of its greater density. *Article:* Volcanic ash.

7.1.457 Pyroxene

An important group of rock-forming chain silicates. *Article:* Igneous rocks.

7.1.458 P-waves

Body waves consisting of moving zones of compression and tension. *Article:* Seismic waves.

7.1.459 Quartz

A mineral consisting entirely of silicate tetrahedra in a lattice structure, so that each oxygen atom of each tetrahedron is shared with one other tetrahedron, giving quartz the chemical formula SiO_2 . *Articles:* Silicate minerals, Igneous rocks.

7.1.460 Quartzite

A metamorphic rock formed by metamorphism of quartz sandstone. *Article:* Metamorphic rocks.

7.1.461 Quartz sandstone

Sandstone of which the sand grains are almost entirely quartz. *Article:* Sedimentary rocks.

7.1.462 Ra

Chemical symbol for the element radium. *Articles:* Chemistry for geologists, U-Th, U-Pa, and Ra-Pb dating.

7.1.463 Racemic

Composed of equal amounts of both forms of enantiomers. *Article:* Amino acid dating.

7.1.464 Racemization

The process by which a collection of chiral molecules become racemic. *Article:* Amino acid dating.

7.1.465 Racemization dating

An alternative term for amino acid dating. *Article:* Amino acid dating.

7.1.466 Radioactive

Having a tendency to radioactive decay. *Article:* Radioactive decay.

7.1.467 Radioactive decay

Any process by which the composition of the nucleus of an atom is changed, such as alpha decay, beta decay, and electron capture. *Article:* Radioactive decay.

7.1.468 Radiocarbon dating

Radiometric dating of organic material by analysis of the isotopes of carbon it contains. *Article:* Radiocarbon dating.

7.1.469 Radiogenic

An atom is said to be radiogenic if it is the product of radioactive decay. *Article:* Radioactive decay.

7.1.470 Radiolaria

A group of single-celled organisms which produce tiny intricate tests, usually siliceous; these form a major component of siliceous ooze. *Article:* Siliceous ooze.

7.1.471 Radiometric dating

A collection of methods of absolute dating which depend on the constancy of radioactive decay rates. *Articles:* K-Ar dating, Ar-Ar dating, Rb-Sr dating, Other isochron methods, U-Pb, Pb-Pb, and fission track dating, Radiocarbon dating, Cosmogenic surface dating, U-Th, U-Pa, and Ra-Pb dating, Absolute dating: an overview.

7.1.472 Rain shadow

A dry area on the lee side of a mountain, caused by the tendency of clouds to burst on the stoss side. *Article:* Deserts.

7.1.473 Raleigh wave

A type of seismic wave which travels on the surface of the Earth rather than through it. *Article:* Seismic waves.

7.1.474 Rank

The degree to which coal has undergone metamorphism. *Article:* Peat and coal.

7.1.475 Ra-Pb dating

Radium-lead dating, a form of radiometric dating. *Article:* U-Th, U-Pa, and Ra-Pb dating.

7.1.476 Rb

Chemical symbol for the element rubidium. *Articles:* Chemistry for geologists, Rb-Sr dating.

7.1.477 Rb-Sr dating

Rubidium-strontium dating, a form of radiometric dating. *Article:* Rb-Sr dating.

7.1.478 Re

Chemical symbol for the element rhenium. *Articles:* Chemistry for geologists, Other isochron methods.

7.1.479 Reaction

In chemistry, a process in which molecules form, break apart, or recombine. *Article:* Chemistry for geologists.

7.1.480 Recrystallization

Change in the texture of a rock. *Article:* Metamorphic rocks.

7.1.481 Recumbent fold

A **fold** in a rock which has been pushed so far over as to lie on its side. *Articles:* Orogeny, Folds.

7.1.482 Redbeds

Sedimentary rocks cemented together chiefly by iron oxide, characteristic of dry climates. *Article:* Sediments and climate.

7.1.483 Red clay

A synonym sometimes used for pelagic clay. *Articles:* Pelagic clay, Marine sediments

7.1.484 Reef

An underwater ridge or mound formed from the calcareous shells of organisms (typically **coral** in the present day, but the term is not restricted to **coral** reefs). Note that the geological usage is more restricted than the nautical usage, in which a sandbar or rock sufficiently near the surface of the water to cause a hazard to shipping would also be considered a reef. *Article:* Reefs.

7.1.485 Reef limestone

Limestone resulting from the intact preservation of hard parts of **coral** or other organisms. *Article:* Reefs.

7.1.486 Reflection seismology

A method of examining the structure of buried **rocks** by studying the reflections of **Seismic** waves produced by artificial explosions. *Article:* Sea level variations.

7.1.487 Refraction

The change of direction undergone by a wave when it passes from a material which permits travel at one speed to a material which permits travel at another speed. A consequence of the principle of least time. *Article:* Seismic waves.

7.1.488 Regional metamorphism

Metamorphism over a wide region, caused by deep burial or wide-acting tectonic forces; as opposed to **contact** metamorphism. *Articles:* Metamorphism, Orogeny.

7.1.489 Regression

An event in which the shoreline moves in a seaward direction; the opposite of a **transgression**. *Articles:* Sea level variations, Walther's principle.

7.1.490 Relative dating

Dating methods which allow us to put **fossils** and/or rocks in order of age, but without telling us how old they are, as opposed to **absolute** dating, which does. *Articles:* Steno's principles, Principle of faunal succession, Index fossils, Geological column, Cross-cutting relationships, Igneous rocks and stratigraphy.

7.1.491 Re-Os dating

Rhenium-osmium dating, a form of radiometric dating. *Article:* Other isochron methods.

7.1.492 Replacement

A process forming fossils in which the original organic material is replaced by **minerals**. *Article:* Fossils.

7.1.493 Residence time

The average amount of time a given type of atom or molecule will spend in the ocean or in the atmosphere. *Article:* Radiocarbon dating.

7.1.494 Residual minerals

Minerals which are not dissolved by chemical weathering. *Article:* Chemical weathering.

7.1.495 Reversed polarity

A condition in which the north and south magnetic poles of the Earth were opposite in orientation to their present position. The opposite of normal polarity. *Article:* Geomagnetic reversals.

7.1.496 Reverse fault

A dip-slip fault in which the hanging wall moves upwards relative to the foot wall. *Article:* Faults.

7.1.497 Rhyolite

A felsic extrusive igneous rock; the extrusive counterpart of granite. *Article:* Igneous rocks.

7.1.498 Rhythmite

A sedimentary rock which display a repetitive vertical succession of types of sediment. *Articles:* Tidal rhythmites and dating, Varves, Milankovitch cycles.

7.1.499 Ring silicates

Silicate minerals in which the silicate tetrahedra are bonded together to form rings. *Article:* Silicate minerals.

7.1.500 Ripple

A very small dune. *Article:* Rivers.

7.1.501 Roche moutonnée

A hump of rock with one side shallow, polished, and striated and the other side steep and ragged, caused by a glacier flowing over the rock. *Article:* Glaciers.

7.1.502 Rock

An aggregate of one or more minerals or mineraloids. *Article:* Rocks.

7.1.503 Rock cycle

The set of processes by which rocks are formed, altered, destroyed, and reformed. *Article:* Rocks.

7.1.504 Rock flour

Extremely fine sediment formed by the grinding action of a glacier. *Article:* Glaciers.

7.1.505 Rock glacier

A glacier consisting mainly of rocks held together by ice. *Article:* Glaciers.

7.1.506 Rock salt

Common salt (NaCl) when it occurs naturally as a rock; a synonym for halite. *Articles:* Minerals, Saline giants.

7.1.507 Rounding

A clast is said to be rounded if its sharp edges and corners have been worn away by erosion. Note that the term does not imply that the clast in question is spherical or near-spherical, just that its shape is smooth. *Article:* Sedimentary rocks.

7.1.508 Rudaceous rocks

Term for conglomerates and breccias. *Article:* Sedimentary rocks.

7.1.509 Rudists

A group of reef-building molluscs that went extinct at the end of the Cretaceous period. *Article:* Reefs.

7.1.510 S

Chemical symbol for the element sulfur. *Article:* Chemistry for geologists

7.1.511 Sabkha

A salt flat found between the sea and a desert above the high-water line. *Article:* Deserts.

7.1.512 Saltation

The motion of a wind-blown or water-borne **clast** along the ground, river bed, sea bed, etc, by a series of short hops, when the particle is too large and the current too weak for it to be transported in suspension. *Articles:* Mechanical weathering and erosion, Rivers.

7.1.513 Salt flat

An accumulation of **minerals** on dry land by the evaporation of water containing dissolved **minerals**. While the commonest **mineral** in salt flats is indeed rock salt, other minerals such as **gypsum** may be deposited. *Article:* Deserts.

7.1.514 Sand

Particles of **sediment** between 1/16mm and 2mm in diameter. *Article:* Sedimentary rocks.

7.1.515 Sandstone

Rock formed by the **cementation** of sand. *Article:* Sedimentary rocks.

7.1.516 Saprolite

Completely **weathered** rock. *Article:* Chemical weathering.

7.1.517 Sapropelic coal

Coal where the original organic material comes from the deposition of algae in lakes; as opposed to **humic coal**. *Article:* Peat and coal.

7.1.518 Satelite Laser Ranging

A system in which ground-based observation stations measure the round-trip time of ultrashort pulses of light traveling to and from satellites. Used by geologists to measure plate motion and isostatic rebound. *Article:* Continental drift.

7.1.519 Schist

A high-grade metamorphic rock exhibiting pronounced **foliation**. *Article:* Metamorphic rocks.

7.1.520 Schistosity

The kind of **foliation** found in **schist**. *Article:* Metamorphic rocks.

7.1.521 Sclerochronology

An absolute dating method based on the study of growth patterns in shells and **corals**. *Article:* Sclerochronology.

7.1.522 Scleroclimatology

The study of past **climates** by the analysis of the composition of shells. *Article:* Scleroclimatology.

7.1.523 Sea floor spreading

The motion of two **plates** away from one another, producing a **rift** which is continuously filled by **magma**, producing fresh oceanic **crust**. *Article:* Sea floor spreading.

7.1.524 Seaload

The **sediments** carried by a wave. *Article:* Mechanical weathering and erosion.

7.1.525 Seamount

A marine mountain which is entirely underwater. *Article:* Hotspots.

7.1.526 Seat earth

The **paleosol** underlying **coal beds**. *Article:* Peat and coal.

7.1.527 Secular equilibrium

A condition in which the rate of production of a radioactive isotope in a rock is exactly balanced by the radioactive decay of the same isotope. *Article:* U-Th, U-Pa, and Ra-Pb dating.

7.1.528 Secular variation

The wandering of the magnetic poles over time. *Article:* Geomagnetic reversals.

7.1.529 Sediment

Particles transported and/or deposited by wind, water, glaciers, precipitation, etc; the constituents of **sedimentary rocks**. *Article:* Sedimentary rocks.

7.1.530 Sedimentary rock

Sediment lithified by cementation and/or compaction, or as a result of simple crystal growth in the case of evaporites. *Article:* Sedimentary rocks.

7.1.531 Sediment trap

A device that collects sediment as it settles. *Article:* Deposition rates.

7.1.532 Seismic tomography

The science of discovering the internal structure of an object (typically, the Earth) by studying the passage of body waves through it. *Articles:* Seismic waves, Structure of the Earth.

7.1.533 Seismic waves

Waves in the body or surface of the Earth generated by earthquakes. *Articles:* Seismic waves, Structure of the Earth.

7.1.534 Seismometer

A device for detecting earthquakes and measuring their properties. *Article:* Seismic waves.

7.1.535 Semidiurnal

Occurring twice daily. *Article:* Tidal rhythmites and dating.

7.1.536 Series

A stratigraphic unit smaller than a system but larger than a stage. *Article:* Geological column.

7.1.537 Serpentinite

A metamorphic rock produced from peridotite in the presence of heat and water. *Article:* Ophiolites.

7.1.538 Serrated margins

Edges of leaves which are not smooth, characteristic of a temperate climate. *Article:* Leaf shape and temperature.

7.1.539 Shale

A sedimentary rock formed from silt or clay which exhibits bedding. *Article:* Sedimentary rocks.

7.1.540 Shear

Stress that causes an object to skew, e.g. the stress that would deform a rectangle into a parallelogram. *Article:* Physical properties of rocks.

7.1.541 Sheeted dikes

Dikes which stand side by side with one another like books on a shelf rather than intruding into some other rock. *Article:* Ophiolites.

7.1.542 Sheet glacier

Synonym for continental glacier. *Articles:* Glaciers, Historical_Geology/Ice_ages.

7.1.543 Sheet silicate

Any silicate in which the silicate tetrahedra bond together to form a sheet. *Article:* Silicate minerals.

7.1.544 Shoestring

A geological feature which is long and thin, e.g. a river or a shoreline. *Articles:* Rivers, Nearshore sediments.

7.1.545 Si

Chemical symbol for the element silicon. *Articles:* Chemistry for geologists, Silicate minerals.

7.1.546 Siderite

The mineral FeCO_3 . *Article:* Banded iron formations.

7.1.547 Silica

Silicon dioxide, (SiO_2). This occurs in many forms, such as quartz, opal, and chert. *Article:* Minerals.

7.1.548 Silicate

A shorter way of saying silicate mineral. *Article:* Silicate minerals.

7.1.549 Silicate mineral

Any of a large and important class of minerals the chemistry of which is based on the **silicate tetrahedron**. *Article:* Silicate minerals.

7.1.550 Silicate tetrahedron

The ion SiO_4^{4-} , consisting of four oxygen atoms arranged around a silicon **atom** in a tetrahedron. Such units can link together with each other by sharing oxygen atoms at their corners to form a variety of structures including sheet silicates, chain silicates and **quartz**. *Article:* Silicate minerals.

7.1.551 Siliceous

Composed of **silica**. *Article:* Siliceous ooze.

7.1.552 Siliceous ooze

Ooze on the sea floor, composed of the siliceous test of **radiolaria** and **diatoms**. *Articles:* Siliceous ooze, Marine sediments.

7.1.553 Sill

A sheet of **intrusive** rock forced between strata. *Articles:* Igneous rocks, Igneous rocks and stratigraphy.

7.1.554 Silt

Clasts between 1/16mm and 1/256mm in diameter. *Article:* Sedimentary rocks.

7.1.555 Siltstone

Sedimentary rock composed of **silt**. *Article:* Sedimentary rocks.

7.1.556 Sinstral

A fault is said to be sinstral if someone standing on one side of the **fault** and looking at the other when there is motion along the **fault** would see the other side moving to the left. *Article:* Faults.

7.1.557 Skarn

A rock produced by **metasomatism**. *Article:* Metamorphic rocks.

7.1.558 Sky island

A mountain habitat which is home to species which are isolated by their inability to cross the drier hotter surrounding plain. *Article:* Ice ages.

7.1.559 Slab

The portion of a plate being thrust into the **atmosphere** during **subduction**. *Article:* Subduction.

7.1.560 Slate

A **metamorphic** rock formed by **metamorphism** of shale, exhibiting pronounced **foliation**. *Article:* Metamorphic rocks.

7.1.561 Slatey foliation

The sort of **foliation** found in slate. *Article:* Metamorphic rocks.

7.1.562 Slickenside

A smoothed and **striated** surface produced by the friction between the two sides of a **fault**. *Article:* Faults.

7.1.563 SLR

Abbreviation for **Satellite** Laser Ranging. *Article:* Continental drift.

7.1.564 Sm

Chemical symbol for the element **samarium**. *Articles:* Chemistry for geologists, Other isochron methods.

7.1.565 Sm-Nd dating

Samarium-neodymium dating, a form of radiometric dating. *Article:* Other isochron methods.

7.1.566 Snell's Law

A law relating the density of two mediums to the angle of refraction undergone by a wave when it passes from one medium to the other. *Article:* Seismic wave.

7.1.567 Soil

Sediment which has been altered by the effects of chemical weathering and the activity of organisms (plants growing in it, burrowing animals such as worms, the addition of humus etc). *Article:* Soils and paleosols.

7.1.568 Sole mark

A mark made in sediment when it is scoured by a current. *Article:* Turbidites.

7.1.569 Solid solution

A mineral in which some positions in the crystal lattice may be filled by different elements. *Article:* Minerals.

7.1.570 Soluble

Capable of becoming dissolved. *Article:* Chemistry for geologists.

7.1.571 Solute

A substance dissolved in a solvent. *Article:* Chemistry for geologists.

7.1.572 Solvent

The medium in which a substance is dissolved. *Article:* Chemistry for geologists.

7.1.573 Sorosilicates

Silicates in which the silica tetrahedra are bonded together in pairs. *Article:* Minerals.

7.1.574 Sorting

Sediment is said to be well-sorted if it consists of particles of about the same grain-size. *Article:* Sedimentary rocks.

7.1.575 Speleothem

A feature in a cave, such as a stalactite or stalagmite, formed by the precipitation of dissolved minerals, typically calcium carbonate. *Articles:* U-Th, U-Pa, and Ra-Pb dating, U-Pb, Pb-Pb, and fission track dating.

7.1.576 Spherical symmetry

A body (in this textbook, invariably the Earth) is said to be spherically symmetric with respect to some property if the value of that property at any given point in it depends only on the distance of that point from the center, and not on the longitude and latitude of the point. *Article:* Structure of the Earth.

7.1.577 Sr

Chemical symbol for the element strontium. *Articles:* Chemistry for geologists, Rb-Sr dating.

7.1.578 Stage

A stratigraphic unit smaller than a series but larger than a zone. *Article:* Geological column.

7.1.579 Stalactite

A speleothem hanging like an icicle from the roof of a cave. *Article:* U-Pb, Pb-Pb, and fission track dating.

7.1.580 Stalagmite

A speleothem in the form of a mound or column rising from the floor of a cave. *Article:* U-Pb, Pb-Pb, and fission track dating.

7.1.581 Standard

A rock of known age used in Ar-Ar dating. *Article:* Ar-Ar dating.

7.1.582 Steno's principles

The principles of superposition, of original horizontality, and of original continuity. *Article:* Steno's principles.

7.1.583 Step heating

A process used in Ar-Ar dating in which a rock sample is heated in steps of progressively higher temperatures. *Article:* Ar-Ar dating.

7.1.584 Strain

The deformation of a solid body as a result of stress. *Article:* Physical properties of rocks.

7.1.585 Stratigraphy

The study of strata, in particular their order of deposition. *Articles:* Actualism, Steno's principles, Way-up structures, Fossils, Principle of faunal succession, Index fossils, Unconformities, Faults, Folds, Geological column, Walther's principle, Cross-cutting relationships, Igneous rocks and stratigraphy

7.1.586 Stratum

A layer of **sedimentary rock** with distinctive mineralogical, structural, or fossil characteristics such that it can be distinguished from the strata above and below it. *Article:* Sedimentary rocks.

7.1.587 Stream-dominated deltas

Deltas with long **distributary** channels reaching seaward; deltas in which the most important factor in their formation is the river discharging via the **delta**. *Article:* Deltas.

7.1.588 Stress

The force per unit area exerted on a surface of a deformable body; also by extension the external pressure which creates the internal force. *Article:* Physical properties of rocks.

7.1.589 Striae

Synonym for striations. *Article:* Glaciers.

7.1.590 Striation

Grooves left by the movement of a glacier over a rock, parallel to the direction of motion. *Article:* Glaciers.

7.1.591 Strike-slip fault

A **fault** in which the blocks on either side of the **fault** move laterally but not vertically with respect to one another in a direction parallel to the fault. *Article:* Faults.

7.1.592 Stromatoporoids

Sponges which secrete a **calcium carbonate** skeleton and so were once important reef-forming organisms. While not actually extinct, they now live only in marginal habitats. *Article:* Reefs.

7.1.593 Subbituminous coal

Coal of a higher rank than **lignite** and a lower rank than **bituminous coal**. *Article:* Peat and coal.

7.1.594 Subduction

The motion of one **plate** beneath another and into the **mantle**. *Article:* Subduction.

7.1.595 Supercontinent

A landmass consisting of most or all of the continental crust joined together to form a single continent. *Article:* Continental drift.

7.1.596 Supercontinent cycle

A process in which **supercontinents** repeatedly form and then rift again into separate continents. *Article:* Continental drift.

7.1.597 Superposed river

A river which exists before the creation by erosion of the hills through which it flows. *Article:* Rivers.

7.1.598 Superimposed river

Synonym for superposed river. *Article:* Rivers.

7.1.599 Suspension

A form of transport of **clasts** by wind or water where the particles are carried above the ground, sea bed, river bed, etc. *Articles:* Mechanical weathering and erosion.

7.1.600 Suture zone

The line along which a continent becomes joined to another continent, microcontinent, or island arc. *Article:* Terranes.

7.1.601 Swamp

An area of waterlogged ground in which the water is shallow enough for land plants to grow. *Articles:* Peat and coal, Sediments and climate.

7.1.602 S-waves

Body waves consisting of waves of shear: that is, of displacement at right angles to the direction of travel of the wave, resembling the waves produced by shaking the end of a rope. *Article:* Seismic waves.

7.1.603 Syncline

Structure formed when rocks are folded downwards. *Article:* Folds.

7.1.604 System

The largest stratigraphic unit. *Article:* Geological column.

7.1.605 Tarn

A lake that forms in the former cirque of a glacier after the glacier has melted. *Article:* Glaciers.

7.1.606 Tectonic window

A place at which a rift in the Earth's crust allows us to see deeper into the crust than is normally possible. *Article:* Ophiolites.

7.1.607 Tectosilicate

A synonym for lattice silicate. *Article:* Silicate minerals.

7.1.608 Tension

Stress that produces elongation of a solid along the direction in which force is applied. *Article:* Physical properties of rocks.

7.1.609 Terminal lake

A lake which water flows into but not out of, usually salty as a result of the accumulation of dissolved minerals. *Article:* Lakes.

7.1.610 Terminal moraine

A moraine deposited at the ablation zone of a glacier. *Articles:* Glaciers, Ice ages.

7.1.611 Terragenic

Having an origin on land. *Article:* Marine sediments.

7.1.612 Terrane

Part of a landmass, bounded by tectonic faults, which is different in many ways from the main landmass to which it is attached. *Article:* Terranes.

7.1.613 Terrestrial

Having to do with the land. *Articles:* Sedimentary rocks, Lakes, Radiocarbon dating.

7.1.614 Test

The shell of a micro-organism such as a diatom or a foraminiferan. *Articles:* Calcareous ooze, Siliceous ooze.

7.1.615 TEX86

A temperature proxy based on measurement of the different varieties of GDGTs in sediments. *Article:* TEX₈₆.

7.1.616 Texture

Physical characteristics of a rock including crystal size (in igneous or metamorphic rocks), and clast size and the degree of sorting and rounding of clasts (in sedimentary rocks). *Articles:* Igneous rocks, Sedimentary rocks, Metamorphic rocks.

7.1.617 Theory of evolution

The explanation of the facts of evolution in terms of such mechanisms as mutation, recombination, lateral gene transfer, genetic drift, and natural selection. The explanation for the principle of faunal succession. *Article:* Principle of faunal succession.

7.1.618 Thermohaline circulation

Deep-water circulation driven by density differences in the temperature and salinity of sea water. *Article:* Climate models.

7.1.619 Thrust fault

A reverse fault in which the angle of the fault is more than 45° from the vertical. *Article:* Faults.

7.1.620 Tide-dominated delta

A delta in which the most important factor in its dynamics is the tide; characterized by the formation of offshore

bars running parallel to the direction of the tide. *Article:* Deltas.

7.1.621 Till

Unsorted and usually unbedded sediment deposited by a glacier. *Article:* Glaciers.

7.1.622 Tidal braking

Slowing of the Earth's rotation as a result of the tidal interaction between the Earth and the Moon. *Articles:* Sclerochronology, Tidal rhythmites and dating.

7.1.623 Tillite

The lithified equivalent of till. *Article:* Glaciers.

7.1.624 Topset beds

The flat beds of sediment deposited on the top surface of a delta. *Article:* Deltas.

7.1.625 Trace fossil

A fossil such as a footprint which is not *of* an animal but which was produced by one. *Article:* Fossils.

7.1.626 Transgression

An event in which the shoreline moves inland; the opposite of a regression. *Articles:* Walther's principle, Sea level variations.

7.1.627 Transitive

A relation is said to be transitive if when A stands in that relation to B, and B stands in that relation to C, then A stands in that relation to C. For example, the relation "is smaller than" is a transitive relation: if A is smaller than B, and B is smaller than C, then A is smaller than C. *Article:* Principle of faunal succession.

7.1.628 Trench

A depression in the sea floor formed along the line where one plate subducts beneath another. *Article:* Subduction.

7.1.629 Trona

An evaporite mineral having the chemical formula $\text{Na}_3(\text{CO}_3)(\text{HCO}_3)\cdot 2\text{H}_2\text{O}$.

7.1.630 Tsunami

A sea-wave caused by any high-intensity, short-duration submarine event, most usually an earthquake. Often colloquially and completely inaccurately known as a "tidal wave". *Article:* Actualism.

7.1.631 Tsunamite

Sediment deposited by a tsunami. *Article:* Actualism.

7.1.632 Tuff

Lithified Volcanic ash *Article:* Volcanic ash.

7.1.633 Turbid

Loaded with sediment. *Article:* Turbidites.

7.1.634 Turbidite

Rock formed from sediment deposited by a turbidity current. *Article:* Turbidites.

7.1.635 Turbidity current

A current which manages to keep from mixing with the medium through which it flows because, being turbid, it is denser than the surrounding medium. *Article:* Turbidites.

7.1.636 Turbidity sediment

Sediment deposited by a turbidity current. *Article:* Turbidites.

7.1.637 U

Chemical symbol for the element uranium. *Articles:* Chemistry for geologists, U-Pb, Pb-Pb, and fission track dating, U-Th, U-Pa, and Ra-Pb dating.

7.1.638 $\text{Uk}'37$

A proxy for temperature based on measurements of the different kinds of alkenones preserved in sediment. *Article:* $\text{Uk}'37$.

7.1.639 Ultrabasic rock

An obsolete and inaccurate term for ultramafic rock. *Article:* Igneous rocks.

7.1.640 Ultramafic rock

A rock which is extremely mafic; that is, particularly low in silicate tetrahedra and high in magnesium and iron. *Article:* Igneous rocks.

7.1.641 Unconformity

A surface between successive strata representing a period of erosion or of no deposition. *Article:* Unconformities.

7.1.642 Underclay

Synonym for seat-earth. *Article:* Peat and coal.

7.1.643 Uniformitarianism

An alternative term for actualism, not used in this textbook because of ambiguities and inconsistencies in its usage. *Article:* Actualism.

7.1.644 Unstable

Prone to radioactive decay. *Article:* Radioactive decay.

7.1.645 U-Pa dating

Uranium-protactinium dating, a form of radiometric dating. *Article:* U-Th, U-Pa, and Ra-Pb dating.

7.1.646 U-Pb

Uranium-lead dating, a form of radiometric dating. *Article:* U-Pb, Pb-Pb, and fission track dating.

7.1.647 Upper plane bed

A flat layered surface formed in a river bed when the river is travelling at too great a velocity to form ripples or dunes. *Article:* Rivers.

7.1.648 Uraninite

The mineral UO_2 *Article:* Banded iron formations.

7.1.649 U-Th dating

Uranium-thorium dating, a form of radiometric dating. *Article:* U-Th, U-Pa, and Ra-Pb dating.

7.1.650 Vail curve

A reconstruction of past variations of sea level based on the study of unconformities in the geological record. *Article:* Sea level variations.

7.1.651 Valley glacier

A glacier which has its accumulation zone on a mountain (typically in a cirque) and which flows down through valleys under both gravity and its own pressure; as distinct from a continental glacier. *Article:* Glaciers.

7.1.652 Valve

One of the two shells of a bivalve. *Article:* Way-up structures.

7.1.653 Varve

A lamina of coarse light sediment grading into fine dark sediment, often found in lakes fed by meltwater from a glacier and representing one year's deposition. *Articles:* Glaciers, Lakes, Varves.

7.1.654 Very Long Baseline Interferometry

A technique in astronomy involving widely separated radio telescopes observing the same object, such as a quasar. Used by geologists to measure the motion of tectonic plates by inferring the motion of the radio telescopes necessary to account for the data. *Article:* Continental drift.

7.1.655 Viscosity

Informally speaking, the reluctance of a liquid to flow; so for example maple syrup is more viscous than water. *Article:* Igneous rocks.

7.1.656 Vis plastica

The name of an imaginary force once thought to cause fossils to grow in rocks. *Article:* Fossils.

7.1.657 VLBI

Abbreviation for Very Long Baseline Interferometry. *Article:* Continental drift.

7.1.658 Volcanic ash

Fine debris formed when a volcano sprays out fine particles of **lava**. Note that the term “ash” is a misnomer, since volcanic “ash” is not a product of combustion. *Article*: Volcanic ash.

7.1.659 Walther’s principle

The principle that if **sediment A** is succeeded vertically by **sediment B** without an **unconformity** between them, then **sediment A** will also be succeeded horizontally by **sediment B** in some direction. *Article*: Walther’s principle.

7.1.660 Wave base

The greatest depth at which the action of a wave has any effect. *Article*: Nearshore sediments.

7.1.661 Wave ripples

Ripples in sand or other sediment caused by the action of the tide. *Article*: Nearshore sediments.

7.1.662 Wave-dominated deltas

deltas in which longshore drift forms barrier islands in front of the **delta**. *Article*: Deltas.

7.1.663 Way-up structure

A geological feature which enable us to discover which way up a **rock** was when it was originally formed. *Article*: Way-up structures.

7.1.664 Weathering

Processes which break up rock but do not themselves transport it, as distinct from erosion. *Article*: Mechanical weathering and erosion, Chemical weathering.

7.1.665 Weathering rind

The outer, **weathered** volume of a **rock** in which the outside has undergone **weathering** but the weathering process has not yet penetrated all the way through the rock. *Article*: Chemical weathering.

7.1.666 Welded tuff

Tuff which forms when a fall of **Volcanic ash** is still hot enough to weld itself together. *Article*: Volcanic ash.

7.1.667 Xenotime

The mineral YPO_4 , useful because it can be used in the radiometric dating of **sedimentary rocks**. *Article*: U-Pb, Pb-Pb, and fission track dating.

7.1.668 Y

Chemical symbol for the **element** yttrium. *Article*: Chemistry for geologists.

7.1.669 Zircon

The mineral ZrSiO_4 , useful for radiometric dating because of its resistance to erosion, weathering, and metamorphism. *Article*: U-Pb, Pb-Pb, and fission track dating.

7.1.670 Zone

The smallest stratigraphic unit. Can also be used in the usual informal sense of a region or area, as in the term “ablation zone”. *Article*: Geological column.

7.1.671 Zr

Chemical symbol for the **element** zirconium. *Article*: Chemistry for geologists.

7.2 Appendix B: Bibliography

This is a partial bibliography. I have not attempted to keep records of the innumerable sources I have read during the preparation of this textbook which told me either nothing I didn't know, or (much more frequently) nothing that I thought belonged in an introductory work.

It is almost certain that I have omitted some sources on which I have in fact relied; I can only apologize to the undeservedly anonymous authors of those valuable books and papers.

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7.3 Appendix C: Chemistry for geologists



Artificial crystal of the element bismuth.

In this article we shall take a look at some of the chemistry necessary to follow this or any other introductory textbook on geology.

Chemistry is of course a large and complicated subject which can hardly be addressed in an article such as this one. I have omitted all inessential details and probably several that a chemist would think are in fact essential; I have recklessly sacrificed accuracy to simplicity in a way

that would be quite unacceptable in the main body of this textbook. To get a full and accurate picture of chemistry, the reader will have to turn to a book devoted to chemistry; the objective of this appendix is to make sure that when chemical terms, notation, and concepts are introduced in this text, the reader will not find it written in a *completely* foreign language. More than that I have not attempted.

Some other less basic information on chemistry is introduced within the individual articles in this textbook as and when it becomes relevant.

7.3.1 Structure of an atom

An **atom** consists of protons, neutrons and electrons. **Protons** have an electrical charge of +1, **neutrons** have no charge, and **electrons** have a charge of -1. The number of protons in an atom is equal to the number of electrons, so the atom as a whole has no charge. The number of protons in an atom is called its **atomic number**, and atoms are classified into **elements** according to their atomic numbers: for example the element sodium consists of all atoms having an atomic number of 11.

The protons and neutrons in an atom sit together at the center of the atom in a cluster called the **nucleus**. The electrons form a cloud surrounding the nucleus. A proton or a neutron has a mass about 1836 times that of an electron. This means that the mass of an atom will be determined almost entirely by the sum of the protons and neutrons it contains; this is called the **atomic weight** of the atom.

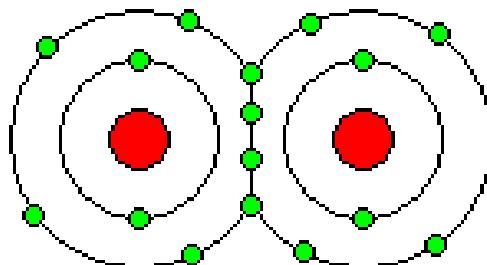
The electrons don't just buzz randomly around the nucleus like flies around a jam jar. Rather, they arrange themselves around the nucleus according to the laws of quantum electrodynamics. The reader will find these discussed in chemistry textbooks; to drag quantum theory into an introductory textbook on geology would be a step too far in reductionist explanation.

Instead, we shall have to content ourselves with the sort of explanation that you would have been offered before the discovery of quantum mechanics, and which is still offered to middle-schoolers; this should be quite sufficient for our purposes.

According to this scheme, the electrons are arranged in **electron shells** around the nucleus. Two electrons can fit into the innermost shell, eight in the second, eight in the third, eighteen in the fourth, and so on in a pattern which we shall have to leave unexplained since we're not going to talk about quantum theory. The electrons arrange themselves in these shells from the inside working out, so that the first shell must have its full quota of electrons before any will go into the second; the second shell must be full before any electron go into the third shell; and so on.

7.3.2 Chemical bonds and molecules

Metaphorically, we might say that an atom is “happiest” when out of all its electron shells that contain electrons at all, the outermost one contains as many electrons as will fit in the shell.



Covalent bonding of two oxygen atoms.

This means that atoms can come to a mutually beneficial relationship in which they share the electrons in their outer shell; this is known as a **covalent bond**. For example, consider the diagram to the right, showing such a bond between two oxygen atoms, with the nucleus represented in red and the electrons in green. Each atom on its own has six electrons in its outer shell, but by sharing electrons they have eight electrons apiece in their outer shells. The two atoms bonded together in this way form a **molecule**: in this case the molecule O_2 (this chemical notation will be explained below).

Alternatively, atoms can form **ionic bonds** in which instead of sharing electrons, one atom outright gives one or more of its electrons to another. So, for example, a sodium atom, with eleven electrons, would prefer to lose an electron, leaving it with eight electrons in what would then be its outermost shell, whereas a chlorine atom, with 17 electrons, would prefer to add an electron to its outermost shell, filling it up. If the sodium atom gives its unwanted electron to the chlorine atom, both are satisfied.

This leaves the sodium atom with a charge of +1, because it has one more proton than it has electrons, and the chlorine atom has one more electron than it has protons, giving it a charge of -1.

7.3.3 Chemical notation

Abbreviations for elements

Each element is represented either by a single capital letter (e.g. H for hydrogen, K for potassium, W for tungsten) or by a capital letter followed by a lower case letter (e.g. Cl for chlorine, Hg for mercury, Na for sodium).

Chemical formulas

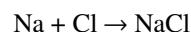
We can use these abbreviations to describe the com-

position of molecules: for example the combination of sodium and chlorine described above can be written as $NaCl$. (By convention Na is written first because the sodium has a positive charge.)

When there is more than one atom of a given element in a molecule, this is written by means of a number written in subscript to the right of the symbol for the element. So, for example, a molecule of water contains two atoms of hydrogen (H) and one of oxygen (O), and is written H_2O ; a molecule of methane has one atom of carbon (C) and four of hydrogen (H), and is written CH_4 .

Chemical reactions

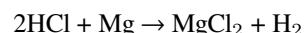
We can write chemical **reactions**, in which molecules form, break apart, and recombine, by using chemical formulas and the addition of a couple more symbols. For example, we can write:



to indicate that sodium and chlorine, to the left of the arrow, will combine to produce $NaCl$, on the right.

Note that the same elements, in the same quantities, appear on both sides of the arrow; in a chemical reaction no atoms are created or destroyed, only the nature of their relationship with one another are changed.

Another convention used in describing chemical reactions is to use full-sized numbers in front of the name of an atom or molecule to indicate how many of them enter into the reaction. So for example we can write:



Here the full-sized 2 indicates that there are two HCl molecules on the left-hand side. Note that the 2 refers to the entire formula HCl to the right of it, not just to the H immediately to the right of it.

Why don't we just write H_2Cl_2 instead of $2HCl$? Because H_2Cl_2 is not a molecule; rather we have two molecules of HCl , and the notation reflects that. On the other hand we write H_2 on the right hand side rather than $2H$ because H_2 is a molecule, and it would be inaccurate to write $2H$.

Ions

If an atom or a collection of atoms has lost an electron, as in the formation of ionic bonds, then it will have a positive charge, and is said to be a positively charged **ion**. We can represent this fact by a plus sign written as a superscript to the right. If instead it had gained an electron it would have a negative charge and would be described as a negative ion; we represent this by a minus sign written in the same place, above and to the right of the symbol for the element.

So for example we could if we wished write the combination of sodium and chlorine as Na^+Cl^- .

If an ion has lost or gained more than one electron, then we can represent this with a number written before the sign of the charge: so, for example, we could write Ca^{2+} to indicate an atom of calcium which has lost both of the electrons in its outer shell, giving it a charge of +2.

It is not compulsory to indicate the ionization. For example, no chemist would actually bother to write Na^+Cl^- , because any chemist would know that this is how sodium and chlorine *must* be ionized in order to form a molecule. However it is useful to have this notation as an option.

Chemical names

There are also conventional ways of giving molecules reasonably pronounceable names: so for example a chemist would look at the formula CaSO_4 and know to call it calcium sulfate. For our purposes it is not necessary to explain the details of this system: in the articles in this textbook I have simply presented the chemical name alongside the chemical formula.

7.3.4 The periodic table

If two elements have a similar situation in their outermost shells, then chemically one will behave much the same as the other. For example, sodium and potassium both have one electron in their outermost shells. This means that just as one atom of sodium can combine with one atom of chlorine to make NaCl , so can one atom of potassium combine with one atom of fluorine to make KCl .

Similarly fluorine (F) is in the same situation as chlorine: each of them is short one electron in their outer shells. This means that fluorine will combine with sodium or potassium just the same as chlorine will, forming NaF or KF , respectively.

This information about elements can be summarized by arranging the elements in the **periodic table**, shown below.

Group →	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18
↓ Period	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18
1	H																He	
2	Li	Be																Ne
3	Na	Mg																Ar
4	K	Ca	Sc	Ti	V	Cr	Mn	Fe	Co	Ni	Cu	Zn	Ga	Ge	Se	Br	Kr	
5	Rb	Sr	Y	Zr	Ta	W	Ts	Ru	Rh	Pd	Ag	Cd	In	Sn	Pb	Bi	Xe	
6	Cs	Ba	Hf	Ta	W	Ts	Os	Ir	Pt	Au	Hg	Tl	Fr	Uup	Fr	Fr	Fr	
7	Rf	Ra																Uuo
Lanthanides			La	Pr	Nd	Pm	Sm	Eu	Gd	Dy	Tb	Ho	Er	Tm	Yb	Lu		
Actinides			Th	Pa	U	Np	Pu	Am	Cm	Bk		Es	Fm	Md	No	Lr		

Elements in the same column are said to belong to the same **group**, and have similar situations in their outer electron shells. For example, the elements in the far left-hand column all have a single electron in their outer shells, which they are anxious to give away; meanwhile those in the far right-hand column all have full outer electron shells, and so will not form chemical bonds, since they

are perfectly happy the way they are.

The facts summarized in the periodic table are important to our understanding of various aspects of geology; for example in our discussion of paleoclimatology we shall make use of the fact that both magnesium and strontium can substitute for calcium; and in discussing U-Pb radiometric dating, it is important to know that uranium can substitute for zirconium in the formula ZrSiO_4 but that lead cannot.

The reader will note by looking at the periodic table that in that last example uranium (U) does not exactly lie in the same column as zirconium (Zr). However, for reasons which can hardly be explained without reference to quantum theory, there is more flexibility for elements colored pink in the table to substitute for one another: so uranium can stand in for zirconium, but (for example) calcium could not substitute for sodium despite the fact that they're closer together in the table than uranium is to zirconium.

7.3.5 Solvents and solutes

A substance (**a solute**) is said to be **dissolved** in another substance (**a solvent**) if it is mixed with it in such a way as to acquire the **phase** of the solvent (i.e. whether it is solid, liquid or gas, and its crystal structure if it is a solid).

Compare, for example, what happens when we add to water a substance which is not soluble in water, such as sand, and a substance which is soluble in water, such as table salt (NaCl). The sand remains a solid; all we have done is mix a solid with a liquid. What's more, it will not be evenly distributed in the water, but will sink to the bottom, being more dense than water; if it was less dense, it would float to the top.

By contrast, salt does dissolve in water, so what we get is not a solid (salt) in a liquid (water), but rather is simply a liquid, salty water; and the salt rapidly becomes evenly distributed throughout the water rather than sinking to the bottom.

Salt (NaCl) is an example of a molecule held together by an ionic bond, as we have noted above. Dissolved in water, however, the bond is also dissolved, as the positively charged sodium is attracted to the negative charge on the oxygen atoms in the water molecules, and the chlorine atoms are attracted to the positive charge on the hydrogen atoms. Instead of describing the dissolved salt as NaCl , it is more accurate to describe it as $\text{Na}^+ + \text{Cl}^-$; this is one occasion on which the notation for ions is useful.

7.3.6 Isotopes

Because the chemical behavior of atoms is determined by the interaction between their electrons, the key figure that determines the chemical properties of an atom is its atomic number; this tells us how many electrons it has,

and the number of electrons determines their arrangement in the electron shells. This is why to chemists the right way to classify atoms is into elements according to their atomic number.

However, we can make a finer distinction between atoms: two atoms which have the same number of electrons (and therefore protons) can have different numbers of neutrons and therefore different atomic weights. So for example we can distinguish between uranium-235 (uranium with 143 neutrons, 92 protons, 92 electrons, and an atomic weight of 235) and uranium-238 (uranium with 146 neutrons, 92 protons, 92 electrons, and an atomic weight of 238).

The notation used for isotopes is to write the atomic weight as a superscript to the left of the symbol for the element: so for example uranium-235 would be written as ^{235}U . (Some texts also write the atomic number as a subscript to the left of the atomic symbol; in this textbook I have not followed this convention.)

Because two different isotopes of the same element still *are* the same element, with the same arrangement of electrons, they have the same chemical properties. So just as one element can substitute for another if they are in the same group, it is even more the case that two isotopes of the same element will be chemically interchangeable. This fact is particularly useful to geologists, as many forms of radiometric dating would not be possible without it.

At this point it would be usual to explain nuclear decay and radioactivity; however, this textbook contains **an article exclusively devoted to the subject** in the chapter on absolute dating.

Chapter 8

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