EPSC 233: Earth and Life History

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Contents

Chapter 1

Introduction to Geology and the Earth System

1.1 The Science of Historical Geology

Historical geology is the study of the Earth's past. This includes the evolution of the Earth (including both its surface and interior) and the life that has inhabited it. It also includes events in the Earth's history, many of which significantly shaped the course of life, the biosphere, and more generally, the *Earth System*. It is virtually impossible to avoid Earth history if one work on rocks, because most rocks are old, and discerning their significance entails knowing how old and in what environment or context they were formed. Earth history simply integrates as much information as possible obtained from all the rocks on the surface of the Earth (and biological and chemical clues they contain) into a narrative of Earth's past. As you will appreciate, this is not an easy task and historical geology courses have no choice but to omit most of the details. Like most Earth history courses, a large component of this course will entail stepping through the geological time scale to understand how Earth formed then changed over the course of its 4.56 billion years. But first it will begin with introductions to the the sub-disciplines employed in studying Earth's past and tools used to interrogate it.

Why study Earth history? To be honest, probably the main reason people do is because it is fundamentally interesting. If you are curious about how the Earth came to be the way it us, then you need to know about its past. That is, the only way to appreciate the oxygen in the atmosphere, the diversity of life, and the present geography of the globe that we inhabit is to delve into Earth's past. Scientists more broadly interested in how planets evolve are best served to begin with our planet, where we have a rich rock record of its history. However, Earth history also has certain practical applications. In light of the current challenges humanity faces due to environmental deterioration and global warming, many Earth historians now rationalize their work in terms of the need to predict Earth's future. Although this argument is overplayed in my opinion, it is true that by studying, for example, past global warming episodes, we learn how the *Earth System* has responded to global warming, most importantly the rates at which it responded and recovered. It turns out there were multiple *hypothermal* episodes of sudden warming over the past 50 million years, many if not all of which were driven by rapid input of greenhouse gases into the ocean-atmosphere system. If we want to learn the consequences and rates of response to the current episode of fossil-fuel and land-use driven rise in CO_2, it is logical to begin by looking at past analogs.

The distribution and extent of petroleum and mineral deposits is also governed by past geological events. Consider the source rock (i.e., an organic-rich sedimentary rock) for a petroleum reservoir in a sedimentary basin. By understanding how and under what *depositional environment* this source rock formed, a petroleum geologist might better predict its distribution in the subsurface and extrapolate these results to other sedimentary basins where similar types of rock might be expected to have been deposited.

Earth history can be subdivided into three broad themes: *Deep Time*, *Plate Tectonics*, and the *Evolution of Life* (Levin, 2013). Compared to the human life span, geological time us incomprehensibly vast. Geology has taught us this, and now a important part of geology is dating and calibrating Earth's 4.56 billion-year-old history. Of particular interest is dating major catastrophic events and the the first appearance or occurrence of certain phenomena. Plate tectonics was not fully accepted by the geological community until the early 1970's, but now it is and integral component of virtually any study of past geological events (sedimentologists and palaeontologists ignore tectonics at their own peril!). And the origins of geology itself are deeply rooted in the study of the fossil record and the documentation and interpretation of the evolution of life. Logically, as we dig deeper into Earth's past, it becomes more and more difficult to divine its stories. The oldest minerals that formed on Earth (that is, after planetary accretion) are about 4.4 billion years old, and the oldest rocks are 4.0 billion years old. Hence, deciphering Earth's earliest history requires a combination of esoteric geochemical techniques, modelling, and an abundance of creativity.

1.1.1 Geology

Geology is the fundamental tool or approach that is used to study Earth's past. Geology encompasses the study of all types of rock found on Earth and on other planetary bodies. The three main rock types are *igneous*, *metamorphic*, and *sedimentary* and all contain important clues about past geological events. Igneous rocks, for example, indicate past thermal events that may have been related to subducting plates, rifting, hot spots, or other processes capable of melting the mantle or the crust. Studying metamorphic rocks can reveal key details about ancient mountain building events, hence continent-continent collisions. The sedimentary record, which comprises both sedimentary rocks and unlithified sediments, also contains important information about past thermal and tectonic events (the weathered detritus from volcanoes and mountain belts is ultimately deposited in sedimentary basins). Sedimentary rocks are also the only of the three major rock types to preserve fossils. Because the study of ancient life is effectively the core of historical geology, Earth History is based heavily on sedimentary geology and palaeontology.

1.1.2 Application of Scientific Method in Geology

Geology is frequently accused of being unscientific by scientists that do not fully understand what geologists do. But it is true that although experimentation is important in certain domains of the Earth sciences, it is not the basis of geology as it is in physics, chemistry, and much of biology. So where is the science in geology, in particular historical geology? They way to look at this problem is terms of testing hypotheses. Earth historians effectively formulate ideas about what happened in Earth's past. Ultimately, these models should offer the most parsimonious explanation of all available data and present concrete ways of being tested. For example, some researchers have argued that the Permo-Triassic extinction, like the Cretaceous-Paleogene extinction (formally known as the K-T), was likely triggered by a meteorite impact. Well, lots of other scientists disagree with this model, but it is testable, at least in theory, because meteorite impacts leave specific traces in the sedimentary record, including unusual accumulations of platinum group elements and unique organic chemicals derived from meteorites.

I find historical geology tricky because it so often lacks definitive answers. Show an outcrop of rocks to ten different geologists, and probably you will hear ten different explanations for why those rocks are there and what they signify. Likely as not, a few of those geologists will argue passionately about their hypotheses, even if there seem to be multiple viable explanations. As a consequence, it is reasonable to question any model that is put forward to explain Earth's past. You might ask why bother if geologists cannot be coerced to agree upon any hypotheses anyway. The answer is because as more data are collected and hypotheses proposed, geologists tend to converge on a given model or hypotheses, to the extent that it then gives way to a theory that is widely accepted by the scientific community.

1.1.3 The Principle of Actualism

The foundation upon which Geology is built is the principle of *Actualism*, which holds that fundamental chemical and physical processes do not change. James Hutton, who has earned the monicker of "Father of Modern Geology", and his great promoter Charles Lyell recognized that process on Earth operate slowly and gradually and surmised that these same processes occurred throughout the history of Earth, giving rise to the rocks and physiographic features we see today. This concept of *uniformitarianism* has come to dominate geological thought and is nicely encapsulated in the famous lines: "No vestige of a beginning, no prospect of an end" and "The present is the key to the past."

When it was proposed, the idea of uniformitarianism was radical because it contradicted the prevailing catastrophic theories for the Earth, with obviously immense implications for the age of the Earth and the legitimacy of the Bible. However, through Charles Lyell's persuasive arguments, the paradigm shift occurred and uniformitarianism has guided geological thought and research for much of the past two centuries. However, like any great new theory in geology, this one was flawed. Specifically, Hutton and Lyell envisioned that only those processes witnessed by humans occurred in the past and that all processes were gradual. Given the immense age of the Earth as compared to the short duration of historical records, this suggestion now seems rather absurd, and it is now well accepted that catastrophic events of the sort never witnessed by humans have occurred episodically throughout Earth history. Indeed, these episodic events, such as meteorite impacts, and massive volcanic eruptions, probably play an outsized role in shaping Earth's surface and the evolution of life and the environment.

Another important point on which Lyell erred was the idea that the nature of the processes and the rocks they formed have not changed since the Earth was formed. We now know, through the study of Earth history, that this is truly not the case. The increase in atmospheric oxygen concentrations, influences of life, and cooling of the interior of the Earth, to name but a few examples, have all led to unidirectional changes in geological processes and the types of minerals and rocks that form on Earth's surface.

1.2 The Earth System

It is in vogue these days to discuss Earth as a system. Indeed, our textbook is called *Earth System History*. So what does this mean? The Earth system concept is rooted in the appreciation that Earth is unique, and the conditions on Earth's surface that sustain life are the result of interaction and balance between various components, or systems, of the Earth, namely the hydrosphere, the atmosphere, the biosphere, the cryosphere the Earth's interior, its exterior, and climate. All of these systems can be subdivided into subsystems, which are also interrelated both to each other and to some lesser extent to subsystems of other systems. You might think of the Earth system concept as a more scientifically palatable version of the *Gaia* hypothesis, which holds that all organisms and their inorganic milieu are integrated to form a single, self-regulating system, not unlike how various cell types, organs, and bacteria in our bodies all work together and interact to sustain our lives.

Earth's systems are dynamic and none of them operates in isolation. That is, the systems are coupled. The habitability of the planet is ultimately maintained by *negative feedback loops* that regulate couplings between the systems. A *feedback* is a response to a perturbation or stimulation of a system. For example, if you deprive yourself of food for a long time, you will become thirsty, which might cause you to take a drink of water to alleviate that thirst. This is an example of a negative feedback. Negative feedback loops tend to diminish disturbances and to maintain *equilibrium states*.

Positive feedback loops occur when a disturbance results in a response that amplifies the disturbance. Unchecked positive feedbacks lead to unstable scenarios and can result in a *runaway*. For example, consider the situation with Arctic sea ice. It is both gradually thinning and decreasing in aerial extent, due to warming of the Arctic. The more this sea ice thins and disappears, the more the ocean can absorb and retain heat energy from the sun, which will make the Arctic warmer and melt more sea ice. Eventually, this feedback loop, if unchecked by some other negative feedback, will runaway, leaving the Arctic ocean sea ice-free.

Equilibrium states may be either stable or unstable. In a stable state, a minor disturbance will result in a response that returns the state to the original equilibrium. In an unstable state, a disturbance will result in a shift to a new equilibrium state or to disequilibrium.

Earth system history is simply an approach to interpreting Earth's history that appreciates the interconnectedness between the different components of the planet and the inescapable fact that Earth as we know it today is the result of its long and unique history. For example, mass extinctions have impacts on the oceans, the atmosphere and the biosphere, and may be triggered by internal Earth processes. Hence, if one wishes to understand mass extinctions, it is not enough simply to study the fossil record to determine the pattern of extinction and recovery. As you will see, mass extinctions are invariably linked to major carbon isotope excursions, which illustrates that the they represent perturbances to the global carbon cycle. Because Earth's climate is also coupled to the carbon cycle, *mass extinctions* might be expected to be associated with climate change–either as a trigger, a consequence, or both. Similarly, the chemistry of the ocean is controlled in part by climate and by the processes that remove CO_2 from the atmosphere over geological time. Hence, ocean chemistry also typically changes across mass extinction events. The coupled changes in the atmosphere, oceans, climate, and the biosphere during mass extinctions exemplify the behaviour of the Earth system and the necessity of considering the systems together when interpreting Earth's history, and its future.

From a somewhat philosophic perspective, Earth history has taught us that the Earth surface environment that we inhabit is both fragile and robust. On relatively short time scales, in can be strongly perturbed with dire consequences for the life that inhabits it. However, life is tenacious and none of these disturbances, from the snowball Earth to huge meteorite impacts, has been sufficient to eradicate it. Over long (non-human) time scales, the Earth system recovers from perturbations, including the current experiment in rapid, human-induced environmental change.

Chapter 2

Minerals and Rocks: The Building Blocks of Earth

Reading: Chapter 2 in *Earth System History* (Stanley, 3rd edition)

2.1 Introduction

Most of what we know about Earth's history comes from the study of rocks. The rocks are made up of minerals, which themselves comprise molecules that form from diverse combinations of elements. Hence, it is useful to review the basics of elements, chemical bonds, minerals, and rocks before jumping into Earth history, which presupposes a certain knowledge of rocks. But even before we begin that discussion, we need to talk about the basic structure of the Earth in order to frame the discussion of rocks.

2.2 Structure of the Earth

Geologists now have a reasonable grasp on the basic physical and chemical structure of the whole Earth. We owe much of this knowledge to *seismology*, through which geophysicists image Earth's interior, via a technique not unlike CT scans used by doctors to imagine the body. When an Earthquake occurs, it sends seismic waves across and through the Earth, the patterns of which are recorded on a global network of *seismographs*. Scientists use this data to pinpoint the location and magnitude of earthquakes, and as we'll see later, the collection of data on the foci of earthquakes was an important piece in the plate tectonics puzzle. Importantly, these seismic waves are also sensitive to temperature and whether the material they pass through is liquid or solid, and hence, reveal the structure of the interior of the Earth, which is one of abrupt boundaries between liquids and solids and denser and less dense layers. You will know doubt learn a lot more about seismology and seismic waves in another course.

The average radius of Earth is 6370 km. Earth can be subdivided into four physical layers:

- the solid iron+nickel inner core (6370–5150 km)
- the liquid iron+nickel outer core (5150–2900 km)
- the *viscoelastic*, silicate asthenosphere (2900 to 40–150 km)

- the elastic lithosphere

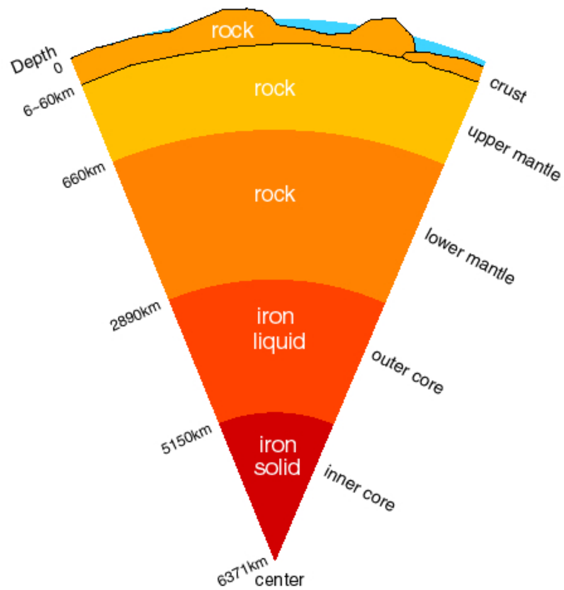


Figure 2.1: The components of Earth's deep interior.

The lithosphere is further subdivided into the *crust* and the *mantle asthenosphere*, where the *mantle* encompasses everything from the base of the crust to the top of the outer core (hence the terms mantle and asthenosphere are often interchanged, even though they are not exactly the same). The boundary between the crust and the mantle is chemically defined as is known as the *Mohorovicic discontinuity*—the *Moho* for short. Rocks below this boundary are significantly denser than the rocks above.

The crust is highly variable in thickness. It is thinnest beneath the deep oceans (7–10 km), where it is made up of basalt and has an average density of about 2.9 g/cm^3. Continental crust has is less dense ($\sim$ 2.7 g/cm^3), is compositionally similar to granite (on average), and varies from about 25 to 70 km in thickness. As you know, the continents stand above the oceans, and in mountain ranges, far above the ocean floor. The continental crust also extends well below the oceanic crust. This geometry is explained simply by *isostasy*, which is the gravitational equilibrium between the lithosphere and the asthenosphere (that is, think of the lithosphere floating on the asthenosphere). The oceanic crust-continental crust dichotomy yields a distinct *hypsometry* (Fig. 2.2), which sets Earth apart from the other planets and moons in the solar system.

2.3 Elements and Isotopes

In order to understand minerals, it is necessary first to step back and understand elements.

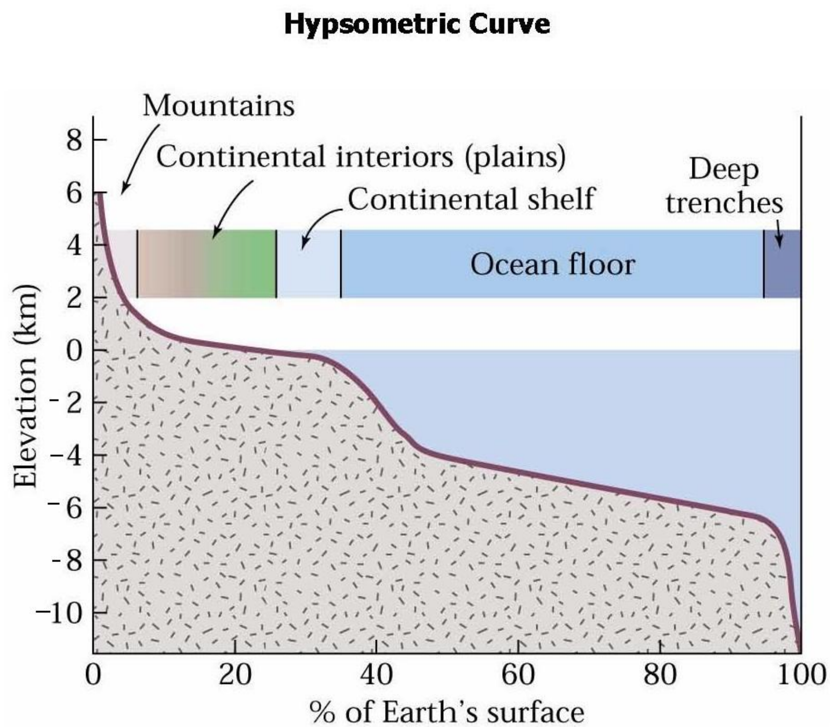


Figure 2.2: The hypsometric curve of Earth shows two distinct modes of elevation (or bathymetry) relative to see level: low continental interiors and the deep ocean floor. Mountains and trenches only cover a small part of Earth's surface.

2.3.1 Elements

An element consists of a unique kind of atom. Atoms comprise a nucleus of neutrons and protons(+), each of which has an atomic mass unit of 1, surrounded by *shells* of electrons(-), which have a minimal mass.

- Elements are distinguished by their number of protons (*atomic number*)
- Protons and neutrons roughly equal in nucleus
- Electrons form *shells* around the nucleus, and to a large extent, are responsible for the chemical behavior of an atom

2.3.2 Isotopes

Although the number of neutrons in the nucleus of a given atom is typically close to the number of protons, particularly in lower mass elements, the number of neutrons varies for many elements. So, for example, whose atomic number is 8, may have 8, 9, or 10 neutrons. Each type of oxygen atom (that is, with 8, 9, or 10 neutrons) is called an *isotope*. The chemical properties of different isotopes of the same element are identical, but due to their different masses, they react at slightly different rates. This phenomenon turns out to be extremely important and is the basis of *stable isotope geochemistry*, which will be discussed in more detail later in the course. Another important phenomenon is that some isotopes are inherently unstable, or *radioactive*. These isotopes decay to form daughter isotopes (which may or may not be stable), and in the process, emit energy in the form of radiation. As you know, radioactive isotopes are an important source of energy, including that for the rover Curiosity which is currently cruising around Gale Crater on Mars. But radioactive isotopes are also important as a tool for dating geological materials, as will be discussed in detail in a subsequent lecture.

2.3.3 Chemical bonds

Minerals are formed by way of chemical reactions, during which two or more atoms of one ore more elements combine to form a *molecule*, which is the basic until of a chemical compound. As a general rule, compounds behave differently than their constituent elements, because their size, electrical charge, and other characteristics are modified.

Ionic bonds are bonds that involve the transfer of electrons from one to another, typically resulting in stable electron shell configurations. This transfer results in opposite electric charges between the two atoms, and it is the attraction between these two oppositely charged *ions* that binds the two elements. The classic example of an ionic bond is NaCl (halite), where sodium gives up one electron from its outer shell and chlorine adds that electron to a shell that contained 7 electrons to round it out at eight. Ionic bonds commonly occur between metals and non-metals (i.e., between elements on the left and right-hand side of the period table, respectively).

Covalent bonds occur where two atoms share pairs of electrons. Common examples include O_2 and CO_2. The mineral diamond, in which carbon atoms share electrons on all sides, is an example of a covalently bonded mineral.

Ionic complexes form where two or more atoms bond but do not achieve charge balance. Rather, then retain a charge. For example, the carbonate ion in seawater: $(CO_3)^{2-}$.

2.4 Minerals

A *mineral* is a naturally occurring, inorganic, crystalline solid with a specific chemical composition and distinctive physical properties. Minerals require charge balance to form.

- Mineral compositions are shown by a chemical formula (e.g. quartz = SiO_2)
- Two different minerals may have the same chemical formula. For example, *calcite* and *aragonite* both have the chemical formula $CaCO_3$. The distinction is in the way the atoms are packed. Calcite tends to form stubby or tooth-shaped crystals wheres aragonite forms prismatic or needle-like crystals.
- *Native elements* are minerals, such as gold (Au) made up of a single element
- Elements commonly substitute for one another in minerals, most commonly when the two elements have similar atomic radii and ionic charges.

Table 2.1: Common mineral groups

Group	**(-) Ion** or **ion group**	**Example**	**Comments**
Silicates	$(SiO_4)^{4-}$	feldspar	main mineral group in most rocks
Carbonates	$(CO_3)^{2-}$	calcite	mainly sedimentary
Sulfates	$(SO_4)^{2-}$	anhydrite	mainly sedimentary
Sulfides	S^{2-}	pyrite	minor constituent of many rock types
Phosphates	$(PO_4)^{3-}$	apatite	minor constituent of igneous and sedimentary rocks
Halides	Cl^{-}, F^{-}	halite	significant only in sedimentary rocks
Oxides	O^{2-}	hematite	mainly sedimentary, but all rock types

  - e.g. feldspar has the composition $(K, Na)AlSi_3O_8$
  - olivine has the composition $(Mg, Fe)_2SiO_4$

- Most minerals form and are stubble under very specific conditions (e.g., pressure, temperature, pH, E_h)
- Ideal minerals (with time and space to grow as they like) will form perfect crystals which planar crystal faces, sharp corners, and straight edges
- Physical properties include
  - Crystal form
  - Cleavage
  - Hardness
  - Specific Gravity
  - Color, luster, streak, etc.

We are mainly interested in the *rock-forming minerals*, that is, minerals that are common constituents of rocks, such as quartz, feldspar, carbonate, mica, hornblende, etc. Most minerals fall into one of the mineral groups, defined by their negative ion or ion groups. A small number of minerals comprise the majority of rocks.

Accessory minerals are those minerals that occur in a rock but are not major constituents. Although they may only comprise a small proportion of a rock, they may be important economically or in determining the characteristics of the rock, such as its magnetic or chemical properties.

2.5 Rocks

I presume most everybody knows what a rock is, but if you want a simple definition: A *rock* is a cohesive amalgamation of minerals. That is, it hurts if someone hits you with it.

2.5.1 Igneous Rocks

Igneous rocks form from cooling *magma*, which consists of a mixture of molten material (liquid), gas (e.g. H_2O, CO_2), and minerals. Igneous rocks are crystalline, made up of interlocking minerals. As a general rule, igneous rocks that cool quickly are finer grained, and those that cool more slowly are coarser grained.

2.5.2 Extrusive igneous rocks

Volcanic (*extrusive*) igneous rocks solidify quickly on the surface from lava. They tend to be mainly *aphanitic*–that is, very fine grained. However, in some cases, minerals begin to crystallize in the magma chamber and are erupted along with the magma. Volcanic eruptions produce gases, steam, lava, and *tephra*, which is fragmented material generate by an eruption and includes ash.

- Lava flows
- Obsidian is volcanic glass that forms when a degassed magma cools extremely quickly
- Pumice forms where magmas rich in gases solidify rapidly and contain significant pore space
- *Pyroclastic* deposits consist of airborne volcanic material. *Tuffs* are the pyroclastic deposits consisting mainly of ash.

2.5.3 Intrusive igneous rocks

Plutonic (*intrusive*) igneous rocks solidify slowly below the surface and are mainly *phaneritic*—consisting of visible crystals. Plutonic rock bodies come in many shapes and sizes.

- Pluton is the general term for a larger body of intrusive igneous rock
- Dykes are vertical (relative to the land surface at the time of emplacement) sheets of intrusive igneous rocks
- Sills are horizontal sheets of intrusive igneous rocks

2.5.4 Classification of igneous rocks

Both extrusive and intrusive igneous rocks are distinguished by their chemical composition, namely by the abundance of SiO_2 and other key elements, such as Na, K, Fe, and Mg. However, there are also very important difference in trace elements between these different rock types. So-called *incompatible* elements, such as U and the rare earth elements (REEs) tend to be concentrated in felsic (i.e., more silica rich and Fe-, and Mg-poor melts), whereas *compatible* elements, such as Ni and Ti, are found preferentially in more mafic (silica-poor and Fe-, and Mg- rich) rocks.

The first step in making an igneous rocks is melting existing rocks. Melting is accomplished by one of three general processes:

- increasing temperature (e.g. mantle plumes)

Table 2.2: Classification of igneous rocks

	Felsic or	**Intermediate**	**Mafic**	**Ultramafic**
	SiO_2-rich			SiO_2-poor
	Na, K-rich			Fe, Mg-rich
Extrusive	rhyolite	andesite	basalt	komatiite
Intrusive	granite	diorite	gabbro	peridotite

- adding water (e.g. subduction zones)
- decreasing pressure (e.g. at mid-ocean ridges)

Melting of a given rock is usually not complete. That is, some residue is left behind. The result is segregation of elements, with the more compatible elements being left behind in the *restite*. A parcel of mantle that has not been previously melted is regarded as *fertile*, which means that it can be relatively easily partially melted and that melt will be relatively rich in incompatible elements. However, once a significant fraction of melt has been extracted from that parcel of mantle, it becomes *infertile*, meaning that it is difficult to melt and what does melt will be strongly depleted in incompatible elements.

2.5.5 Sedimentary rocks

Sedimentary rocks form through the lithification of sediment. *Sediment* includes unconsolidated minerals and rock fragments derived from the *weathering* and *erosion* of rocks and chemically precipitated minerals and shells.

- *weathering* is the physical breakdown or chemical dissolution of rocks and minerals. These two types of weathering are intimately linked.
- *erosion* is the transport of sediments
- *deposition* is the laying down and burial of sediments
- *lithification* is the process of turning sediments into rocks, usually through compaction and cementation of grains

Within the context of Earth history, sedimentary rocks are particularly important because they often preserve textures, structures, or chemical signatures of the environment in which they formed.

Detrital (or siliciclastic) sedimenatary rocks

Consist of fragments of other rocks which have then been transported and deposited

Chemical sedimentary rocks

Formed by chemical and biological precipitation of minerals.

- *Carbonates* (limestone = $CaCO_3$; dolomite = $CaMg(CO_3)_2$ may be inorganic or biogenic. These usually form in clear, shallow, warm waters, hence on tropical platforms away from rivers or other sources of clastic sediment.

Table 2.3: Classification of detrital sedimentary rocks

Sediment	Rock
gravel (grains > 2 mm)	conglomerate (rounded), breccia (angular), diamictite (grains in matrix
sand (1/16 - 2 mm)	sandstone (quartz sandstone, feldspathic sandstone, lithic sandstone)
silt (1/256 - 1/16 mm)	siltstone
mud (silt + clay)	mudstone or shale (finely laminated)
clay (< 1/256 mm)	claystone
mud matrix + sand and/or gravel	wacke

- *Evaporites*, including halite (salt) and gypsum ($CaSO_4$)•$2H_2O$ form by evaporation of concentrated solutions. Significant evaporite deposits usually form in the subtropics.
- *Chert* consists of of microcrystalline quartz. Whereas most marine chert that has formed in the past few hundred millions of years is derived from microscopic, siliceous shells (diatoms or radiolaria) or sponge spicules, older cherts were most likely inorganic (although their formation may have been biologically mediated in cases).
- *Banded iron formation* or BIF, is layered, iron-rich rock
- *Coal* is the compressed and altered remains of land plants

2.5.6 Metamorphic rocks

Metamorphic rocks form through the solid state alteration of other rocks (igneous, sedimentary, or metamorphic) due to changes in temperature, pressure, and/or fluids. Metamorphism involves the slow crystallization of new minerals that are thermodynamically stable under the changed temperature/pressure/fluid conditions. For example, the minerals muscovite and quartz react under *amphibolite* facies (see below) pressure-temperature (P-T) conditions to form orthoclase, sillimanite, and water (Fig. 2.3).

$$KAl_3Si_3O_{10}(OH)_2 + SiO_2 = KAlSi_3O_8 + Al_2SiO_5 + H_2O \tag{2.1}$$

- *Contact metamorphism* is driven by heat (± fluids) from an adjacent igneous body
- *Burial metamorphism* occurs through the depositional burial of sedimentary and volcanic rocks
- *Regional metamorphism* effects large areas that are buried and heated, usually in mountain-building events
- *Hydrothermal metamorphism* is the result of interaction of rocks with hot, percolating fluids (most commonly on mid-ocean ridges)
- *Shock metamorphism* occurs when an asteroid impacts Earth

Metamorphism often occurs under differential pressures (i.e. mountain belts), which results in the development of distinct textures which, in combination with the *mineral assemblages*

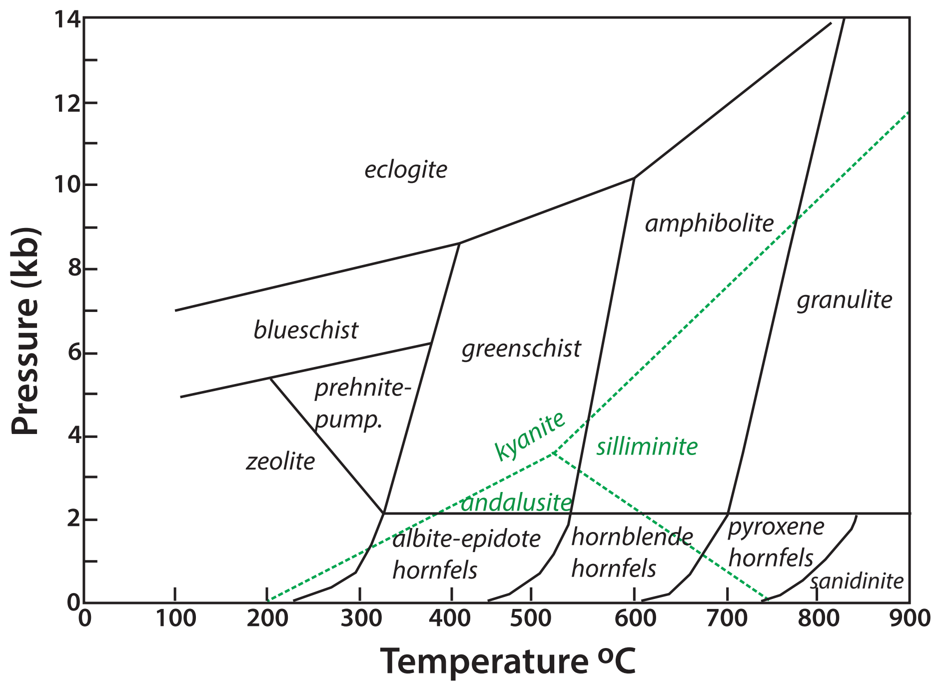


Figure 2.3: Metamorphic facies plotted in P-T space, along with several key metamorphic reactions. Note that metamorphic reactions can proceed both directions: *prograde* under increasing P-T conditions, and *retrograde* under decreasing P-T conditions.

can used to reconstruct the geological evolution of these rocks.

Many metamorphic rocks are *foliated*, resulting from the preferential orientation of platy and elongate minerals under differential pressure conditions. The progressive metamorphism of a mudstone or claystone will yield:

- *slate*, a low-grade metamorphic rock with fine-grained minerals aligning to form a platy *cleavage* that makes the rock *fissile*
- *schist*, a medium-grade metamorphic rock with larger minerals (mainly mica) that align to form a *schistose foliation*
- *gneiss*, a high-grade metamorphic rock where minerals have intergrown in dark and light wavy bands form
- *migmatite* a very high-grade metamorphic rock which has started to melt

Other metamorphic rocks include

- *Marble*: metamporphosed limestone
- *Quartzite*: metamorphosed quartz sandstone where the quartz grains are interlocking
- *Greenschist*: metamorphosed, dark green rocks of mafic igneous origin

- *Blueschist*: high pressure, moderate temperature, altered basalt (in subduction zones)
- *Eclogite*: very high pressure, moderate–high temperature
- *Granulite*: formed under extremely high temperature conditions

Chapter 3

Plate Tectonics

Reading: Chapters 8–9 in Stanley

3.1 Introduction

Plate tectonics describes the motion of Earth's lithosphere, which is broken into a series of plates that move relative to one another. Plate tectonics is driven by mantle convection and explains many distinct observations that the continents drifted across the globe over geological time, episodically crashing into one another. A half a century into acceptance of plate tectonics, it is now almost baffling to think that theory took so long to take hold – one only has to look at a bathymmetric map of the oceans to see the work of plate tectonics in all of its glory.

3.2 Continental Drift

In 1912, Alfred Wegner, a German meteorologist, proposed the hypothesis of continental drift to explain massive evidence the he and previous scientists had assembled that Earth's continents had once been part of a single continent (*Pangea*) that had subsequently broken apart. British geologist Arthur Holmes and South African geologist Alexander du Toit provided additional geological and paleontological evidence for continental drift and the ancient supercontinent.

- Geography: the shorelines of the continents (in particular eastern South America and western Africa) fit together like pieces of a puzzle
- Correlation: sedimentary sequences from many continents now separated from each other were very similar in composition and age
- Fossils: Many fossils of extinct organisms from scattered continents converged to a single geographic zone when the continents are restored to their supercontinental configuration
- Paleoclimate: Indicators of ancient ice flow directions related to similar aged glacial deposits, including in many areas currently outside of the polar latitudes, all consistent with a large ice sheet centered on the southern part of the supercontinent. A similar pattern is seen in northern hemisphere coal seems.

Although a small subset of geologists were convinced by continental drift, most geologists harshly opposed Wegener's hypothesis. The prevailing hypothesis to explain the major structural features on Earth was through vertical motions in the crust, driven by a cooling and contracting Earth. One of the main reasons typically offered for this refusal of what in hindsight seems like incontrovertible evidence is that Wegener was not able to furnish a plausible model for why the continents should drift. Wegener died on the Greenland ice sheet in 1930, without having converted many geologists to mobilism. Arthur Holmes subsequently offered a mechanism (mantle convection), but he was a lone voice in the wilderness, and it was not until two decades later that geophysicists furnished enough evidence in support of continental drift that plate tectonics began to be accepted. It would take longer for many geologists to come around to plate tectonics.

3.3 The Plate Tectonic Revolution

Starting in the late 1940s and continuing through the 1960s, many disparate pieces of evidence and new hypotheses began to emerge that supported Wegener's theory of continental drift. Many *fixest* geologists held out against continent drift. Finally, in 1962, Harry Hess published a paper in which he proposed that the entire crust moves, with new oceanic crust being made at *mid-ocean ridges*, then spreading away like a conveyer belt. A year later, PhD student Fred Vine and Lawrence Morley separately[1] found robust evidence for sea-floor spreading in the form of magnetic stripes on the seafloor, providing a resoundingly positive test of plate tectonic theory. The principle pieces of the puzzle they eventually combined to demonstrate plate tectonics include:

- Apparent polar wander
- Magnetic reversals on the seafloor
- The lithology, age, and bathymetry of the ocean basins
- Transform faults
- The location of earthquakes along the mid-ocean ridges and oceanic trenches

3.3.1 Paleomagnetism

Early evidence support plate tectonics and the vindication of the sea floor spreading hypothesis were accomplished through *paleomagnetism*, the study of the record of Earth's magnetic field as recorded in rocks. Magnetic minerals (mainly iron oxide minerals) tend to align with the Earth's magnetic field when they crystallize (igneous rocks) or when they are deposited (sedimentary rocks). Hence, in theory, if we can measure the orientation of the magnetic field locked in rocks, we can determine the nature of the magnetic field at the time they were deposited. In practice, this is more difficult because the magnetic signature in many rocks has been overprinted during tectonic, hydrothermal, and other events (even lightning strikes!). But in many places, the original magnetic signal remains, and we can

[1]Morley, a geologist, submitted his findings first, but they were rejected by both *Nature* and *Journal of Geophysical Research*. Later that year, Vine and Drummond Matthews (Vine's PhD supervisor) published their results in *Nature*. The consequence was that for thirty years or so, Vine and Matthews received much of the credit for the positive test for plate tectonics. Now that credit is deservedly shared with Morley.

measure both the *declination* (the deviations from true north) and *inclination*, the tilt of the magnetic mineral. This inclination reflects latitude because magnetic field lines are normal to the magnetic poles and curve around earth, such that if you use a magnetic compass, the needle will be perfectly horizontal at the (magnetic) equator and vertical at the magnetic poles. However, the magnetic north and south poles do not align perfectly with the true axial north and south poles. Furthermore, the magnetic poles migrate about the polar regions. The upshot of this is that over timescales of tens of thousands of years, these wayward magnetic poles average out to be the same as the true poles, a phenomenon known as the *geocentric axial dipole*. By the early 1900s, the French scientist and Bernard Brunhes and the Japanese geophysicist Matonari Matuyama recognized that two polarities were preserved in rocks, suggesting that the earth's magnetic field had reversed in the past.

In the 1950's, paleomagnetists noticed that through successively older rocks, the magnetic pole (even averaged out) seemed to drift significantly away from the geographic poles. This pattern was consistent with continental drift, but the entrenched opposition to continental drift led scientists at first to suggest that the magnetic pole wandered, i.e., *apparent polar wander*. However, magnetic evidence mounted against this hypothesis. First, the apparent polar wander paths from different continents, while suggesting wander, were different. Instead, the paths are consistent with continents drifting apart from one another, such that when that drift is accounted for and restored, the apparent polar wander paths converge. So in fact, these apparent polar wander paths can be used to deduce when continents collide and split up.

The other great victory of paleomagnetism came in the aforementioned magnetic stripes on the seafloor. Because Earth's magnetic field flips from time to time, the polarity of magnetic minerals in rocks also changes. Harry Hess had proposed that new ocean crust was formed at the mid-ocean ridges and then drifted away as if on a conveyor belt. The prediction then is that strips of oceanic crust parallel to the mid-ocean ridge should have either normal or reversed polarity in their magnetic fields. Vines, Morley, and others observed evidence for this be measuring the magnetic field across north-south oriented mid-ocean ridges. What they found were a series of valleys and ridges in field intensity. The ridges were positive anomalies resulting from the addition of the magnetic field from the rocks to the Earth's magnetic field (normal polarity), and the valleys were negative anomalies resulting from the subtraction of the magnetic field from the Earth's magnetic field (reversed polarity).

3.3.2 The ocean basins

The ocean basins are underlain by oceanic crust and mantled in sediments. Harry Hess pointed out that if the ocean basins were as old as the Earth, then even assuming a minimal sedimentation rate, they should be deeply buried in sediments (up to, say, at least 15 km). Not only are they not covered in so much sediment, this mantle of sediments thins towards the mid-ocean ridges, which is consistent with the mid-ocean ridges being young and the oceanic crust being no more than several hundred million years old. Various other lines of evidence, including the profile of the mid-ocean ridges and the formation of *guyots*, supported the notion that new oceanic crust is made at the ridges and then carried away, as if on a conveyor belt.

Dating of seafloor has beautifully confirmed Harry Hess's original hypotheses. The youngest and most buoyant oceanic crust is invariably found along the mid-ocean ridge. The oldest oceanic crust in the open oceans is about 180 million years old. Where plate boundaries cut across seafloor time lines, that seafloor is being destroyed–in subduction zones.

3.3.3 Transform faults

Canadian geologist and geophysicist J. Tuzo Wilson saw positive evidence for seafloor evidence in the *transform faults* that occur ubiquitously in the ocean basins and offset the mid-ocean ridges. It turns out that the relative motion on the segment of the faults between mid-ocean ridges is the opposite of the apparent sense of offset between the mid-ocean ridge segments. Beyond the mid-ocean ridge segments, there is no relative motion on these faults, although the seafloor on either side is at a different depth. These patterns can only be explained by new oceanic crust forming at the spreading ridges and subsequently drifting away from the ridges.

The motion of a rigid plate on a sphere is defined by identifying its axis of rotation and angular velocity. The axis of rotation, known as an Euler pole, intersects the surface of the Earth in two places. It turns out that we can use the transform faults to identify the Euler pole for the relative movement of oceanic crust on either side of a mid-ocean ridge by drawing lines tangent to the transform faults: the intersection of these tangent lines is the Euler pole.

3.3.4 The location of earthquakes

The plate boundaries on Earth can be identified by looking at a map of earthquake sources, because these are tectonically active zones. The Japanese geophysicists Kiyoo Wadati recognized in the 1930's that near the deep ocean trenches (the zones of the deepest seafloor on Earth), these earthquakes are also deep and they occur along a line that is inclined away from the trench. Hugo Benioff of CalTech also recognized these earthquake zones and was the first to hypothesize that these were zones of descending seafloor—that is, subduction zones.

3.3.5 The Hawaiian-Emperor Seamount

J. Tuzo Wilson also proposed the first way to measure the absolute motion of plates. He noted that the age of the Hawaiian-Emperor volcanic-seamount chain increased away from present day Hawaii, where the volcanoes are still active. This chain is kinked, and the age of the seamount at that kink is about 43 million years. The Emperor chain disappears at the Alleution trench, where the seamounts are 75 million years old. This chain of volcanoes formed above a *hot spot*, which is currently under Hawaii. *Hot spots* are the surface expression of vertical temperature anomalies in the mantle, which reflect narrow plumes of up going asthenosphere that generate high temperatures melts resulting in volcanism at the Earth's surface independent of tectonic boundaries. The chain of volcanoes that has drifted across a hot spot is known as a *hot spot track*, and the ages and distances of this hot spot track can be used to calculate the absolute plate motion (direction and velocity) of the Pacific Plate in this case, assuming that the hot spot has remained in the same place. The kink in the hot spot track 43 million years indicates a wholescale shift in the absolute plate motion

3.4 An overview of plate tectonics

In the simplest of terms, the lithosphere moves laterally as a result of convection within the mantle, itself a consequence of the Earth cooling itself off. The lithosphere consists of a finite number of rigid *plates* that move relative to one another and whose boundaries are tectonically active: convergent, divergent, and transform boundaries. Ocean floor is relatively young, is created at mid-ocean ridges (divergent), and is destroyed in subduction zones (convergent).

3.4.1 Folds and faults

The most obvious manifestation of plate tectonics is in the form of folding and faulting of rocks in the lithosphere. Virtually all rocks have experiences some deformation, and knowledge of at least the basic elements of *structural geology* is necessary if we are going to discuss the geological record. We can map out the occurrence of folds and faults by measuring the *attitudes* of beds and fault surfaces: that is, their orientation. The *strike* of a planar surface is defined as the azimuthal orientation (in degrees) of the intersection between a planar surface and the surface of the earth. The dip of the surface is defined as the angle between it and the horizontal earth surface, such that your bed should dip at $0°$ and your walls at $90°$. The *dip direction* is geometrically constrained to be $90°$ from the orientation of the strike. By convention, we define the strike direction such that we arrive at the dip direction by rotating $90°$ clockwise from the strike. This is known as the *right-hand rule.*

Folds

Most folds form in convergent environments due to compressive forces, just as you might fold a phonebook (wait, what are those?) by pressing on its two sides. *Synclines* are bowl-shaped folds, where layers slope downward on both sides to a low point, which lies on the *fold axis. Anticlines* are roof-shaped folds where layers inline upwards from the two sides and converge at a high point. Most synclines air paired to an anticline and vice versa. In map view, synclines are distinguished as displaying younger strata in their cores, wheres anticlines display older strata.

The fold axes are effectively the horizontal trace of a fold hinge on the earth surface. Most folds are not perfectly upright, but rather are tilted such that the *hinge lines* actually tilt. The orientation of these hinge lines is called the plunge, which can be measured and recorded in the same way as strike and dip, but where the dip is measured in the same direction that the hinge is plunging. In map view, a syncline opens in the direction of plunge, whereas an anticline closes.

Faults

Faults can be classified as *dip-slip*, *strike-slip*, or *oblique*, where the slip direction is not the same as either the strike or dip direction. Whereas many faults have an oblique component, we tend to describe them as either dip-slip or strike-slip.

Normal faults are faults that involved down-dip slip, such that the *hanging wall* drops down relative to the *foot wall.* These are also referred to as *extensional* faults and they juxtapose younger rocks above on older rocks below (with intervening strata removed locally). The total vertical displacement on a normal fault is known as the *throw*, whereas the horizontal displacement is known as the *heave*.

Reverse faults are faults in which the hanging wall moves upward relative to the footwall. These juxtapose older rocks on top of younger rocks and so have the effect of duplicating stratigraphy. *Thrust faults* are simply low angle reverse faults. These may have heaves of many tens or even hundreds of kilometres.

Strike-slip faults are vertical faults where rocks slide past each other horizontally.

3.4.2 Divergent boundaries

You now know that new oceanic crust is made at mid-ocean ridges, or *spreading ridges*. These spreading ridges are plate boundaries, because the plates on either side are moving apart from each other; that is, they are diverging. These spreading ridges form in the middle of the oceans basins stand out as underwater mountain ranges because they are buoyed up by hot, upwelling asthenosphere below. As this new oceanic crust moves away from the spreading center, it gradually cools, and hence 'sinks'. The result is a depth to ocean floor which is proportional to $\sqrt{age}$.

New oceanic crust formed at spreading ridges has a typical depth profile: the top is sediments, followed by *pillow basalts*, sheeted dikes, and layered gabbro. The base of the layered gabbro is the base of the crust (the Moho), and is so underlain by mantle.

Continental rifting

Divergence can also occur in continental settings. *Rifting* is the spreading apart of continental crust. It is the result of tension on the continental crust, which thins the crust. In the upper crust, this thinning is *brittle*, meaning it is accomplished by a series of faults known as *normal* faults, where one side (the *hanging wall* slips down relative to other side *foot wall*. At depth, the thinning occurs in a ductile fashion, meaning that the crust is pulled apart like taffy.

This normal faulting gives rise to a series of fault bound *rift basins*, where the downthrown blocks are the basins and are bound by relatively upthrown blocks, which provide a source of sediment to the basins. The East African Rift system is a classic example of a *continental rift*. Here, like many places, this rifting is accompanied by upwelling mantle, resulting in a broad topographic high, despite the fact that thinning of the crust alone should lower the elevation of the top of the crust as a result of isostasy. Magmatism is concentrated along these faults, and gives rise to *bimodal* volcanics, comprising mixed basalts and rhyolites in and adjacent to the rift basins. Continental rift basins are typically filled in by sediments (being derived from nearby highlands), which are often coarse (e.g. conglomerates and sandstones) and red (due to an arid, continental environment). Alluvial, fluvial, and lacustrine sediments are common, and these sediments tend to be highly discontinuous. If rifting continues, the central part of the rift basin may eventually begin to subside below sea level, in which case they are initially highly restricted seaways where evaporites are prone to be deposited (namely, gypsum and halite).

Passive margins

If continental rifting continues until two continental fragments are completely separated, then a new spreading centre forms. The zone of divergence between the now separated plates becomes a new plate boundary. Newly formed rift basins have uplifted shoulders, such as the highlands that border the Red Sea, which continue to be a major source of sediment to the basins. But eventually, the rifted continental margins, which have been thinned and heated up by the upwelling mantle, begin to cool and subside. The result is long basins known as *passive margins*, where the cooling of the upwelled mantle and new sea floor acts as a sinker on the edge of the continent, pulling it down and making room for a vast amount of sediment to be deposited. The eastern margin of the United States and southern Canada is a passive margin, which originated with the opening of the Alantic Ocean. Effectively, passive margins point to sites of previous rifting of continents, which is why the the trace of the Mid-Atlantic Ridge in the South Atlantic closely resembles the eastern shoreline of South America and the western shoreline of Africa. The sedimentary package on a passive margin is sigmoidal in cross section, first thickening away from the continent, then thinning gradually towards the deep ocean basin.

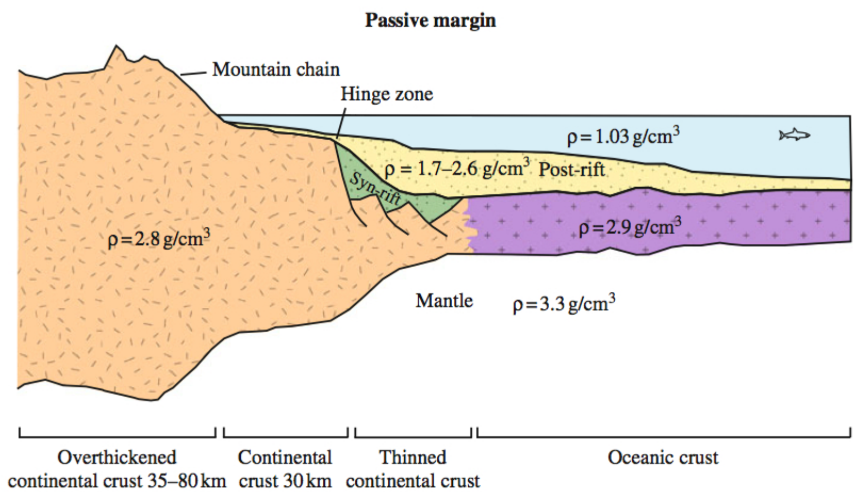


Figure 3.1: Cross-section across a passive margin illustrating different densities and compositions of crustal components involved. Note both the thinned edge of the continent and the oceanic crust are subject to subsidence as the upper mantle beneath them cools (hence converting to lithosphere). From Bjorlykke (2010).

Triple junctions and failed rifts

Continental rifting often occurs above zones of upwelling mantle (e.g. hotspots). Where this happens, the divergence is initially localized on three axes oriented 120° from one another. This is precisely what is happening near the Horn of Africa, where the Red Sea and Gulf of Aden comprise two axes of a triple junction, and the less well developed East

African Rift the third arm. If this rifting continues, and East Africa separates from mainland Africa, then the point where these three spreading centers intersect will become a *triple junction*. In fact, triple junctions are any points where three plate boundaries intersection, but triple rifting junctions are arguably the most important and spectacular.

In many cases, the third arm of a triple junction does not continue to completion, in which case it becomes a *failed rift*. These are distinct in the geological record

3.4.3 Convergent boundaries

Convergent plate boundaries are those where two plates collide. There are two types of convergent plate boundaries: subduction zones and continent-continent collision zones. Subduction zones occur where an oceanic plate thrusts beneath another plate and into the asthenosphere. The lithosphere of the other plate may be either continental or oceanic. Where a subduction zone occurs beneath continental lithosphere, a continental arc–that is, a chain of volcanoes–develops. This volcanism is the result of melting of the *mantle wedge* between the subducting *slab* and overlying crust, driven by the release of fluids from the slab. An excellent example of this type of convergent plate boundaries is along the western margin of South America, where the Andes are the volcanic arc. Continental arcs tend to be dominated by igneous rocks of intermediate composition, hence the term *andesite*. However, both felsic and mafic volcanism also occurs in continental arcs. Where oceanic lithosphere subducts underneath oceanic lithosphere, an *island arc* develops. An example is the Aleutian Island arc, where the Pacific plate is subducting underneath the North American plate.

Long, linear sedimentary basins form on either side of the volcanic chain of mountains. These basins form as the result of the weight of the volcanic arc and associated fold and thrust belts on the edge of the overriding plate. The basin facing the subducting plate is known as the *forearc* basin. During the process of subduction, sediments and seafloor from the subducting plate are commonly scraped off in the *trench* and accreted to the margin of the overriding plate, forming *accretionary complexes*. These accretionary complexes typically mostly comprise a chaotic jumble of metamorphosed and deformed fine-grained sediments (commonly bedded cherts) and oceanic crust called *mḿelange*. The Franciscan complex, which makes up much of the west coast of California, is a mélange.

Another feature of subducting, convergent margins are *ophiolites*. Ophiolites are slivers of oceanic upper mantle and crust that have been thrust up on to continental margins. Because oceanic crust is more dense than continental crust, ophiolites are not omnipresent on convergent boundaries. However, where present, they are clear indicator of a convergent boundaries involving at least some oceanic crust.

3.4.4 Transform boundaries

Transform plate boundaries occur where plates slide past one another, without either converging or diverging (On plate tectonic scales). Oceanic transform faults are just offsets of the mid-ocean ridge. Large plate boundaries may occur on the continent, as does the San Andreas Fault in California. Transform plate boundaries are not as widespread as the other two types of plate boundaries, but they are nonetheless important. Where there is a sharp jog or offset on a transform fault, it generates either small zone of compression (a small mountain range) or extension (a strike-slip basin).

3.5 Vertical Motions in the Mantle

Plate tectonics elegantly explains most of the major features on Earth and utterly changed geology. But like most sweeping theories, it is not perfect an cannot explain all of the features on Earth. Other processes, namely vertical motions in the mantle, may have profound effects on the Earth's surface. We have already discussed hot spot tracks, most of which appear to record mantle plume volcanism. But upwelling mantle, event without volcanism, may cause the lithosphere to buoy upwards, hence generating 100s to 1000s of meters of topographic relief. It is widely believed that the high average elevation of southern Africa is due to one such upwelling cell of the mantle. Similarly, downwelling mantle can pull the whole lithosphere down with it, even generating new sedimentary basins in the process. This *dynamic topography* is typically transient, because the plates are in motion and eventually drift across and away from these zones of upwelling and downwelling.

Chapter 4

Geological Time and the Age of the Earth

Reading: Chapter 3 in Stanley (Chapter 4 in Wiccander and Monroe)

4.1 Introduction

Time is the axis of Earth history. The age of the Earth, as determined by biblical scholarship, did not allow enough time for either Hutton's gradualism or Darwin's evolution. Beginning in the late 1700's, however, scientists began to experiment with ways of estimating the age of the Earth that invariably led to an older Earth. Today, we have impressively precise means of dating old materials and the know the age of the Earth to be precisely 4.54 billion years. The ability to produce *absolute ages* is a huge boon for Earth scientists (and evolutionary biologists). Even so, geologists are still heavily dependent on the original frame of reference for dating rocks, which Nicolas Steno first elaborated: *relative ages*.

4.2 Relative Ages

The earliest stratigrapher (and also a bishop an anatomist), Nicolas Steno, established various principles with regards to the deposition of strata:

- Principle of superposition
- Principle of original horizontality
- Principle of lateral continuity

Along with the the *principle of cross-cutting relationships*, these basic laws enable geologists to work out the relative ages of rocks and geological structures in given region. While simple in essence, these are powerful laws and underlay a large part of what field geologists do.

4.2.1 Principle of faunal succession

In the late 18th century, an English surveyor and engineer by the name of William Smith, who worked to build canals, observed that the suite of fossils within the Paleozoic strata of southern and central England varied up-section, and that this *assemblage* never repeated itself. Hence he concluded, in what is now the basis of biostratigraphy, that the succession of fossils varies systematically, in a reliable order. This *principle of faunal succession* is an extraordinarily useful tool for working out the relative ages of rock not just in one region, but between regions and globally. Indeed, this principle is the basis for how the *geological time scale* was born: the first defined geological intervals were based on the assemblages of fossils within specific bodies of rock.

4.2.2 The fossil record reveals extinctions

At about the same time that William Smith was working out faunal succession, the *incontournable* French anatomist Georges Cuvier was making important discoveries about extinction. In dissecting a fossil mammoth, he realized that these bones did not belong to a dead elephant, but rather a mammal that had previously inhabited the Earth, but no longer does. He also observed evidence for extinctions in the fossil record of the Paris basin. In fact, horizons at which many extinctions occur, and which therefore marked a turnover in fossil assemblages, have naturally become geological boundaries. Cuvier did not live to read Darwin's great treatise and never believed in evolution, but his contributions to paleontology were no less fundamental.

4.2.3 The geological time scale

The principal of faunal succession, along with the observation of extinctions, laid the groundwork for the development of the geological time scale. The basic structure of the geological time scale was established relatively early on in the history of geology, based largely on the fossil record and acceptance of the principal of faunal succession. In this way, the major systems (rock packages distinguished by their fossil assemblages) of the Phanerozoic were established and themselves subdivided. It was not until much later that radiometric dating techniques permitted geologists to pin ages on these boundaries (Fig. 4.1).

4.3 Absolute Ages

Absolute ages are mostly made thanks to *radiometric dating*, which exploits the natural and statistically consistent change of certain intrinsically unstable *nuclides* into new nuclides. Once an age has been established in one place, it can, in some instances, be applied elsewhere through *correlation*. For example, the principle of faunal succession allows us to identify confidently the Cambrian-Ordivician and Permian-Triassic boundary in successions containing strata of those ages. If that boundary can be dated in one place, then that age can be applied elsewhere as well. There are many available tools for correlation, as will be discussed in a subsequent section.

Of course, there are some other ways of determining absolute ages, but these really only apply to the fairly recent past. For example, very precise chronologies can be made counting tree rings or layers in ice cores. Another method is through the exploitation of known periodic events. For example, it is now well accepted that the Pleistocene ice ages are modulated by the *Milankovitch cycles*, whose periodicities are well constrained. Hence, geological records that record these climatic fluctuations can be tuned to the Milankovitch

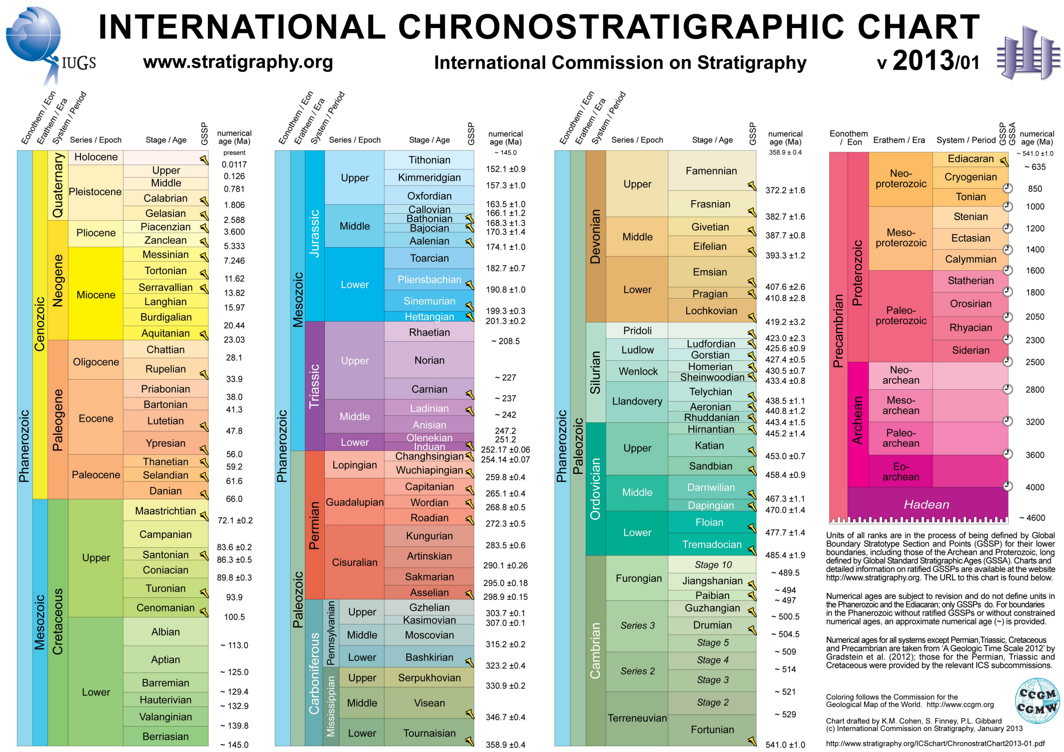


Figure 4.1: The most recent edition of the Geological Time Scale, from Gradstein et al. (2012).

cycles to produce time scales for sediments and ice deposited over the past few million years.

4.3.1 The age of the Earth

One of the great accomplishments of mid-20th century geology was the ability to precisely date the Earth. This was first accomplished by C. C. Patterson, who dated the Canyon Diablo Troilite meteorite using the Pb-Pb *geochronometer*. Since that point, many different meteorites have been dated using a variety of isotopic dating techniques, and they consistently yield the same result. You might ask, why do geologists date meteorites rather than rocks formed on Earth? Think about this question for now, and you should be able to answer it following a subsequent lecture on the earliest Earth. For now, suffice it to know that the oldest dated material that formed on Earth is about 4.4 billion years.

Before digging into the details of radiometric dating, let us first consider other methods early scientists used to estimate the age of the Earth. While these methods may seem naive in light of modern science, they represent great advances in Earth science and reflect the sort of creativity and ingenuity that drive progress in the natural sciences.

- The Archbishop Usher (1640)
  - Earth was created on Sunday, October 23, 4004 BC

- Georges-Louis Leclerc de Buffon (1774)
  - Empirically determined the Earth was 75,000 years old by studying how long it took iron balls of various sizes to cool off
- Charles Walcott (1893) 35–80 million years
  - Time required to deposit the Paleozoic, based on counting layers
- John Jolly (in 1908)
  - Estimated age of 90 million years based on concentration of salt in ocean
- Lord Kelvin (William Thompson) (1866)
  - Approximately 100 m.y., although he revised it downward
  - Based on a heat-loss model (conduction)
  - Assumed internal temperature of 3870° C
  - Surface geothermal gradient of 35° C/km
  - Thermal diffusivity from experiments (0.01178 cm^2/sec)

Lord Kelvin's estimate of the age of the Earth held incredible sway at the time because he was a highly respected physicist and had produced the age through rigorous mathematical calculations based using the best available empirical data. However, this age was at serious odds with the prevailing view of geologists at the time, which was that Earth must have been much older to explain both the rock record and the patterns of evolution. This tension between physicists and physical geologists has largely persisted since then, with physicists often unimpressed by the lack of quantitive support for geological models, and geologists suspicious of physicists calculations based on unrealistic (not ground-truthed) assumptions. In this particular case, the geologists eventually won the argument. Ernest Rutherford later offered the conciliatory explanation to Lord Kelvin that it was because radioactivity had not yet been discovered. And in fact, this has been the standard explanation for why Kelvin got it wrong ever since. However, this explanation is wrong.

4.4 Radioactive dating

Building on recent discoveries of the evidence and nature of radioactivity by Roentgen, Becquerel, and the Curies, New Zealander Ernest Rutherford and Englishman Frederick Soddy (both at McGill University at the time) formulated the general theory of radioactive decay in 1902. The key elements of their theory were

- That radioactivity involves conversion of one element into another
- Radioactivity is proportional to number of *parent* atoms
- Radioactivity declines exponentially at a time scale determined by a characteristic *decay constant*

They identified three types of radioactive decay

- *Beta particle* decay: neutron converted to proton and an electron emitted
- *Electron capture*: proton converted into a neutron by capture of an electron
- *Alpha particle* decay: a particle consisting of two protons and two neutrons (i.e. a ^{4}He nucleus) is emitted

Each of these types of decay emits *gamma radiation*, which is a high frequency, electromagnetic radiation. The gamma rays are what make radioactive minerals potentially dangerous to our health.

4.4.1 The law of radioactive decay

Given that the theory of radioactive decay states that the rate of decay of a radioactive parent nuclide to a stable daughter product is proportional to the number of parent atoms N, at any time t:

$$-\frac{dN}{dt} = \lambda N \tag{4.1}$$

Where λ is the *decay constant*. This equation can be integrated to yield

$$N = N_0 e^{-\lambda t} \tag{4.2}$$

where N_0 is the original number of parent atoms. It is convenient to cast the decay constant in terms of half-life ($t_{1/2}$), the time required for half of the original parent atoms to decay:

$$t_{1/2} = \frac{ln(2)}{\lambda} \tag{4.3}$$

Equation 2 can be re-written by substituting the number of daughter atoms (D) and original number of daughter numbers (D_0) for N and N_0:

$$D = D_0 + N(e^{\lambda t} - 1) \tag{4.4}$$

This is the fundamental equation used in radioactive dating. Now, let's consider an actual radioactive decay scheme, such as

$$^{87}Rb \rightarrow {}^{87}Sr + \beta^- \tag{4.5}$$

We can substitute these isotopes into the equation 4:

$$^{87}Sr = {}^{87}Sr_0 +^{87} Rb(e^{\lambda t} - 1) \tag{4.6}$$

This equation, in theory, is all we need to use the Rb-Sr radioactive decay scheme to date rocks! However, it isn not quite this easy, because it turns out measuring precise concentrations of individual nuclides is very difficult. In practice, it is much easier to measure ratios. Fortunately, algebra allows us to incorporate this analytical requirement into the equation, which we do in this case by dividing through by ^{86}Sr:

$$\frac{^{87}Sr}{^{86}Sr} = (\frac{^{87}Sr}{^{86}Sr})_0 + \frac{^{87}Rb}{^{86}Sr}(e^{\lambda t} - 1) \tag{4.7}$$

Table 4.1: Commonly used radioactive decay schemes for rocks and minerals

Parent	**daughter**	**half-life (years)**	**time frame**	**comments**
Rubidium-87	Strontium-87	49 x 10^9	>100 my	felsic igneous rocks
Samarium-147	Neodymium-143	110 x 10^9	>500 my	igneous rocks
Uranium-238	Lead-206	4.5 x 10^9	>50 my	widely used
Uranium-235	Lead-207	0.7 x 10^9	>50 my	widely used
Potassium-40	Argon-40	1.3 x 10^9	>0.1 my	widely used
Rhenium-187	Osmium-187	41.6 x 10^9	>100 my	black shales, sulfide ores
Carbon-14	Nitrogen-14	5,370	<40 ky	archeology, climatology

4.4.2 U-Pb zircon dating

Of all the geochronometers available to geologists, the U-Pb technique, in particular as applied to zircons, is the gold standard. One tremendous advantage of the U-Pb method is that it has two separate decay systems with different decay constants:

$$^{238}_{92}U \rightarrow {}^{206}_{82}Pb \tag{4.8}$$

$$^{235}_{92}U \rightarrow {}^{207}_{82}Pb \tag{4.9}$$

This duplication of the decay scheme offers a unique way to verify the reliability of data. The mineral zircon ($ZrSiO_4$) is also uniquely well suited for U-Pb radiometric analysis, because it incorporates trace U (which replaces the Zr atom), but does not incorporate any Pb. Furthermore, it is a widely occurring mineral (formed early during the cooling of intermediate to felsic magmas) and is very robust. Indeed, zircons are retained in sands populations (hence sandstone) and commonly used as a way of tracing the source (*provenance*) of sediments. Zircon crystals can now be dated very rapidly by in-situ methods, and this has resulted in the burgeoning of *detrital zircon geochronology* as a geological tool with wide-ranging applications (LaMaskin, 2012).

There are three common analytical techniques for measuring zircon minerals, but in all cases, the data are plotted on a *concordia diagram*, in which the $^{206}Pb/^{238}U$ are plotted against the $^{207}Pb/^{235}U$. If zircons remain closed relative to the U-Pb isotope systems, they will evolve along a single curved line on the concordia diagram. If on the other hand, the zircons were perturbed somehow, and either lost or gained Pb or U, then this will be immediately apparent. But even in this case, not all is lost! Another remarkable thing about the U-Pb chronometer is that where zircons have only been perturbed once (resulting in lead loss) in their history, we can often not only still date the age of crystallization, but also the time at which they were disturbed, because a suite of similarly disturbed zircons will plot along a straight line known as the *discordia*. The upper intercept of the discordia line and the concordia curve is the age of the zircon, and the lower intercept dates the lead loss event (usually metamorphism).

4.5 Other Chronostratigraphic Techniques

Chronostratigraphy deals with strata with respect to time. That is, it involves dating and correlating strata (base on age). Obviously, producing precise ages is the most reliable way of determining the age of a rock, but in most cases, this is not so easy. Fortunately, mother Earth has given us a variety of other ways to determine ages. We have already mentioned that we can use certain annual phenomena, such as tree rings and layers and ice cores to produce geochronologies. But these have limited applicability to the whole of Earth history. Most chronostratigraphic techniques that do not involve directly dating rocks entail *correlation*. We will describe correlation as it pertains to strata more when we discuss the sedimentary record. Here, we are more concerned with correlation of distinctive patterns or signatures that can be extracted from stratigraphic records. We will discuss just a few of the more important techniques.

4.5.1 Biostratigraphy

We have talked about biostratigraphy already and we will discuss it in more detail when we talk more about the fossil record. Suffice it for now the work of Smith (Priniciple of Faunal Succession) and Cuvier (extinctions) laid the groundwork for using the fossil record to make age correlations of fossil-bearing rocks around the world. It goes without saying that this method is only widely applied to Phanerozoic and latest Precambrian rocks.

4.5.2 Magnetostratigraphy

As discussed in a previous lecture about the magnetic stripes on the seafloor, Earth's magnetic field switches polarity from time to time (on average, every million years or so). This can be seen not only in the stripes on the ocean floor, but also in the stratigraphic record, because the switch in polarity shows up in the magnetic signatures carried by iron minerals in volcanic and sedimentary rocks. The result is a sort of bar-coding of the geological record. Because the magnetic polarity record has been precisely calibrated for the Cenozoic Era (past 65 million years), reasonably calibrated for the Mesozoic Era (c. 250 to 65 million years ago), and somewhat calibrated for the Paleozoic Era (541 to 250 million years ago), a magnetostratigrapher can in principle date a sequence of strata by matching the bar coding pattern in those strata with the calibrated reversal record. You will be savvy enough to realize that the preservation of similar patterns is dependent on the reliability of the sequence of strata that is being analyzed. For example, where there are highly variable sedimentation rates, depositional hiatuses, or worse, intervals of erosion, then the magnetic reversal patterns will be heavily distorted. It follows that the most reliable strata to which to apply magnetostratigraphy are relatively quiet (and preferably deep) water deposits with slow but regular sedimental accumulation rates.

4.5.3 Sr isotope stratigraphy

We will talk more about isotopes and the Earth system in later lectures. However, strontium isotopes are uniquely well suited for determining the age of carbonate (or evaporite) rocks that cannot otherwise be dated reliably. As you already know, radiogenic ^{87}Rb decays to the stable nuclide ^{87}Sr. Sr has several stable isotopes, of which ^{86}Sr has a similar abundance as ^{87}Sr. Consequently, the ratio ^{87}Sr/^{86}Sr of the bulk Earth has progressively

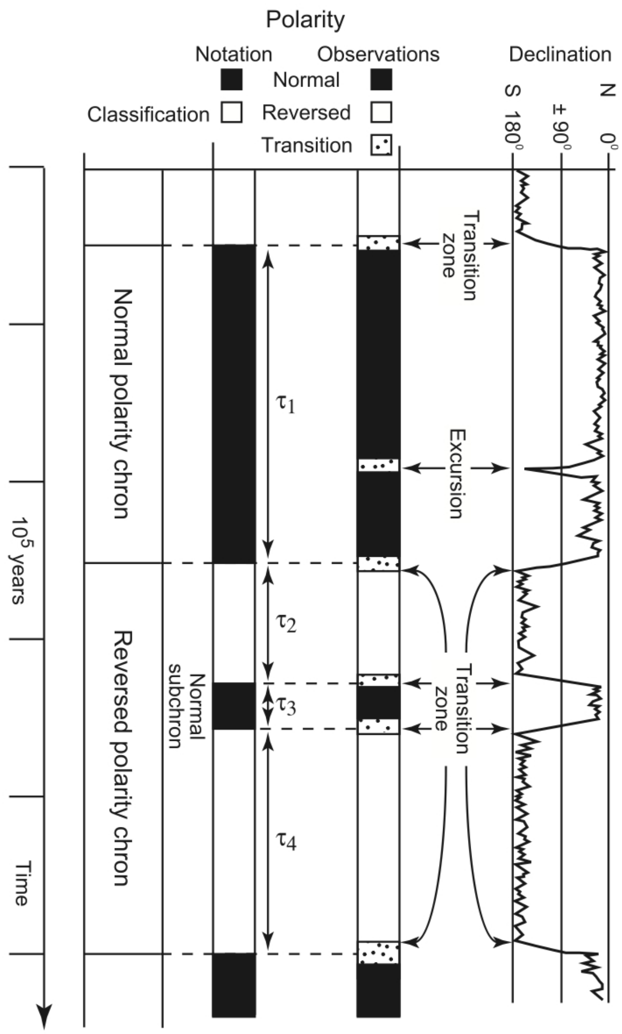


Figure 4.2: Graphical explanation of polarity chrons. From Ogg (2012).

increased over the course of history as a result of *radiogenic ingrowth* of ^{87}Sr.

Rubidium is a *large ion lithophile* element, meaning that because of its large atomic radius, it is incompatible and concentrated in the crust. Consequently, basalt has lower Rb concentrations than granite, and so basalts tend to have much lower ^{87}Sr/^{86}Sr ratios than granites. The seafloor has ^{87}Sr/^{86}Sr ratios equal to that of the mantle, from which it is derived. Strontium is reasonably abundant in the oceans, and the ^{87}Sr/^{86}Sr ratio of seawater largely reflects the balance between low ^{87}Sr/^{86}Sr (unradiogenic) Sr derived form hydrothermal alteration and high ^{87}Sr/^{86}Sr (radiogenic) Sr derived from weathering of the continents (delivered via rivers). Hence, at an given time, the ^{87}Sr/^{86}Sr ratio of seawater reflects the competing influences between hydrothermally derived and continental derived strontium, plus the background effect of a gradual increase in the ^{87}Sr/^{86}Sr ratio of the bulk Earth

Strontium is quite similar in ionic radius and charge to Ca, and so is incorporated into CaCO3 minerals precipitated from seawater, including shells. The ^{87}Sr/^{86}Sr ratio of carbonates then is a *proxy* for seawater composition at the time the mineral was precipitated. Compilations of the ^{87}Sr/^{86}Sr of carbonates of different ages show systematic fluctuations in seawater strontium isotope compositions, most notably a major rise over the past 40 m.y. from about 0.7078 to 0.7091 (this may not sound like much, but is in fact one of the most impressive rises in ^{87}Sr/^{86}Sr in Earth's history). Consequently, if you were to measure the ^{87}Sr/^{86}Sr ratio of a calcite shell of some critter that lived in the past 40 million years, you could determine its age (in this case, within about 1 million years).

4.5.4 Tephrachronology and event stratigraphy

Felsic volcanic rocks are useful for dating because they often contain the mineral zircon, which you now know is the golden child of Earth's minerals. However, very young volcanic rocks are difficult to date (because not much Pb has yet evolved), and, as it unfortunately turns out, many volcanic rocks do not contain *primary* zircons (that is, zircons that date from the age of the magma). Fortunately, these same volcanic rocks, in particular volcanic ashes, can be deposited over vast swaths of land because they erupt high into the atmosphere. Huge eruptions are sufficiently rare that such *tephra* layers can usually be traced to a specific source and correlated, yet sufficiently frequent that you can expect to find a few of them in sedimentary layers that are being deposited downwind from massive volcanoes.

In the same way that huge volcanic eruptions can be used to establish chronostratigraphic marker horizons, so too can other catastrophic events in Earth's history. The most obvious examples of such events are *mass extinctions*, which leave an imprint on the fossil and stratigraphic record that can be used to correlate rocks globally. Some mass extinctions are thought to be linked to meteorite impacts, which leave even better chronostratigraphic markers (e.g. impact eject, shocked quartz, chemical anomalies), because these correspond to a geologically instantaneous event, whereas mass extinctions might be more drawn out.

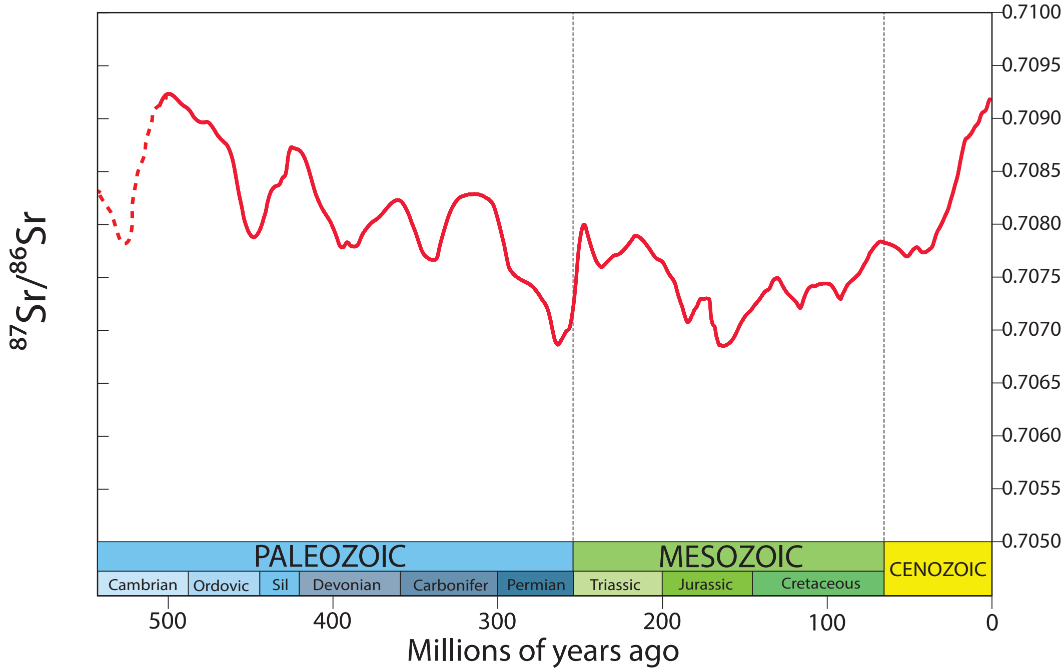


Figure 4.3: The record of seawater $^{87}Sr/^{86}Sr$ over the past billion years. From Halverson (in press). Strontium isotopes can be a powerful means of dating and correlating rocks, but because most $^{87}Sr/^{86}Sr$ are non-unique, additional information on the broad age of the rocks you are trying to date with this method are required.

Chapter 5

The Stratigraphic Record and Sedimentary Environments

Reading: Chapter 5 in Stanley; Chapter 6 in Wiccander and Monroe

5.1 Introduction

Geologists use all rock types to piece together the history of the Earth, but sedimentary rocks are especially important for several reasons. First, sedimentary rocks cover 75% of the continental surface and are highly visible and relevant to geologists. Second, fossils are almost exclusively preserved in sediments and sedimentary rocks, and hence the history of life on Earth has largely been extracted from the sedimentary record. Third, sedimentary rocks archive information about past environments through their physical make-up and structures, the fossils they contain, and their geochemical signatures.

Sedimentary Geology encompasses any field of science dealing with sediments or sedimentary rocks – that is rocks formed by Earth surface processes that include physical settling of grains or chemical precipitation of minerals from air or water. Sediments are an archive of information about environmental, tectonic, and biological conditions that prevailed at the time they were laid down: their very existence in some regions is revealing to the interested geologist. Sediments are largely derived from older sedimentary, igneous and metamorphic rocks, and themselves may be metamorphosed or melted to form other rock types. So they represent one arc in the continuous, and long cycle of shaping and reshaping of the earth's landscape, as envisioned by James Hutton.

5.2 Stratigraphy

Stratigraphy is the study of stratified rocks, which include both sedimentary rocks and layered volcanic rocks. Stratigraphy involves determining the ages (or relative ages), origins, distribution, and lateral relationships of strata.

Steno's Laws+

- Law of Superposition

- Law of Original Horizontality and Lateral Continuity
- Law of Cross-cutting Relationships
- Law of Inclusion

5.2.1 Stratigraphic contacts

- conformity
- unconformity
- angular unconformity
- disconformity
- nonconformity

5.2.2 Lithostratigraphic subdivisions

Lithostratigraphy is the study, description, and correlation of layers of rock and usually involves identifying individual units based on the physical nature of the sediments comprising them (e.g. a sandstone versus a limestone, or a conglomerate versus a shale).
The fundamental lithostratigraphic unit is the **formation**, which is defined as a *a mappable body of rock that is identified by its lithological characteristics and stratigraphic position.* But there is a whole hierarchy of lithostratigraphic units that are commonly used:

- Supergroup
- Group
- Formation
- Member
- Bed or Flow

Traditionally, formations and other rock units are defined based on *lithology*, which is the the physical nature of the rock (e.g., well-sorted sandstone versus conglomerate or shale). One of the results, as we'll describe later, is that formation boundaries commonly do not make very good time boundaries.

5.3 Describing and interpreting detrital sedimentary rocks

5.3.1 Properties of sedimentary rocks

The first step in studying sedimentary rocks is to describe them in the field (or in a core library), followed, often, by observation in thin-section. Of course, some of the easiest things to describe are some of the most useful: color (both in fresh and weathered surfaces), their degree of *induration* (how well cemented are they), and how they weather (e.g., blocky or rounded). But most of the important features require a closer look and systematic descriptions.

Composition and texture

- Grain size
- Grain composition
- Grain rounding
- Grain sorting
- Porosity
- Matrix

Bedding and grading

Sedimentary units almost always consist of individual beds that represent a single depositional cycle (perhaps a storm event, high tide, a flood, etc.). Beds are separated by *bedding planes* that represent periods of non-deposition or even erosion or a change in depositional process. Features to notice about bedding are

- Thickness of layers (beds vs. laminae)
- Lateral continuity of beds
- Nature of bedding planes (planar, curved, wavy, erosional, etc.)
- Type, if any, of *cross-bedding*
- Grading

Graded beds are beds that show a systematic variation in grain size through the bed. Beds whose grain size decreases *upsection* (i.e., from older to younger) are called *normally graded*, whereas those that exhibit an increase in grain size are *reverse* or *inverse graded*.

5.3.2 Sedimentary structures

Sedimentary structures are physical features in sedimentary rocks that formed during or shortly after deposition and often provide some clue as to the environment in which the sediment was deposited. Some important sedimentary structures are

- ripples (or ripple marks)
  - asymmetric ripples (formed in currents)
  - symmetric ripples (formed by waves)
- channels
- mud cracks
- halite casts
- sole marks
- biogenic structures (trace fossils)

5.3.3 Sedimentary basins

Sediments accumulate in *basins*, which require some form of *subsidence* — that is, a means of creating space for those sediments to fill. That space is referred to as *accommodation space*. For example, isostatic considerations tell us that rifting generates subsidence just from the thinning of the continental crust. It turns out that isostasy also enables a much greater thickness of sediments to accumulate than the original amount of subsidence, because the weight of the sediments (and the water) causes the crust to sink even more. Sediments also compact after they are buried, and this compaction generates additional space for sediments to fill. Hence, once accommodation space is produced, it is easier to produce more space (sort of like having a lot of money makes it easier to make money!).

There are three main ways of generating a basin:

- thinning of the continental crust (rift basins)
- cooling of underlying mantle or crust (passive margin)
- depression of the crust under the weight of mountains (foreland basin)

Another way to produce accommodation space is through *eustasy*, which is global fluctuation in sea level. The best understood way of driving global sea level fluctuation are through climate (namely, waxing and waning of glaciers, but also through the effects of thermal expansion/contraction of seawater), which causes changes on the the time scale of 10's to 100's of thousands of years. Another commonly invoked means of changing global sea level is through total volume of the ocean basins, which might change with changing globally integrated rates of production of oceanic crust.

5.3.4 Depositional environments

Depositional environments are areas of active sediment accumulation characterized by a distinct combination of physical, chemical, and biological processes, that result in a unique *facies*, that is, a suite of sedimentary features that point to a specific depositional environment. Depositional environments change in space and time due mainly to changes in water depth, which can be driven by local, basinal factors (such as tectonics or sediment infilling) or actual variations in global sea level. Climate and tectonics are the main factors that drive these changes in sea level.

By their very nature, shorelines (of lakes or oceans), and the affixed depositional environments, are unstable. Imagine a beach. That beach will only remain in the same geographic location over time if the rate of production of accommodation space (basically subsidence plus space generated or lost by rising or falling sea level) is exactly balanced by sediment accumulation rates. Where that balance is not achieved, the shoreline will either migrate continent-ward, in what is called a *transgression*, or ocean-ward, in what is known as a *regression*. Transgressions are associated with an increase in water depth, whereas regressions result in a decrease in water depth (or complete exposure of a basin to subaerial weathering and erosion).

Transgressions and regressions result in laterally migrating depositional environments. *Walther's Law* states that a vertical succession of facies in a sequence of sedimentary rocks reflects such lateral changes in sedimentary environment. For example, if a beach facies is overlain by back-beach aeolian facies, this indicates that these two environments were juxtaposed and that this sequence experienced a regression. Walther's Law is the basis by which geologists can convert the sedimentary record into a paleonvironmental map of a sedimentary basin.

The continental environment

- **Fluvial.** Commonly coarse-grained channel fill/bars with fine-grained flood-plain deposits. May be associated with coals. Important archives of terrestrial environments.
- **Lacustrine.** Varved (rhythmic) laminae common. May be either open or closed, they latter commonly forming chemical precipitates via evaporation.
- **Desert.**
  - alluvial fans are the coarse grained aprons that form on the margins of mountain ranges
  - dunes are deposited by blowing sand and have characteristic bimodal grain size distribution and large-scale cross-bedding
- **Soils.** Not extremely common in the sedimentary record, but highly informative because soil formation is sensitive to so many environmental parameters.
- **Glacial.** Unsorted tills/morraines (diamictites), ice-sculpted landforms (i.e. drumlins), striations and grooves

Coast lines

The coast line is the *marginal marine* environment that occurs along the boundary between the ocean and the continents. It is influenced by a combination of rivers, wind, waves, tides, and other shallow marine processes and can be subdivided into a several sub-environments.

- **Deltas.** Dominated by sediment supply from rivers. Form prograding lobs of sand and mud, commonly arranged in *clinoforms*
- **Estuaries.** Flooded river mouths. Brackish water, significant muds.
- **Beaches and barrier islands.** Sandy, commonly cross-bedded with heavy mineral laminations, multi-directional currents.
- **Tidal flat.** Significant muds, commonly mud-cracked, interbedded with sands with distinct structures and bedding types associated with tidal currents.

The open marine environment

Open marine environments are spatially the most expansive because they include most of the continental shelves and the deep ocean basins.

- **Continental shelf.** Influenced by storm events, pelagic sedimentation

- **Continental slope.** Dominated by *gravity flow* processes (mud and debris flows, turbidity currents)
- **Deep ocean.** Pelagic sedimentation: dust and tests of micro-organisms, such as diatoms and foraminifera.
- **Glaciomarine.** Heavily controlled by processes at the *ice-grounding line*. Commonly poorly stratified sediments with outsized and *exotic* clasts.

Carbonate platforms

Carbonate sediments preferentially accumulate in shallow, tropical marine settings with limited siliciclastic sediment input. A *carbonate platform* is a broad region of shallow water dominated by carbonate sedimentation. The Bahamas banks off the east coast of Florida is a well known example of a carbonate platform. Here, carbonate sands (which include *ooids*) and lime muds are abundant.

Reefs constitute another important site of carbonate sediment accumulation in the oceans. The most prominent are barrier reefs, such as the Great Barrier Reef off the coast of northeastern Australia. Barrier reefs are relatively common in the stratigraphic record where they form the outer edge of a carbonate platform and provide protection to the inner part of the platform (a *lagoon*), which enables carbonate sediments to accumulate. Whereas most modern reefs consist largely of corals (with significant contributions from sea grasses and carbonate-secreting algae and cyanobacteria), a variety of different organisms have formed reefs throughout Earth history, notably *stromatolites* in the Precambrian.

Limestone precipitation is concentrated in the tropics because calcium carbonate solubility is lower in warmer waters. This mineralogically unusual situation arises because of the somewhat complicated speciation of dissolved inorganic carbon in seawater. But what it boils down to is that processes that remove CO_2 from seawater drive those waters towards carbonate precipitation. Because warmer water can hold less dissolved CO_2, this is where carbonate precipitation is focused.

- continental
- marine

Carbonates also do occur in the deep sea, but only above the *carbonate compensation depth* (the depth at which carbonates dissolve in seawater; this will be discussed in more detail in a subsequent chapter). Carbonates found on the deep sea floor are almost entirely derived from the sea surface or from shallow water carbonate platforms and reefs (transported by gravity flows, such as turbidites).

Chapter 6

Life, Fossils, and Evolution

Reading: Chapters 3, 7 in Stanley; Chapter 7 in Wiccander and Monroe

6.1 Introduction

The vestiges of ancient life are called *fossils*. Fossils come in various forms, including moulds, casts, imprints, impressions, burrow traces, recyrstallized shells, and replaced hard and less commonly, soft parts. Molecular fossils (*biomarkers*) are molecules preserved in the organic matter of sedimentary rocks that can be used indicators of specific groups of organisms. The study of ancient life is broadly known as *paleontology*. Though some ancient scholars contemplated the meaning of fossils, most did not appreciate what they represented. It was not until William Smith, Georges Cuvier, and a few others in the early 19th century began to make sense of the fossil record was the field of paleontology born. Today, this field is broad and includes many subdisciplines, such as evolutionary biology, biostratigraphy, paleoecology, and molecular paleontology, to name but a few. Paleontology is an increasingly interdisciplinary field, often combining phylogenetics with organic and isotope geochemistry.

The word *Paleontology* is derived from the Greek *palaeo* = ancient, *ontos* = being, and *logos* = knowledge. Modern paleontology is the overlap between biology, geology, ecology, paleogeography, geochemistry, and increasingly, molecular biology. It can be defined as the study of ancient life in the context of environment and evolution. Classical paleontology involves making many measurements of various dimensions of innumerable fossil specimens (*biometrics*), then trying to make sense of these data through statistical analyses. We do not have the time to get too deeply involved in this aspect of paleontology.

6.1.1 Subdisciplines of paleontology

- paleozoology
- paleobotony
- invertebrate paleontology
- vertebrate paleontology

- micropaleontology

The western history of palaeontology, like many sciences, goes back to Aristotle, but really got started by Da Vinci. Modern paleontology was born around the end of the 18th century, when geology first took roots. Four 18th and 19th century scientists are largely responsible for the development of paleontology and how we read the paleontological record.

- Georges Cuvier (1769–1832) recognized that fossils represented extinct organisms
- William Smith (1769–1839) first applied biostratigraphy
- Charles Darwin (1809–1882) and Alfred Wallace (1823–1913) demonstrated evolution and proposed the theory of *natural selection*.

6.2 Fossils

Fossils, quite simply are the preserved remains or traces of organisms in rocks. Fossils provide the framework for how we understand ancient life. However, sedimentology, evolutionary biology, ecology, and increasingly molecular biology fill in the gaps in this framework and by now have provided has with a rich knowledge of ancient life. Almost all fossils are found in sediments and sedimentary rocks that have not been heavily metamorphosed.

6.2.1 Types of fossils and preservation

Fossils come in many shapes and forms and may be preserved in a variety of manners.

- Various factors favor fossilization:
  - abundance of organisms
  - minimal disturbance after death
  - rapid burial or 'entombment'
  - small size
  - marine organisms are more likely to be preserved than terrestrial organisms
- Hard parts much more likely to be preserved than soft tissues
  - in vertebrates: bone, enamel, keratin (fibrous structural proteins)
  - in invertebrates: chitin (a polymer of glucose used, e.g., in exoskeletons), shells (carbonate or silica)
  - in plants: cellulose (a polysaccharide) or lignin (a polymer of ring alcohols that provides structure to plant cell walls)

6.2.2 Modes of preservation

Soft parts

The soft parts (tissues, hair, etc.) of animals are rarely preserved, but certain exceptions do occur and unaltered remains are preserved (permafrost, amber, mummification, tar, peat). Invariably, the preservation of soft parts requires a hermetic tomb, that is immediately and permanently sealed off to oxygen.

Molds

Molds are the negative imprint of an organic structure. They commonly arise through the dissolution of the structure during diagenesis and require that the space left behind is retained, either permanently, or for long enough for some other material to fill it, and hence preserve the original shape of the structure

Impressions

Impressions are delicate surface markings in soft sediments that preserve the flattened outline of certain soft or semi-hard organisms.

Changes to organic components

- carbonization: release of volatile constituents; nearly pure C is left behind as a film
- pyritization: mineral replacement of soft parts

Changes in the inorganic substance of hard parts

- unchanged hard parts are most likely Cenozoic
- recrystallization
- petrifaction (by percolating fluids)
  - Most commonly calcite, silica, and iron minerals
- solution of hard parts
  - external mold: a shape in the sediment that reflect that outer shape of the organism. Imagine your footprint in wet sand on the beach.
  - internal mold: sediment fills an open cavity in an organism: for example, a clam shell that is filled with sand.
  - *casts* fill molds: e.g. the sediment or minerals that replaces the cavity when a clam shell dissolves

6.2.3 Trace fossils and other indirect fossils

Ichnology is the study of trace fossils - that is, of the tracks, burrows, footsteps and other traces left behind by organisms.

- Trace fossils are useful biogenic sedimentary structures
  - may indicate sedimentary environment/paleodepth
- *bioturbation* is the burrowing of organisms in unconsolidated sediments
  - destroys other sedimentary features, such as bedding and structures
- Often preserved in sediments otherwise not conducive to fossil preservation
- Always *in situ*

- Preserved in various ways
    - most commonly as sand or mud cast of a burrow or track

6.2.4 Lagerstattën

A sedimentary deposit that contains exceptional preservation of fossils is referred to as a *lagerstattë*. Lagerstattën (plural) may form by a variety of means and span from the late Neoprotroterozoic to the present. They commonly (but not always) include fossils of preserved soft tissues, and so provide extraordinary insight into ancient life. Well known examples of lagerstattë include the Cambrian Burgess Shale in B.C. and the Jurassic Solnhofen limestone in Germany.

6.2.5 Biomarkers

Biomarkers are molecular fossils. You can think of a biomarker as the altered chemical trace of some previously living organism. Biomarkers can be measured in rocks and sediments, and the key biomarkers are molecules that can be linked to precursor molecule produced by a specific organism, only lacking functional groups (which are quickly stripped off during diagenesis. The field of biomarker geochemistry has largely been developed my the petroleum industry who use biomarkers to link petroleum to source rocks, among other applications.

6.3 Biostratigraphy

Biostratigraphy is the application of the fossil record to stratigraphic correlation and determining the relative ages of strata.

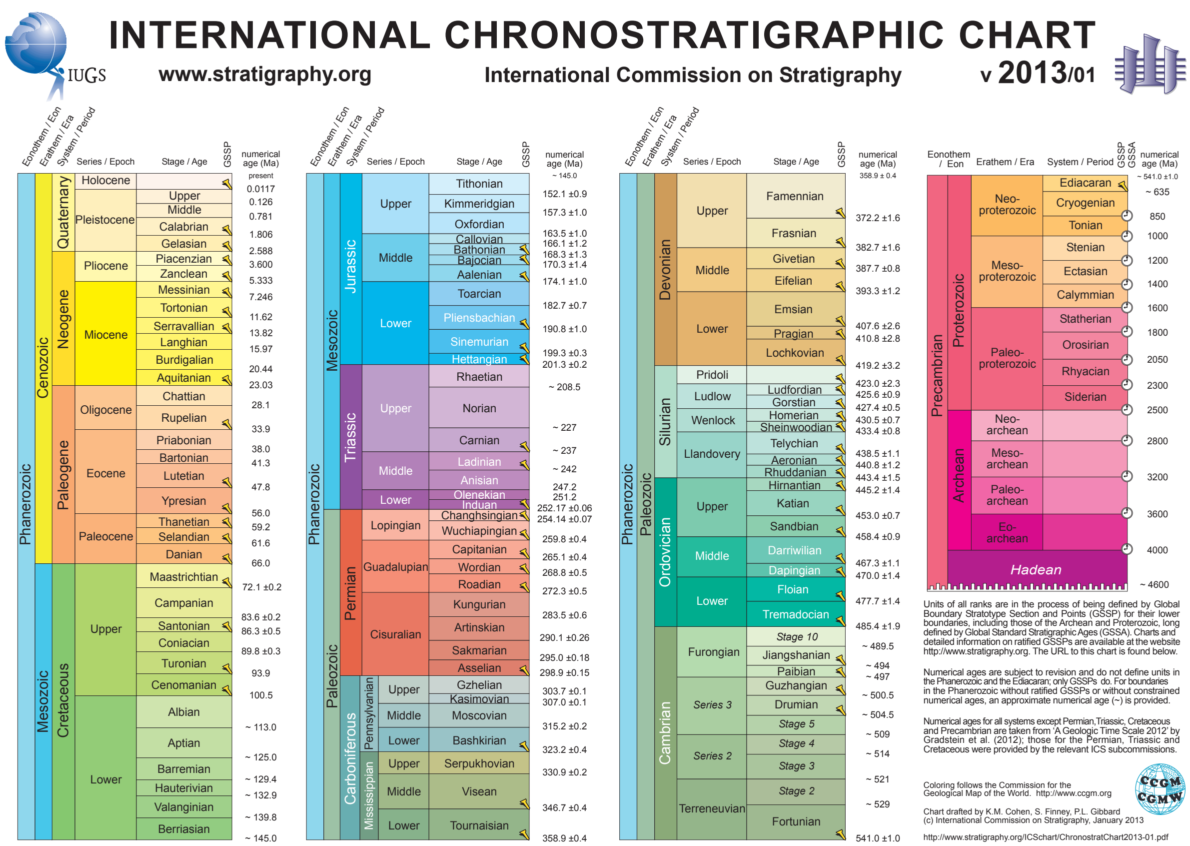
- Guided by the *principle of faunal succession* - the observation that sedimentary rocks contain fossils that succeed one another vertically in a consistent order and that these successions can be recognized over wide long distances
- Biostratigraphic zonation
- Stages and zonation
    - range zone: The body of strata representing the known range of a particular taxon
    - biozone: A body of strata delimited by a characteristic assemblage of taxa
    - concurrent range zone: a body of strata in which a specific suite of taxa all occur (and certain other don't occur)
    - an *index fossil* is a fossil with a short range and a wide geographic distribution
    - first and last appearance of fossils
    - The *Signor-Lipps effect* is the principle that, due to the inherent incompleteness of the fossil record, neither the first nor the last organism of a given taxon will be preserved in the fossil record.

The text-book demonstration of the Signor-Lipps effect: the coelacanth, thought to have gone extinct in the late Cretaceous (based on last occurrence).

6.4 The Geological Time Scale

The origin of the geological time scale can be traced to the work of German geologist Abraham Werner who divided the rocks on the surface of the Earth into the *Primary, Secondary, Tertiary, and Quaternary.* Werner went on to argue that these rocks were precipitated and deposited during successive stages of the great flood. Although this basis for the subdivision of the geological time scale is long since discredited, the terms Tertiary and Quaternary persist today, although *Tertiary* is being phased out of use (replaced by the combination of *Neogene* and *Paleogene*.

The modern geological time scale can be traced to early 19th century geologists, mainly in Britain and France, who recognized that rocks of certain ages showed distinctive assemblages of fossils. They lumped these into "systems", which eventually became what we call *periods*, such as the Cambrian and the Cretaceous. It was only much later that relative order of these "systems" was determined, and only with the advent radiometric dating that they came to be calibrated. Based on the fossil-centric method of establishing the time scale, the majority of Earth's history, during which there were no animals to leave fossils, was lumped into the *Precambrian*.



6.5 Systematics and Taxonomy

Systematics is the study of the structure and diversity of life, in particular as it has evolved. Although similar in some aims, it should not be confused with *taxonomy*, which is more strictly concerned with the identification, description, and nomenclature of organisms and fossils. Hence, the role of systematics is to understand the pattern presented by taxonomy.

A *taxon* is any named grouping of organisms. Palaentologists, like biologists, use the linnaean, *binomen* hierarchical classification system to name fossil species:

- **Superkingdom:** Eukarya
- **Kingdom:** Metazoa
- **Phylum:** Chordata
- **Class:** Mammalia
- **Order:** Primates
- **Family (-idae, -aceae):** Hominidae
- **Genus:** *Homo*
- **Species:** *Homo sapiens*

6.5.1 Classification of fossils

Taxonomy is not simple business

- originally classified based solely on morphology and function (that is, *phenotypical expression*
- *Evolutionary taxonomy* was mainstream in the first two thirds of the 20th century
  - Substantially aided by *ontogeny*: the study of the origin and development of animals, with an emphasis on embryonic development. Ontogeny demonstrated close evolutionary linkages between taxa that are not as obviously similar in their adult phenotypes
  - *Phylogeny*: The origin and evolution of a set of organisms. This is how we place organisms within an evolutionary frameowork
- *Numerical taxonomy*: use of objective, statistical similarities to develop phylogenetic relationships

6.5.2 Cladistics

Cladistics is the classification organisms by their evolutionary relationships rather than their morphology. While this seems like an obvious way to classify organisms and fossils, it was not always so. This method was first championed by the German entomologist Willi Hennig (1913–1976). The underlying assumption in cladistics is that two taxa share the same trait because they are both derived from the same ancestor that also had that trait. This is referred to as a *primitive* trait. *Derived* traits (*synapomorphies*) appear later in the cladistic succession.

- *Cladograms* are tree-like diagrams that illustrate the phylogenetic relationships between organisms
- A *clade* is any subgroup (or subtree) of organisms along a branch of the tree. Clades **cannot** join parts of two separate branches
- Two taxa on either side of a branch point are referred to as *sister groups* or *sister taxa*
- The last common ancestor is the most recent common ancestor of a group of taxa
- The breakdown of groups into hierarchical levels (e.g., class, order, family) is rather arbitrary, but inevitable
- The actual classification is not simple either, but is typically guided by the principle of *parsimony*
  - Simplest scenario (i.e. with fewest evolutionary events)
- *Homologs* are shared traits

6.5.3 Molecular Phylogeny and the Three Superkingdoms

Molecular phylogeny is based on variations in homologous molecules (mainly in RNA and DNA) between organisms, which accumulate by random mutations through time.

- Most useful in molecules that are *homologous* among all organisms
- Has completely changed the biological tree
- Has also fundamentally changed palaeontology
- The most common molecules for molecular phylogeny are
  - ribosomal RNA
    - 18S ssu rRNA - evolves slowly
    - 16S used early on
  - rDNA

Molecular phylogeny has demonstrated that there are three main domains, or superkingdoms of life: *Bacteria*, *Archaea*, and *Eukarya*. The bacteria and archaea are commonly lumped together as *prokaryotes*, although most phylogenies show the Archaea and Eukarya sharing a single branch (sister superkindgoms). Prokaryotes are exclusively single-celled organisms, and they lack nuclei and organelles. Eukaryotes, on the other hand, have more complex cell structures, including nuclei and other organelles.

Archaea

The Archaea include many groups that are capable of enduring harsh environments, such as high temperatures, acidity, and salinity.

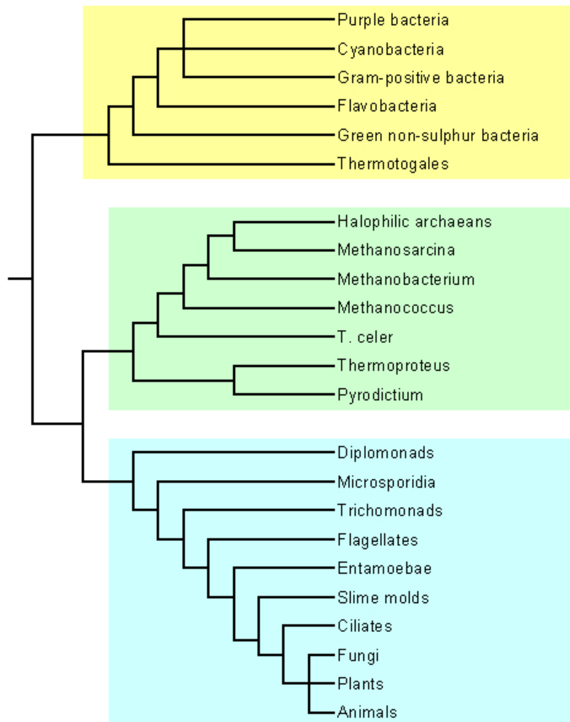


Figure 6.1: An early version of the tree of life produced by molecular phylogenetics. One of the first things to come to light in this approach was the breakdown of life into the three superkingdoms Bacteria, Archaea, and Eukarya.

Bacteria

The bacteria include many groups that pollute and cause diseases, but also many groups that decompose—that is, they break down previously synthesized organic matter. They also include the *cyanobacteria*, which are photosynthetic, and have both played a huge role in the evolution of Earth's environment through their important role in the global carbon and oxygen cycles.

Eukarya

Eukaryotes may be either unicellular or multicellular. The Eukaryotes are subdivided into four kingdoms:

- Plants: exclusively multicellular and photosynthesizing
- Fungi: mostly multicellular, but some unicellular
- Protists: very simple organisms and a diverse group. Some multicellular (the so-called *fleshy algae*, which lack tissues), but mostly unicellular. These include both autotrophs (unicellular photosynthesizers, such as diatoms and calcareous nanno-plankton) and heterotrophs (foraminfera and radiolaria)
- Animals: I think you know what these are

6.6 Evolution

About half a century before Darwin's epic voyage aboard the HMS Beagle II, Jean-Baptiste Lamarck (1744–1829) proposed the first well-developed hypothesis for evolution. Commonly referred to as *Lamarckian evolution*, this hypothesis posits that new traits in organisms arise during their lifetime as the result of some need, and these traits can be passed along to their descendants. We may snicker at the seeming naivety of this idea now, but in fact it was a tremendous leap forward in biology at the time.

6.6.1 Evolution happens

Charles Darwin (1809–1882) and his near contemporary Alfred Wallace (1823–1913) established that evolution accounts for the origination of new species and the diversity of life on Earth. Charles Darwin was trained in geology and had a copy of the Lyell's just-published *Principles of Geology*. As such, the observations on his voyage on the HMS Beagle II (1831–1836) and the interpretations he made to explain them were framed in a geological context. Importantly, he not only made observations, but also collected the detailed data and notes required to convince others of evolution and the theory of natural selection. Some of the bedrock observations that led him to his conclusion:

- The fauna of the Americas was completely different from that of Europe and Africa, even in comparable environments
- Marine species are strikingly dissimilar on either side of the Isthmus of Panama
- Small islands >5000 km from continents lacked endemic land mammals
- Shells of Galápagos tortoises unique on different islands
- Finches most closely resembled those of South America; unique adaptations of beaks to grazing habits
- Similarity in the embryos of vertebrates
- *Homology*: similar body parts with different functions
- *Vestigial organs*: organs, such as the appendix, that serve no purpose

Fundamentally, evolution is driven by random mutations that occur during sexual reproduction. Most such mutations are not advantageous (perhaps why being called a "mutant" is not a compliment) and do not propagate. Where evolution occurs is where a mutation actually imparts an advantage on organism within its milieu. This process is further complicated by the fact that environments themselves are not static. So whereas white fur is an evolutionary advantage to polar bears when they are hunting for seals on the sea ice, it will be less of an advantage and probably a disadvantage as they are forced to find food on bare land.

Much of what Darwin and Wallace observed in the (then) modern world is also observed in the fossil record. Quite simply, it is unavoidable to conclude that organisms evolve through time. Whereas *evolution* as articulated by Darwin may have qualified as a theory at the time, it is a well observed fact today, and the "theory of evolution" is not about whether it occurs but rather why evolution happens.

6.6.2 The theory of natural selection

Darwin was well aware that artificial selection, by dog or plant breeders, was capable of producing great variability among given species. He was also deeply convinced by Hutton and Lyell's gradualism, and the enormity of time. The theory of *natural selection* was perhaps not that great of a leap once he had observed evolution. The challenge in proposing the theory, and the reason he waited over twenty years to do so following his return from the Beagle voyage, was that natural selection is utterly at odds with religion and the notion of a benevolent god. Indeed, it was only the realization that Wallace had also stumbled upon the theory based on his research in southeast Asia that motivated Darwin to publish his evidence for evolution and theory of natural section to explain it.

The English cleric Thomas Malthus (1766–1834) observed that organisms produce more offspring than can survive, as part of a natural impetus to increase population. Both Darwin and Wallace (1823–1913) recognized that this resulted in competition within species to survive. For example, certain species of finches that Darwin observed in the Galapagos Islands bear two chicks, even though only one of those chicks ultimately survives. In fact, in the brutal struggle to survive, the weaker of the two chicks is often killed by his or her sibling. It is in this way that individuals that are stronger or more capable will preferentially survive and pass along their genes, while the weakest individuals will be killed off and their genes eliminated from the gene pool. This is a brutal, but efficient way for species to propagate and adapt to their environment (given, of course, sufficient time).

In a similar way, different species struggle against one another. They may be in direct competition for the same resources, or one species may prey on the other, leading to a so-called evolutionary arms race, where one group of organisms will evolve to better prey on another group of organisms, triggering them to evolve better defence mechanisms, and so forth. It has often been argued that the Cambrian Radiation records the first major evolutionary arms race, after which, effectively all modern body plans had been achieved (at the phylum level) and from there, evolution largely preceded at lesser (order, family, etc.) levels. Under this process of natural selection, taxa evolved gradually (*anagenesis*) as the result of long-lived environmental pressures.

6.6.3 Patterns and rates in evolution

Evolutionary radiation

Anagenesis is the gradual evolution of one taxon, but we know that taxa also diverge. The splitting of a single taxon into multiple taxa (branches), is known as *cladogenesis*. Both processes are observed in the fossil record, which, despite its inherent weaknesses, provides us a measure of the timing and rates of origination of species (or other clades). Apparent in the fossil record are periods when certain taxa produce many new genera or species. These *evolutionary radiations*, which include the famous Cambrian radiation, show up spectacularly in the fossil record. These radiation events typically result in certain groups of organisms adapting to new modes of life, which enable them to expand and populate new ecospace.

These radiation events may result from the *extinction* of other taxa that dominate available environments. For example, we now know that mammals coexisted with dinosaurs through much of the Mesozoic, but they remained small and subordinate. Only after extinction of dinosaurs did mammals flourish. The radiation of mammals over the last 66 million years has been nothing short of extraordinary (look at how different bats are from whales, for example!).

Other radiations were triggered by *adaptive breakthroughs*, whereby a taxon develops certain features that enable it to survive and thrive in available environments, outcompeting other organisms. One such example are the *hexacorals*, whose porous structure allows them to grow much quicker than other calcite-secreting sessile organisms because they need much less $CaCO_3$ per unit volume.

Molecular clocks

A *molecular clock* is a technique of dating divergence times between different groups of related organisms, based on molecular phylogeny. It is particularly useful for estimating divergence times between extant organisms that have left no or incomplete fossils records. But of course, it has wide-ranging applications.

- Needs to be calibrated
- Based on the assumption that rates of divergence are constant (at least between calibration points)
- Assumes no lateral gene transfer
- Inherently less reliable the further back in time one goes

Evolutionary convergence

Evolutionary convergence occurs when separate taxa evolve to have similar traits. Common examples of convergence include the various types of anteaters, flying squirrel-like creatures, and cactus and cactus-like euphorbia. Convergence attests to the strong control of environment on evolution.

Horizontal gene transfer

In case cladistics and evolution is not sufficiently difficult, it turns out that in some cases, genetic information has been transferred between different species in a process known as *horizontal gene transfer*. This usually occurs among prokaryotes as a result of bacteria or viruses invading a cell of organism, then transferring its genetical material to another victim.

Extinction

As mentioned before, the first-order structure of the geological time scale reflect the pattern of major *extinctions* in Earth history, that is intervals marked by the abrupt disappearance of fossil taxa. Extinctions result from a variety of factors, including predation, disease, competition for resources, and of course, natural disasters.

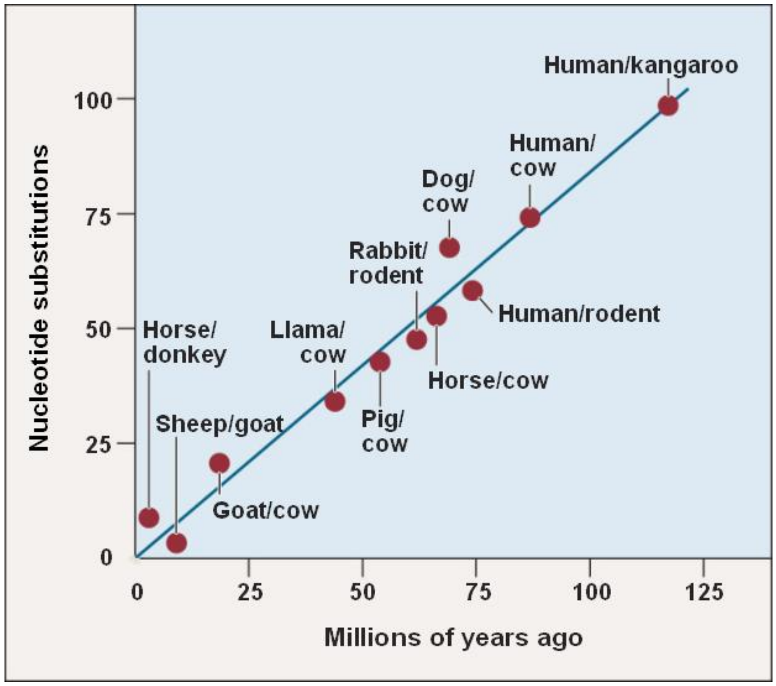


Figure 6.2: A plot of divergence times versus divergence (nucleotide substitutions) for 17 mammalian species illustrates that at lease for this clade, divergence occurs at a constant rate.

Extinction is a normal part of the fossil record, and different taxa exhibit very different characteristic rates of extinctions. For examples, about half of all mammal species goes extinct every million years, whereas among marine molluscs, the rate is an order of magnitude slower.

Despite the difference in rates of extinctions among different groups, global trends in extinct are apparent. Specifically, plots of either total number of species, genera, or classes at a given time, or simply percentage of species or genera going extinct through time highlight several *mass extinction* events, where an unusually large percentage of existing taxa go extinct. The big five of mass extinctions are the

- end-Ordivician event
- late Devonian event
- Permo-Triassic event
- late Triassic event
- Cretaceous-Paleogene event

We will discuss the individual events more in late lectures.

Evolutionary trends

Animals evolve towards larger body size. This is mainly a result of competition to breed, with larger males tending to win the fight for females, driving a tendency towards larger and larger animals. This pattern highlights an important point about evolution, which is that 'more evolved' does not imply any sort of overall superiority. This tendency among animals is fundamentally unsustainable, because animals can grow only so large before their size and nutrient requirements become a serious problem. Indeed, this tendency probably contributes to the high extinction rates among mammals, where it occurs relatively quickly. Large animals, you can imagine, would be highly susceptible to perturbations to the food chain.

Rates of evolution vary greatly. For the most part, they are slow, with individual species ranging for millions or tens of millions of years. However, speciation can happen quickly in cases where a simple genetic changes have major consequences. The fossil record also abounds with examples that suggest that taxa experience intervals of *stasis*, followed by rapid evolutionary expansion, as discussed in the following section.

Finally, evolution is irreversible. Because evolution is the result of a series of genetic changes, it is effectively impossible to undo, even if subsequent evolution might produce a creature that resembles any ancestor. This unidirectionality in evolution is known as *Dollo's law*.

6.7 Gradualism Versus Punctuated Equilibrium

6.7.1 Phyletic gradualism

Evolution by natural selection, as originally conceived by Darwin and Wallace, occurs gradually, requiring vast tracts of geological time. Darwin was very aware of this requirement of his theory, and for this reason, was greatly troubled by both Lord Kelvin's stingy estimate of the age of the Earth (ca. 100 m.y.) and by the Cambrian explosion, which appeared to record a very sudden initial evolution of animals. In fact, the fossil record shows many cases where speciation appears to occur instantaneously, in geological terms, and the question is whether or not this feature can be accommodated by Darwinian evolution.

- Darwin himself recognized this feature of the fossil record
- But *phyletic gradualism* reigned as the preferred model for evolution until the latter half of the 20th century

In this model, new species evolved largely through *anagenesis*, the gradual transition of one species into another. This sort of evolutionary pattern gives rise to *pseudo extinctions* in the fossil record, because specific species disappear over time, but because they have changed not because some catastrophic event or other change in the environment wiped them out.

6.7.2 Punctuated equilibrium

In the early 1970's, Niles Eldridge and Stephen Jay Gould published a landmark paper in which they argued that most species change little for most of their history (*stasis*), and when evolution occurs, it does so in rapid branching events (*cladogenesis*). This hypothesis was fundamentally rooted in the observation that the fossil record rarely demonstrates gradualistic evolution and that new taxa appear rapidly. The explanation for this pattern is based on observations that

- large populations dilute *phenotypic* changes—that is, changes in the physical manifestation of the genetic code
- geographically isolated (peripheral) populations are more likely to change and form new species (*allopatric speciation*)
- however, because the allopatric species are marginal, these adaptations are usually wiped out
- the record of speciation is actually very rare in the fossil record—it occurs when chance changes in environment favor a peripheral population and are detrimental to the parent population

At its core, the punctuated equilibrium model doesn't contradict the gradualistic model. Rather, it argues that gradualism alone cannot explain the fossil record. Importantly, this hypothesis emphasizes the importance of ecological context within paleontology.

Chapter 7

The Environment and Chemical Cycles

Reading: Chapter 11 in Stanley; Chapter 4 in Wiccander and Monroe

7.1 Introduction

An important goal in Earth history is reconstructing ancient environments (*paleoenvironments*). Earth historians do this in part because the detective work involved in figuring out past environments is intrinsically interesting. But we do this also to understand how Earth's surface environment has evolved and responded to major perturbations in the past. Indeed, by studying the history of past climatic perturbations, we gain important insight on how the current Earth system will respond to our current experiment in global warming.

The modern environment is also the best tool we have to understand ancient environments. However, it is insufficient and inaccurate to compare the ancient Earth solely to the modern environment, because the Earth has changed considerably over its history, and certainly many ancient environments have no analog in the present world. For example, we live in a glacial epoch (despite the fact that we are in a glacial minimum), which is relative rare in Earth history. Consequently, the present climate is not comparable to climate that prevailed through most of Earth's history.

First order clues of the nature of ancient environments can be obtained directly from the sedimentary record. For example, sedimentologists can readily discern between deltaic, beach, fluvial (rivers), and deep sea environments based the physical characteristics of the strata. Other important clues may come from the chemistry of the rocks. At the heart of studies of paleoenvironments during the Phanerozoic (541 million years ago to the present) is the study of the fossil record. Fossils may serve to help understand a particular environment better (for example, certain types of trace fossils are specific to certain environmental settings). In turn, studies of the sediments might help to unravel the environment and life habits of the organisms that left certain fossils. And by examining the whole suite of fossils within a package of rock, a paleontologist can develop an outline of the entire ecosystem.

Because environments and the behaviour of the modern Earth system are covered in detail in many other courses, we will only briefly review modern environments here.

7.2 Ecology

Habitats, as the word implies, are those parts of the Earth's surface and near surface environments that are inhabited by life. We can broadly break down Earth's habitats in terrestrial, marine, and freshwater.

Most organisms can only live within a relatively limited range of environmental conditions. In order to survive and perpetuate, organisms require a source of food (or energy), and physical conditions within certain thresholds, which define the limits of its ecological *niche*. The way that an organism performs its basic functions in that niche, such as feeding and reproduction, is its *life habit*.

A *population* is a group of individuals of a species that inhabits a region, and the ensemble of species in that area is a *community*. The community and the physical environment it occupies constitute an *ecosystem*. Most species compete with other species within an ecosystem for resources, and many species are a source of food for other species or themselves eat other species, giving rise to *food chains*, and the more complex *food webs*, in which multiple species occupy *trophic* tiers. *Diversity* in both modern environments and in the geological record refers to the number of species that inhabit a community. We have good, if biased records of diversity over the Phanerozoic Eon, that show tremendous fluctuations, but an overall increase in diversity through time. Of course, at present, we are facing potentially catastrophic loss of diversity.

7.2.1 Food webs and nutrient supply

A *food webs* depicts the feeding connections in an ecological community, and so is a more holistic depiction of a food chain. The foundation of a food web is made up of the *primary producers*, which are those organisms that harvest sunlight, or less abundantly, chemical energy, to survive and build tissue. All other trophic tiers ultimately rely on these primary producers, even if they themselves occupy a lofty tier. That is, a lion is a *carnivore* and might eat *herbivores* like wildebeests and gazelles, and not grass, but without grass, herbivores cannot survive, and so won't be around to nourish the lions. *Parasites* are unsavoury members of a food web that feed on living organisms, but do not tend to cause death because their own life depends on the life of their unwitting host. *Scavengers* such as vultures, on the other hand, feed on already dead organisms. All trophic levels in the food web are subject to disease and will decay after death, with the help of decomposers (mostly, but not entirely, bacteria).

A food web is fundamentally limited by the supply of nutrients to nourish the primary producers. The principal *macronutrients* are C, H, O, P, K, N, S, Ca, Fe, and Mg. Lesser amounts of the *micronutrients* Si, Cl, Na, Cu, Zn, and Mo are also necessary to sustain a food web. Sunlight (or energy-rich chemical species) can also be considered a sort of nutrient. In any community, primary production, and hence the biomass of the entire food web, is limited by some nutrient or nutrients. In the oceans, for example, iron is the limited nutrients in some regions around the southern ocean, and N can become limiting in anoxic settings. However, ultimately, P is the principal limiting nutrient to primary productivity in the surface ocean over geological time.

Because the primary producers lie at the base of food webs, any perturbation to them would effect the entire ecosystem. The populations in the upper tier of the food web tend to be larger, have higher energy requirements, and live longer. Hence they take much longer to recover (if they recover at all) from any major perturbation to the ecosystem.

7.2.2 Biogeography

The distribution of groups of organisms across the globe is its *biogeography*. This range is limited by a variety of factors, the most important of which is temperature. However, many other factors also control biogeography, such as calcite and aragonite saturation for $CaCO_3$-secreting organisms, availability of fresh water, and nutrient supply. History also controls biogeography, for species that originate in a certain area may be prevented by geographic barriers or hostile environments from migrating to other areas that have the right conditions for them to survive. Humans, for example, which originated in Africa, migrated into Europe and Asia relatively early on, but only made their way to Australia and the Americas much later, even though these continents were just as inhabitable as Africa.

7.3 The Atmosphere

The atmosphere is the thin envelope of gases that surrounds Earth. While there is no clearly delineated upper boundary to the atmosphere, 97% of it is confined to the lower 30 km above the Earth's surface. The atmosphere play two key roles in the Earth system: it regulates climate, and it is a reservoir, or market place if you will, of elements and molecules that are used by organisms.

7.3.1 Composition

The atmosphere is made up overwhelmingly of nitrogen (N_2) and oxygen O_2. But many other gases occur in the atmosphere, a number of which are crucial to life as we know it (Table 7.3.1). Water is of course an important component of the atmosphere, but its concentration is highly variable, and it is generally just treated as being along for the ride. You all know that oxygen is important, but so also are the various greenhouse gases, which help maintain reasonable temperatures on Earth.

7.3.2 Greenhouse gases

The composition of the atmosphere has varied through Earth's history. For example, early in Earth's history carbon dioxide and methane concentrations were likely much higher. Most Earth scientists agree that CO_2 levels have decreased gradually over the planet's history to compensate for increasing solar output. That is, solar evolution models predict that the early Sun was about 30% dimmer than it is today and has increased in luminosity roughly linearly since. The geological record, however, indicates that there has been liquid

Table 7.1: The principal and biologically important gases in Earth's atmosphere (excluding H_2O.

Gas	Concentration	Function
N_2	78.1%	source of N for N-fixing bacteria
O_2	20.9%	source of oxygen for aerobic organisms
Ar	0.9%	geochemical analyses via plasma source
CO_2	400 ppmv	greenhouse gas, nutrient for photosynthesizers
Ne	18.2 ppmv	inert gas
He	5.2 ppmv	inert gas
CH_4	1.8 ppmv	strong greenhouse gas
H_2	0.6 ppmv	tends to escape from atmosphere
O_3	>0.07 ppmv	shields ultraviolet radiation

water on Earth at least as far back as 3.5 Ga, and probably as early as the Hadean. There is evidence for late Archean (c. 3.2 Ga) and Paleoproterozoic (c. 2.3 Ga) glaciation, but for the most part, the early Earth was not frozen. This contradiction was termed the *faint young sun paradox* by Carl Sagan.

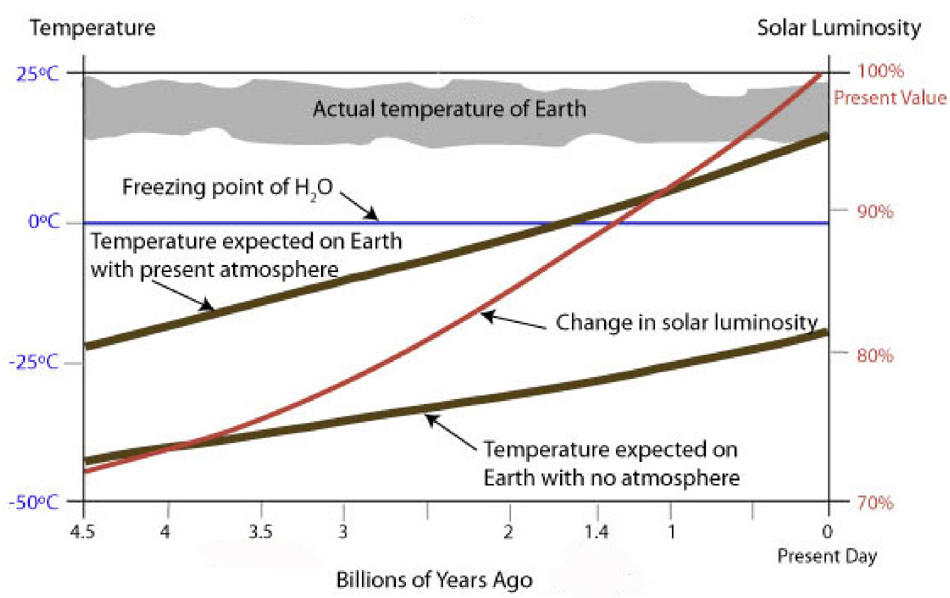


Figure 7.1: Illustration of the Faint Young Sun paradox, which shows that given the current atmosphere, the Earth's surface should have been frozen from

Earth's energy balance

The reason the early Earth was not frozen is simple: the cool sun was compensated for by elevated concentrations of greenhouse gases, namely CO_2 and CH_4, the latter of which must have mostly disappeared from atmosphere by about 2.45 Ga when O_2 first started to accumulate in the atmosphere at greater than trace levels (we'll discuss this topic more in a subsequent lecture). The role of greenhouse gases and solar luminosity is easily understood in terms of the *Energy Balance Equation*:

$$4f\sigma T_e^4 = S(1 - A) \tag{7.1}$$

where,

f: greenhouse transmission factor

σ: Stefan-Boltzmann constant

T_e^4: Average surface temperature on Earth (in K)

S: Incident solar energy (at the surface of the atmosphere)

A: Average planetary albedo (a number between 0 and 1)

The silicate weathering feedback

CO_2 levels are regulated over geological time scales (>10's or 100's of thousands of years) by the *silicate weathering feedback*

- $CO_2 + H_2O \rightarrow H_2CO_3$
- $CaSiO_3 + H_2CO_3 \rightarrow Ca^{2+} + 2HCO_3^- + SiO_2 + H_2O$
- CO_2 regulated through direct temperature effect and indirect precipitation effect on weathering of silicate minerals on the continents
- The geological carbon cycle
  - Mantle/metamorphism source of CO_2
  - Carbonate and sedimentary organic matter ultimate sink

7.3.3 Atmospheric oxygen and organic carbon burial

Earth's atmosphere did not always contain free oxygen. So where did this lovely O_2 come from? Most of it came from photosynthesis, which, as you well know, converts CO_2 into *reduced* C in the form of organic matter, which consists of mainly proteins, lipids, and carbohydrates. But we are all familiar with the smell of decaying leaves and know that animals and bacteria alike eat plants and other animals, resulting in the reoxidation of that organic matter through respiration. We can summarize this two way street of C reduction and oxidation as

$$CO_2 + H_2O \rightleftharpoons CH_2O + O_2 \tag{7.2}$$

You, your dog, trees, and all organic material exposed to the atmosphere are inherently thermodynamically unstabe—that is, out of equilibrium with a highly oxidizing atmosphere, which wants to convert us back into CO_2. So how then does O_2 manage to accumulate in the atmosphere? The answer is in the equation: we need to remove the CH_2O from the the cycle. The earth does this for us by burying a very small fraction of all organic material that is produced (<0.1%) in sediments, mostly in shales. Some of this organic matter subsequently reinvents itself as oil or natural gas (collectively known as *petroleum*), which is why petroleum geologists like shales (think the Bakken and Marcellus shales). Most of this petroleum will eventually be reoxidized, either gradually as it migrates through seeps to the oceans or atmosphere, or much more quickly by airplanes and automobiles. The most *recalcitrant* of the organic matter remains in sediments as *kerogen*, and this is really the counterweight to the O_2 in our atmosphere.

The O_2 content of the atmosphere is strongly regulated by *negative feedbacks* in the oxygen cycle. The more O_2 there is around, the harder it is for organic matter to go unnoticed by oxygen and succeed in being buried. If the Earth was to find some way to start pumping new O_2 into the atmosphere, forests would eventually start to combust spontaneously ! On the flip side, if O_2 levels go down a bit (as they are currently, due to fossil fuel burning), it becomes easier to bury organic matter, and hence to build up O_2 again. Other important feedbacks regulate oxygen through nutrient recycling.

7.3.4 Carbon isotopes

The oxygen cycle and the history of O_2 in the atmosphere is both more complicated and much more interesting than all of this. We will spend more time discussing atmospheric O_2 in subsequent lectures. But before we move on, we should first discuss carbon isotope ratios, which are a powerful tool for querying the ancient carbon cycle, as well as carbon flow at many different spatial scales. Carbon has two stable isotopes: ^{13}C and ^{12}C, of which the latter is by far more abundant. The ratios of ^{13}C/^{12}C are variable in different C-bearing molecules, and this difference arises through *fractionation*, whereby a physical, chemical, or biochemical process results in differences in this ratio between one region and another or between reactants and productions. This fractionation arises because the heavier isotope is more sluggish when both isotopes are freely bouncing around, and because it also tends to form stronger bonds that are harder to break.

Life being intrinsically lazy, much prefers to use to use ^{12}C when it synthesizes organic molecules. The result is a very strong fractionation between the oxidized carbon pool it uses (atmospheric CO_2 or dissolved CO_2 in waters) and the organic materials it makes.

By convention, we do not refer to carbon isotope by the raw ^{13}C/^{12}C ratios. Rather, we use the *delta* notation, through which we normalize them a *reference* ratios and then multiply by a thousand, so that we can express them in *per mil* units. The result is numbers that are much easier to deal with than the ratios themselves:

$$\delta^{13}C_{VPDB}(‰) = \left(\frac{^{13}C}{^{12}C}_{sample} / \frac{^{13}C}{^{12}C}_{vpdb} - 1\right) \times 1000 \tag{7.3}$$

where VPDB referes to the reference ^{13}C/^{12}C ratio we use for normalization.

Organic matter, such as me and most *sedimentary organic matter* (SOM), has very low $\delta^{13}C_{vpdb}$ values (probably about -25‰) compared to the average CO_2 emitted by volcanoes (the ultimate source of CO_2 in the atmosphere-ocean system), which is about -5‰. dissolved inorganic carbon (DIC), in the oceans, which is close to 0‰. The reason the oceans are *heavier* than the source of carbon is that about 20% of the carbon removed from the oceans is SOM, which by virtual of being isotopically so negative, buoys the isotopic composition of the remaining reservoir DIC. The result of the carbon is removed from the ocean as carbonate, which has nearly the same isotopic composition as the seawater from which it precipitates. Consequently, we can measure the δ^{13}C of carbonates as a proxy for the δ^{13}C of the ancient ocean and broadly track fluctuations in the amount of SOM being buried.

Changes to the the carbon isotope composition of seawater can be either steady state, result from a change in the parameters controlling the isotopic composition of inflow and out flow of carbon, or non-steady state, in which the volume of the reservoir changes. These produce very different isotopic responses, the former typically being more gradual and the latter, typically more abrupt, followed by a gradual recovery.

7.3.5 Circulation

Wind happens, which means that the atmosphere is mobile. Wind is the result of pressure differences, facilitated by the low density and viscosity of the atmosphere. On a global scale, circulation is driven by heating in the low-latitudes, which drives air upwards, followed by cooling and sinking in the higher latitudes. However, the fact that the Earth rotates complicates what would otherwise be a very simple convective cell. Rotation gives rise to the *Coriolis effect*. Quite simply, this deflects the trajectory of air masses clockwise in the northern hemisphere and counterclockwise in the southern hemisphere. The result is that the hot air masses that rise in the tropics gradually deflect further and further to the right as they drift north (and to the left in the southern hemisphere), such that they cannot make it to the poles and instead eventually sink (as cool, dry air) in the sub-tropics. There, the air masses diverge north and south. In the northern hemisphere, the air moving north deflects towards the east giving rise to the *westerlies* in the mid-latitudes, and the air moving south deflects westwards as it approaches the equator, generating the *trade winds*.

The combined effect of convection and the Coriolis effect produces a general circulation of the atmosphere dominated by three convective cells in each the northern hemisphere and southern hemisphere, known as *Hadley cells*. These Hadley cells exert a major control on climate, accounting for the large amount of rainfall in the tropics in the *intertropical convergence zone* (or ITCZ as it is commonly abbreviated), where warm moist air rises, and the arid conditions in the subtropics, where cool, dry air descends.

7.4 The Terrestrial Realm

As you know from the hypsometric curve, most of the terrestrial realm sits at relatively low elevations above sea level. One stark feature of the present terrestrial realm is a large temperature gradient between the poles and the equator. It is this temperature gradient that gives rise to the atmospheric circulation, but this is not enough to even it out. This gradient also begets a diversity of climates. The type and diversity of vegetation is closely coupled to climate, and because vegetation forms the base of of community food webs, there is a strong geographic control on the major terrestrial communities on Earth:

- Rain forests

- Deserts
- Mediterranean; chaparral
- Savannahs
- Temperate grasslands
- Temperate forests
- Mountains (and glaciers)
- Northern coniferous forests
- Arctic tundra

The major atmospheric circulation patters we learned about in the previous section have the greatest control on precipitation patterns, which along with temperature, control vegetation types. Hence, the sub-tropics, where dry air descends and warms on its way down, are dominantly desert and largely sparse in vegetation. Most rain forests, on the other hand, straddle the equator, in the ITCZ.

The other major control on precipitation patterns is mountains, in particular those along coast lines. Mountains deflect passing air upwards, causing it to cool and drop its moisture. The air then descends down the other side of the mountains, warms up again, and is dry. Hence, *rain shadows* form on the leeward side of mountain ranges (that is, on the downwind side). This is why Victoria, BC is relatively dry and sunny compared to Seattle and Vancouver, and why the west side of the Sierra Nevadas are significantly wetter than the east side.

Seasons also exert a strong control on precipitation patterns. Mediterranean climates have hot and dry summers and cool and wet winters. Savannahs tend to receive their rainfall in the warm, summer months, as the ITCZ shifts. Similarly, the great Asian monsoon occurs during the summer, when the ITCZ migrates northward, and prevailing winter air pattern of offshore winds shifts to strong onshore winds, that deliver extraordinary amounts of humidity that drops as those air masses warm and rise over the continent.

Mountains (or better yet, elevation) also give rise to climatic zonation, that parallels, at a much smaller scale, the changes in climates from the low to high latitudes. The change in temperature, with elevation, is called the *lapse rate*, and is about 6.4°C/km, on average (a bit lower under very humid conditions, and a fair bit higher under dry conditions). So you can see it does not take terribly high mountains to produce arctic-like conditions, even at low latitudes.

Some of the best clues we have to determining ancient climate zones from the geological record come from the fossil records of plants. This record only stretches back to the middle Paleozoic, and is only augmented by flowering plants for the past 80 to 90 million years. Nevertheless, plants provide a powerful means of quantifying ancient climates, because they display patterns that are a function of climatic conditions. For example, the percentage of smooth leaves (versus jagged-edged leaves) correlates almost perfectly with temperature. Similarly, the size of leaf stomata reflect CO_2 abundance and humidity.

7.5 The Marine Realm

Most sediments accumulate in the oceans, and those that do accumulate in the oceans have a much better chance of escaping subsequent erosion than do sediments deposited in the terrestrial realm. For this reason, our knowledge of ancient environments and the fossil record are heavily skewed towards the marine realm. This said, deep water sediments (i.e. those deposited directly on ocean crust) are not well preserved in the geological record, because they tend to subduct. However, we have access to such sediments dating back to about 200 Ma through deep sea sediment cores and the odd sliver of deep sea sediments preserved elsewhere in mountain belts.

7.5.1 Ocean currents

Surface ocean currents are largely driven by the wind, with modification by the Coriolis effect. For this reason, major ocean currents parallel major wind patterns, with strong easterly equatorial currents westerly mid-latitude currents. But unlike the atmospheric circulation, the ocean currents are constrained by the location of continents. Hence, when these currents collide with the continents, they deflect counterclockwise in the southern hemisphere and clockwise in the northern hemisphere. They also generate gravitational counter currents.

Subtropical gyres form in all of the ocean basins straddling the equator, and their sense of rotation is again governed by the Coriolis effect. The centre of these gyres tend to be relatively barren, because they are cut off from sources of nutrients. A branch of the North Atlantic gyre, known as the *Gulf Stream*, flows northward towards western Europe and eventually the North Atlantic. The current carries warm, salty water northward and accounts for the relatively temperate climates in northwestern Europe (compared to equivalent latitudes in North America).

The warm, salty water of the Gulf Stream eventually cools off in the polar regions, and because it is saltier than surrounding waters, becomes denser and sinks. This formation of *North Atlantic Deep Water* sets up the great conveyer belt of seawater known as the *thermohaline circulation*, which delivers deep waters throughout the Atlantic Ocean. There is only week formation of deep waters in the northern Pacific Ocean, and for this reason, the deep waters in the Pacific are much older (and more acidic) than those in the deep ocean.

The Antarctic *Circumpolar Current* flows clockwise around Antarctica. This strong current effectively cuts of the *Southern Ocean* from the Atlantic, Indian, and Pacific oceans, which helps account for why Antarctica is so much colder than the northern polar region.

7.5.2 Climatic oscillations

Certain quasi-periodic climatic phenomena result from oscillating atmospheric air pressures, sometimes coupled to fluctuating oceanic conditions. These are known as oscillations and can play an important role in regional weather from year to year.

El Niño-Southern Oscillation (ENSO)

El Niño refers to an interval of unusually warm surface waters in the eastern tropical pacific. It corresponds with high atmospheric pressures in the western tropical Pacific and low pressures in the eastern Pacific. This pressure gradient drives easterly wind currents that drag warm western Pacific water eastward. The result is unusually high temperatures and abnormally high levels of precipitation in the eastern tropical and even sub-tropical Pacific.

La Niña refers to the opposite scenario, where atmospheric pressures are higher over the eastern Pacific, setting up easterly winds. The drag from these winds causes cold, deep Pacific water to upwell, resulting in anomalously cool surface waters in the eastern Pacific, and correspondingly cooler weather and lower than average rainfall.

These two states tend to oscillate back and forth with a periodicity of about 5 years, hence the term *oscillation*.

North Atlantic Oscillation

Another oscillating climatic phenomenon is the North Atlantic oscillation (NAO), although this is more an atmospheric feature. It is defined by oscillating air pressure differences as recorded in the Azores and Reykjavík, Iceland, which lie at the centre of permanent air pressure highs and lows, respectively. The NAO results in weakening and intensification of this pressure difference with a periodicity of about 10 years. When the pressure difference are higher, they strengthen the westerlies, which transport warm, most air to western Europe. The result is milder winters and summers. When the pressure differences diminish, the westerlies diminish, resulting in more extreme weather over western Europe.

7.5.3 Marine environments

We commonly subdivide the marine realm physiographically into the *continental shelf*, *continental slope*, *continental rise*, and *abyssal plain*, which is the deep sea floor. As you know the mid-ocean ridges stand tall above the deep sea floor, and are underwater mountain ranges where new oceanic crust is made.

Epicontinental seas are shallow, typically somewhat isolated or restricted seaways that spread out over large swaths of low-lying continents during period of high sea level. Because we are in a glacial epoch, there are few notable epicontinental seaways. An excellent ancient example is the *Western Interior Seaway*, which covered much of central North America about 100–70 million years ago, and connected the Gulf of Mexico to the Arctic Ocean.

There are various different zones in the marine realm that are distinguished based on their hydrographic, biological, and biogeochemical conditions.

- Supratidal zone
- Intertidal zone
- Shallow subtidal zone

- Storm wave base
- Below storm wave base
- Photic zone
- Oxygen minimum zone (OMZ)

The photosynthesis that take place in the photic zone is mostly carried out by single-celled, cyanobacteria and algae, that are collectively referred to as *phytoplankton*, meaning they are small photosynthesizes that drift in the surface ocean. The production of primary biomass by photosynthesizers tends to deplete the surface ocean in nutrients. After death, this organic material begins to settle through the water column, effectively exporting carbon and nutrients to the deeper ocean. Much of this carbon is consumed or decays before it reaches the seabed, releasing CO_2 and nutrients into the deeper ocean. At the same time, this decay consumes oxygen, leading to the development of an *oxygen minimum zone* where rates of O_2-consumption peak in the water column.

The nutrients released by decaying phyotoplankton and the organisms that consume them are returned to the surface in zones of upwelling. Such upwelling zones, such as the margin offshore Peru, tend to be highly fertile and productive.

Zooplankton are small animals, including many crustaceans, that drift in the surface ocean and feed upon the phytoplankton. The *nekton* include the animals that swim, and these include animals that both feed on phytoplankton and on zooplankton. Together, the plankton and nekton constitute the *pelagic* life in the oceans, mean they live and move about in the water column. The group of animals that live on the seafloor (or in shallow sediments) are known as the *benthos*. These include both mobile organisms, and *sessile* organisms, such as sponges and corals, which are attached to the seafloor.

Chapter 8

Origin of the Earth and the Hadean

Reading: Chapter 11 in Stanley; Chapter 8 in Wiccander and Monroe

8.1 Introduction

Earth's history is commonly broadly separated into two temporally unequal halves: the *Precambrian* and the *Phanerozoic*, which roughly correspond to the time before and after the first appearance of large eukaryotes. The Precambrian, which is not an official name in the Geological Time Scale, is subdivided into the Hadean, Archean, and Proterozoic. While what we know about Earth's history is overwhelmingly biased towards the Phanerozoic (the past 541 million years), the Precambrian encompasses the lion's share of geological time. Before digging into Early Earth's history, we will begin with a brief review of the origin of the solar system.

8.2 Origin of the Solar System

Earlier in this course, we discussed the age of the Earth, which we now know quite precisely from the age of meteorites. These meteorites reveal extraordinary information about the early history of the solar system and Earth's own early history. For example, some meteorites show evidence of *differentiation*, with one class of meteorites representing the iron-rich core of a differentiated body (iron meteorites) and other meteorites more closely resembling magmatic rocks (stony meteorites). Undifferentiated, *chondritic* meteorites, on the other hand, provide an indication of what Earth's bulk chemical make-up might have been like when it formed. Hence, geochemists often compare the geochemistry of certain chemical reservoirs in the Earth or rocks with average chondritic values as a measure of the differentiation that has occurred on Earth.

Our sun and the planets in our solar system formed from the leftover material from a *supernova*, an exploding star that left a wake of material in the form of a slowly rotating nebula. As this nebula cooled, condensed, and flattened, it began to spin faster due to the conservation of angular momentum. This rotating disk segregated into a central sun and rings of material that rotated around it and gradually coalesced into planetesimals.

During this process, the lighter and more volatile elements were concentrated in the outer ring, which went on to form the gaseous planets, like Jupiter and Saturn, while the denser and less volatile elements were concentrated in the inner solar system, forming the rocky planets like Earth and Venus. Based on the abundance of certain daughter isotopes of short-lived radioactive isotopes, we know that the the sun and the planets formed within about 50 million years of one another, and possibly even less.

8.3 Differentiation and the origin of the Moon

The proto-Earth was a hot place, due to the kinetic energy resulting from the frequent meteorite impacts as the proto-planet grew, along with the heat released by radioactive decay, which would have been much greater early in the solar system's history than it is today. This heat melted the early Earth, and the denser parts of this melt, most notably iron and nickel, migrated to the core, while the less dense part rose to the top, forming a proto-mantle. The act of differentiation alone generated enormous amounts of heat through the conversion of potential energy, and this helped sustain a molten early Earth.

Earth's moon is thought to have formed very early in Earth's history, probably as the result of a collision with a Mars-sized planet. Such a model conveniently accounts for Earth's unusually fast spin (for its position in the solar system), and probably remelted Earth, after it cooled initially. We surmise that the moon formed after differentiation, because its bulk chemistry and density differ from the bulk composition of Earth, but quite closely resemble Earth's mantle. Consequently, it would also have blown away any atmosphere that the Earth and moon might have developed by that time as a result of early cooling and differentiation. Indeed, the near absence of water within the crystal structure of lunar minerals attest to the loss of water during the collision.

Strong support of the impactor hypothesis emerged when Clayton and Mayeda (1996) showed that the oxygen isotopic composition of the moon and Earth are indistinguishable. A recent paper on Zn isotopes in lunar rocks (Paniello et al., 2012) had seemed to have confirmed the Mars-impactor hypothesis for the origin of the moon. However, these similarities are not necessarily consistent with the impactor model after all, based on recent modelling of the impact scenario, which suggests that significant isotopic variations should have resulted from the standard mars-sized impactor scenario. So this once staid and accepted model for the origin of the moon is again under scrutiny—an excellent example of how science and our knowledge of how the Earth and solar system evolved is never settled.

8.4 The Late Heavy Bombardment

The Hadean was originally defined as the period of time dating from the origin of the solar system 4.56 Ga to the age of the oldest terrestrial rocks, 3.8 Ga. This time also corresponds to the *Late Heavy Bombardment*, when frequent large impacts would have remelted most early crust that had formed. How do we know that meteorites continue to impact the Earth for hundreds of millions of years after the moon formed when we have no record of these impacts on Earth? The answer is in the moon, which is heavily pock-marked by craters formed by metrorite impacts. The ages of the largest craters range from 3.8 to about 4.5

Ga, and studies of the pockmarks suggest that the rate of impacts was about 1000 times greater in the early Hadean than it is today. In fact, we owe it to Jupiter that the Earth and moon were not more thoroughly peppered by meteorites, for it deflects most would-be impactors of doom. Nevertheless, these impacts were sufficiently large and frequent to add significant material and heat to the early Earth, and presumably sterilize it frequently. It is also due to the early clean-up of meteorites from the solar system that the planets have tilted spin axes.

Rocks and minerals older than 3.8 Ga have now been dated, but we still know very little about the first 0.75 Ga of Earth's history.

8.4.1 Origin of the ocean and atmosphere

The earliest atmosphere was likely composed mostly of H and He (not unlike the sun), but most of these original gasses were lost during meteorite impacts and gradual escape to space.

- Hydrosphere derived from volcanoes and comets
- Salts derived from weathering of seafloor, then continents

Earth's earliest ocean was not particularly salty. Modeling suggests that it would have reached roughly modern salinity by the early Archean, and since that time, the flux of salts into the ocean has been balanced by the deposition of salts.

- Atmosphere derived from volcanic outgassing
- Early atmosphere largely lost from meteorite impacts
- A reducing atmosphere (H_2O, HCl, CO, CO_2, N, CH_4, NH_3)

8.4.2 Origin of continental crust

The earliest 'crust' that would have formed as the magma ocean cooled at its surface would have been basaltic in composition. This proto-crust was simply cooled from the magma ocean, and so was mafic in composition. Hence, there were no continents initially. The formation of distinct continental crust required *differentiation* to form lighter, less dense, felsic rock.

The formation of felsic magmas could only have occurred through the melting of basalt. Hence the the formation of the first continental crust would have resulted from the descent of early, mafic crust back into the mantle—subduction. The earliest evidence for the formation of felsic crust comes from the ages of the oldest zircons in the world: the Jack Hills zircons. These zircons, obtained from the Jack Hills of Western Australia, are dated at 4.404 Ga and are the oldest Earth materials ever dated (Wilde et al., 2001). These zircons appear to suggest that proto-plate tectonics had begun within 150 m.y. of the origin of the solar system—that is, basaltic crust had begun to subduct and remelt, forming felsic magmas. The rocks that would have formed from these early felsic melts would have been sodium-rich, because sodium is a highly incompatible element.

The earliest bits and pieces of granitic crust would have probably been pushed and shoved about on the surface of the early ocean by the still vigorously convecting mantle. Over time, they would have begun to coalesce to form *protocratons*: early, relatively stable continental crust that would eventually assemble to form the earliest continents. These *Precambrian* shields are at the core of all continents.

The Jack Hills zircons are interesting not only because they are really old, but also because they contain oxygen isotope signatures that suggest the involvement of water in the melting of the magmas. Some scientists have argued that this is evidence that an ocean already existed by about 4.4 Ga, although it is really impossible to constrain what the nature of this early ocean might have been like or whether or not it was permanent.

The oldest rock: the Acasta Gneiss vs. the Nuvvuagittuq Greenstone

The oldest actual terrestrial rock ever dated by an uncontroversial method is the Acasta Gneiss, which occurs on the Slave craton in northern Canada and was dated by Bowring and Williams (1999).

- Oldest U-Pb zircon discordant age of 4.06 Ga
- A gneiss, so *protolith* was even older

The same zircons in the Acasta Gneiss that show that the *protolith* formed from a felsic magma also record metamorphic events at 3.75 and 3.60 Ga (Bowring and Williams, 1999). Similar aged gneisses occur elsewhere in the world, notably in Issua Greenland, where rocks as old 3.8 Ga preserve what closely resembles modern oceanic crust. Hence, it is clear that by about 3.8 Ga, not only had continental and oceanic crust similar to that found today already begun to form, but also these rocks were participating in collisional events that drove regional metamorphism.

The prize for the oldest rocks on Earth was challenged a few years ago when then McGill PhD student, Jonathon O'Neill argued that a patch of rocks in Ungava Québec might be as old as 4.3 Ga, based on a new dating method that uses an extinct Nd isotope system (O'Neil et al., 2008). This technique and the suggestion that these are the oldest rocks is controversial. The oldest zircons in these rocks date to 3.8 Ga, which still means they are extremely old. To the extent that they are in fact older, they provide a rare glimpse into conditions on the earliest Earth.

It is clear that differentiated continents had already begun to drift about by the late Hadean or early Archean. Whether or not this constituted plate tectonics as we know it today remains a hotly debated question. As you will see in the following chapter, certain rock associations that are common in Archean terranes are in fact quite distinct from those that occur in younger rocks. These *greenstone belts* are one of several lines of evidence that skeptics of modern-style (horizontal) Archean plate tectonics (e.g., Bédard et al., 2003) highlight in their arguments. In fact, these arguments beg the question: what are the essential criteria for plate tectonics?

Chapter 9

Earth's Earliest Record: the Archean

9.1 The Archean

The Archean Eon spanned from 3.8 to 2.5 Ga. The Earth's interior was still very hot during this time, and as a consequence, it seems that no large continents like those found today had formed. Nevertheless, cratons stabilized during this time and recognizable plate tectonic processes were active, although noticeably different from those today. The Archean also preserves the first putative and first indisputable evidence of life on Earth. The Archean Eon ended with the Great Oxidation Event, during which free O_2 accumulated in the atmosphere for the first time.

9.1.1 Greenstone belts

Greenstone belts are elongated terranes with a synclinal structure that consists of mafic to ultramafic volcanic rocks and mafic-derived sedimentary rocks (turbidites and mudstones), and sometimes *banded iron formation* (see below) that have been metamorphosed to *greenschist* facies (hence their name; green, because they contain abundant metamorphic *chlorite*) and are commonly intruded by granite. Greenstone belts tend to be sandwiched between granite-gneiss complexes (typically highly metamorphosed gneiss of granitic origin).

- Restricted to Archean (and possibly Hadean)
- Often occur in parallel belts, perhaps forming in arc, back-arc, or rift settings
- Often include *komatiites*—ultramafic lava flows, which are rare in post-Archean world
- Reflect hotter interior of the Earth and early version of plate tectonics (although precisely how they formed remains a mystery)
- Earliest record of subaqueous sedimentation

The origin of greenstone belts has long intrigued geologists and continues to be a source of much debate. One only has to read several different Earth and Life History textbooks to see the differing views. Whereas Stanley's book seems to support an intercontinental drift model, others advocate for an Arc model to account for greenstone belts. I personally think that back-arc basins are the most likely tectonic setting for greenstone belts. *Back-arc basins* form on the opposite side of a subduction zone when the subducting slab begins to *roll-back*. This essentially means that the forward motion of the subducting slab does not outpace the sinking motion. Sinking of slabs would have been quicker in a hotter Archean mantle, and so roll-back would have been common. This shift to slab roll-back places the arc into an tensional regime, and the weak back-arc region is where the stretching occurs, eventually leading to volcanism and the opening of a small ocean basin. The *supracrustal* sequence of ultramafic then mafic volcanics, then sediments, sometimes with interbeddd felsic volcanics, would have been deposited in this environment. As the foundering slab sinks, it would have melted, and this melt would potentially be over quite a broad area, accounting for the widespread intrusion of granitoids into the supracrustal sequence. Subsequent collision would have folded the basin and enveloping granitoids, much of which were probably still quite warm, creating the synforms.

9.1.2 Banded iron formation

Banded iron formations, or BIFs, are layered, iron-rich sedimentary rocks that are a common component of the Archean and Paleoproterozoic sedimentary record. Archean BIFs are volumetrically less significant than their Paleoproterozoic counterparts and commonly formed in close association with mafic volcanics (e.g. in greenstone belts). This relatively restricted BIF is commonly referred to *Algoma-type* BIF, in contrast to the more extensive, *Superior-type* BIF that presumably formed on open marine continental shelves.

- Mostly consist of alternating layers of chert and the iron-oxides hematite and magnetite. It now seems that magnetite in BIFs is all secondary, having formed either through diagenesis (where sufficient organic matter was present to re-reduce some of the iron) or metamorphism.
- Also include iron carbonates, and, where metamorphosed, iron silicates
- Fundamentally require anoxic waters to store large masses of Fe^{2+} in solution
- But that iron has to be oxidized to form the iron-oxides:
  - Oxidation by photosynthetically derived O_2
  - Photochemical oxidation
  - Bacterial iron oxidation

9.1.3 Growth of continental crust

Most geologists would agree that significant continental crust existed by the Archean, but there remains significant disagreement over the tempo and timing of crustal growth. Some would have nearly an equal area to today formed already by the beginning of the Archean, whereas authors would argue that these levels were not achieved until the latter Proterozoic or even the Phanerozoic. Why is the timing and tempo of crustal growth important? Well, for one reason, weathering of the continental crust plays a first order role in regulating both climate and seawater chemistry, including nutrient availability. Recently, a variety of isotopic approaches are beginning to converge on a model whereby continual crust grew in volume linearly through the Hadean and most of the Archean at a relatively rapid rate,

and then slowed down, but continued to grow linearly, after about 3 Ga (Dhuime et al., 2012). This model is compelling because various different lines of evidence appear to point to something rather significant happening in terms of mantle convection and plate tectonics at this time.
Also about 3.0 Ga, the first known cratons began to stabilize.

- *cratons* are the stable cores of ancient continents
- oldest known stable craton is about 3.0 Ga, represented by mixed carbonate-clastic Witwatersrand and Pongola groups in South Africa and Swaziland. The development of a carbonate platform requires a stable carton
- sediments contain abundant structures indicating shallow water and even glacial deposition (oldest evidence of glaciation, c. 2.9 Ga; Young et al., 1998).

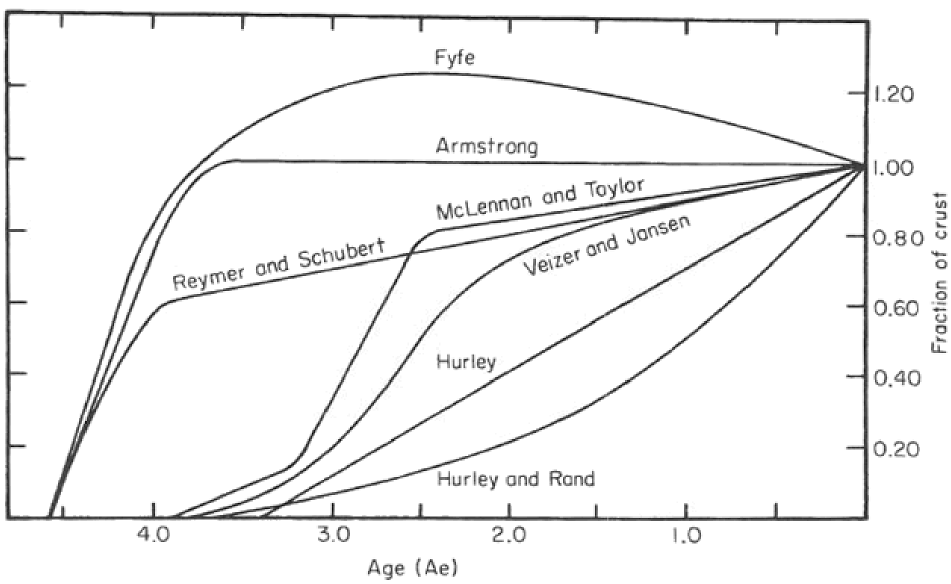


Figure 9.1: Various models for the tempo of crustal growth (plotted as fraction of continental crust relative to the present through time), from Taylor and McLennan (1985)

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Precambrian shields are regions of dominantly Precambrian-aged cratonic rocks exposed at the Earth's surface. The Canadian Shield is perhaps the most impressive example, and as would be expected for such a large tract of very old rocks, consists mostly of crystalline rocks. The Precambrian shield is stitched together by various ancient orogenic belts, that represent collisions between the ancient cartons.

9.1.4 Archean life

It is impossible to say for certain when life first evolved (some think it may have done so multiple times in the Hadean, only to be annihilated by large meteorite impacts. Before even looking at the evidence for early life, however, it is worth considering what is meant by life.

- Able to metabolize
- Have some barrier from outside world
- Able to reproduce
- Able to produce complex molecules

This definition of what is required for life leads us to a list of essential components for life to originate:

- Amino Acids, which are relative simple organic compounds, which strung together form the more complex proteins that are catalyze chemical reactions and act as building materials
- Nucleic acids, large and complex acids: ribonucleic acid (RNA), which is simpler but less stable, and deoxyribonucleic acid (DNA), which is more complex and more stable
- Organic phosphorus compounds. Adenosine triphosphate (ATP) transforms energy from light or chemicals into the energy required by cells. Phospholipids provide structure to cell walls and create a semi-permeable membrane

It is conceivable that the building block organic materials required for life to originate were delivered by comets and meteorites during the Late Heavy Bombardment. Stanley Miller showed in the 1950's that sparks of electricity could actually stimulate the formation of amino acids and a variety of other somewhat complex organic molecules from the inorganic compounds that likely existed in the early atmosphere.

How did amino acids aggregate to poem proteins? Some scientists have suggest that clays provided a necessary template through the metal ions on their surfaces. These could have attracted amino acids and arranged them in an orderly way.
Many people believe life originated in extreme environments because some extremophiles occur on the deepest branches of the tree of life. However, others, including Andrew Knoll, think that life most likely originated in 'normal' environments and later moved into the extreme environments through environmental pressures.

- Required a source of appropriate building blocks for organic molecules
- A source of energy

The earliest indirect evidence for life comes from carbon isotope compositions of concentrated carbon in a highly metamorphosed gneiss in a 3.8 Ga Akilia Gneiss in southwest Greenland

- Low δ^{13}C values imply organic synthesis
- Debate over whether this is the only process that can produce low carbon isotope values
- Debate over whether or not the protolith was sedimentary or granitic

The earliest direct evidence for life on Earth comes from 3.5 Ga stromatolites and microfossils in Western Australia

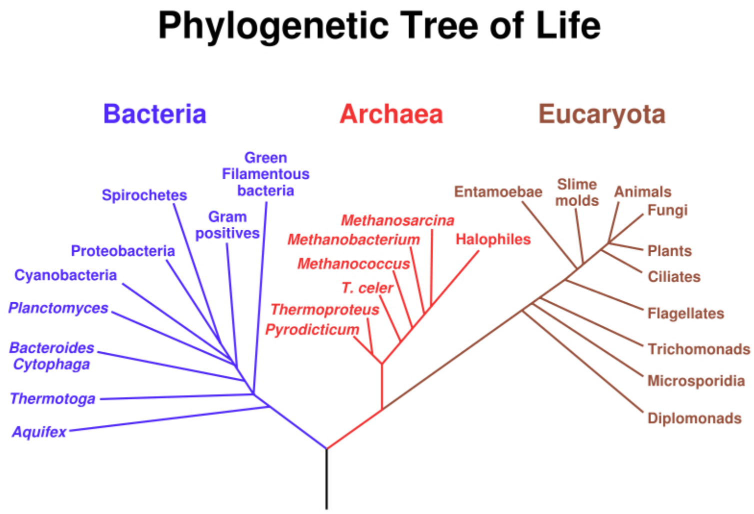
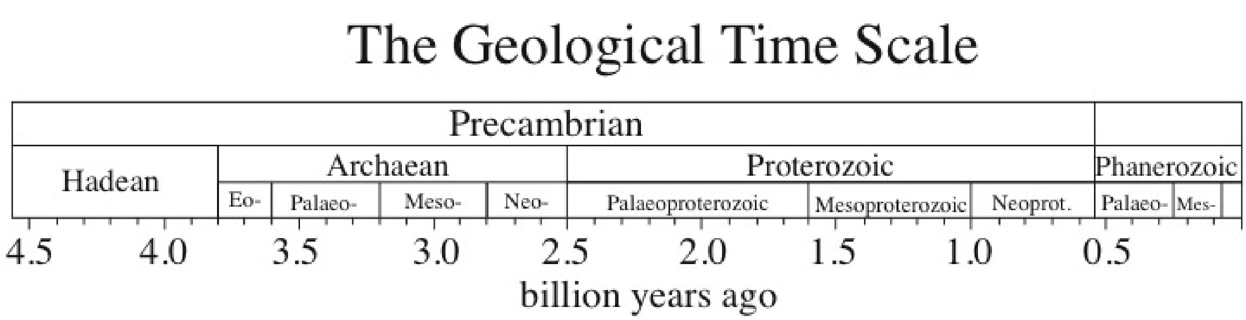


Figure 9.2: A simplified tree of life that suggests that extremophiles are amongst the lowest branches

- (Grotzinger and Knoll, 1999) defined *stromatolites* as "attached, lithified sedimentary growth structures, accretionary from a point or limited surface of initiation"
- Biogenic stromatolites can be distinguished from non-biogenic stromatolites
- Biogenic origin of microfossils also debated, but most agree that there are biogenic structures that represent early prokaryotic cells

There is controversial evidence that Eukaryotes first appeared in the Archean, c. 2.7 Ga based on the occurrence of sterane *biomarkers*.



Chapter 10

The Great Oxidation Event

10.1 The Great Oxidation Event

The *Great Oxidation Event* was the appearance of free oxygen in the atmosphere about 2.45–2.35 Ga. Whereas geologists have long recognized this event based on the geological record (and some theoretical considerations), the onset of this event is now virtually unambiguously pinpointed in the stratigraphic record by sulfur isotope data (Fig. 10.2).

10.1.1 Geological evidence for an early anoxic atmosphere

By the late 1950's, geologists and geochemists like Preston Cloud, Heinrich Holland, and Robert Garrels recognized that the geological record from early Earth's history was significantly different from that of later in Earth's history and that this was due, in part, to different atmospheric oxygen levels. The occurrence and pattern of banded iron formation was an important clue that Earth's early environment lacked sufficient oxygen to ventilate the deep oceans. It was the anoxic oceans that permitted ferrous (Fe^{2+}) iron to accumulate in the oceans. Massive iron formation disappeared from the geological record by about 1.8 Ga, and it was long argued by Holland and others that this marked the oxygenation of the deep oceans.

In addition to the BIF record, there is other geological evidence that pointed specifically to an Archean atmosphere with no free oxygen and geochemical evidence that suggest an important change in ocean geochemistry sometime in the Paleoproterozoic.

Redox sensitive minerals

Fluvial sediments in the Archean contain certain detrital minerals that are not stable under an even mildly oxidizing environment: pyrite, siderite, and uraninite. By the early Proterozoic, these distinctive minerals had disappeared from the sedimentary record.

Soil profiles

Archean soil profiles are noticeably depleted in iron, which could only have occurred if there was no free oxygen around. Otherwise, iron would have been retained in the soils as iron oxide and oxyhydroxide minerals. By the Paleoproterozoic, iron had begun to be retained in soil profiles, indicating free atmospheric O_2.

10.1.2 Growth of the marine sulfate reservoir

Sulfate is abundant in the modern ocean, but this has not always been the case. Various lines of evidence indicate that Proterozoic oceans had much less sulfate overall, and that the Archean ocean had almost no sulfate. Geologically, this is seen in the complete absence of calcium sulfate evaporite deposits (gypsum or anhydrite) of Archean age. The first significant sulfate deposits are found in the ca. 2.3 billion-year old Gordon Lake Formation in the Huronian Supergroup in Ontario and possibly equivalent rocks in the Lucknow Fm. of South Africa (Rasmussen et al., 2013). However, sulfate deposits remain scarce overall through much of the Proterozoic.

In the late 1990's, Don Canfield published a now famous plot of the sulfur isotopic composition of sulfieds (mainly pyrite) deposited in marine sediments through time. This dataset showed a substantial increase in the spread of these values around the Archean-Proterozoic boundary. In the modern ocean, large differences in the $\delta^{34}S$ values between seawater sulfate and sedimentary pyrites are the result of sulfur isotopic fractionation by bacterial sulfate reducers, which use sulfate as an electron acceptor during metabolism. That is, they reduce the sulfate to sulfide. Oxygen recycling of the reduced sulfur can lead to even bigger differences between sulfate and sedimentary sulfides. Consequently, Canfield interpreted this increased spread in sulfur isotope values around 2.3 billion years ago (Fig. 10.1) to herald growth in the marine sulfate reservoir, consistent with higher atmospheric oxygen concentrations and in increase in oxidative weathering on the continents (Canfield and Teske, 1996).

The Canfield Ocean

Shortly after he published his sulfur isotope curve, Canfield proposed a radical new model for deep ocean chemistry during the Proterozoic. He argued that rather than becoming oxygenated following the end of BIF deposition around 1.8 Ga, the oceans instead remained anoxic, but became sulfidic rather than iron-rich as they had been for the previous two billion years. The word used to describe such conditions is *euxinia*, and the possibility of euxinic oceans had important implications for seawater chemistry and eukaryotic evolutions. The interval of the *Canfield Ocean* ($\sim$ 1.8 to 0.8 Ga) has otherwise been referred to as the *Boring Billion* because it lacks both iron formation and significant carbon isotope anomalies, reflecting a relatively infertile era in biospheric evolution. The Canfield Ocean model seems at first blush to be consistent with this observation, because euxinic conditions would have resulted in vanishingly low concentrations of Mo and some other key nutrients (Anbar and Knoll, 2002). After Canfield first proposed this model, several different datasets bearing on the nature of sulfur and iron chemistry in seawater were published from Mesoproproterozoic (1.6 to 1.0 Ga) which seemed to support the euxinia model (e.g. Shen et al., 2003; Brocks et al., 2005)

The Canfield Ocean understandably received great attention and a large following. However, it has recently been challenged both by new data sets that suggest that some were ferruginous rather than sulfidic (Planavsky et al., 2011), and simple modelling that indicates that the low trace metal concentrations observed during this interval in time do not in fact require widespread sulfidic or even anoxic conditions (Reinhard et al., 2013). I

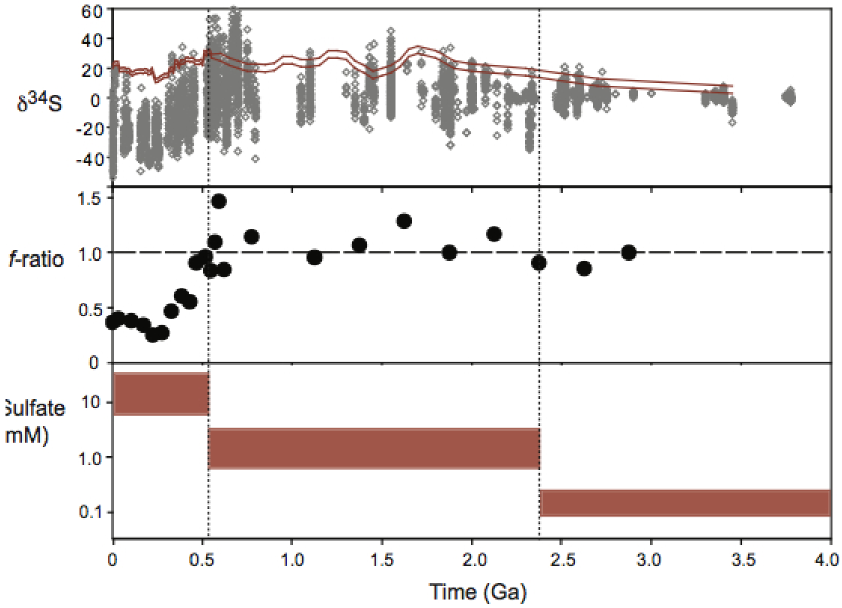


Figure 10.1: The δ^{34}S record of sedimentary pyrite through time (top panel) shows two almost stepwise increases in spread of values around the Great Oxidation event and in the Neoproterozoic. These are though to reflect increased marine sulfate concentrations (bottom panel), the first related to the GOE and the second perhaps related to a second oxidation event or alternatively to the effect of bioturbation on sulfur cycling (Canfield and Farquhar, 2009). f_{pyrite} (middle panel) is a measure of the relative proportion of S buried as pyrite through time (analogous to f_{org}). That it remains close to 1 throughout most of the Proterozoic implies little to no burial of gypsum during that time.

would further argue that the evidence found to support the Canfield Ocean model is also heavily biased by sampling, which is mainly restricted to sedimentary rocks deposited on continental shelves, with no real proxy for open deep ocean conditions. Hence, it is entirely possible that we have completely missed records for what true deep ocean chemistry was like during the *Boring Billion* years, and that even though euxinia may well have been widespread, it was not necessarily pervasive.

10.1.3 Mass independent isotope evidence of the Great Oxidation Event

The clinching evidence for an oxygenation event at ca. 2.45 Ga comes for the record of *mass independent fractionation* (MIF) in sulfur isotopes, specifically the Δ^{33}S record from sedimentary sulfur-bearing minerals. First published by Farquhar et al. (2000), this data set showed that MIF was the norm in sedimentary sulfur minerals in the Archean, but that this unique signature shut off just after the Archean–Proterozoic boundary. MIF occurs today to a very small degree when volcanoes inject sulfur into the upper atmosphere. There, above the UV-shielding effects of ozone, photochemical reactions partition the ^{33}S isotope between reduced and oxidized gaseous species in a non mass-dependent fashion. Although this style of isotopic fractionation is exceedingly rare today, it was apparently the norm in the atmosphere in the Archean due to the absence of ozone. These atmospheric sulfur gases with the MIF signal were then eventually incorporated into sedimentary sulfur minerals by a combination of biological and non-biological processes.

Because ozone in the upper atmosphere is tied to oxygen concentrations, it stands to reason that the MIF sulfur signature is a first-order and unequivocal recorder of an increase in atmospheric oxygen levels beyond some threshold where the MIF signal shuts off. In a clever atmospheric chemistry modelling study, Pavlov and Kasting (2002) showed that this threshold is approximately 10^{-5} times the present atmospheric level (PAL) of O_2.

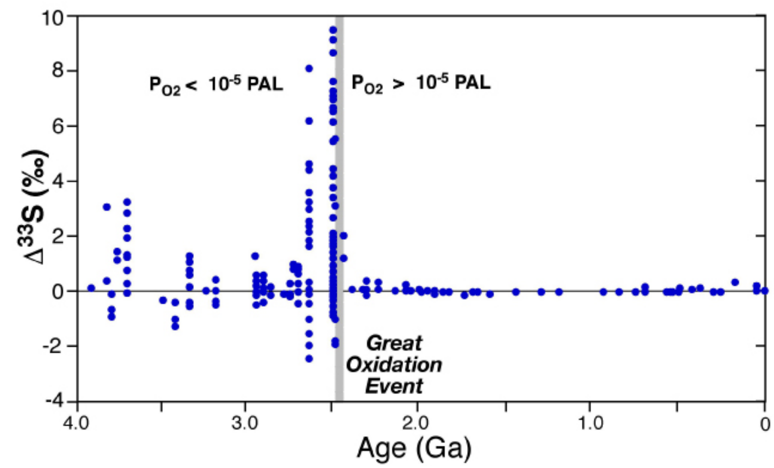


Figure 10.2: The Δ^{33}S record from sedimentary sulfur-bearing minerals shows a rise in P_{O2} over the critical threshold of 10^{-5} present levels (Farquhar et al., 2007)

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Other proxies for oxygen abundance also show increasing oxygenation of Earth's environment. However, it is currently *virtually impossible to quantify P_{O2} during the Proterozoic.*

10.1.4 Source of oxygen and cause of the oxidation event

If the MIF signature in sedimentary sulfur pinpoints where in the geological record oxygen concentrations rose beyond some key threshold, what drove this rise in oxygen in the first place? The most logical answer is that it resulted from the steady production of O_2 by photosynthesizes, who we know inhabited the oceans for at least several hundred million years prior to the GOE. Another possible source of atmospheric oxygen was hydrogen escape, whereby photodissociation of H_2O in the upper atmosphere, followed by escape of the light hydrogen left behind a more oxidizing atmosphere (Catling and Claire, 2005). However, whereas these are presumably the source of that oxygen, why did the GOE not happen earlier?

Once possible answer is that the *sinks* for oxygen outweighed the input of oxygen from photosynthesis and hydrogen escape. We have already learned that banded iron formation (BIF) contains a significant amount of iron oxide and that oxidation of sulfide is the main source of sulfate to the oceans. These would have constituted tremendous sinks for oxygen. Another important sink would have been the flux of volcanic gases to the atmosphere from volcanoes. Many scientists have argued that the oxidation state of these gases might have evolved through time as the result of gradual oxygenation of the mantle or a shift from dominantly subaqueous to increasingly subaerial volcanism (Kump and Barley, 2007).

10.1.5 Early Paleoproterozoic glaciation

A detailed look at the geological and geochemical records of the GOE reveals that it closely corresponded with a series of three glaciations in the early Paleoproerozoic (between 2.4 and 2.25 Ga), at least one of which might have been global in extent (i.e., a snowball glaciation Kirschvink et al., 2000). Whereas one argument has been put forth that snowball glaciation could have triggered oxygenation through the buildup of CO_2 in the atmosphere and subsequent high nutrient deliver to the oceans in the aftermath (Kirschvink et al., 2000), Kopp et al. (2005) proposed instead that oxygenation triggered the snowball by rapidly depleting the atmosphere of methane, which had maintained its greenhouse.

The exact relationship between glaciation and oxygenation remains unclear, but recent dating and correlations between Paleoproterozoic glacial deposits in South Africa and North America suggest that they are too interconnected to be a mere coincidence (Rasmussen et al., 2013).

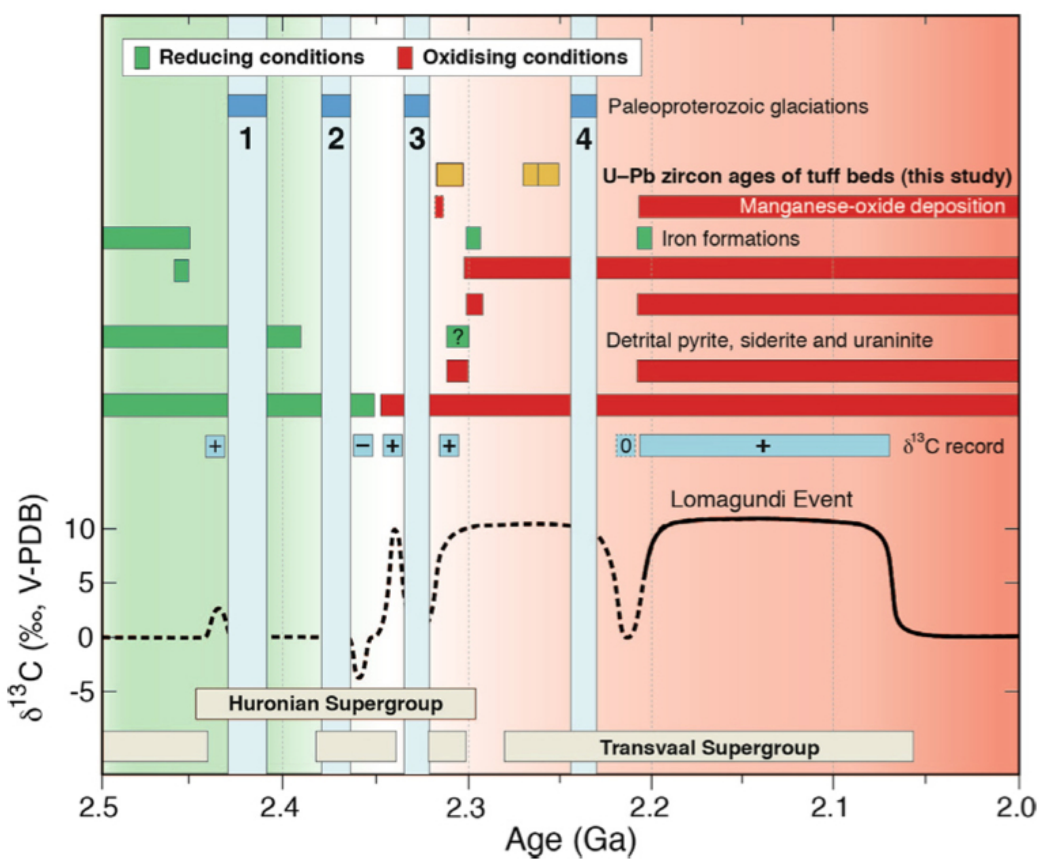

Figure 10.3: Cartoon illustrating the temporal overlap between evolution of atmospheric oxygen (GOE), Paleoproterozoic glaciation, and the δ^{13}C record. From Rasmussen et al. (2013).

10.1.6 The Lomagundi Event

Shortly after the last Paleoproterozoic glaciation, δ^{13}C values in marine carbonates began to rise, and they stayed unusually high ($> 8‰$) for about 150 million years (Fig. 10.3. Long intervals of high δ^{13}C are commonly associated with input of oxygen to the atmosphere due to the implied high rates of organic carbon burial. However, it is interesting to note that the Lomagundi event started after the initiation of the GOE and was over long after the GOE had begun. This paradox highlights that fact that it was not increased carbon burial alone that drove the onset of the GOE, and also suggests that just what the GOE represents perhaps is not so easy to describe or quantify. For example, the MIF-sulfur record tells us when pO_2 surpassed a certain minimum threshold, but it does not actually say anything about continued oxygenation after that. It seems highly likely that oxygen levels continued to climb after the start of the GOE for some several hundred million years, perhaps approaching levels not dramatically different from those of the late Proterozoic or early Paleozoic. New trace metal data and isotope data also suggest that atmospheric oxygen, and by extension, marine oxygen and sulfate levels may have declined abruptly following the Lomagundi event (Planavsky et al., 2012; Partin et al., 2013).

Chapter 11

Early–Middle Proterozoic Life and Environment

Reading: Chapter 12 in Stanley

The Proterozoic spans from 2.5 to 0.542 Ga and is subdivided rather arbitrarily into the Paleo- (2.5–1.6), Meso- (1.6–1.0), and Neoproterozoic (1.0–0.542 Ga). The onset of the Proterozoic conveniently closely corresponds with the so-called *Great Oxidation Event*, during which time free O_2 first accumulated in the atmosphere, and its end closely corresponds with the origin and diversification of animals, which presumably indicates that atmospheric oxygen levels had risen to at least many hundredths, if not, tenths of the modern level. The first indisputable Eukaryotes appear in the Paleoproterozoic, and they diversify in the early Neoproterozoic, although cyanobacteria (and stromatolites) reigned supreme through the Eon. The Earth was glaciated multiple times in the both the Paleoproterozoic and the Neoproterozoic, some of these glaciations being global in extent. Banded iron formation (BIF), which was already being deposited in the Archean, became more widespread in the early Paleoproterozoic (vast *Superior-type* BIFs), and all but ended for good by the end of the Paleoproterozoic. *Supercontinents* were born and died multiple times in the Proterozoic.

11.1 The Paleoproterozoic Glaciations

The early Paleoproterozoic was punctuated by a series of glaciations, at least one of which may have been global in extent, between 2.45 and 2.2 Ga. Well represented north of Lake Huron within the *Huronian Supergroup*.

- Closely followed assembly of an early supercontinent (*Kenorland*)
- PPz glacial deposits found in N. America, Finland, South Africa, and Australia
- In Great Lakes area: Mt. Ramsay, Bruce, and Gowganda formations
- Followed by the *Lomagundi Event* – high $\delta^{13}C$ carbonates – a prolonged carbon burial event

11.2 Life in the Proterozoic

Andy Knoll has dubbed the Proterozoic the "Age of Cyanobacteria" because the cyanobacteria were the dominant life from in terms of energy flow in Proterozoic oceans. They most certainly account for the increase in oxygen levels and most organic carbon that was buried in Proterozoic sediments. However, one of the big questions in Precambrian paleobiology is when the Eukarya branched from the Archaea. Up until recently, it was believed that this branch point was 2.7 Ga or older based on biomarker data, but these data are now called into question. So the best evidence we have is that the earliest eukaryotes appeared in the Proterozoic.

11.2.1 Eukaryotic evolution

Eukaryotic cells are larger and more complex than prokaryotic cells:

- internal nucleus, which store genetic information in *chromosomes*
- most contain other organelles

Most eukaryotes are aerobes and many are multicellular. Thus, they represent a fundamentally huge step in Precambrian evolution. It is widely regarded that Eukaryotes evolved, in part, as the result of symbiosis between different bacteria: an invading or predatory bacteria and a host. Lynn Margulis, who was also known for supporting the Gaia hypothesis, was the first biologist to recognize and argue for the symbiotic origin of Eukarya—a hypothesis that is now widely accepted. She argued that the symbiosis eventual resulted in dependence between the invading and host bacteria, with the invader performing a specific function related. For example, mitochondria, which are the engines of eukaryotic cells, resemble certain extant bacteria, and they have distinctive RNA and DNA from the remainder of the cell. Similarly, chloroplasts in plants closely resemble certain cyanobacteria, and it has hard to imagine that cyanobacteria and eukaryotes separately evolved such similar platforms to carry out photosynthesis.

In order to have engulfed bacteria, early eukaryotes must have first evolved a *cytoskeleton*, which is a set of fibres beneath the outer membrane of the cell that give it rigidity and shape. The timing of the origin of the eukaryotes is a subject of great debate and significant research, but as of yet, little consensus.

Was the eukaryotic evolution simply a matter of time or was it nudged along by environmental change? Answering that question fundamentally requires knowing when eukaryotes first evolved. As discussed earlier, some circumstantial evidence (some of it now discredited) has been offered to suggest they evolved in the Archean, prior to the Great Oxidation Event. However, the fossil record does not preserve any unambiguous fossils pre-dating the GEO.

- 2.1 Ga macrofossils of colonial (eukaryotic?) algae, Francevile Series, Gabon
- 2.1 Ga *Grypania*, a colonial bacteria (or eukaryote?) from Negaunee iron formation, Michigan
- 1.9 Ga Gunflint chert: Microfossils of iron-oxidizing bacteria (i.e, recognizable modern counterpart)

- 1.8 Ga oldest *acritarchs*: "closed, organic-walled microfossils of uncertain systematic affinities" (Knoll et al., 2006a), but many are thought to be eukaryotic
- 1.5 Ga *Tapannia*, an acritarch that likely represents a eukayrote with a complex cytoskeleton (thought by some to be an early fungus)
- 1.2 Ga *Bangiomorpha pubescens* (bangiophyte red algae) with differentiated holdfasts
- Early Neoproterozoic: Eukaryotic diversification
- c. 0.65 Ga: the first sponges? biomarkers, macrofossils, and molecular clocks

11.2.2 The age of stromatolites

As previously mentioned, life on the Proterozoic Earth was dominated by cyanobacteria, who are the dominant organism involved in the construction of stromatolites. Stromatolite diversity peaked in the late Mesoproterozoic, when they were a major component of carbonates rocks. Throughout the Proterozoic, stromatolites were major reef-builders, filling the biogeochemical-ecological role that corals would eventually occupy. Stromatolite biodiversity began to decline precipitously in the middle Neoproterozoic and they all but disappeared by the end of the Ediacaran Period, presumably as the result of the expansion of grazing organisms.

Chapter 12

Supercontinents and the Assembly of North America

12.1 Tectonics and the Supercontinental Cycle

Archean greenstone belts offer a glimpse of how plate tectonics operated on a much hotter Earth (that is, in the interior). By the 3 Ga, the first modern-scale continents had formed, as represented by the stable continental platform onto which the Witwatersrand and Pongola supergroups (South Africa) were deposited.

It was long a matter of debate whether or not modern-style plate tectonics operated at all in the Precambrian, and many prominent geologists argued that it did not. However, it is now well established that plate tectonics resembling the familiar rifting and mountain building processes taking place today did in fact operate in the Precambrian. Paul Hoffman's pioneering work on the c. 1.9 Ga Wopmay orogen in northern Canada clearly established that the modern mode of tectonics had been established by the late Paleoproterozoic.

12.1.1 Precambrian supercontinents

The cratons appear to assemble periodically into large continents (supercontinents) that include many to most of the cratons. The best known example of a supercontinent is *Pangea*, which finished assembling the in the late Paleozoic. Not surprisingly, reconstructing older supercontinents is difficult business and amounts to putting together highly fragmented puzzles with poorly preserved pieces.

- Geological tie point
  - lithology
  - metamorphic grain
  - geochemistry
  - geochronology
  - stratigraphic architecture (e.g. rift–drift)
- Paleomagnetic data (apparent polar wander paths)

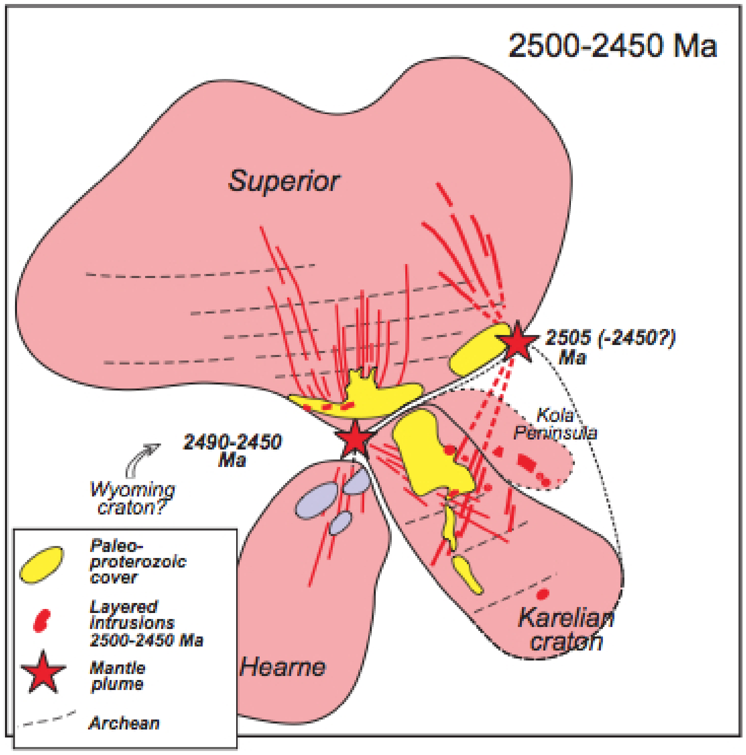

Figure 12.1: Reconstruction of the ca. 2.5 Ga connection between the Superior, Hearne, and Karlian cratons based on the distribution of dike swarms and related layered intrusions (from Ernst, 2007).

One of the most useful tools for establishing ancient continental connections is through matching radiating dike swarms (Ernst et al., 2008).
The reconstruction of Precambrian supercontinents is highly controversial and consensus is hard to reach, particularly on older supercontinents. Nevertheless, it is well that established the orogenic belts seem to cluster in age, which strongly supports the notion of periodic supercontinental assembly. Probably the best established past supercontinents are:

- Kenorland (c. 2.45 Ga)
- Columbia (c. 1.8 Ga)
- Rodinia (c. 1.1 Ga)

12.2 The Assembly of North America

12.2.1 The Wopmay orogen

The ca. 2.0–1.8 Ga Wopmay *orogen* on the western margin of the *Slave craton* records a full *Wilson Cycle*, that is, the opening and closure of an ocean basin:

- Rift: Coarse-grained siliciclastics and pillow basalts

- Drift: Passive margin, development of a stable carbonate platform
- Collision: Outer metamorphic belt, central fold-thrust belt and foreland basin, internal undeformed craton

12.2.2 Proterozoic evolution of North America (Laurentia)

The core of North America is an ancient continent that consists of Archean cratons sutured by Paleoproterozoic orogens. Precambrian North America is known as *Laurentia* and includes Greenland, Svalbard, and northwest Scotland which were separated from North America during Cretaceous rifting.

Archean cratons

The Canadian shield comprises at least 7 Archean cratons: Superiour, Wyoming, Nain, Rae, Slave, Hearne, Burwell. These largely assembled between about 2.0 and 1.8 Ga, such that the boundaries between the distinct cartons are orogens with associated arcs.

Yavapai-Mazazatl Terranes

Following stabilization of Laurentia, southern and eastern Laurentia was the locus of active crustal growth from about 1.8 to 1.6 Ga, through the accretion of oceanic terranes and arc volcanism. This phases of *juvenile* crustal growth was followed by a prolonged episode of *post-orogenic* igneous activity and emplacement of vast volumes of granite between about 1.5 and 1.3 Ga

The Mackenzie Dike Swarm

North America also experienced major plume activity and continental rifting in the Mesoproterozoic. The Mackenzie Dike swarm was emplaced about 1.27 Ga, and is one of the largest dike swarms known on Earth. *Dike swarms* are curved or radiating clusters of mafic dikes that likely formed as the result of plume activity and constitute the roots of widespread volcanism. They often all that remains of what are commonly referred to as *large igneous provinces.*

The Mid-continent Rift

Later in the Mesoproterozoic, continental extension began about 1.2 Ga, forming the Keeweenawan, or mid-continent rift. A series of large grabbens, filled with basalts and sediments, formed at this time, but rifting ultimately ceased, and the Laurentian continent remained in tact. Lake Superior roughly occupies one of these former rifts, but the associated volcano-sedimentary succession is exposed on either side of the lake.

Chapter 13

Neoproterozoic Snowball Earth and Rodinia Break-up

Reading: Chapter 12, pages 276-277, in Stanley

13.1 Introduction

The Neoproterozoic Era (1000–542 Ma) experienced many important events and changes. At the onset of the Neoproterozoic, by all counts the biosphere was still rather simple. Cyanobacteria were the dominant *primary producer* and Eukarya had evolved, but not diversified. The oceans were largely anoxic, and probably at least in part sulfidic (= *euxinic*). The continents were arranged in the supercontinent Rodinia, and climate was probably fairly steady. By the end of the Neoproterozoic, Rodinia was gone and a new supercontinent had already begun to form. The Earth had experienced a series of glaciations so severe that the tropics were covered in ice. Yet Eukarya survived, and by the late Neoproterozoic (if not somewhat earlier) animals populated at last parts of the oceans. Most of the major animal phyla may even have already evolved. Oxygen was probably fairly abundant, even if the deep oceans were not yet entirely oxygenated. Algae had replaced cyanobacteria as the core of the marine food chain. In short, the planet was eminently habitable and had begun to resemble the modern world in many important ways.

- Break-up of the low-latitude supercontinent Rodinia
- onset of amalgamation of the supercontinent Gondwana
- High average δ^{13}C values punctuated by negative excursions
- At least three glaciations, two of which were global in extent
- Reprise of iron formation deposition
- Oxygenation of the environment
- Emergence and diversification of animals

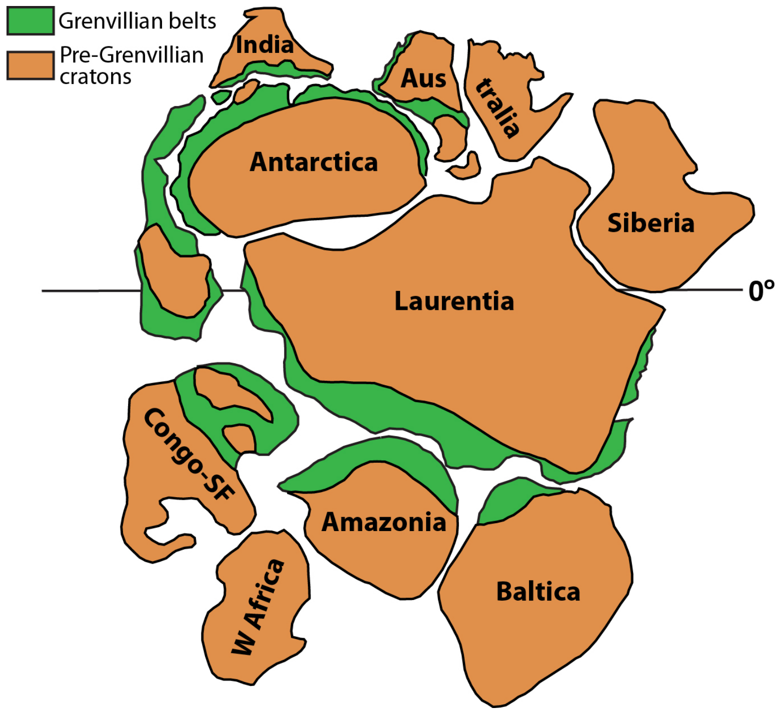

Figure 13.1: An early version of the proposed Rodinia supercontinent (Hoffman, 1991), based in large part on the locations of *Grenvillian orogens* and geological tie points. Additional evidence that Laurentia was at the centre of this supercontinent comes from the fact that it is almost entirely encircled in latest Proterozoic to early Paleozoic *passive margins*, which require continent rifting and subsequent drift to form.

13.2 Rodinia break-up

The supercontinent Rodinian finished assembling around the end of the Mesoproterozoic, perhaps the earliest Neoproterozoic (Li et al., 2008). It begin to drift towards the lower latitudes and show at least initial signs of breaking up around 850–800 Ma. By about 720 million years ago, break-up seemed to be well underway and the supercontinents continued to disintegrate until about 600 Ma with the opening of the *Iapetus* (Paleo-Atlantic) ocean.

The break-up of Rodinia appears to be closely associated with the emplacement of a large number of Large Igneous Provinces (LIPs), which are indicated by a combination of a flood basalt and/or a large radiating dyke swarm. Large LIPs notably occur around 830 (Willouran-Guibei), 780 (Gunbarrel), and 720 Ma (Franklin), and much of the evidence for these is clustered on cratons that are widely thought to have been contiguous in Rodinia: Laurentia, India, Australia, and South China. Hence, it has been proposed that these LIPs were are related to a single *super plume* event.

Both the emplacement of LIPs and the paleogeography during early break-up would have had a large influence on global climate, for multiple reasons.

13.2.1 Paleogeographic influence on global climate

Even before the pieces of Rodinia had entirely disaggregated, *Gondwana* begin to assemble with continental collisions encircling Africa (the so-called *Pan-African orogens*).

- Laurentia surrounded by passive margins
- Rifting associated with plumes and emplacement of several *large igneous provinces* (LIP's). In North America these include the 780 Ma Gunbarrel event and the 720 Ma Franklin event.
- Unique paleogeography played a strong role in regulating climate:
  - Response of silicate weathering feedback to low-latitude, rifting continents, abundant mafic rocks
  - Affect of low-latitudes on albedo

13.3 Geochemical Proxies

13.4 Record of Neoproterozoic glaciations

It has long been recognized that the Earth was glaciated in the late Precambrian. By the late 19th century, *Infracambrian* glacial deposits had been documented on several continents. But it was not until two eminent geologists working in the middle of last century championed their existence and pointed out their widespread occurrence did these rocks take on importance in Earth history: the Cantabridgian, Brian Harland worked mainly in Europe, Svalbard, and Greenland, while the erstwhile Antarctic explorer Douglas Mawson worked in Australia. The Neoproterozoic glacial deposits are now well known, figure prominently in text books, and are the subject of much controversial, but also exciting, interdisciplinary research.

It now seems that there were at least three glacial epochs:

- Sturtian ($\sim$ 720–660 Ma)
- Marinoan ($\sim$645–635 Ma)
- Ediacaran ($\sim$ 580 Ma)

Of these, the first two appear to have been the most severe, with clear evidence that the were long-lived and entailed low-latitude continental glaciation. These two are hypothesized to have been *snowball Earth* glaciations. Several different observations regarding the glaciations underlie the snowball Earth hypothesis:

- Glaciations widespread (all continents)
- Carbonate platforms glaciated
- Evidence that glaciers reached sea level in the tropics
  - Elatina tidal rhythmites

  - Paleomagnetic data

- Iron formation deposited during at least one glaciation
- Glaciation lasted on the order of 10 m.y.
- Glacial deposits sharply overlain by unusual *cap carbonates*
- Glaciation associated with large negative δ^{13}C anomalies and other geochemical perturbations

13.4.1 Models for Neproterozoic glaciation

- High obliquity hypothesis
- Orbiting ice rings
- Snowball Earth hypothesis

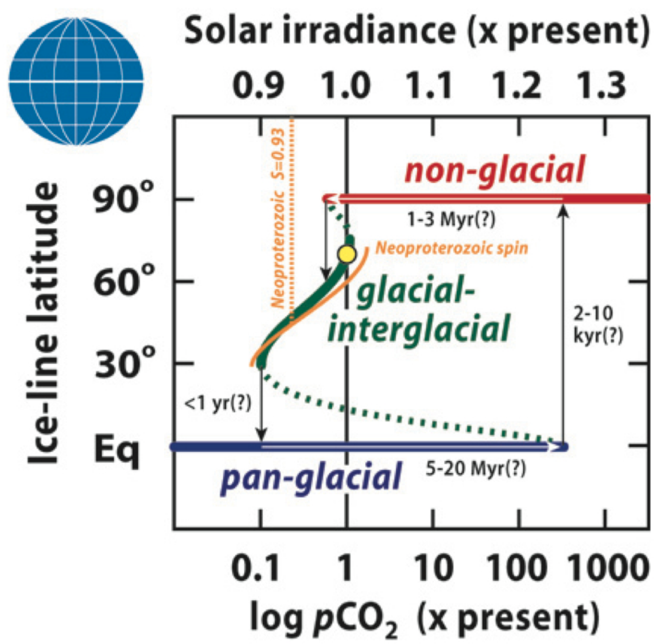

Figure 13.2: The three stable climate states that are predicted by simple energy balance models. Note that a fourth climate state, the so-called *Jormungand* state has recently been proposed, but is not yet widely accepted.

13.5 The Snowball Earth hypothesis

Quite simply, the snowball Earth hypothesis aims to explain all of the features associated with the Neoproterozoic glaciations, which can be done if the entire Earth froze over.

- Requires a proximal trigger for glaciation

- Once glaciers surpass the mid-latitudes, ice-albedo feedback triggers climatic collapse: runaway glaciation
- Hydrological cycle would largely shut down on a frozen-over Earth
- Ocean would be cut off from atmosphere
  - Ocean would become anoxic
  - CO_2 would accumulate in the atmosphere
- Would take about 10 million years to accumulate enough to compensate for high albedo
- Catastrophic melting: ice-albedo feedback in reverse
- Post-glacial ultra-greenhouse
- Cap carbonates represent flux of alkalinity to the ocean
- Intense silicate weathering draws down CO_2 over a few million years

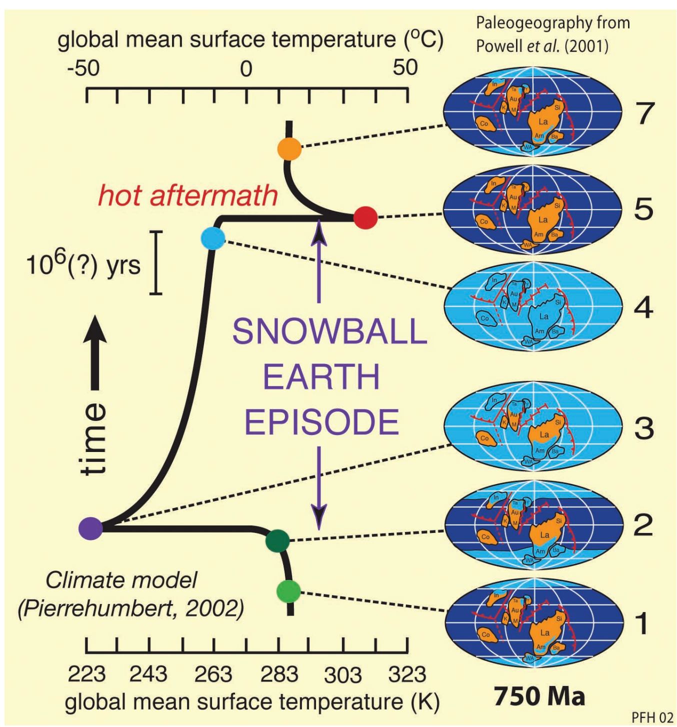

Figure 13.3: The snowball cycle.

13.5.1 Problems with the snowball

- How did life survive (refugia)?
- The glacial record

  - Thick glacial deposits reminiscent of Pleistocene glacial deposits
  - Evidence of cyclicity
  - Evidence of open water

- Issues raised by modeling

13.5.2 Geochemical support for snowball

- Iridium and other platinum group elements
- Spike in Sr isotope ratios after glaciation
- Boron isotopes
- Mass-independent oxygen isotopes ($\Delta^{17}O$)

13.5.3 Possible consequences of the snowball

- Trimming the tree of life
- Evolutionary radiation
- Extreme greenhouse, weathering
- Some of the earliest petroleum systems?

Chapter 14

The Ediacaran Period and the Origin of Animals

14.1 Introduction

The Ediacaran Period was the first named period in the Precambrian

- Just named, with the GSSP in Brachina Gorge, South Flinders (Knoll et al., 2006b)
- Base of the Ediacaran Period is the base of the Nuccaleena Formation
- Base of the Cambrian is formally defined by the first appearance of the trace fossil *Phycodes pedum* (now known as *Treptichnus pedum*).

The Ediacaran Period began 635 Ma with the end of the snowball Earth, which by most accounts was followed by a catastrophic greenhouse. At this time, no unambiguous animals had yet populated the Earth, although several lines of circumstantial evidence suggest that early animals may have already diverged from the eukaryotic lineage. By the end of the Ediacaran Period (541 Ma), the biosphere was quite diverse and may have included most if not all of the major animals clades. A lot happened in the 94 million years in between, and so it is not surprising that the Ediacaran Period draws so much attention. One of the prevailing questions among researchers today is the extent to which early animal evolution was driven by an increase in oxygen in the environment. The underlying principal behind this view of *permissive evolution* is that most large animals require a significant amount of oxygen to metabolize, and hence could not have evolved until atmospheric oxygen reached a reasonably large fraction of present atmospheric level (PAL). The implication is that early animal evolution was sort of waiting around for this oxygenation to occur. This view is at least partly supported by molecular clock data that place several important early metazoan divergences prior to the Ediacaran Period, 10's to 100's of millions of years before the first macroscopic animal fossils (Erwin et al., 2011)(Fig. 14.1).

A countervailing view is that early animal evolution was not triggered so much by a changing environment, but by the eventual acquisition of the genetic toolkit that allowed them to evolve, which took hundreds of millions of years (Butterfield, 2009). Proponents of this view of early animal evolution posit that it was the animals themselves that triggered the major change in the environment, specifically, oxygenation. That is, animal evolution trig-

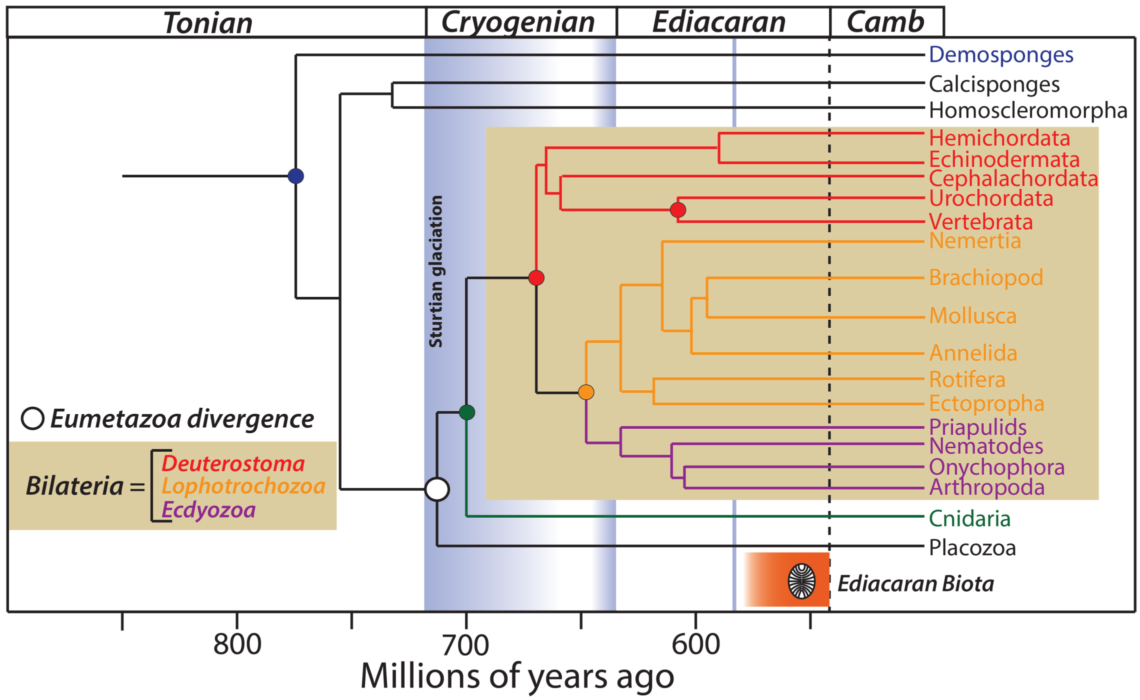

Figure 14.1: Early animal evolution based on molecular clock data, modified from Erwin et al. (2011). Note that the Ediacaran biota, which likely include at least several phyla of early animals, do not appear until after the c. 580 Ma Gaskiers glaciation.

gered environmental revolution.

I tend to believe that the answer must lie somewhere in between. One research group (Sperling et al., 2013) has recently pointed out that certain animals can live and thrive in modern environments that are impressively oxygen deficient (i.e. oxygen minimum zones) and that early, simple grade animals (the diploblasts) would have had much more modest O_2 requirements (similar to other Eukarya) than you, me and the nautilus (Fig. 14.2). Eukaryotes have been around since at least 1.8 Ga, and quite possibly longer. So just maybe, the long slog of evolution and acquisition of the genetic tool kit happened between then and about 750 or 650 million years ago, when the first simple animals evolved. These would have had another 100 million years or so to evolve and acquire the more sophisticated developmental genes that enabled them to diverge into the metabolically complex animals that we tend to think of when we think of animals. At the same time, these early animals would have impacted biogeochemical cycles through consumption and supply of nutrients and inevitable development of trophic tiers. This biogeochemical revolution would have continued as new clades evolved, perhaps triggering further oxygenation, which in turn, permitted predation to develop and evolve, which in turn triggered a rapid radiation, which in turn had a strong impact on the global carbon and oxygen cycles. As often proves to be the case, the answers to such fundament questions like "how did the first animals evolve?" are nuanced and lie somewhere between the long entrenched camps.

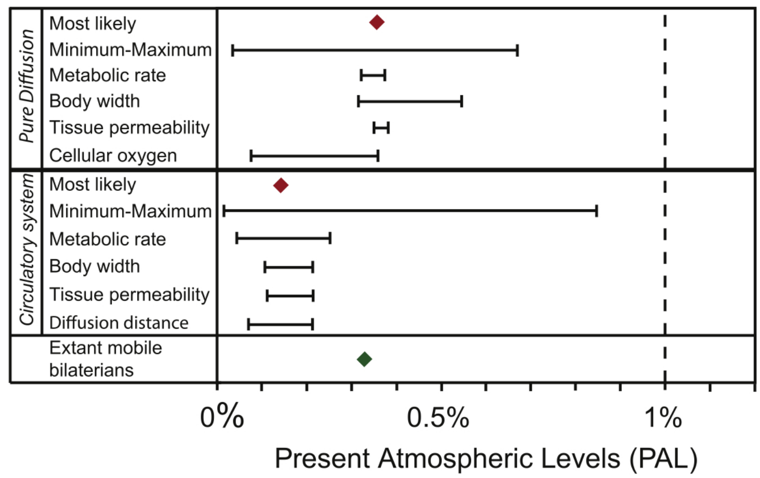

Figure 14.2: Summary of minimum oxygen requirements for two hypothetical last common ancestors of the bilaterians: one in which is limited by diffusion, and another which is limited by the circulatory system. Note that these are quite similar and not that far off from the current lower limit for extant mobile bilaterians. From (Sperling et al., 2013).

14.2 Chronology of Early Animal Evolution and Ediacaran Events

- ~700 Ma: The earliest sponges?
- 635 Ma: The end of the last global glaciation
- ~632 Ma: Possible early microscopic animals?
- 595 (?) Ma: The Doushantuo biota (fossilized animal embryos)
- ~ 590 (?): Acraman impact
- ~ 590 (?): Turnover in the acritarch record
- 582 Ma: The Gaskiers glaciation
- 580 - 565 (?) Ma: Shuram-Wonoka negative δ^{13}C anomaly
- 575 Ma: first appearance of the Ediacaran biota
- <555 Ma: first appearance of *Kimberella* (mollusca?) and *Dickinsonia*
- ~ 545 Ma: fist appearance of lightly calcified metazoa (LCM)
- 542 Ma: Precambrian-Cambrian boundary: the first animal extinction event?
  - Disappearance of the LCM
  - Negative δ^{13}C anomaly
  - Some Ediacara continue?

14.3 Early Sponges? and the Molecular Clock Record of Metazoan Diversification

Recently, several lines of evidence have emerged suggesting the divergence of *demosponges* (sponges whose skeletons are made of spicules of spongin or silica) prior to the last snowball glaciation. If correct, this discovery has important implications for the snowball Earth and its role (and generally, the role of the global environment) in regulating the evolutionary tree.

- *Demosponges* appear to have diverged before the *calcisponges*
- Molecular clock data place this divergence at c. 665 Ma
- Biomarker data also suggest that there were sponges prior to the last snowball glaciation
- Putative fossil evidence for sponges prior to the last snowball glaciation

Other molecular data suggest that sponges (*Porifera*) are paraphyletic, meaning they actually belong to multiple clades instead of just one, and that the calcisponges are more close related to Eumetazoa (basically, all animals with a gut) than they are to the demosponges (Sperling et al., 2007). Whereas the monophyletic grouping of the sponges would have required that the Eumetazoa diverged prior to the last snowball glaciation, the paraphyletic grouping allows that they may have diverged afterwards, which we might argue is more consistent with the fossil record.

Following the evolution of the gut, the next key evolutionary step gave us the bilaterians, which are all triploblastic and diverged from the diploblasts, which include the cnidarians. Bilaterians include both the proterostomes (think mollusks and arthropods), and the deutorostomes (you and the echinoderms). This divergence, and presumably several other animal divergences within bilaterians, occurred during the Ediacaran Period.

14.4 The Acritarch Extinction and Radiation

Acritarchs are enigmatic, organic-walled microfossils that are incomparable to any modern organisms, but likely mostly represent eukaryotic phytoplankton. Much of the focus on acritarchs in the Ediacaran Period has centered on evidence for an extinction of the simple, spherical *Leiospheric* acritarchs shortly after the Acraman impact and the subsequent divergence of the more complex, spiny *Acanthomorphic* acritarchs (Grey et al., 2003). However, recent studies suggest that fairly complex acritarchs occur in early post-snowball rocks and that these may in fact represent cysts formed during the diapausal resting stage of early invertebrates, which originated as an evolutionary response to prolonged oxygen-depletion (Cohen et al., 2009). This tantalizing hypothesis add further fuel to the idea that early animal evolution was closely linked to snowball Earth and may have significantly preceded the first appearance of Ediacaran fossils.

14.5 The Ediacara Biota

The *Ediacaran Biota* encompasses a group of enigmatic, mostly tubular and frond-shaped fossils that occur in the latter half of the Ediacaran Period. Many of these biota appear to have been *sessile* (attached to a substrate). The Ediacara biota include fossils which resemble modern phyla and others that cannot confidently be ascribed to any known clade. The first Ediacara appeared in the geological record around 575 Ma, in deep-water sediments in eastern Newfoundland.

- Represent the earliest known complex multi-cellular organisms
- Most consist of impressions left in sandstone and commonly below ash beds
- Other important Ediacaran fossil locations include South Australia, southern Namibia, and northern Russia

One group of Ediacaran fossils includes simple discoid fossils. These were formally interpreted as jelly-fish like organisms, and thus related to the Cnidarians, but this taxonomic affiliation remain unestablished, and at least some of these certainly belong to other clades, quite likely *stem groups*.

- *Cyclomedusa*
- *Mawsonites*
- *Medusinites*
- *Ovatoscutum*
- *Aspidella*

Some of these discoid fossils are now reinterpreted as *holdfasts* of a another group of Ediacaran fossils, known by some as the *Vendobionts*. The Vendiobionts consist of complex, quilt-like, frond structures up to a meter tall (or long). These include

- *Dickinsonia*
- *Rangea*
- *Pteridinium*
- *Swartpuntia*
- *Ernietta*

Dickinsonia is the best known and most controversial of the Ediacaran fossils. It is variably interpreted as a relative of the Acoel flatworm, a jelleyfish, a coral, a lichen, a chordate, or a colonial organism (most likely a Cnidarian). The inimitable Dolf Seilacher 2003 has argued that the Vendiobionts represented a failed experiment in multicellular life (animal or not).

14.5.1 Ediacaran trace fossils

The Ediacaran fossils include spaghetti-like traces that suggest that they were formed by bilateria - that is, organisms with bilateral symmetry, like most animals on Earth today.

14.5.2 More complex Ediacaran fossils

Another class of Ediacaran fossils exhibit bilateral symmetry and loosely resemble some Phanerozoic organisms, such as arthropods and molluscs.

- *Spriggina* (arthropodoid?)
- *Praecambridium* (arthropodoid)?
- *Kimberella* (a mollusc?)
- *Parvancorina* (a *stem group* trilobites?

Ediacaran fossils can be subdivided into three separate assemblages, which include some overlapping fossils, but represent somewhat different sedimentary environments and modes of preservation.

14.5.3 The Avalon assemblage

The oldest Ediacaran fossils are found in Newfoundland, in deepwater sediments (turbidites, shales, and tuffs) and date as far back as about 580 Ma (Narbonnne and Gehling, 2003). These include the frond structures, which appear to have formed relative complex, ecologically tiered communities (Fig. 14.5.3. These fossils are preserved as impressions on bedding planes, commonly draped in volcanic ash (which helps account for the spectacular preservation). Whereas some scientists claim these fossils include early animals, it remains to be demonstrated unambiguously that these were early animals. What is certain, however, is that they represent a bold eukaryotic experiment in multicellularity, but one which, ultimately, failed because most of these fossils are at best stem groups. Furthermore, they inhabited deep waters, which seems to suggest these deeper waters were more environmentally friendly for colonization be early multicullar eukaryotes (maybe animals).

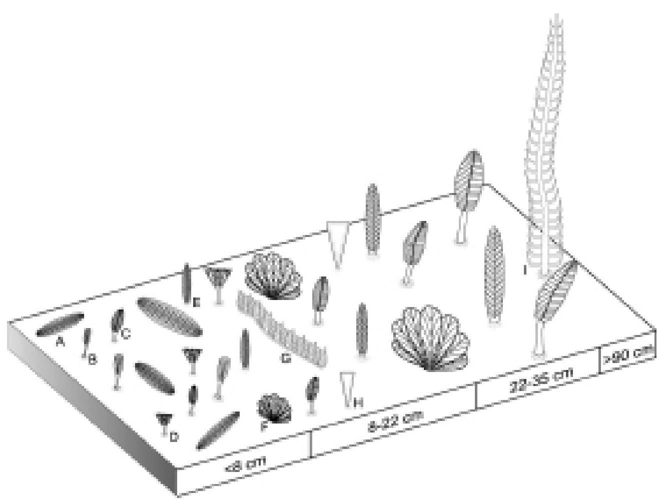

Figure 14.3: Guy Narbonne and colleagues interpret many of these fossils to represent organisms that lived in tiered-communities consisting of *sessile* organisms attached to the seabed. From (Narbonnne and Gehling, 2003).

14.5.4 The Ediacaran Hills assemblage

Ediacaran fossils were first discovered in Australia by Reginald Sprigg, a South Australian geologists and student of the renowned Douglas Mawson. These fossils are found in the Ediacaran Member of the Rawnsley Quartzite in and around the Flinders Ranges, and were at first regarded as Cambrian jelly fish. Ediacaran fossils preserved here occur in fine sandstones in a shallow, sub-tidal, pro-deltaic environment. They occur mainly as impressions, and their preservation seems to be closely linked to the occurrence of microbial mats, which the organisms were apparently feeding upon at the time of burial.

Jim Gehling, a palaeontologist at the South Australian museum in Adelaide argues that these microbial mats were critical in the preservation of this style of Ediacaran fossils. These *death masks*, as he called them, would have formed as the result of early diagenesis after feeding Ediacaran biota (along with the mats) were smothered by fine sand (Gehling, 1999). Precipitation of iron minerals during this early diagenesis would have provided a mechanism for preserving these fossils in a sediment type that is otherwise poorly suited for preserving large fossils.

14.5.5 The Nama assemblage

Ediacaran fossils are found in the Nama Group, a foreland basin sequence in southern Namibia related to the formation of Gondwana. This fossils occur in somewhat sands in a deltaic environment, and the preservation is largely three dimensional. It includes mostly fronds (the Vendiobionta), and this style of preservation has led some to suggest that these organisms were *epifaunal*–that is, that they lived buried in the sediments. However, others disagree.

14.5.6 The big questions

The Ediacaran biota raise a lot questions about the origin and early evolution of the Metazoa

- Were the Vendiobionts truly a failed experiment in complex multicellular life? If so, what was there metabolism? Some researchers argue that they were chemosynthetic, or at least, symbiotic with chemosynthesizers.
- Were certain Ediacaran fauna simple ancestors of modern lineages, for example, of the Arthropods, the Cnidarians, and the Molluscs? If they represent *stem groups*, where are the *crown group* fossils?
- If so, major divergences in the animal phylogenetic tree must have occurred well before the Precambrian-Cambrian boundary. This seems to be consistent with the most recent molecular clock data (Erwin et al., 2011).
- Where is the fossil evidence of all of the other clades?
- Were the earliest Metazoa microscopic? The extraordinary preservation of what are regarded as bilaterian animal embryos, dating to some time prior to 580 Ma, suggest that this idea may not be so far fetched

14.6 Weakly Calcified Metazoa

In the latest Ediacaran Period, the first calcified metazoans appeared. These are known colloquially as the *weakly calcified metazoa*, or *WCM*.

- Not as heavily calcified as the small shelly fossils that appeared in the Cambrian, but they seem to represent a first pass a development of skeletons of some sort.
- Very limited range: ca. 548–542 Ma
  - appear to have gone extinct at the Precambrian-Cambrian boundary
- Concentrated in microbial mats (i.e. *benthic*)
- Two principal groups:
  - *Namacalathus*
    - only positively documented in southern Namibia and western B.C.
    - concentrated in *thrombolites*
  - *Cloudina*
    - cosmopolitan

Chapter 15

The Cambrian Explosion

Reading: Chapter 13 in Stanley; Chapter 10 in Wiccander and Monroe

15.1 The Early Cambrian

The Cambrian and subsequent Ordovician periods were recognized and defined in the 19th century in Great Britain. Even before the time scale was formalized, it was recognized for two centuries that the fossil record changed abruptly in the early Cambrian, and the so-called *Cambrian explosion* continues to captivate palaeontologists and be a major topic of research. Charles Darwin was well aware of this feature of the fossil record, which seemed to be at odds with his own theory of natural selection and gradual evolution. Charles Walcott, a North American geologist who discovered the Burgess Shale, suggested that the apparent abruptness of the Cambrian radiation could be ascribed to failure in the stratigraphic record to preserve the early evolution of animals; he proposed that a vast stretch of timing was actually missing from the geological record at the Precambrian-Cambrian boundary (the so-called *Lipalian interval*). Stephen J. Gould, who was fond of the idea of episodic, very rapid rates of evolution, believed that the Cambrian radiation represented just that — a burst in evolution due to fortuitous environment conditions. Of course, many geologists do not agree with this thesis, and increasingly, even as the Cambrian time-scale is being shrunk by new radiometric ages, the timing of early animal diversification is being pushed back.

15.1.1 The Cambrian (541–485.4 Ma) Period and Time Scale

The Cambrian is known for the diversification of animal fossils, but if animal fossils date to the Ediacaran, what sets these two periods apart? Mainly, even though animal fossils do occur in the Ediacaran, they are limited to a few regions of exceptional preservation, and many discoveries of these fossils have required extraordinary perseverance by dedicated paleontologists. The fossils record in the Cambrian, on the other hand, is widespread, such that you and I could readily identify the remains of what were sure early animals in many outcrops of Cambrian age across the globe. This great diversification of animal life is effectively restricted to the oceans. Whereas we cannot rule out the possibility that there were freshwater early animals, there is no evidence in the sedimentary record to support this argument. Even within the oceans, the life habits of the earliest Cambrian animals

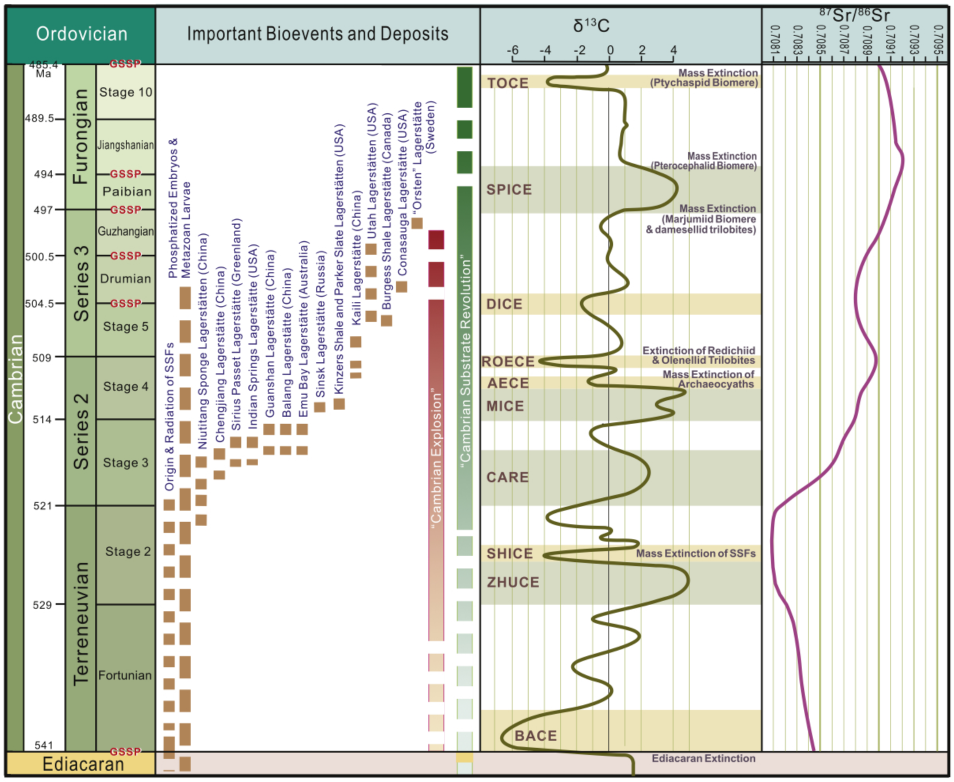

Figure 15.1: The Cambrian Time scale and its corresponding epochs and stages, along with the key evolutionary and other bioevents and the carbon and strontium isotope records. From Peng et al. (2012).

were relatively restricted. Most of the earliest animals appear to have been benthic, and apparently could not yet manage in the more challenging marginal marine environments (intertidal zones, tidal flats, etc.). The earliest animals were mainly deposit feeders (obtain nutrients from organic material in sediments) and suspension feeders (filtering fine particulate organic matter and dissolved organic matter from seawater).

The Cambrian Period marks a transition in the behaviour of the global carbon cycle. Where the extreme positive and negative δ^{13}C compositions of seawater that characterize the Neoproterozoic Era become somewhat muted, significant fluctuations continue, about an average of about 0‰. Both prominent positive and negative carbon isotope spikes in the Cambrian Period (Fig. 15.1 are useful for correlation and dating purposes.

- The basal Cambrian GSSP is defined by the first appearance of *Trycophytes pedum* (also known as *Treptichnus pedum* or formally known as *Phycodes sedum*), a branched trace that is more complex than the simple burrows of Ediacaran time.
- The lowermost Cambrian (Fortunian; 541–529) has sparse fossils
  - hard parts had not yet become important yet
  - most distinguished by increasingly complex traces

- The Cambrian radiation really begins in the Stage 2 (latter Terreneuvian Epoch; 529–521.1 Ma)
  - First appearance and diversification of the *Archaeocyathans*
  - SSF's diversify
  - First Arthropod traces
  - The earliest crown group sponges and likely brachiopods
- And explodes in the Atdabanian (521.1–518.7 Ma)
  - Trilobites and Echinoderms by the middle Atdabanian
- The Chengjiang fauna span from the late Atdabanian into the early Botomian (521.1–513 Ma)
  - Includes most of the animal phyla, e.g. Porifera, Mollusca, Brachiopoda, Arthropoda, Priapulida, Echinodermata, Annelida, Hemichordata, Chordata

15.1.2 The Early Cambrian substrate revolution

During the Ediacaran Period, microbial mats likely still covered much of the sea floor, providing a sort of protection to the substrate and maintaining a sharp sediment-water interface (Bottjer et al., 2000). At the same time, the simple trace fossils suggests there were no burrowers as yet. This changed rather abruptly in the early Cambrian, when animals began to burrow in a big way and the extent of microbial mats likely dropped precipitously due to grazing. It is not clear if the early burrowers were looking for food or escaping from predators, but the consequences were the same: sediments in habitable parts of the ocean were irrigated. It is perhaps the destruction of this substrate, which seems to have been so important for Edaicaran life habit and fossil preservation, that led to their disappearance (both literally, and in the fossil record).

One clear manifestation of the substrate revolution that accompanied the evolution of animals is the sharp decrease in stromatolite abundance spanning the Precambrian-Cambrian boundary. Stromatolites still formed in the Cambrian, but were restricted to zones not yet inhabited by animals, such as intertidal zones. And many of the microbial build-ups in the Paleozoic lack the well defined laminations of stromatolites. These *thrombolites* are probably a close cousin of stromatolites, their difference being that the were subject to grazing and burrowing, which obliterated much of the original texture.

With the advent of burrowing, the uppermost sediment column would have been soupier and the sediment-water interface more diffuse, resulting in a discrete *mixed layer* between seawater and sediments. This would have had a profound impact on early benthic organisms, which would have had to seek alternative methods for attachment to substrates (for sessile organisms) and who have adapted to a completely different style of sediments and food supply. As pointed out by Canfield and Farquhar (2009), the onset of vertical bioturbation would also have had a profound effect on seawater chemistry by irrigating surface sediments and introducing oxygenated seawater. These authors argue that the substrate revolution resulted in a sharp decline in the proportional burial of sedimentary pyrite and compensatory increase in seawater sulfate concentrates and gypsum precipitation. At the same time, bioturbation should have made it much harder to sequesterer organic matter in sediments, hence imposing a negative feedback on environmental oxygen abundances.

15.1.3 The Small Shelly Fauna

The first fully shelled metazoans are known as the *small shelly fauna* (SSF). They appear within 5 Ma of the pC-C boundary and gradually radiate through the Nemakit-Daldynian. They are tiny skeleton fragments (1-5 mm diameter), comprising tubes, spines, cones, and plates. Some are actually small shells of animals, whereas other are *schlerites*, disarticulated hard parts of animals. Whereas their affinity had long been mysterious, it now appears that the oldest SSF schlerites are disarticulated elements of lophotrochozoan affinity, likely early molluscs (Caron et al., 2006; Morris and Caron, 2007). The diversity of forms suggest a parallel genetic and phyllogenetic diversity. Importantly, the SSF's are commonly phosphatic (although some are also calcitic and others siliceous) and are concentrated in phosphate deposits. Phosphate deposits are relatively common in both the Ediacaran and the Cambrian, and the temporal restriction of the SSF's to the Tommotian suggests that animals did not evolve the capability of building phosphatic skeletons until the Cambrian.

15.1.4 The Archaeocyathans

The *Archaeocyatha* were the first major reef-building Metazoa, although their reefs were relatively small and patchy compared to later Paleozoic and modern reefs.

- Appeared in Early Cambrian—ca. 530 Ma
- Extinct 10–15 million years later (by the Middle Cambrian)
- Wide geographic distribution. Hence superb index fossil for the early Cambrian
- Built reefs up to 10s of meters tall and hundreds of kilometers wide
- Unknown affinity
  - either a separate phylum
  - or an extinct lineage of sponge (phylum Porifera)
- Distinctive morphology
  - sessile, inverted cone-shape cup
  - high Mg calcite skeleton
  - double-walled, walls connected by *septa*

The disappearance of the Archeaocyatha so soon after their appearance on the scene is probably due to the fact that they appeared during a time of *aragonite seas*, that is, when Mg/Ca ratios in seawater were relatively high, which is why they evolved to make high Mg calcite skeletons. Aragonite seas gave way to calcite seas soon after, and this most likely accounts for their sudden decline. Subsequent reef-builders, like the stromatoporoids, calcite sponges, and corals were longer lived, presumably because they could make their skeletons out of low Mg calcite.

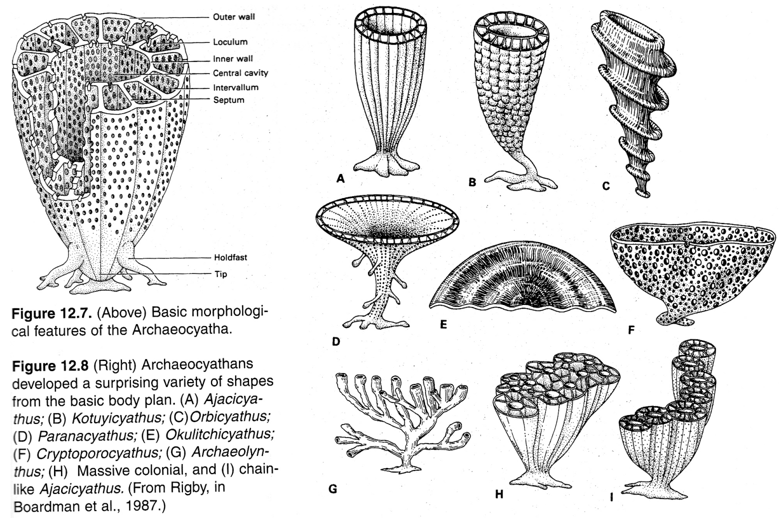

Figure 15.2: Morphological features of the Archaeocyatha and various Archaeocyatha shapes, from (Black, 2005).

15.1.5 Ocean Chemistry and Skeletal Mineralogy

Whereas the overall chemistry of seawater has varied only modestly over the Phanerozoic, a number of elements have varied sufficiently to have had important affects on the mineralogy of carbonate sediments and the types of shells precipitated by carbonate-secreting organisms. Specifically, the Mg/Ca ratio of seawater has fluctuated, with an impact on the style of $CaCO_3$ precipitation. The Mg^{2+} and Ca^{2+} ions are sufficiently similar in size that Mg can replace Ca in the calcite mineral lattice (up to a weight percent), but the Mg^{2+} ion is too small to fit comfortably into the aragonite lattice. Even though Mg^{2+} can fit into the high-mag calcite lattice, Mg/Ca ratios inhibit normal, *low-magnesium calcite* precipitation, such that at Mg/Ca > 2, aragonite and *high-magnesium calcite* are the preferred $CaCO_3$ minerals. Conversely at low Mg/Ca, normal (low-mag) calcite is typically precipitated. Because certain carbonate-secreting organisms can only secrete one type of $CaCO_3$, fluctuations in Mg/Ca (typically attributed to variations in sea-floor spreading rates) have played an important role in the existence and abundance of different carbonate-secreting taxa.

By the same token, palaeontologist Susannah Porter (UCSB) has argued that Mg/Ca ratios played an important role in early Cambrian evolution, because seawater shifted abruptly from high Mg/Ca (aragonite sea) to low Mg/Ca (calcite sea) in the early Cambrian (Porter, 2007). She notes that the first appearances of many taxa that exhibit strong mineralogical preferences appeared during the corresponding high and low Mg/Ca seas in the late Ediacaran to Ordovician (Fig. 15.3). A similar argument can be levied for the dominant reef builders throughout the Phanerozoic.

15.1.6 Trilobites

The first trilobites likely appeared by the end of the Terreneuvian or in the early Series 2 epoch, ~521 Ma. Like many other when other large animals that appeared around this

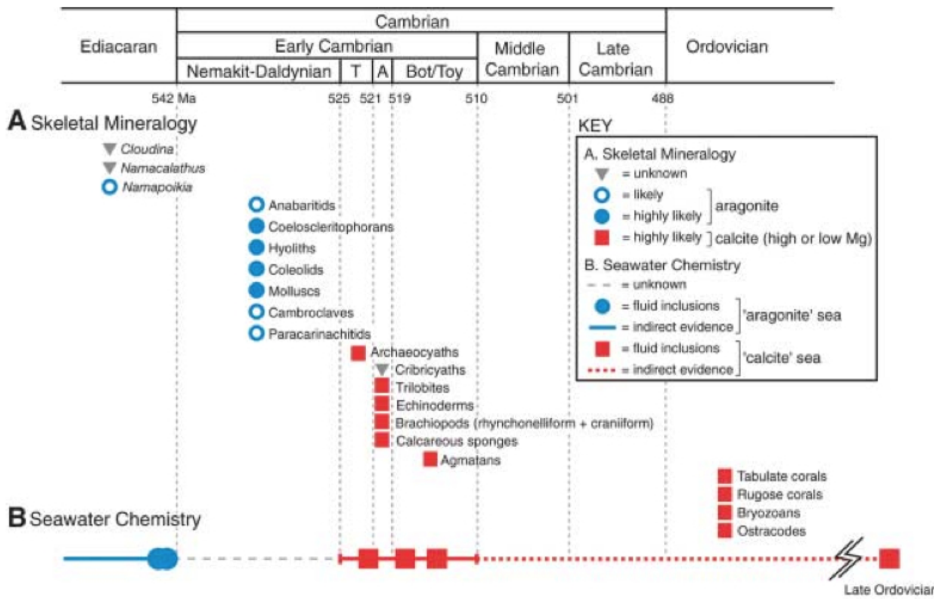

Figure 15.3: First appearances of carbonate skeletal taxa from the late Ediacaran to the early Ordovician that show a strong preference for aragonite or calcite, compared with the timing of aragonite (high Mg/Ca) versus calcite (low Mg/Ca) seas. From (Porter, 2007).

time, they subsequently diversified in a big way. The first appearance of trilobites coincides with a broader shift in animal types, and a turnover from the simpler fauna to a diverse and more complex fauna. The Cambrian explosion was officially under way.

Trilobites are rather symbolic of the Cambrian explosion, because the earliest examples are already quite complex, and with the possible exception of Ediacaran fossil *Parvanchorina*, there are no signs of intermediate ancestors. Trilobites were arthropods (invertebrates with segmented bodies and jointed legs), and thus had chitinous exoskeletons. The name derives from their three longitudinal lobes (left *pleural*, axial, and right pleural). However, we more commonly discuss their body types in terms of their there sections: *cephalon* (head), *thorax* (mid-section), and *pygidium* (tail). The cephalon shows considerable diversity among trilobites, which we will not delve into here. However, it is worth noting that trilobites had compound eyes made of calcite, meaning that eyesight is a rather primitive trait. The thorax consists of between 2 and 16 segments, which span the three lobes. The edges of the pleurae on each segment are sometimes spiny. The pygidium, which can be difficult to distinguish from the thorax, contains a number segments that were fused together. The size of the pygidum is a useful way of distinguish different groups of trilobites.

Early trilobites were benthic, deposit feeders with relatively week jaws and small mouths. Their segmented bodies were rather rigid, leaving them little means of protecting their inner, soft bodies. Rapid diversification gave rise to trilobites that could curl up and protect themselves from predators. Some of these later trilobites were planktonic, meaning they lived in the surface water column. The trilobites suffered their first cataclysm in the Middle Cambrian owing to increasing predation, followed by a major extinction event at the end of the Cambrian, which marked a major change in the trilobite record. They suffered subsequent major extinctions at the end of the Ordovician and in the late Devonian, after which they were trimmed down to one order (Proetida). They eventually succumbed to complete extinction at the end of the Permian.

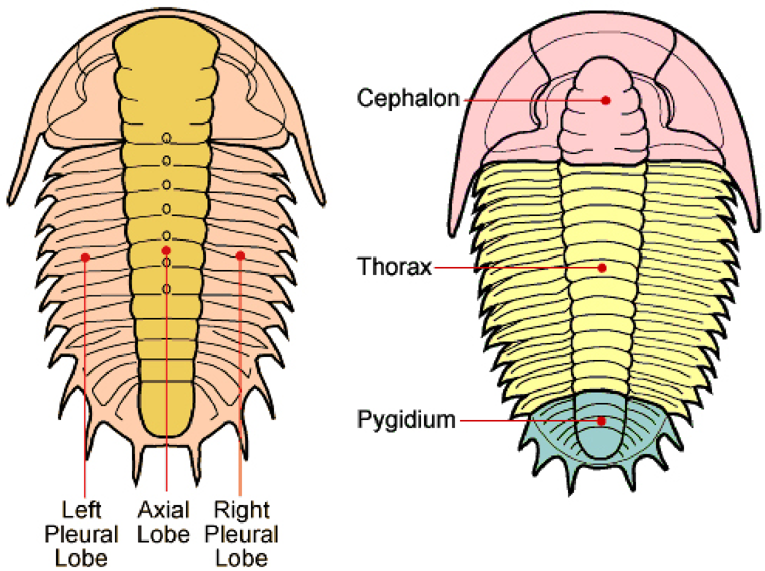

Figure 15.4: Morphological features of a trilobite.

15.2 The Explosion

The *Cambrian Explosion* refers to the geologically abrupt appearance in the fossil record of complex metazoa representing most of the extant phyla. The 'explosion' is probably less explosive than once thought, but is still spectacular insofar as the earliest Cambrian was marked by low diversity and subdued fossil evidence for animals, and by the middle Cambrian, diverse and complex animals clearly inhabited the entire ocean.

- Begins in the late Terreneuvian and accelerates early in Series 2
- Documents a massive radiation in hard parts/skeletons—probably partly the reason for the abrupt appearance of skeletal fossils in the geological record
- All major body plans were present by this time; 80% of skeletal elements recognized in modern phyla were present by this time
- Subsequent evolution characterized by diversification at the species to class level

15.2.1 Early-Middle Cambrian Lagerstätte

The Burgess Shale

The Burgess Shale, the most famous *lagerstätte*, is 505 Ma

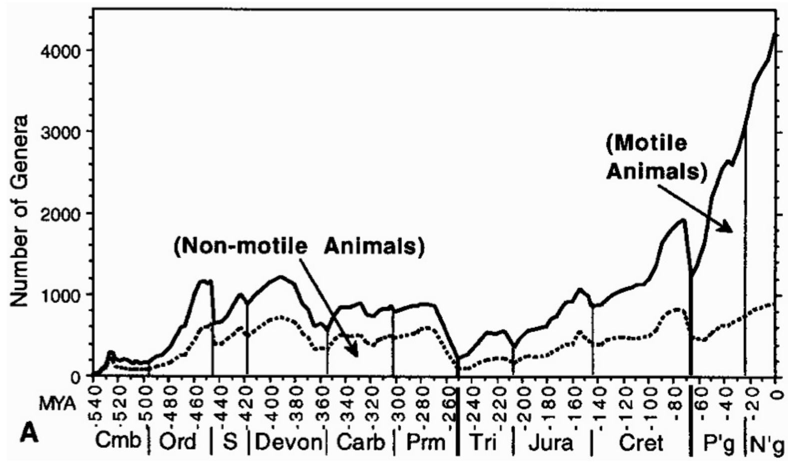

Figure 15.5: Plot of diversity of motile and non-motile organisms through time, at the genera level. Note the limited diversity in the Cambrian. From Bambach et al. (2002).

- Changed the way we look at early animal evolution
- It is no longer viewed as a window into early animal radiation, but still offers interesting clues about life in the Cambrian (older lagerstätte show similarly impressive diversity)
- S.J. Gould (*It's a Wonderful Life*), regarded some of the Burgess Shale fossils (e.g. *Opabinia*) as extinct phyla.
- But maybe they were just *stem groups*, namely of the arthropods (Knoll, 2003).

Among the many fantastic fossils in the Burgess Shale are the *Pikaia*, which are now regarded as the oldest stem group chordate (Conway Morris and Caron, 2012), and *Anomalocarus*, the largest and most fearsome Cambrian predator, which likely made quick lunch of trilobites.

15.2.2 The Chengjiang Biota

The Chengjiang (Maotianshan) shales represent the oldest Cambrian lagerstätte (c. 520–525 Ma), predating the Burgess Shale by some 20 million years. Importantly, it contains many of the same fossils as the Burgess Shale, thus pushing back their origins to the Early Cambrian. It is dominated by arthropods (many of them without hard parts), but contains divers animal phyla, of which over 10% cannot be assigned a systematic position.

Other important Cambrian lagerstätte include the Sirius Passet fauna in Greenland and the Emu Bay Shales in South Australia, the latter of which appears to contain some fossils types that appear transitional between the Ediacaran fauna and the Burgess Shale fauna.

15.2.3 The Hypotheses

The Cambrian radiation has long posed a problem for theories of evolution. Indeed, Charles Darwin recognized early on that it was a serious problem. Many hypotheses have been offered for the Cambrian explosion, and as of yet, no consensus has been reached.

1. No explosion at all—just an artifact of the record
2. Driven by oxygen or some other external environmental factor
3. Exponential filling of available evolutionary space
4. High sea level and shallow seas (environmental space)
5. Development of Hox genes, which govern the development of the body plan (Davidson and Erwin, 2006)
6. Awaiting assembly of the genetic toolkit (timing)

15.3 Cambrian Biodiversity

15.3.1 Experiments, body plans, animal phyla

Just as the the Ediacaran biota present a wide range of unusual morphologies in soft-bodied organisms, so too do early Cambrian animals present a large diversity of shapes and body plans. While the overall diversity in raw numbers of species and genera was low in the Cambrian compared to later in the Paleozoic, a large number of lower order clades evolved, possibly including phyla that are now extinct. This was animal life's early experimentation with a diversity of body plans, as it filled the breadbasket of ecological space at its disposal. Many Cambrian fossils are recognizable ancestors of modern day counterparts, but are strange and long since extinct. These early days of experimentation yielded a rich fossil record, and also established all of the basic body plans still used by animals today. Hence, already by the middle Cambrian, animal evolution at the phylum level was largely complete, and to some extent, much of the remainder of evolution was fine tuning at higher order taxonomic levels.

- Brachiopoda
- Mollusca (e.g. snails, ammonites, clams, squids)
- Arthropoda (e.g. trilobites and insects)
- Priapulida (marine worms)
- Echinodermata (e.g. crinoids and sand dollars)
- Annelida (e.g. earthworms)
- Chordata (e.g. you)
- Hemichordata (e.g. graptolites)

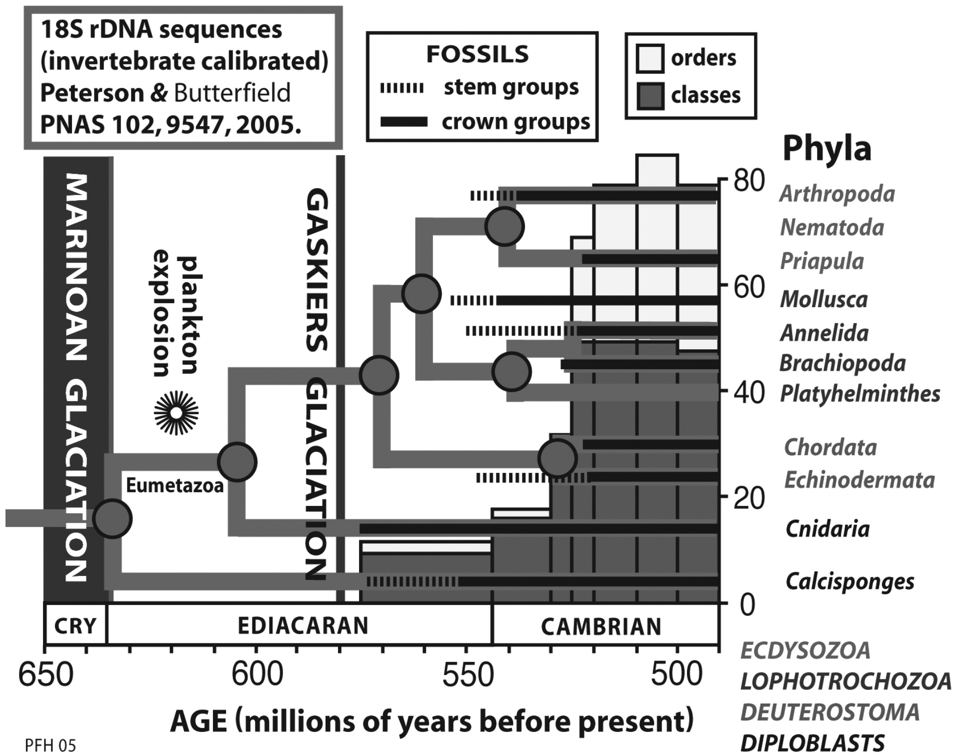

Figure 15.6: Divergences of the various Metazoa phyla (modified from Knoll and Carroll, 1999).

15.3.2 Conodonts – a mystery solved

Small, tooth-like phosphatic fossils formerly known as *conodonts* (and now referred to as *conodont elements*) occur in fossils from Early Cambrian to late Triassic in age. Although they had been extensively studied and heavily used by the petroleum industry (both in biostratigraphy and as a tool for determining thermal maturity), their actual biological and systematic affinity remained unknown until recently. They are now known to be an extinct taxa of chordates (stem-group lampreys).

Chapter 16

The Early Paleozoic: Age of the Invertebrates

Reading: Chapter 3 (65-70), Chapter 13 in Stanley; Chapter 12 in Wiccander and Monroe

16.1 Orodovician life

If the Cambrian is known for the incredible radiation in animal life and the emergence of all extant phyla, the Ordovician Period (485.4 to 443.8 Ma) might be best distinguished by extraordinary diversification at higher order taxonomic levels and the origin of many animal clades that dominated Paleozoic marine life. Even though the early chordates had appeared already by the middle Cambrian, marine life in the early Paleozoic was dominated by the *invertebrates*, which include the proterostomes (lophotrochozoa and ecdyozoa) and the echinoderms.

The trilobites suffered a major extinction at the end of the Cambrian, but retained a strong foothold in the benthic marine environment in the early Ordovician. The brachiopods and gastropods, both benthic organisms that originated in the Cambrian, also began to diversify at this time. At the same time, in the pelagic environment above, two other groups that appeared in the Cambrian also begin to diversify: the nautiloids (cephalopods) and graptolites (colonial Metazoa).

Although the early Ordovidian saw the start of a recovery from the end Cambrian extinctions, the pace of recovery in overall diversity was slow. It was not until later in the Ordovician that the pace of diversification picked up, and then it did so with a bang. The number of animal families increased by a factor of three as animals ramped up there presence on and within (epifaunal) sediments and in the overlying water column. Reef building went into recession in the latter half of the Cambrian, but reemerged in the Ordovician in a massive way, with corals (rugose and tabulate) and stromatoporoids constructing vast reef tracts.

Stromatolites, which had already suffered great indignities during the Ediacaran and Cambrian, continued to decline in abundance, and by the end of the Ordovician, were largely relegated to the relatively harsh environments (e.g., hypersaline lagoons and lakes) that they continue to inhabit today. While microbial life on the sea floor took a major hit, plants did begin to take a foothold on land in the Ordovician.

The great Ordovician diversification was punctuated by a series of extinctions at the close of the period, collectively known as the late Ordovician mass extinction.

Below is a review of the invertebrate clades the emerged in the Cambrian and Orodivian Period.

16.2 Simple animals

16.2.1 Porifera

The phylum Porifera (sponges) is the most primitive (deeply branched) of the multicellular eukaryotes

- Similar to colonial choanoflagellates (flagellate protozoa)
- Sessile, simple organisms - lack tissues
- Filter feeders
- Live in water depths from intertidal down to over 5000m
- Most basic body plan: inverted cup- or vase-shaped
- Commonly with spicules - basis for subdivision into classes
  - Calcarea (calcium carbonate) - the simplest form
  - Hexactenelida: silica spicules with six rays
  - Demospongiae (spongin)
- Probably also include the Archaeocyatha
- Stromatoporoids?

The fossil record of sponges is spotty

- First appeared in the Neoproterozoic?
- few are cosmopolitan
- oldest sponge borings occur in Ordovician
- More prominent than usual in the Devonian, lower Carboniferous, Jurassic, and Cretaceous

16.2.2 Stromatoporoids

The stromatoporoids (affinity really unknown) resemble a cross between a stromatolite (morphology) and a tabular coral (laminations). Like both, they were important reef builders. Prominent in Ordovician to middle Devonian reefs, but ranged from the Cambrian toe Cretaceous.

16.2.3 The Cnidaria

The Cnidaria are also very primitive and represent the next major divergence of the animal branch

- First appeared in the Ediacaran Period
- Radial symmetry (radiata)
- Single gastrovascular cavity
- Some have have sensory organs
- Stinging cells; feed by catching prey, or through symbiosis (i.e. corals)
- Four main classes:
  - Anthozoa (corals)
  - Scyphozoa (jelly fish)
  - Cuozoa (box jellies)
  - Hydrozoa (Portuguese Man o' War)

Of the four Cnidaria classes, Anthozoa is the most important in terms of the fossil record, because corals secrete calcium carbonate skeletons.

- Always attached to a substrate
- Polyps, either solitary or colonial
- Exclusively marine
- Subdivided into two subclasses
  - Alcyonaria (8-way symmetry) (include soft corals and sea pens)
  - Zoantharia (6-way symmetry of which three classes are particularly important in the fossil record
    * Rugosa (horned corals)
    * Tabulata (tabulate corals)
    * Scleractinia (stony corals)

Rugosa

The rugose, or horned-corals, are solitary or compound Palaeozoic corals that went extinct at the P-T extinction

- Calcareous skeleton
- Fossil range from Middle Ordovician to Late Permian
- Widespread by the Silurian
- Important reef builders

Tabulata

The Tabulata, or tabulate corals, are almost all colonial and were the major reef builder in the Palaeozoic

- Calcareous skeleton
- Honeycomb-like colonies of individual hexagonal cells (corallites)
- Distinguished by well-developed horizontal partitions
- Smaller than rugose corals, but more variable in morphology
- Ranged from the Late Ordovician to Late Permian; most prominent in the Silurian and Devonian

Scleractinia

The Scleractinia (stony corals) are the most important reef-builders today

- Solitary or colonial aragonitic skeletons
  - The colonial scleractinians occur in shallow, tropical waters, build reefs
  - The solitary scleractinians do not build reefs, may live in deep and cold waters
- First appeared in middle Triassic

16.3 The Bilateria

16.3.1 Brachiopoda

The brachiopods emerged in the Early Cambrian and persist today.

- solitary, benthic, filter-feeding, marine bilaterians with an external shell of either calcite (class Articulata — hinge) or alternating phosphate and chitin (class Innarticulata — no hinge)
- distinguished from bivalves by their plane of symmetry, which bisects the valves

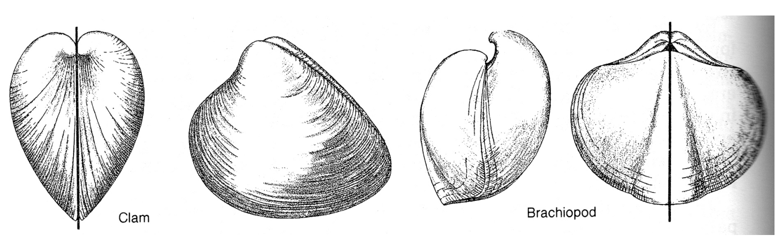

Figure 16.1: Bivalve versus brachiopod symmetry.

Brachiopod Morphology

- the shell encloses the body

- two valves: the pedicle (ventral) and the brachial (dorsal)
- pedicle emerges from posterior (hinge) margin

The brachiopods were prominent in the Palaeozoic, but declined in importance during the Mesozoic and Cenozoic.

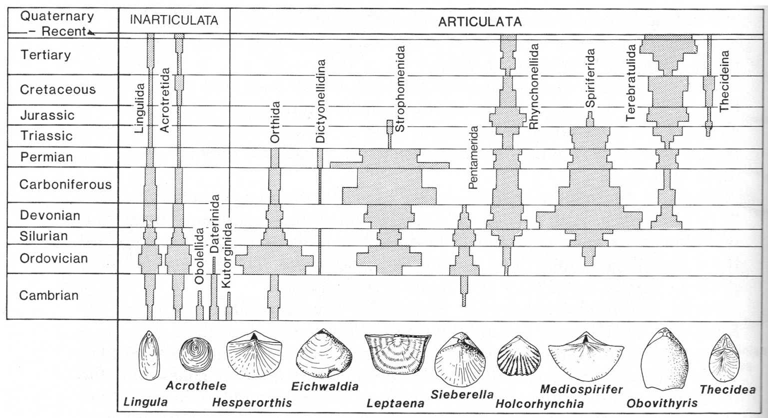

Figure 16.2: Brachiopod diversity

16.3.2 Arthropoda

The Arthropods include insects, spiders, crustaceans, and the trilobites

- Largest phylum of animals
- Probably emerged in the late Ediacaran Period
- Characterized by segmented body with appendages on each segment
- Chitin exoskeleton
- The trilobites (class Trilobita) are the most important in the Palaeozoic fossil record

The arthropods include the *eurypterids*, which were lobster like ancestors of the scorpion that appeared in the Ordovician, proliferated in the middle Paleozoic, and went extinct in the late Permian.

The trilobites were the most diverse metazoa in the Early Cambrian and had evolved by the late Terreneuvian (see previous chapter). They are named after there three lobes (parallel to axis of symmetry), but also subdivided into three regions from anterior to posterior.

- Extensive fossil record — because of their wide geographic distribution, broad range of environments, and hard exoskeleton
- 17,000 known species
- Rapid radiations and turnovers — useful for biostratigraphy

- Both *benthic* (scavengers, grazers, and filter feeders) and *pelagic* (eat plankton)
- 9 orders, all but one of which appeared by the late Cambrian
- Numbers began to decrease in Silurian and Devonian (predation by sharks?)
- All but the order Proetida died out during the Late Devonian extinction
- Remaining trilobites went extinct by the Late Permian (Permo-Triassic extinction)

16.3.3 Mollusca

The phyllum Mollusca is diverse and includes many familiar organisms, such as snails, squids, and oysters

- triploblasic protostomes (bilaterians)
- Some 112,000 species
- Closely related to the annelids (sister group)
- Probably first appeared in the late Ediacaran
- Diversified in the early Cambrian
- The most important in the fossil record include the bivalves, gastropods, and cephalopods

Bivalvia

The class Bivalvia evolved by the Middle Cambrian and diversified in the Ordovician

- Distinguished as typically having two shells that are more or less symmetrical
- Calcareous shells
- Filter-feeders: feed and excrete through siphons
- Most groups are either epifaunal or infaunal
- Include oysters, mussels, cockles, and clams

Gastropoda

The class Gastropoda (snails) first appeared in the Late Cambrian and have since been an important and diverse class of animals

- Highly diverse (over 60,000 known living species)
- Freshwater, marine, and terrestrial
- Includes snails, slugs, abalone, conches
- Most have a coiled or spiraled aragonitic shell
- Fairly robust during mass extinctions

- Earliest gastropods were quite distinct from extant gastropods
- Not until the Mesozoic did the ancestors of the modern gastropods appear
- Become especially important in the Cenozoic

Cephalopoda

The class Cephalopoda first appeared in the Late Cambrian and was important among marine animals through much of the Paleozoic and Mesozoic

- Distinguished by the large number tentacles that surround the head (e.g. squid, octopus)
- Important fossil groups (subclasses) include the Nautiloidea, Ammonoidea, and Coleoidea

The Nautiloids where the first of the Cephalopods to appear.

- Cone-shaped shell, typically curved or coiled
- Shell made of aragonite
- Consists of a series of chambers used to regulate buoyancy
- Important in the early Paleozoic, when they were the main predatory animals (eating trilobites)
- Highly diverse in fossil record, but sparse today
- Sister group to the Ammonoids

The Ammonoids appeared in the late Early Devonian and persisted until the end of the Cretaceous

- Very similar morphologically to the Nautiloids
- Planar-spiralled, complexly sutured shells
- May be loosely or tightly coiled
- Aragonite
- Hit hard by extinction events in Late Devonian, Late Permian, Late Triassic
- Extinct at the K-T boundary

The Belmnites are an extinct order of the Subclass Coleoidea, which include the squid and cuttlefish

- Very similar to the cuttlefish
- With notable exception that they contained an internal shell, located at the back end of the creature
- Largest of the belemnites up to 3 meters long
- Appeared in the Early Carboniferous; prominent in Jurassic and Cretaceous
- Extinct at the K-T boundary
- Useful index fossils

16.3.4 Echinodermata

The phylum Echinodermata is a diverse phylum that appeared in the fossil record in the Early Cambrian

- Deuterostomes (sister group of the hemichordates)
- Evolved from animals with bilateral symmetry but have radial symmetry
  - larvae are ciliated, bilaterally symmetric, free swimmers
  - later develop five-fold symmetry
- All echinoderms have an internal skeletan made of small, calcified plates and spines
  - provides structural support within tissues
- Exclusively marine
- Six living classes
  - Asteroidea (star fish)
  - Concentricycloidea (sea daisies)
  - Crinoidea (sea lillies, crinoids)
  - Echinoidea (sea urchins, sand dollars)
  - Holothuroidea (sea cucumbers)
  - Ophiuroidea (brittle stars)

The starfish (or sea stars) (Asteroidea) first appear in the fossil record in the Ordovician

- Possess five or more arms
- Poorly preserved as fossils because they do not posses a coherent skeleton
- Most live on sandy or rocky sea-floors

The echinoids first appeared in the Late Ordovician

- Benthic and *gregarious*
- Echinoid skeleton is rigid and may be a hemispheroid, disc-shaped, or heart-shaped
- Outer surface covered in spines

The crinoids first appeared in the Middle Cambrian

- Gregarious, live in shallow to deep, cold to warm waters
- Filter-feeders
- Attached to the sea floor by a stem (as juveniles)
- Pieces of the stem, which form disks, are prominent in the fossil record and can comprise a major component of limestones
- Prominent in the middle Paleozoic
- Suffered a serious die-off during the Permo-Triassic extinction
- Radiated again in the Mesozoic

16.4 Colonial Metazoa

16.4.1 The Bryozoa

The phylum Bryozoa, sometimes referred to as the moss animals, appear in the fossil record in the Early Ordovician

- In the same superphylum as Mollusca and Brachiopoda (Lophotrochozoa)
- Colonial — individual animals (zooids) are tiny and perform specific functions
- Typically form calcareous skeletons
- Most common in warm, tropical waters, where they are important reef-building organisms
- Most are sessile, on hard substrates
- Common constituent of middle Paleozoic limestones
- Many species are short-ranged and cosmopolitan - thus useful index fossils
- Most Palaeozoic orders went extinct during the Permo-Triassic extinction

16.4.2 Graptolites

The Graptolithina are an extinct class of the Hemichordata; they first appeared in the fossil record in Late Cambrian

- Colonial organisms (rhabdosomes), consisting of branches (stipes) emanating from a single organism (sicula)
- Diverse morphology, but characteristically form hook-like or serrated structures
- Form skeleton of collagen
- Typically preserved in shales, but also in limestones and cherts
- Went extinct in the Carboniferous
- Make superb index fossils

Chapter 17

The Early–Middle Paleozoic: Paleogeography, Stratigraphy, and Environments

Reading: Chapters 13 (301–313) and 14 (331–339) in Stanley; Chapters 10 and 11 in Wiccander and Monroe

The transition from the Precambrian to the Paleozoic is best known for the advent and diversification of animal life, but it was also an important time tectonically, with the large continent *Gondwana* forming from fragments of other continents. The Cambrian was likely an unusually warm period in Earth's history, but Earth experienced a series of glaciations in the Paleozoic. Some of these are closely connected to impressive mass extinction events and tectonic reorganizations, which triggered significant changes to the biosphere.

17.1 Paleozoic Tectonics and Paleogeography

The land masses were largely separated into four continents of variable size in the Early Paleozoic, including Gondwana, Laurentia, Baltica, and Siberia. Most, if not all of the continental fragments would eventually collide to form the next supercontinent, Pangaea, at the end of the Paleozoic. As we already learned, North America (or Laurentia) was one of these continents that was freed from the center of Rodinia in the late Proterozoic.

17.1.1 Gondwana

Gondwana was a large, composite continent that formed from several continental fragments that broke out of Rodinia: Australia, India, Antarctica, and most of present day Africa and South America. Just as Rodinia appears to have been centred on Laurentia, so also was Gondwana centred on Africa. Gondwana began to form around 600 Ma, through a series of orogenic events that surround the various cratonic fragments of Africa.

Pan-African Orogenies

These orogenic events are collectively referred to as the *Pan-African orogenies*, because the various cratons that make up present-day Africa were at the centre of Gondwana, and so they have late Proterozoic to early Paleozoic orogenic belts on all sides. The largest of these was the *East African Orogen*, through which Mozambique and India collided with Africa, and along which the *Arabian-Nubian Shield* (a large, juvenile craton) formed. The result of the East African Orogen was likely a mountain belt on the scale of the Alpine-Himalayan chain, which presumably had profound impacts on global climate and biogeochemical cycling. One likely consequence of the Pan-African Orogenies is a peak in $^{87}Sr/^{86}Sr$ ratio of marine seawater at this time.

17.1.2 Laurentia

Much of eastern, southern, and western North America, which occupied a tropical to sub-tropical latitude, was blanked in a series of transgressive (sandstone-shale-carbonate)-regressive sedimentary sequences during the Paleozoic that can be recognized across the continent. The Cambrian and Ordovician, in particular, were time of expansive, shallow seas on the passive margins of Laurentia, which was bordered to the east by the Iapetus Ocean. These continental scale flooding events were punctuated by orogenic events on both sides of North America.

Paleozoic sedimentary sequences

You will recall that Laurentia broke out of Rodinia through a series of rifting events on the western and eastern margins of North America that began in the latest Precambrian. The results was that by the early Cambrian, present day North America was largely surrounded by passive margins and much of the continent, was in fact, flooded by shallow seas. The ocean basin that opened on the east margin of Laurentia at this time is known as the *Iapetus Ocean*.

- Sauk sequence (Cambrian) — a continent-wides transgression; much of North America flooded by shallow seas. Followed rifting on the eastern and western margins of the North America
- Tippecanoe sequence (Ordovician) — extensive carbonate reefs in North America and a foreland basin in the east (the carbonates and shales around Montreal). Capped by the Queenstown Delta clastic wedge. Subsequently exposed due to glacial eustasy and Taconic orogeny
- Kaskaskia sequence (Silurian–Devonian) — passive margin to foreland basin (Acadian orogeny) — capped by the *Catskill clastic wedge*. Much of North America was flooded at this time (an *epeiric sea*). Extensive development of barriers reefs in western North America (Devonian) at this time (now important petroleum reservoirs)
- Absaroka sequence (Late Carboniferous to Permian)

Paleozoic orogenic events in eastern North America

The overriding tectonic event(s) of the Paleozoic was the amalgamation of the supercontinent Pangaea, which was largely complete in the Permian when Laurasia (Laurentia plus Baltica) collided with Gondwana. But despite the fact that it was large continent-continent collisions that stitched Pangaea together, there were many smaller orogenies that deeply effected the paleogeographic evolution of Laurentia. The series of Paleozoic collisions along eastern Laurentia make up the *Appalachian orogenic system*.

- Taconic orogeny: Late Ordovician-early Silurian (440 Ma) collision of island arc with northeastern US/eastern Canada (east-dipping subduction—beneath the Taconic microcontinent)
- Acadian orogeny: Late Devonian (375 to 325 Ma) collision of the Avalon Terrane (which rifted off of northern Gondwana and opened the *Rheic Ocean* behind it) with New England and eastern Quebec (east-dipping subduction)
- Alleghenian orogeny: Carboniferous (350 to 300 Ma) collision with northwestern Gondwana (NW Africa), from SE US to Newfoundland
- Ouachita orogeny: Middle Carboniferous collision of part of South America with southeastern Laurentia; extension of the Alleghenian orogeny, which began in the north and progressed southward

Paleozoic orogenic events in western North America

Western North America remained a stable margin through the early Paleozoic. By the Devonian, a long belt of carbonate reefs rimmed the margin in what is now western Canada. These reefs were flooded in the late Devonian and blanketed in mud as foreland subsidence began in response to the impending collision of the Eastern Klamath island Arc (west-dipping subduction). This orogenic event is referred to as the *Antler orogeny* and was the only major orogenic event to effect western Laurentia in the Paleozoic.

17.1.3 Other important Paleozoic orogenic events

The Caledonian orogeny

The Scandian-Grampian phase (425–395 Ma) of the Caledonian orogeny was essentially an extension of the Acadian orogeny and extended through the northern British Isles and eastern Greenland to western Scandinavia, where Baltica collided with NE Laurentia to form the continent *Euramerica*. The Baltic margin had previously collided with at least one island arc to the west and with Avalonia to the south. Along with the Acadian, the Caledonian orogeny closed the Iapetus Ocean.

The Hercynian orogeny

The Carboniferous Hercynian orogeny (Carboniferous) is the extension of the Alleghenian orogeny in North America into Europe. It resulted in complex patchwork of deformation that resulted from the collision of Gondwana with Euramerica.

17.1.4 Pangaea

The final amalgamation of Pangaea occurred in the Permian, when Siberia collided with Euramerica (eastern margin of the Baltic craton) to form the Ural Mountains. Pangaea at this time was a C-shaped supercontinent, with a large ocean basin dubbed the *Paleotethys* lying within the C (between Gondwana and *Laurasia*). This ocean basin closed soon thereafter when the Cimmerian (much of present day Turkey and Iran) ribbon terrane rifted off of NE Gondwana and collided with Laurasia. The *Neotethys* ocean basin opened in its wake.

17.2 Paleozoic Climate

The early Paleozoic is distinguished by high pCO_2 levels (10–20 $\times$ present CO_2) and a warm climate that persisted through most of the Cambrian and Ordovian. CO2 had declined to more typical Phanerozoic levels by the Carboniferous, and this was closely related to the burial of vast amounts of coal in the Carboniferous and the onset of the Permo-Carboniferous glaciation (discussed in detail in a subsequent chapter), the last and probably largest of three glacial epochs during the Paleozoic. This carbon burial event is also clearly revealed in the marine carbon isotope record of the late Paleozoic.

17.2.1 Late Ordovician glaciation

The late Ordovician (Hirnantian) glaciation followed a decline of about 40% in atmospheric CO_2 levels. This was probably driven by increased burial of organic carbon, based on a coeval positive excursion in carbonate isotopes. But it appears that CO_2 levels were still very high, which long posed an enigma to geologists and climate models.

Based on oxygen isotope evidence, the glaciation appears to have lasted only about 1–2 million years. Sufficient continental ice formed to lower sea level significantly and bring the the high sea levels of the Sauk sequence to and end. Global cooling and loss of habitat were two likely key factors contributing to the first of two phases of the *late Ordovician mass extinction*. The second phases, which was not as severe, occurred at the end of the glaciation and was presumably the result of rising temperatures and pCO_2.

Ice sheets during the late Ordovician covered what we usually think of as northern Gondwana (North Africa and northeastern South America), which were likely close to the South Pole at the time.

17.2.2 Late Devonian glaciation

The Silurian and much of the Devonian were again warm period in Paleozoic Earth history. This is perhaps best seen in the high global sea level and widespread development of reef systems. However, another glaciation, with ice sheets centred on Brazil, occurred in the late Devonian.

Like the previous glacial epoch, the late Devonian glaciation appears to have been relatively short-lived, and associated with mass-extinction. It may also have occurred during a time of relatively high pCO_2, although carbon dioxide levels must still have declined to help trigger the glaciation. The drop in pCO_2 is likely related to the early expansion of land plants. Not only would trees have become an important sink for organic carbon, but also land plants accelerate silicate weathering in soils by increasing CO_2 concentrations in the soils (via decay of organic detritus) and releasing organic acids, which help to break down silicate minerals.

Like the Orodovician glaciation, the Devonian glaciation appears to have been closely linked to one of the big five mass extinctions, which will be discussed in more detail in a subsequent chapter.

Chapter 18

Paleozoic Vertebrates and the Origin of Land Plants

Reading: Chapters 13 and 14 in Stanley; Chapter 13 in Wiccander and Monroe

18.1 Origin of the Vertebrates

As we learned earlier, the recognition that conodonts, dating to the early Cambrian, belong to a group of lampreys, which belong to the phylum chordata, established that chordates had evolved already by the early Cambrian. Some Ediacaran fossils may represent stem group chordates, which would push this divergence back in Earth's history, consistent with molecular clock data.

Chordates are animals with a notochord and tail (at least at some point in the their development. Ontogenetically, chordates are distinguished by radial cell cleavage in early embryo development, which contrasts with the spiral cleavage in all invertebrates. It is for this reason that chordates are closely aligned with the echinoderms (within the *deuterostomes*), who also show radial cleavage.

The *vertebrates* are a subphylum of the the *chordates* and dominate the chordata phylum in terms of diversity. They are distinguished as having a backbone. The record of the early evolution of the chordates and the divergence of vertebrates is poor, due in large part to the dominantly soft-bodies of the early chordates. These include both jawless fish and the jawed vertebrates, the group to which we belong.

18.2 Fish

The earliest vertebrates were jawless fish (*Agnatha*; e.g. lampreys and hagfish), which first appeared in shallow seas in the Late Cambrian. Most were likely bottom dwellers, feeding directly off the seafloor. Fish diversified significantly in the Devonian, which is often referred to as the *Age of Fish*.

- The *ostracoderms* were early, armored, jawless fish that prospered in the Silurian and Devonian but went extinct by the late Devonian extinction

- The jaw was an important evolutionary innovation that fundamentally changed the marine ecosystem and eventually terrestrial ecosystems.
  - The jawed fish appeared in the early Silurian
  - originated from the first two gill bars, possibly originally for enhanced respiration
  - teeth originated from the *denticles* on the skin that covered the gill bars
- The *acanthodians* likely gave rise to the armoured jaw fish, cartilaginous (i.e. sharks and rays) and bony fish. They contained several key features that distinguish modern fish, such as paired fins and scales instead of plates.
- Armored jaw-fish (*placoderms*) fierce predators, abundant in Silurian and Devonian, disappeared in Permian. The placoderm *Dunkleosteus* grew to be up to 7 m long!
- Cartilaginous fish (*chronodytes*), which include sharks and rays, appeared in the Early Devonian
- Bony fish subdivided into the ray-finned and the lobe-finned fish
  - Ray-finned (*Actinopterygii*) fish include most common fish types (i.e. tuna and salmon); appear to have evolved in fresh waters in the Devonian
  - Lobe-finned fish (*Sarcopterygii*) appeared in the Devonian. Include lungfish and coelacanths
    - muscular fins with articulated bones
    - strong swimmers (perhaps evolved to swim up rivers)
    - amphibians likely evolved from the *crossopteryginians*, which had backbones, limbs that could be used for walking, and primitive lungs
- Appearance of fish likely placed significant evolutionary pressure on invertebrates (for example the trilobites)

18.3 Amphibians

The earliest vertebrates to walk on land were ancestral amphibians that evolved from the lobe-finned fish in the Late Devonian. One important question is why the lobe-finned fish evolved limbs in the first place. Paleontologists believe they served to help navigate shallow water environments, like tidal flats, streams, and swamps. The earliest land-dwelling animals had to overcome many significant barriers to living out of the water:

- desiccation
- respiration
- gravity
- reproduction

The recently discovered *Tiktaalik* ("fishapod") represents the transition between the crosspteryginians and the earliest tetrapods (the early amphibians).

- 375 Ma

- has gills and lungs
- broad head with eyes on top
- large rib cage

18.3.1 Tetrapods

The tetrapods represent the first truly land animals and are the evolutionary bridge between the amphibians and the reptiles. They first appeared in the latest Devonian.The tetrapods also had a strong backbone, four fully developed legs, and pelvic and pectoral girdles, which made them well suited for traversing on land. They diversified rapidly in the late Paleozoic, before giving way to the reptiles. However, the remained limited by their requirement to lay their eggs in water.

18.4 Reptiles

The key evolutionary innovation of the reptiles was the development of the *amniote egg*, which is filled by a liquid-filled sack, with both yolk (food) and allantois (waste) sacs. The shell provided protection from the environment, importantly preventing desiccation and removing the need to lay eggs in the water. The amniotic egg also eliminated the need for a larval stage.

- Likely evolved from ancestors to the labryinthodonts
- Appeared in early Late Mississippian (Carboniferous)
- Oldest known reptile fossil (*Westlothiana*) found in Late Mississippian (middle Carboniferous) in Scotland
- Earliest reptiles part of the groupe *protorothyrids*—small, agile animals
- Diversified in the Permian, displacing the amphibians on land
- *Pelycosaurs*, became the dominant reptiles, and included both herbivores and carnivores.
  - Went extinct in late Permian
  - Ancestors of the *therapsids*
    - Mammal-like reptiles that appeared in the Permian and dominated the terrestrial vertebrates by the P-T extinction
    - Fewer skull bones, fused; larger jawbone; differentiated teeth
    - More vertically-oriented limbs
    - Possibly *endothermic*
    - Found from low to high latitudes

18.5 Land Plants

The first land plants appeared in the middle-late Ordovician and evolved progressively from a marine to freshwater environment, where they were able to adapt gradually to differing osmotic pressures. They are probably derived from the green algae, which are the only algae to have colonized fresh water. Land plants subdivided into *non-vascular* and *vascular* plants.

Non-vascular land plants include the *bryophytes* (liverworts and mosses) that are mostly restricted to moist environments.

18.5.1 Vascular plants

Vascular plants likely appeared sometime in the Silurian. They are distinguished from the non-vascular plants in that they have tissues made from specialized cells that enable them to transport water and nutrients and provide rigidity to the plant. Other key innovations of the vascular plants include

- *lignin* and *cellulose*, which provide additional strength
- *cutin*, which inhibits desiccation, oxidation, effects of ultraviolet light, and parasites
- roots to collect water and nutrients
- leaves for more efficient photosynthesis

Early vascular plants

The earliest vascular plants of the genus *Cooksonia* date from the Middle Silurian and have small, y-shaped stems.

- Seedless
- Lacked a true root system (water and nutrients from the *rhyzome*)
- Lived in low, marshy, freshwater
- Diversified in the Late Silurian and Early Devonian
- *Heterospory*—two types of spores that give rise to female and male plants
- Achieved great size (up to 10 m) by the Late Devonian — first forests
- First ferns appeared in Middle Devonian
- Major source of the Carboniferous coal, e.g. *lycopsids* and *sphenospids* (rushes)

Seed-bearing vascular plants

The first seed-bearing plants likely evolved in the Late Devonian. Seeds reduced the need for plants to remain in moist environments, thus allowing them to diversify and populate dry environments. This key evolutionary innovation led to the first full-fledged forests, which began to spread across the terrestrial landscape in the late Devonian.

Seed ferns appeared in the Carboniferous and eventually gave rise to the gymnosperms (think conifers), which came to be the dominant trees by the late Permian.

18.5.2 Consequences of evolution of land plants

The evolution of land plants and their colonization of the continents had huge implications for the biosphere, climate, and soils.

Land plants spur animal evolution

Early trees would have provided protection from direct sunlight to early animals. Hence, the first amphibians arose about the same time that the first forests spread across the landscape, in the late Devonian.

18.5.3 Soils and erosion

Root systems provide trees with both better access to water and nutrients in the soil and stabilize that part of the plant the projects upward from the ground surface (big trees need big root systems). The emergence of root systems had a profound effect on the development of soils and landscapes, because they anchored soils in place. Hence, richer soils developed, both because nutrient rich minerals are held in place and because the plants contribute organic matter to the soil, replenishing nutrients in a more bioavailable package.

Some scientists content that the development of vascular plants and root systems lead to a change in the style of rivers. Whereas rivers appear to have been exclusively braided (that is, dominated by erosion) prior to the Devonian, meandering rivers emerged afterwards, as the result of the stabilizing affect of plants.

18.5.4 Climate

The origin of land plants would have also impacted climate in two important ways. First, it would have been a new route for organic carbon to be produced and buried. Evidence for this new style of carbon burial is seen in the vast coal beds that formed in the latest Paleozoic (the aptly named *Carboniferous* Period). Two consequences would have been greater removal of CO_2 from the atmosphere and increased O_2 concentrations in the atmosphere.

The development of land plants would also have greatly changed the way silicate weathering works. Respiration in soils tends to concentrate CO_2 in soils, such that concentrations are much higher than in the ambient atmosphere. Plants also produce a variety of organic acids, which took over from CO_2 some of the role of driving silicate weathering (hydrolysis) reactions. It is likely no coincidence that two of the Phanerozoic glaciations corresponded to major evolutionary developments in vascular plants.

Chapter 19

The Late Paleozoic and the Permo-Triassic Extinction

Reading: Chapter 14 in Stanley; Chapters 12 in Wiccander and Monroe

19.1 Pangaea

As you will recall from previous lectures, the great Paleozoic landmass Gondwana had effectively finished assembling by the early Paleozoic. Throughout the remainder of the Paleozoic, Laurentia grew through accretion of terranes on both the eastern and western margins, finally colliding with Baltica (Caledonian orogeny) in the Devonian, to form Eurasia. Collision between the Siberia and Baltica formed the Urals and gave rise to the large continent known as Laurasia.

At about the same time as the collision of Siberia with Baltica, Gondwana began to collide with Laurasia in what is known as the Alleghenian-Hercynian orogeny (think Appalachians). This resulted in the nearly complete amalgamation of the continents into the supercontinent Pangaea, which was a C-shaped landmass with a large ocean basin dubbed the *Paleotethys* lying within the C (between Gondwana and *Laurasia*). This ocean basin closed soon thereafter when the Cimmerian (much of present day Turkey and Iran) ribbon terrane rifted off of NE Gondwana and collided with Laurasia. The *Neotethys* ocean basin opened in its wake. The *Panthalassa* ocean bordered the western side of Pangaea.

19.2 Late Paleozoic Climate

19.2.1 Permo-Carboniferous glaciation

The onset of late Paleozoic glaciation in the middle Carboniferous (about 315 Ma) roughly coincided with the Alleghenian-Hercynian orogenies, but was probably most closely related to spread of large vascular land plants in the Carboniferous, including the gymnosperms, which formed immense forests that give rise to vast peat and coal deposits. The combination of CO_2 sequestration through the formation of these peat and coasl, and an increase in silicate weathering due to the formation of organic acids by these new vascular plants, drove CO_2 to its lowest levels in the Paleozoic.

Evidence of this 'carbon burial' event are seen in the unusually high carbon isotope ratios in the Carboniferous. This high rate of burial of carbon was balanced by high O_2 concentrations—probably the highest in Earth's history.
At the same time, Gondwana drifted over the south pole. The result was the most severe glaciation of the Phanerozoic, lasting 30 million years. Glacial deposition appeared to have occurred diachronously on Gondwana as it drifted across the pole, and on Siberia, which was in the northern high latitudes at the time.

19.2.2 Carboniferous coal and cyclothems

The Carboniferous, as its name implies, was a time of widespread coal deposition, which followed the spread of large vascular land plants, in particular varieties that had developed thick mantles of bark (lignin), which is difficult to decompose. Carboniferous coals are widespread in North America and Europe, but also occur in northern Gondwana (where they formed just to the north of the limit of continental ice sheets). The vast swamps in which the precursor plants to the coal grew probably owed their existence to glaciation, which had lowered sea levels and created vast, lowland swampy regions on the continental margins, not unlike the Florida everglades. The Carboniferous coal dominantly occurs as seams a few meters thick within *cyclothems*, which are repetitive, terrestrial-marginal marine cycles that likely record glacioeustatic fluctuations in sea level.

19.2.3 Late Permian drying

The formation of large, moisture-capturing mountain ranges and a supercontinent with much land area far removed from the oceans resulted in aridification of much of interior Pangea. The resulting decrease in coal formation likely contributed to rising pCO_2 and the end of glaciation. Coniferous trees, which are quite tolerant of dry conditions, spread and floras were generally provincial due to sharp climatic gradients from the coasts to inland. Large volumes of continental red-beds and non-marine and marine (for example, the Zechstein basin) evaporites were deposited during the Permian.

19.3 Late Paleozoic Life

19.3.1 Plants

Plants continued to diversify in the late Paleozoic, and their proliferation accounts for the name of the Carboniferous Period. This was a wet and warm area over much of Pangea and Laurasia, and newly evolved trees and ferns thrived in the expansive wetlands. The spore plants, like the lycopods, that had evolved earlier, continued to thrive under these conditions. Among the spore plants, the segmented *sphenopsids* were particularly important. However, both spore-producing and *seed ferns* also diversified during this time.

The gymnosperms, which include conifers, also diversified during this time and came to dominate large vegetation. One group of gymnosperms, the *Cordaites*, formed large trees that populated forests in dry regions.

19.3.2 Animals

Both vertebrates and invertebrates continued to diversify in the late Paleozoic, driven by diversification and population of new environments by land plants, coupled to high ambient oxygen levels.

19.3.3 Insections

The insects enjoyed a particularly profound role in late Paleozoic ecology, due in large part to the acquisition of flight among some clades. The early flying insects include dragonflies and mayflies; the latter grew upwards of a meter in wingspan. Indeed, many large insects evolved at this time. Some millipede-like insects grew to nearly 2.5 m long! And spiders the size of medium dogs roamed the forests.

By the Permian Period, insects with folding wings (which include most modern flying insects) had appeared, indicating a major diversification of insects prior to the end of the Paleozoic.

19.3.4 Amphibians and Reptiles

Amphibians had begun to populate marginal marine/coastal settings and were large creatures, compared to those of today. Amphibians were the dominant land animal through most of the Carboniferous, but began to give way to the reptiles by the late Carboniferous. One particular advantage that the reptiles evolves was a strong jaw. The were also faster and generally more agile than the amphibians.

As mentioned, the key innovation in reptiles was the development of the amniote egg, which allowed reptiles to remain on land when giving birth. Another innovation of later reptiles was a much more advanced, stronger jaw, which along with new blade-like teeth, enabled them to slice up their pray rather than eat them whole, as early reptiles and amphibians were forced to do.

The *pelycosaurs* were a fin-backed reptile that thrived atop the carnivorous food chain in the Permian. Their skull structures and other skeletal features resemble those of mammals, and they likely gave rise to the therapsids.

19.3.5 Therapsids

The *therapsids*, the progenitors of mammals, evolved late in the Permian, thus marking a break from the lineage that would give rise to the dinosaurs. They had legs that sat more squarely underneath their bodies and developed complex and powerful jaws. Some of the therapsids may even have been *endothermic* (warm-blooded). Some appear to have had hair for insulation. The *ectothermic* reptiles were likely much slower than their therapsid cousins, and not well adapted to deal with the cooler conditions that prevailed in the late Carboniferous and Permian.

19.4 Extinctions in the fossil record

Although the diversity of organisms on Earth has generally increased over the Phanerozoic (at least, at the family level and below), diversification in the fossil record is nearly matched by extinction. Thus, extinction is a regular part of the fossil record throughout Earth history (at least, that part of Earth history for which we have a reasonable fossil record) and of the biosphere, as recognized already the early paleontologists Georges Cuvier and William Smith. Extinction is most readily observed at the lower levels of the systematic groupings, namely family, genus and species, each of which reveals something different about the nature of extinction.
Extinction events are sharp and short-lived increases in the the rate of extinction above the background extinction rate.

19.4.1 Mass Extinctions

What is a *mass extinction* event? Sepkoski (1981) (1981) originally defined mass extinctions as intervals which suffered a prominent loss in diversity at the family level. An alternative approach is look for peaks in the extinction intensity (Fig. 1). In any case, the difference between a mere extinction event and a mass extinction event is somewhat arbitrary, even though most people agree that there were 5 big extinction events and one to two dozen important but smaller extinction events. Overall, the severity of extinction events has decreased over the Phanerozoic (Fig. 1). The most significant extinction events involve loss at the family level both in the ocean and on land. Generally, extinctions effect tropical organisms most severely.

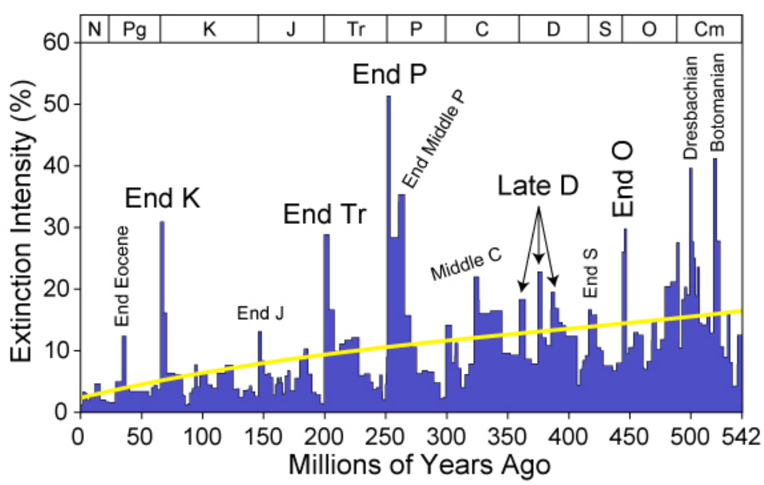

Figure 19.1: Mass extinctions may be discerned in the fossil record by unusually high extinction rates (relative to background extinctions). From Wikipedia (Extinction Events).

Extinctions are important in the course of evolution and Earth history because they clear the way for subsequent radiation events, when surviving groups of organisms rapidly fill the newly vacated ecological space left by the extinct organisms. In this way, the make-up of the biosphere has shifted abruptly many times, with certain groups reigning prominent during one calm interval, only to decline in importance after an extinction event or vice

versa. For example, the Trilobites, which were so such an important component of early Paleozoic seas, were decimated in the late Devonian extinction and finally went extinct during the Permo-Triassic extinction.
Many extinction events are thought to result from a combination of long-term environmental stress and a proximal trigger, which ultimately kills them off. A significant source of confusion when discussing mass extinctions is differentiating between the trigger and the kill mechanisms.

- The *trigger mechanism* is the geological event that leads to extinction events
- The *kill mechanisms* is the actual environmental condition (resulting from the trigger mechanism) that is responsible for killing off groups of organisms

For example, in the case of an asteroid impact, the impact itself, although it will certainly lead to many extinctions of spatially restricted groups of organisms that might be in the region of the impact, it is the follow on effects from the impact that most likely are responsible for the majority of extinctions.

19.4.2 Causes of mass extinctions

Identifying the causes of mass extinctions is not easy, for these causes need to be consistent with available paleontological evidence (e.g., including all groups of organisms that survived and those that went extinct) and geological evidence (e.g. ocean acidification, fluctuations in sea level, etc.). Probably, many mass extinctions were the result of multiple overlapping causes.

- Fall in sea level (exposure of continental shelves)
- Global cooling
- Global warming
- Flood basalt volcanisms
- Meteorite impacts
- Ocean anoxia
- Sulfidic deep waters

19.5 The Big Five Mass Extinctions

Extinction event	Genus loss (%)	Species loss (%)
End Ordovician	60	85
Late Devonian	57	83
Permo-Triassic	82	95
End Triassic	53	80
Cretaceous-Tertiary	47	76

19.5.1 The End Ordivician Event

The first of the big mass extinctions occurred at the end of the Ordovician Period (*End Ordovician* event): 447-444 Ma

- 49% of marine genera
- 100 families
- Brachiopods and Bryozoa hit hard
- Also trilobites, graptolites, bivalves, and echinoderms (in particular, corals) conodonts
- Temporally associated with the Late Ordovician glaciation and drop in sea level (end of the Tippecanoe sequence)

19.5.2 The Late Devonian Event

The next big mass extinction occurred around the Frasnian-Famennian (364 Ma) boundary

- Likely a series of extinction events over a few tens of millions of years
- Selectively decimated shallow, warm-water marine organisms (tropical reef crisis)
  - Stromatoporids, rugose and tabulate corals hardest hit
  - Trilobites, brachiopods, ammonites, conodonts, and fish also hard hit
  - Land plants largely unaffected
- Also associated with glaciation/global cooling

19.5.3 The Permo-Triassic Event

The End-Permian event (ca. 251 Ma) was the big boy of mass extinctions.

- 52% of marine families
- 70% terrestrial vertebrates
- Many extinctions of major groups of organisms
  - Rugose and tabulate corals, Trilobites, Blastoids, Graptolites
- Just about all other major classes heavily hit
- Only known extinction event to hit the insects
- Bivalves not quite as severely as the others
- Extinction gradual (over about 5 million years)
- Final event seems to overlap with extrusion of the Siberian Trap flood basalts
- Anoxia, acidification, severe perturbations to the carbon cycle
- Delayed recovery
- Cause heavily debated (Flood basalt volcanism, impact...)

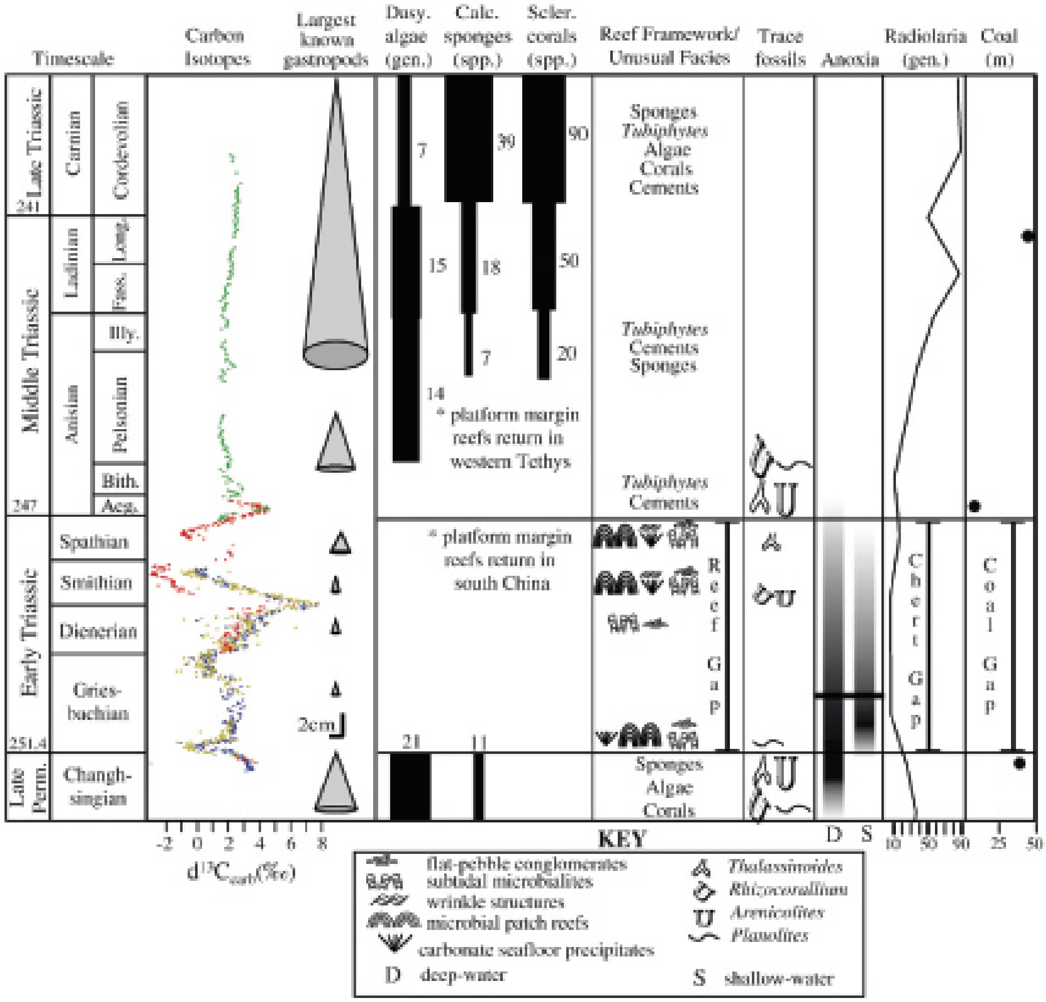

Figure 19.2: Summary of the geological record following the End-Permian extinction (Knoll et al., 2007).

19.5.4 End-Triassic Event

The end-Triassic extinction occurred at 210 Ma

- Killed off 20% of marine families
- Ammonoids, bivalves, gastropods, and brachiopods hard hit
- Occurred very rapidly (10,000 years)
- Opened up the ecological niche that the dinosaurs filled

19.5.5 The K-T Event

The *End Cretaceous* (65.5) Ma event is the best known mass extinction event

- Killed off the dinosaurs

- As well as the ammonoids and rudist clams
- Planktonic marine species especially hard hit
- Land plants and many terrestrial vertebrates not heavily affected
- Associated with an immense meteorite impact, but other related kill mechanisms
- The boundary layer - Ir anomaly, shocked quartz, tektites
- The Chicxulub Impact on the Yucatan Peninsula was contemporaneous

Chapter 20

Mesozoic Life

Reading: Chapter 16 in Stanley; Chapter 15 in Wiccander and Monroe

20.1 Recovery from the Permo-Triassic extinction in the marine realm

The biosphere was decimated at the beginning of the Mesozoic as a result of the Permo-Triassic extinction, and life had a distinctly different flavor in the Mesozoic. Recovery from the extinction took many millions of years, probably the result of the continued inhospitable environment (anoxic, and maybe sulfidic deep oceans) that caused or at least heavily contributed to the extinction.

- *Reef gap* delayed recovery of reef-forming organisms
- *Coal gap* apparently delayed recovery of large trees in swampy areas
- Stromatolites enjoyed a brief resurgence
- Only mollusks and ammonoids reasonably abundant in Early Triassic

By the middle Triassic, the recovery had begun. Whereas many important Paleozoic biota were lost to extinction (the rugose and tabulate corals, the trilobites, fusilinid foraminifera), other groups thrived in the Mesozoic.

20.1.1 Benthic biota

- Bivalves expanded and became hugely important in the benthic environment
  - Rudist bivalves important reef builders (calcite)
  - Important diversification in Cretaceous
- Sea urchins thrived after the early Triassic
- Gastropods diverged and took on a more "modern" appearance, particularly in Cretaceous with appearance of the *Neogastropods*
- Early reef builders were sponges and algae

  - eventually the hexacorals (*Scleractinia*) took over; at first, small reefs, then large, diverse reefs
  - some groups lived in deep waters (i.e., no symbiosis with algae)

- Brachiopods and crinoids declined, particularly in late Mesozoic

20.1.2 Pelagic biota

- Ammonoids thrived after near extinction (just two genera survived)
- Belemnites (evolved in late Paleozoic) diversified and were important predators
- Conodonts disappeared by the Jurassic
- Fish changed dramatically
  - Early ray-finned fish still rather primitive compared to most extant examples; they had, for example, triangular, non-overlapping scales.
  - Evolution of the swim bladder
  - Some modern sharks evolved (e.g. the tiger shark) by early Mesozoic
  - *Teleost* fish, which included many modern fish, evolved by late Cretaceous
    - Overlapping, circular scales
    - Symmetric tails and specialized fins
    - short jaws
- Some reptiles returned to the seas
  - *nothosaurs*
  - *placodonts*
  - *plesiosaurs*
  - *ichthyosaurs*, known as the "fish-lizard" appeared somewhat later in the Mesozoic and are distinct from the other swimming reptiles in that they gave birth to live young
  - Crocodiles also emerged in the Mesozoic, although later than the other groups. Thet diverged from the last common ancestor to the dinosaurs, and hence are a sister group to the dinosaurs.
- Dinoflagellates (which include photosynthesizers) and foraminifera (heterotrophic protists) diversified
- Expansion of calcareous nannoplankton, such as the *Coccolithophora*, and calcareous forams
  - profound influence on carbonate sedimentology
  - carbonate precipitation in open ocean (not constrained to shallow water platforms); shifted the locus of carbonate deposition away from continental margins and into the deep sea.
- Diatoms appeared in Cretaceous, with a large impact on the silica cycle. Now undersaturated in the surface ocean.

20.2 Life on land in the Mesozoic

20.2.1 Gymnosperms: early Mesozoic dominance

The Permo-Triassic extinction hastened the demise of the late Paleozoic terrestrial flora and thus sped up the transition to a much different Mesozoic flora. Terrestrial flora did not begin to fully recover until the Middle Triassic, at which point the Lycopods were the dominant land plant.

- Ferns particularly abundant during the early–middle Mesozoic
- Gymnosperms thrived. Group characterized by exposed seeds, which rest on projecting scales of cones. Relied on wind to carry pollen. The gymnosperms dominated by a few groups
  - Ginkgos — gymnosperm that resembles a hardwood tree
  - Cycads — tropical, palm-like trees that grew in tropical environments
  - Cycadeoids — extinct, sister group to cycads
  - Conifers diversified and took over by the Cretaceous

20.2.2 Angiosperms: Darwin's abominable mystery

Angiosperms are an evolutionary success story. In fact, so abrupt and impressive was their diversification in the middle Mesozoic that they posed a special problem to Darwin and his theory of natural selection. In fact, Darwin invoked a special hypothesis to account for the sudden and diverse appearance on the terrestrial scene that is not unlike Gould and Eldridge's punctuated equilibrium: he postulated that angiosperms first evolved and diversified on an isolated tropical island.

It is now well recognized that the rapid diversification and great success of the angiosperms is due to a pair of important evolutionary adaption. First, *double fertilization*, which distinguishes angiosperms from gymnosperms, involves two sperm, one of which fertilizes an egg cell to form an embryo, and another of which leads to the development of *endosperm*, which provides nutrients to the developing embryo. The results is that angiosperm embryos can develop rapidly from seed to plant, in many cases reaching reproductive maturity in weeks or months. Fortunately for us, the endosperm produces nutrients that can nourish us as well.

Angiosperms produce nectar, fruits, and colourful flowers, which attract insects. These insects cross-fertilize angiosperm plants rapidly and efficiently, in contrast to gymnosperms, which are at the mercy of the wind. Perhaps more importantly, the reliance on insects has established a process of co-evolution, where insects and angiosperm plants evolve together in a symbiotic way, to the extent that certain plans rely on specific species of insects to reproduce, and certain insects only feed on specific species of plants. As a result, not only did the angiosperms greatly diversity during the middle Cretaceous, so also did the insects, which such key groups the bees and butterflies first appearing then.

This major diversification event, which began in the middle Cretaceous, led to a great diversity of angiosperms by the close of the Cretaceous. As part of this diversification and evolution in the late Cretaceous, vein patterns on leaves became increasingly organized. By the close of the Cretaceous, recognizable ancestors of the walnut, birch, palm, oak, and sycamore, among other groups, had evolved.

20.2.3 Early Mammals

The therapsids, which evolved from reptiles, survived the Permo-Triassic extinction and thrived during the Triassic. However, by the early Jurassic, they were in serious decline (due to the dinosaurs) and they were extinct by the end that Period. However, they gave rise to the mammals by the end of the Triassic. Mammals in the Mesozoic were small, largely tree-dwelling animals that remained relatively low diversity and ecologically peripheral throughout the remainder of the Mesozoic.

20.2.4 The Dinosaurs

The Mesozoic is often called the *Age of the Dinosaurs*. Whereas many other animal groups thrived during this time, it is true that the dinosaurs probably dominated the terrestrial animal landscape, having benefited from a few evolutionary head starts that made them large and agile compared to many of their vertebrate brethren on land. Dinosaurs evolved from the *Dinosauromorpha*, a clade of archosaur (Fig. 20.1). The crocodiles and the pterosaurs, the first flying vertebrates, also evolved from this clade in the Mesozoic.

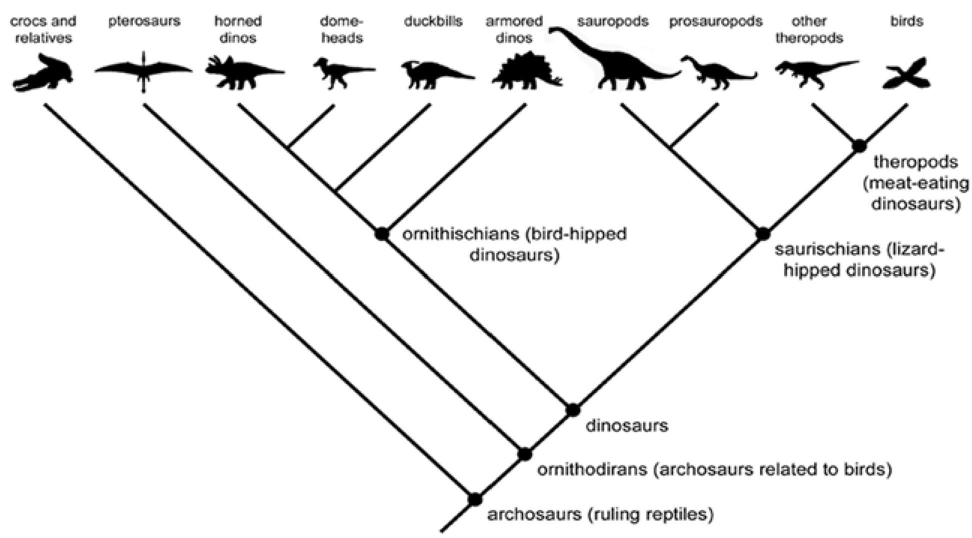

Figure 20.1: Evolutionary tree showing relationship of different groups of dinosaurs and their ancestors and sister groups.

- Dinosauromorpha included agile groups that moved on two legs, which were close together and directly below their upright body, enabling them to run rapidly.
- Dinosaurs had broader skulls and more developed teeth, but had different skulls and more developed teeth than the dinosaurmorphs.
- Early dinosaurs were mostly small; they did not become large until Jurassic

- Diverged into two main groups distinguished by pelvic structure
  - Ornithischian (bird-hipped) — herbivores (but do not include the ancestor to the birds)
  - Saurichia (lizard-hipped) — herbivores and carnivores
- Co-existed with therapsids, dinosauromorphs, amphibians, and reptiles until the end-Triassic extinction, at which point the dinosaurs took over the show
- Extinction wiped out the therapsids, opened the ecological door to dinosaurs
- Highly diverse within a 100,000 years of extinction
- Ruled the land until the end of the Mesozoic. Highly successful because
  - Many were highly agile
  - Complex system of ordering eggs
- Diverse, dinosaur-filled terrestrial ecosystems by the Cretaceous
- Matched in size by other large animals, like crocodiles (up to 15 m in length)

Evidence for endothermy among dinosaurs

In the past 20 years, evidence has mounted that the dinosaurs were not cold-blooded, lumbering creatures as popularly depicted in animations and in older textbooks.

- Must have had high rate of metabolism to suppress mammals for 140 m.y.
- Small percentage of carnivores (high metabolism requires lots of food, can't support large number of carnivores)
- Active locomotion (versus lumbering)
- Abundance of blood vessels in bones
- Evidence of brooding (like warm-blooded birds)
- High growth rates
- Some dinosaurs had feathers (which would have served to insulate)
- Dinosaurs in the high latitudes

Fly vertebrates

The Pterosaurs include the pterodactyl and other flying dinosaur-like creatures. Despite their superficial resemblance to birds, the Pterosaurs branched off the main dinosaur lineage prior to crown-group dinosaurs (Fig. 20.1), and hence were not closely related to birds. The pterosaurs had immense wing spans, and biomechanics studies of their fossils suggest that they did not regularly flap their wings, but rather most glides on air currents.

Birds emerged from the Saurichia line of dinosaurs, and the evidence for the close link to dinosaurs is mounting, with lots of recent evidence for feathered dinosaurs and perhaps flying birds with ostensibly bird-like features.

20.2.5 Frogs and Turtles

These two groups of small vertebrates originated in the Triassic

20.2.6 The Mammals

Mammals evolved from the still reptile-like therapsids in the latest Triassic. The therapsids, which had managed to suppress the dinosaurs, went extinct in the End-Triassic extinction. Mammals survived, but they remained small and relatively unimportant in Mesozoic ecology due to the subsequent dominance of the dinosaurs.

- earliest mammals resemble the modern shrew
- Key evolutionary changes
  - Differentiated and specialized teeth
  - Enlarged brain
  - Modification of jawbones into ear bones
  - Single jawbone
  - Secondary palate (to allow continue breathing while eating)
  - Suckled young
- Remained small due to predatory pressure by dinosaurs
- Many were nocturnal
- Many were tree dwelling
- First herbivores in the Jurassic
- Differentiated into the placentals and marsupials by the Late Cretaceous

20.3 The End Triassic Mass Extinction

The end-Triassic extinction occurred at 210 Ma and overlaps with the emplacement of the enormous Central Atlantic Magmatic Province (CAMP), which was related to the opening of the central Atlantic ocean.

- Killed off 20% of marine families and half of all genera
- Ammonoids, bivalves, gastropods, and brachiopods hard hit
- Occurred very rapidly (10,000 years)
- Therapsids went extinct
- Opened up the ecological niche that the dinosaurs filled

Chapter 21

Mesozoic Earth history

Reading: Chapters 16 and 17 in Stanley; Chapter 14 in Wiccander and Monroe

At the beginning of the Mesozoic, the continents were mostly amalgamated in the supercontinent Pangaea. Sea level was generally high, and carbonates were deposited in much of Europe. The Tethys Sea was a large, mostly tropical ocean basin the separated eastern Gondwana from southeastern Asia. The geology of the Mesozoic was dominated by the break-up of Pangea, starting in the the Triassic, not long after it finished forming. By the end of the Cretaceous, many of the cratons that had assembled to form Gondwana and Laurasia were again separated. Whereas most of the Paleozoic orogenic events to affect North America occurred along its east coast, Mesozoic orogeny exclusively affected western North America. With opening of the Atlantic, the east coast of North America became a passive margin.

21.1 Breakup of Pangaea

Pangaea breakup began in the late Triassic with the onset of rifting between Laurasia and Gondwana, and proceeded slowly into the Jurassic. Break-up, particularly along the Atlantic margins, was accompanied by extensive plume magmatism

- Opening of the Atlantic basin
- Associated with the 200 Ma Central Atlantic magmatic province (flood basalts)
- Tethys Sea extended between Laurasia and Africa
- Gulf of Mexico opens and North America and South America separated by the Late Jurassic
- Pacific, Atlantic, and Tethys connected by deep, narrow basins
- Vast salt deposits, important seals in many petroleum provinces
- Greenland separated from North America by Late Cretaceous

Gondwana began to rift in the Late Triassic and was mostly disaggregated by the end of the Cretaceous

- Antarctica and Australia first separated from South America and Africa; associated with emplacement of the Karoo-Ferrar flood basalt (183 Ma)
- India rifted apart from eastern Gondwana and begin to drift to the north
- South America and Africa began to rift apart in the late Jurassic; associated with Paraná-Etendeka flood basalt (130 Ma)
- Australia and Antarctica rifted in the late Cretaceous

The breakup of Pangaea had profound effects on ocean circulation and regional climate

- Pangaea interior typically arid, global climate dominated by large Pacific ocean
- Tethys realm tropical during the Mesozoic; site of most coral reefs
- Temperature gradients low during first half of Mesozoic – mild climates to 60N
- Drift of continents to the high latitudes increased global temperature gradients
- Rigorous and complex ocean circulation

21.1.1 The Cretaceous

The Cretaceous was a time of continental fragmentation and the development of narrow seaways. Ocean circulation changed, particularly when the Tethys seaway connected to the Pacific ocean. Evaporite were deposited on the margins of many of the newly developed seaways.

The Tethys Seaway

The Cretaceous Tethys Seaway was much different from its predecessor, the Tethys Sea, which was a large ocean basin nearly encircled by Pangaea. The trade winds drove a vast ocean current through the seaway, distributing the warmth of tropical ocean waters to the southern Europe and eastern North America.

21.2 North America in the Mesozoic

In the early Mesozoic, sedimentation in North America was dominated by terrestrial deposition in the continental interior and along the recently formed Appalachians.

21.2.1 East Coast

- Onset of breakup between North America and Africa in Late Triassic
- Deposition of coarse, poorly sorted, red sediments in non-marine rift basins (Newark Group)
- Intrusion of dikes and sills (for example the Palisades sill along Hudson River)
- East coast passive margin established by about 180 Ma
- Gulf of Mexico opened as a shallow, restricted basin. Site of thick salt deposits

- Opened to ocean and deepened by late Jurrasic
- Widespread *rudist* (reef-forming bivalves) reefs (superb reservoirs)

21.2.2 West Coast

Sonoman orogeny

A second island arc system collided with western North America around the Permian-Triassic boundary as a result of west-dipping subduction during the *Sonoman orogeny*. This orogenic event was similar to the Antler orogeny, but in this case, the volcanic arc (the Golconda arc) was formed on the edge of an *exotic* continental terrane (Sonomia) rather than on oceanic crust (as in the Klamath arc), such that the orogeny added a significant amount of new material to western North America. After this collision, the western margin of North America was again a stable, passive margin.

The Classic Model: Subduction polarity switch

Following docking of the Sonoma terrane, the west coast was tectonically quiescent through the Triassic. The traditional interpretation of the Mesozoic tectonic evolution of North America holds that following the resumption of tectonism in the Jurassic is related to foundering of the oceanic crust on the trailing (western) margin of Sonomia terrane (now North America), leading to the onset of east-dipping subduction: a subduction polarity switch. From the Jurassic onward, the west coast was an active continental margin, dominated by east-dipping subduction of the Farallon plate (part of the ancestral Pacific ocean basin) and the accretion of island arcs and microcontinents brought in on the subducting plates. The tectonics of western North America in the Mesozoic mirrored the tectonics of eastern North America in the middle-late Paleozoic.

Cordilleran orogenesis

The *Cordilleran orogeny* refers to the series of orogenic events the influenced the western margin of North America from the middle Jurassic to the early Cenozoic, after the switch in subduction polarity

- Nevadan orogeny — middle Jurassic to early Cretaceous
  - Sierra Nevada arc and widespread, eastward-migrating batholith emplacement in western North America. The present day Sierra Nevadas, which are composed mainly of granodiorite, are an erosional relic of this arc.
  - Deposition of Franciscan complex (mélange) and Great Valley Group in the forearm
  - Addition of a large, composite exotic terrane to western Canada and Alaska
- Sevier orogeny — Cretaceous
  - result of ongoing eastward subduction (thus east-west compression of the margin)
  - Alaska to Mexico

- Laramide orogeny — late Cretaceous to early Paleogene
  - Effected areas east of the Sevier orogeny and involved thick-skinned, block faulting
  - Responsible for much of the present day Rocky Mountains (think Colorado and Wyoming)

The Canadian Cordillera

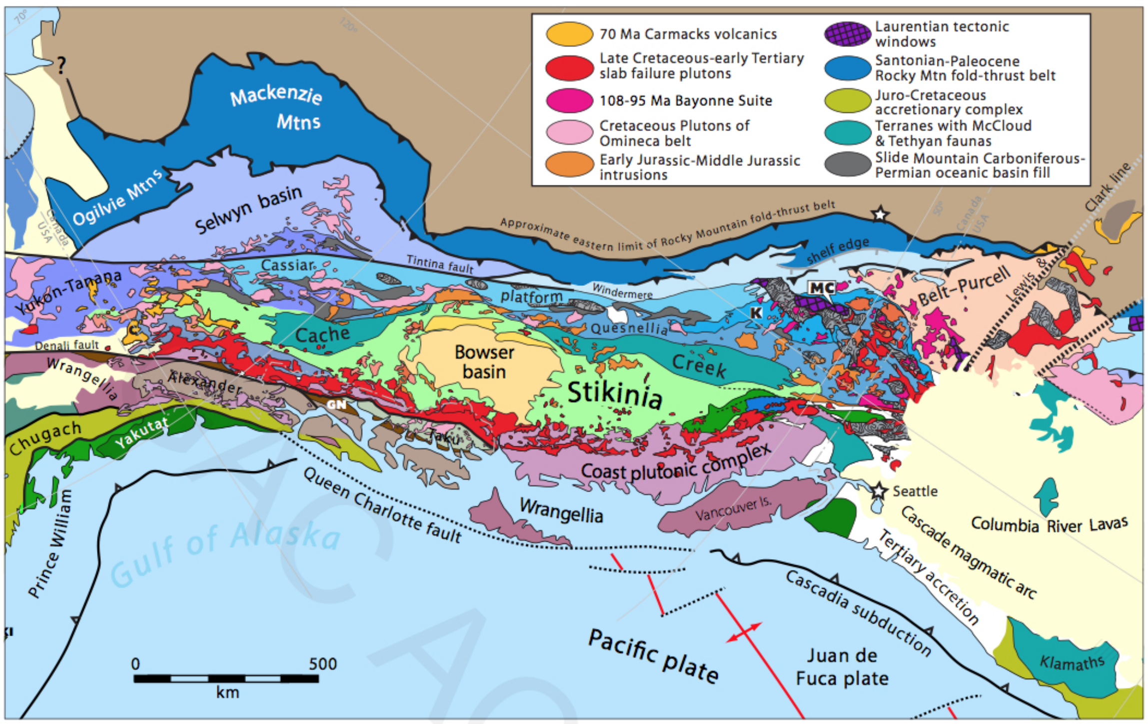

Figure 21.1: Geological map of the major tectonic terranes (bounded by major faults) of the Canadian and Alaska Cordillera. From Hildebrand (2014).

The Canadian Cordillera is a mosaic of different terranes that presumably were accreted to western North America in the Mesozoic and Paleogene. Such terranes, derived from elsewhere, are commonly referred to as *exotic* terranes implying an origin that is very different from the current geographic location. The terranes are bounded by major faults that reflect sutures (where two separate plates were joined) and large transform faults.

The alternative model of the Mesozoic tectonics of western North America

Robert Hildebrand, a former Geological Survey of Canada scientists, has recently upturned the classic model and argued that there was no subduction polarity switch at all, and that instead west-dipping subduction simply continued (Hildebrand, 2014). The model has much of what is now western North America, including the Sierra Nevada (which would have been the west-dipping arc) colliding with western North America in the Mesozoic. In this model, the Nevadan orogeny occurred on the overriding plate, which was separated from North America at the time by an ocean basin. This continent grew as a result of a series of collisions during the middle Mesozoic prior to beginning to collide with western North America in the early Cretaceous (Sevier orogeny) and culminated in the late Cretaceous (Laramide orogeny). Many of the terranes that clog up the Canadian Cordillera (Fig. 21.1) then translated northward along a major strike slip system, consistent with the *Great Alaskan Terrane Wreck* model proposed by Stephen Johnston (Johnston, 2001), which holds that many of the terranes along the Cordillera are fragmented pieces of single ribbon continent that collides with western North America.

21.2.3 Mesozoic sedimentation

Non-marine sediments were deposited across much of the southwestern US in the first half of the Mesozoic. These include many of the famous, pastel-colored units found on the Colorado Plateau, such as the Moenkopi, Chinle, and Navajo Sandstone formations which filled a foreland basin that formed in response to the Nevadan orogeny.

Following brief marine incursions by the Sundance Sea, the colorful late Jurassic Morrison Formation ($\sim$150 Ma), which is famous for its dinosaur fossils, was deposited across much of what is now the Rocky Mountain area of the western U.S.

21.2.4 Cretaceous Interior Seaway

Orogenic activity along the Cordillera resulted in the development of a long foreland basin along the eastern margin of the mountain front. As a result also of long wavelength subsidence of interior North America due to the sinking *Farallon slab* beneath it, the ocean transgressed the mid-continent from the north beginning in the Early Cretaceous. By the Late Cretaceous, this seaway was continuous with the Gulf of Mexico and much of the interior of the continent was underwater. Widespread late Cretaceous coal was deposited along coastal plains as the sea retreated in the late Cretaceous.

21.3 Mesozoic Climate

21.3.1 Early Mesozoic

In the Triassic, Pangaea was still largely intact. Much of the continental interior was far from a source of moisture, and hence dry. Redbeds and evaporites were deposited in many regions. This interior realm would have experience significant seasonality, as well, and probably moisture was heavily dependent on monsoons.

By the Jurassic, Pangaea was fragmenting and the widespread arid climate began to give way to a warm and broadly equable climate, characterized by relatively low temperature gradients between low and high latitudes. Nevertheless, coral reefs were largely restricted to the margins of the Tethys Ocean.

21.3.2 Cretaceous

Ocean Anoxic Events

Carbon dioxide concentrations were unusually high during the Cretaceous, which resulted in high average temperatures. One result of this climate was a generally warm, poorly ventilated ocean. At several times during the Cretaceous organic-rich black shales were deposited globally. The black shales deposited during these *ocean anoxic events* are important petroleum source rocks.

The Cretaceous ocean anoxic events are closely linked temporally with an intense episode of oceanic crust production, which was a direct result of break-up of Pangea. At the same time, Earth's magnetic field was unusually stable, giving rise to the *Cretaceous Superchron* normal polarity interval.

Cretaceous glaciation?

Despite the generally greenhouse climate of the Cretaceous, there are reports of minor glacial episodes, detected mainly in the oxygen isotope records of marine carbonates. However, glaciation, if it did occur, was likely restricted to small areas near the south pole, such as Australia, where some evidence of Cretaceous glacial conditions has been found.

Chapter 22

The Cretaceous-Paleogene Mass Extinction and the Paleogene

Reading: Chapters 17 and 18 in Stanley; Chapters 16 and 18 in Wiccander and Monroe

The Cenozoic, which is split into the Paleogene (65.6 to 23 Ma) and Neogene (23 Ma to present) began 65.5 with the mass extinction event the killed off the dinosaurs, the last of the giant, aquatic reptiles, and the ammonoids, and many other animals groups. By that time, Pangaea had mostly broken up and western North America was undergoing active orogenesis as the Rocky Mountains were forming along the length of the Cordillera. Mammals took over from the dinosaurs as the dominant animals on land and even recolonized the oceans. Our records of Cenozoic Earth history are much more detailed, thanks to a combination of deep sea sediment records and many terrestrial environmental records, such as lake bed sediments and cave deposits.

22.1 The K-Pg Extinction

The *End Cretaceous*, or K-Pg mass extinction (formerly known as the K-T event, for Cretaceous-Tertiary) is the best known mass extinction event because it is the most recent and understood. Most importantly for museums, children's books, and life as we know it on Earth, it killed off the dinosaurs, paving the way for an extraordinary radiation of mammals.

- Dinosaurs, giant aquatic reptiles, and giant turtles went extinct
- Ammonoids went extinct; belemnoids nearly extinct
- Planktonic marine species especially hard hit; 90% of all nannoplankton and planktonic foraminifera went extinct
- Land plants, including both gymnosperms and angiosperms, heavily hit, particularly in North America, but the angiosperms fared better
- Many terrestrial vertebrates not heavily affected

The almost certain cause of the K-Pg extinction was discovered by accident by the Nobel Prize-winning physicist Walter Alvarez, and his son, Luis Alvarez. They were analyzing

Ir concentrations in deep sea clays exposed at Gubbio, Italy for a totally different project (they were trying to constrain sedimentation rates), when they discovered an extraordinary peak in Ir concentrations at the K-Pg boundary. Ir is rare on Earth's surface, but relatively abundant in meteorites, and so, logically, they proposed that the extinction was directly linked to a meteorite impact. The anomaly has since been found globally, and along with it, three other key ingredients have been found in the K-Pg *boundary clay* layer around North and central America that point to a meteorite impact:

- Shocked quartz grains (fractured and welded grains that are formed by a sudden, high pressure shock)
- Microspherules, which are small droplets of glass formed by the rapid cooling of molten rock within the atmosphere
- Microdiamonds, which form only in space or through the sorts of shocks generated by impacts

The K-Pg boundary layer is also found by a high concentration of fern spores, which indicate that other plants died off, and ferns opportunistically repopulated much of the land surface (Vajda et al., 2001). Other evidence in support of the impact hypothesis came from tsunami deposits and concentrations of soot in the boundary clay. But the true triumph for the Alvarez family and for geology came when the K-Pg impact crater was discovered no long afterwards. The Chicxulub crater occurs in the subsurface on the northwestern edge of the Yucatan Peninsula in Mexico, and Ar-Ar dating of glass within impact site yields an age of precisely 65.5 Ma (within the range of error of the measurements).

If the impact hypothesis has now been largely settled (minus a few stubborn holdouts), the kill mechanism or mechanisms remain debated. Whatever they are, they must account for the unique extinction pattern.

- global wildfire triggered by the impact and subsequent heat generated by falling debris
- impact winter, results from the immense quantity of dust in the atmosphere
- acid rain, from generation of nitrous oxides in atmosphere and collision with a sulfate-rich target
- global warming, from collision with a carbonate-rich target (evidence for a peak in CO_2 concentrations from leaf stomata)

The K-Pg extinction nearly coincided with the emplacement of the Deccan Traps flood basalt in India, and some scientists, most notably Greta Keller of Princeton University argue that this was really the main cause of the extinction, at least of the dinosaurs. She argues that dinosaurs appear to started going extinct a few hundred thousand years earlier, and that this is consistent with the volcanic trigger.

22.2 Life in the Paleogene

Life on land changed dramatically after the K-Pg extinction, due in large part to the extinction of the dinosaurs and the diminished role of the gymnosperms. Angiosperms flourished in the early Paleogene, and mammals diversified rapidly to fill the many niches that the dinosaurs and large reptiles left open.

22.2.1 Marine life in the Paleogene

Benthic life survived the extinction fairly in tact, and the teleost fish, mollusks, and benthic foraminifera we see today closely resemble those that existed on the sea floor at the beginning of the Paleogene. The planktonic foraminifera, on the other hand, suffered immensely, and only three species are known to have survived. However, these three species diversified extremely rapidly (due in large part to the empty ecospace in the aftermath of the extinction). Likewise, the calcareous nannoplankton diversified rapidly.

The rudist clams died off during the extinction whereas the corals did not. But the corals did not immediately rebound to take over reef-building duties in the Paleogene. This might have been due to still somewhat low Mg/Ca ratios of seawater. In the Oligocene, after Mg/Ca ratios rose, the hexacorals rebounded and begin to build large reefs again.

Whales

Just as reptiles returned to the sea from land in the Mesozoic, so too did some mammals return to the water in the Cenozoic. The whales evolved from carnivorous land mammals, and in the oceans where other large predators were absent, quickly gained the upper hand. So too did giant sharks become prominent in the early Cenozoic. The cetaceans (whales) are a sister group to the artiodactyla. The earliest whales appeared in the Early Eocene and still had front limbs that enabled them to support their weight on land.

- Land-dwelling ancestors are the raoellids
- Front limbs modified to flippers
- Rear limbs lost
- Nostrils to top of head
- Large tale fluke
- Baleen and toothed whales appeared by Oligocene

22.2.2 Life on land in the Paleogene

Flowering plants diversified rapidly in the late Mesozoic. One of the keys of their success was the advent of *double fertilization*, which produced both a seed and a food supply for that seed (in contrast to gymnosperm seeds, which require much more time to furnish these nutrients, hence their slow reproductive cycle). The flowering plants were also able to attract insects, which helped with fertilization. In fact, the symbiotic relationship that developed between insects and flowering plants resulted in closely coupled evolution among some groups. For example, if a particular plant developed a new scent or colour of flower, it might attract different insects than its ancestors. As a consequence, the new variety would become genetically isolated from its close relatives, and quickly evolve to a new species.

Grasses also gained ground in large part due to the development of continuous growth, which enable them to recover quickly from grazing. The evolution of C4 plants, which perform photosynthesis in a slightly different and more efficient manner, emerged somewhat later, probably in response to falling CO_2 levels.

Land mammals diversify

The most impressive change in life on land in the Paleogene was the result of extraordinary mammalian diversification. The animal class Mammalia includes all warm-blooded vertebrates with hair and mammary glands. Mammals evolved by the late Triassic, co-existed with the Dinosaurs, and had branched into the monotremes, placentals and marsupials by the end of the Cretaceous. Many mammals went extinction during the Pleistocene, as a result of northern hemisphere glaciation, and to an uncertain extent, human predation.

- Monotremes (egg-laying mammals)—poor fossil record, appear to have distinct evolutionary history from mammals
- Marsupials — give birth to live, but embryonic young
- Placentals — amnion fuses with walls of uterus, providing for development of young before birth

The placentals account for 90% of all fossil and extant mammals. The main placental clades originated in the Mesozoic, but most of the extant placental orders originated in the early Cenozoic.

Climatic control on mammalian evolution

The earliest primitive mammals of the Paleocene largely disappeared by later in the Paleogene. Early predators took advantage of abundant forests to hide and ambush their prey. This begin to change later in the Paleogene. Indeed, the overwhelming control on mammal evolution during the Paleogene was the effect of climate. As will be discussed further below, the Eocene-Oligocene boundary marked the onset of southern hemisphere glaciation. With this cooling came the decrease in forests and expansion of grasslands. This first-order change of habitat forced both the carnivores and their prey out into the open, where speed became an increasingly important asset. The carnivores developed sleek and low-crouching bodies, while the herbivores gave rise to the ungulates, whose hands evolved to add what amounts to an extra bone in their limbs, enabling them to run faster.

By the Oligocene, many archaic mammals had gone extinct, while most modern groups of most animals families emerged, including primates. Some extremely large land mammals had also evolved, such as the *Paraceratherium*, which stood over 5 meters tall at the shoulders! Carnivorous cats and canines were also present by this time, including the

Andrewsarchus, a giant, hyaena-like creature.

Ungulates

The rapid diversification of mammals continued in the Eocene, by which time the number of mammalian families equals that today. The hoofed herbivores appeared and diverged into the *odd-toed ungulates* (e.g. rhinos and horses) and the *even-toes ungulates* (most farm animals, antelopes, camels, etc.). Early elephants also emerged in the Eocene.

The Ungulates (hoofed animals) include *Artiodactyls*–even-toed mammals– and *Perissodactyla*–odd-toed mammals.

- Most have long palm bones and toes
- Artiodactyls appeared in Eocene; most numerous today
- Most Artiodactyls are ruminants
- Perissodactyla include horses, for which there is a super fossil record (North America) showing complex diversification

Patterns of Paleogene mammalian evolution

- Paleocene fauna mostly small
- Primates first appeared by the Paleocene (possibly earlier)
- Order Carnivora originated in Paleocene; varied specialized adaptations (teeth, speed, etc.)
- First giant terrestrial mammals appeared in Eocene, along with first large primates
- Many mammalian groups went extinct with shift from warm, humid climates of the Eocene to cool, dry climates of the Oligocene
- Oligocene mammals mostly similar to current mammals

Birds

The birds greatly diversified during the first half of the Cenozoic, then again during the Pleistocene.

- Closely related to dinosaurs and include many dominant, land-dwelling, flightless beasts in the Paleocene
- Flying birds occupy diverse habitats, but little variation in skeletal structure

Gigantism also emerged in the birds, which at first seems surprising, but should not be given their close relation to the dinosaurs. The giant, carnivorous flightless diatrymas were up to 2.5 meters tall and would have terrorized smaller mammals in much the same way T-Rex did several millions years before. Flying birds also achieved significant size. Songbirds did not emerge until later in the Cenozoic.

22.3 Paleogene Tectonics

22.3.1 Alpine-Himalayan orogen

The *Alpine-Himalayan* orogenic belt stretches from Spain in the west to southeast Asia. This extensive orogenic belt represent collision of Africa and India with Europe and Asia as the Tethys ocean basin closed, beginning in the late Mesozoic. The eastern Mediterranean is a vestige of the former Tethys ocean. The extensive orogenesis is responsible for many separate mountain chains, such as:

- Pyrenees
- Alps
- Caucasas
- Zagros
- Hindu Kush
- Pamir
- Himalayas

The Indian plate drifted northward, during the Cenozoic (by as much as 15–20 cm yr^{-1} in the early Cenozoic), colliding with the Asian margin beginning about 50 Ma. It was pulled northward (at least in part) by northward subduction underneath Asia, which resulted in the emplacement of a large volcanic arc and associated batholiths. Orogeny is ongoing, as seen in the intensity of seismic activity in the region.

22.3.2 Paleogene tectonism in North America

The Laramide Orogen

The semicontinuous mountain building that begin in the early Mesozoic on the western margin of North America continued into the Paleogene with the Laramide orogen. This also continued the trend of eastward migration of the orogenic front, which is though by many to be closely linked to an eastward migration in magmatism. Most likely, this is a result of a flattening subducting slab, which need to travel further in order to reach the necessary depth to start melting.

Laramide deformation produced impressive fold and thrust belts in western Canada—the northern US (Idaho and Montana) and in southern New Mexico and Mexico. The style of deformation in these areas (classic fold and thrust belts) was quite different in the more central part of the Rocky Mountains in the US, where it was manifested by broad basement uplifts. These uplifts give rise to adjacent sedimentary basins in which vast volumes of terrestrial sediment were deposited. The Green River basin is the largest known paleo-lake system well represented in the sedimentary record and contains abundant oil shales.

The Laramide orogeny ended by about 50 m.y., at which point the highlands the developed at this time were heavily eroded and planed off, filling the adjacent sedimentary basins. The modern Rocky Mountains reflect subsequent uplift of much of these earlier highlands.

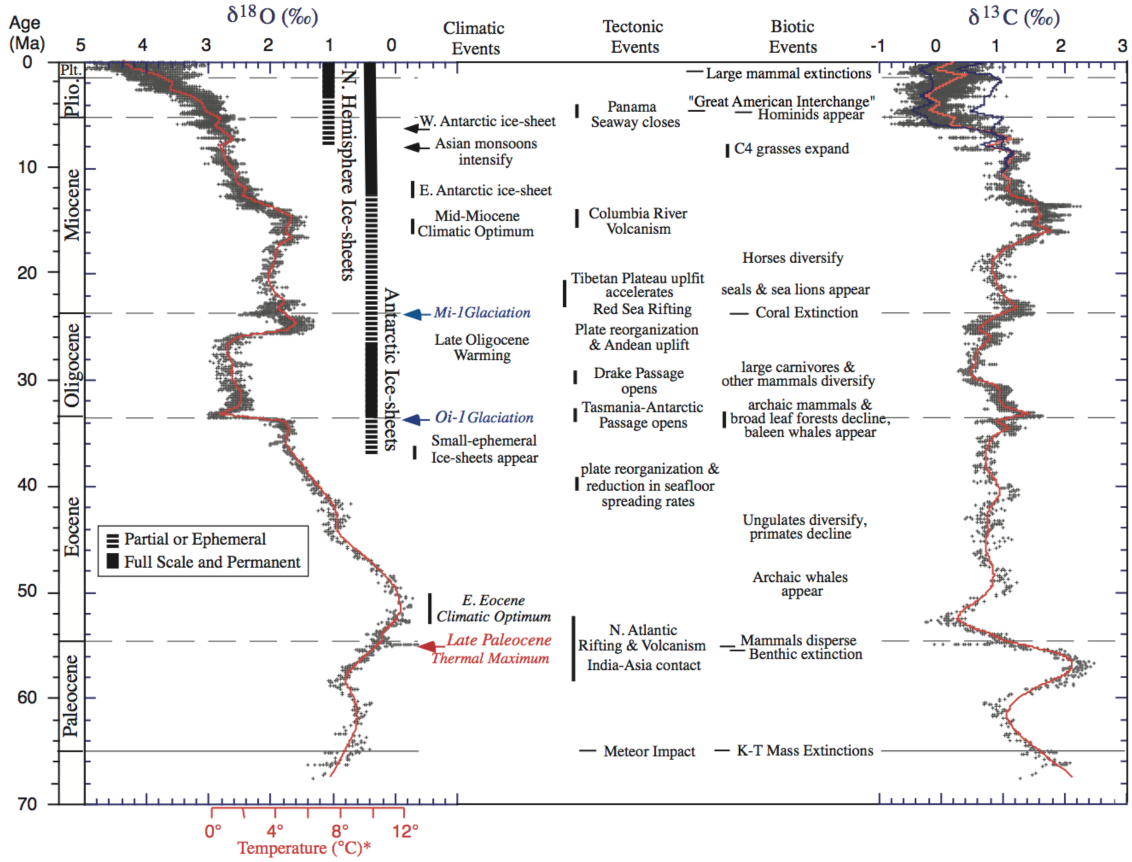

Figure 22.1: Cenozoic benthic δ^{13}C δ^{18}O records and major tectonic, climatic, and biological events (Zachos et al., 2001).

22.3.3 Rifting and volcanism

A continental rift began to form in New Mexico in the late Oligocene. The Rio Grande Rift developed along previously developed Laramide structures, and resulted in massive volcanic episodes that produced the San Juan Mountains and the famous Bandelier Tuff.

Somewhat earlier, in the Eocene, massive and episodic volcanism began in the area that is now occupied by Yellowstone Park. The catastrophic eruptions in this region buried as many as 20 forests in lavas, volcanic ash, and mud, as seen today in the petrified forests in the region.

22.4 Cenozoic Climate

The Cretaceous is generally regarded as a warm interlude in Earth history (although there is much debate about a possible series of minor glaciations). Warm conditions continued into the early Cenozoic.

22.4.1 pCO_2

CO_2 concentrations have been reconstructed from a variety of proxies, including boron isotopes in carbonates and leaf stomatal density. Atmoshperic CO_2 generally declined through the Cenozoic, interrupted by a few increases. A precipitous decline occurred between about 50 and 30 Ma, when pCO_2 was roughly in the ball park of today's level. The decline in pCO_2 is generally attributed to increased silicate weathering, perhaps related to the Himalayan orogen and/or Antarctic glaciation.

22.4.2 Paleocene-Eocene warmth and hyperthermals

Climate generally warmed from the Late Paleocene to the early Eocene, and the Paleocene and Eocene was generally a warm time Earth history. The warming trend spanning this Late Paleocene and early Eocene was punctuated by a series of *hyperthermals*, during which temperatures appear to have risen abruptly. The best known among these is the *Paleocene–Eocene Thermal Maximum* or *PETM*.

- occurs at P-E boundary (55.8 Ma)
- sharp negative $\delta^{13}C$ anomaly — massive input of isotopically light C to the oceanic (DIC) reservoir
- accompanied by negative $\delta^{18}O$ anomaly — global warming
- associated with ocean acidification, oxygen depletion in deep waters, and extinction of benthic forams

The other particularly warm periods are recognized in the Eocene (the Early-Eocene and Mid-Eocene climatic optima). During this Epoch, temperatures appeared to have been unusually uniformly warm across the globe (i.e. the equator to pole temperature gradient was about half of what it is today). Evidence for temperate forests in northernmost Canada.

22.4.3 Onset of Antarctic glaciation

Climate cooled through the latter Eocene, and the first Antarctic glaciers may have formed at this time. A virtually stepwise increase in the size of Antarctic glaciation and essentially the beginning of Cenozoic glaciation occurred at the Eocene-Oligocene boundary at 33.9 Ma, as seen in a sharp positive shift in oxygen isotopes in marine carbonates.

- Decline in pCO_2
- Opening of the Drake passage

Chapter 23

The Neogene

23.1 Introduction

The Neogene Period is the most recent in Earth's history and spans from just 23 million years ago to present. It includes the Miocene, Pliocene, Pleistocene, and Holocene epochs. Earth's paleogeography has not changed a great deal since the Paleogene, although the western coast of North America continued to evolve rapidly, as did the Mediterranean. The present Rocky Mountains and Himalayas also developed during this time. Mammals continued to diversify during the Neogene, but the most profound change on Earth has been climatic, with the continuation of cooling that first initiated in the late Eocene.

23.2 Neogene Tectonics

23.2.1 Circum-Pacific orogens

The Pacific ocean is surrounded by orogens — mostly active volcanic arcs where the various plates that make up the Pacific ocean basin are subducting beneath continents or other ocean basins. A series of trenches is associated with intense volcanic activity:

- Peru-Chile trench (Andes)
- Middle-America trench
- Cascade trend
- Aleutian trench
- Kurile trench
- Japan trench
- Marianas trench

23.2.2 Western North America

The tectonics of western North America changed about 20 Ma with the onset of subduction of a mid-ocean ridge beneath the western margin. There were many results of this tectonic event, the most important of which was that the western boundary of the North American

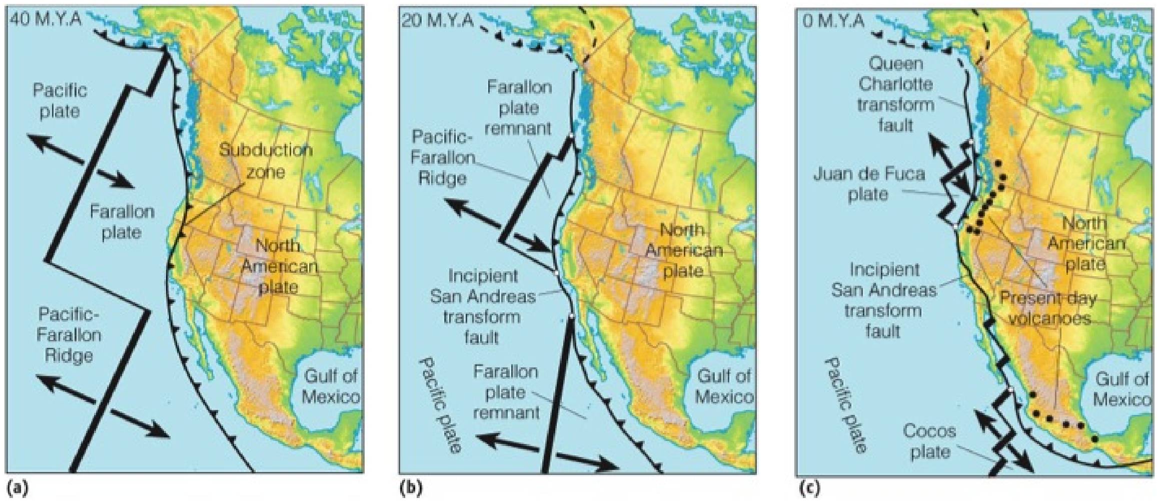

Figure 23.1: Tectonic evolution of the western North American plate boundary over the past 40 million years.

plate in California turned into a transform fault. The reorganization of tectonics that accompanied this event had several other important consequences, such as uplift of the Colorado Plateau, Basin and Range extension, and extrusion of the Columbia River flood basalts (16.6 Ma) and formation of the Snake River Plane and Yellowstone volcanic centre (hot spot?). These and several other physiographic provinces in the western half of North America developed during the Neogene.

The San Andreas Fault

The San Andreas Fault owes its origin to the subduction of a spreading ridge below North America. As the northeast-southwest trending ridge subducted beneath southern California, it resulted in terranes in western California rotating and shifting northwestward. Eventually the modern-day San Andreas Fault formed, which along with the Mendocino fracture zone, links the Juan de Fuca ridge offshore northern California with the northern extension of the East Pacific rise in the Gulf of California.

Subduction and volcanoes

Subduction beneath western North America and associated compression continued through the Neogene, but gradually the southern limit migrated north as the San Andreas fault migrated north. The Cascades are the volcanic record of this Cenozoic subduction.

Colorado Plateau

The Colorado Plateau is a broad, oval shaped region of generally high average elevation ($\sim$1800–2000 m) centred on the Four-Corners region of the American southwest. It began to rise about 6 million years ago (Levander et al., 2011), based on incision rates in the Grand Canyon, which was carved as a result of uplift of the plateau. What makes it distinct from most of the other high elevation country in western North America is that it is not intensely deformed. In fact, most of the Phanerozoic strata on the Colorado Plateau are flat-lying, with only some minor folds here and there. It appears that the anomalous uplift is the result of progressive, inward delamination of the mantle lithosphere, in part as a result of upwelling asthenosphere and emplacement of dense mafic igneous rocks related to mid-Cenozoic volcanism (Levander et al., 2011).

The Basin and Range province

The Basin and Range Province extends from the edge of the Colorado Plateau in the east (effectively the Wasatch Fault in Salt Lake City) to the Sierra Nevada Mountains in the west. It is distinguished by a series of linear (N-S) mountains and intervening valleys that formed as the result of extensional block faulting. This arid region of high average elevation formed as the result of doming and extension of the continental crust, beginning about 18 million years ago.

The present Sierra Nevadas

Although the Sierra Nevadas comprise mainly igneous rocks formed by middle Mesozoic subduction beneath North America, their present high elevation is due to the extension that generated the Basin and Range province. The eastern edge of the Sierra Nevadas is a major, east-dipping normal fault (the mirror image of the Wasatch Fault in Salt Lake City).

Columbia River flood basalt and Snake River plain

A hot spot began to spill out large volumes of basalt in the northwest about 17 million years ago. This Columbia River Flood Basalt covers large areas of present day eastern Washington and Oregon and western Idaho in *scablands*. Volcanism continued after the initial flood basalt volcanism, and as North America drifted to the west, this ongoing volcanism formed the Snake River Plain, which cuts across southern Idaho like a deep wound. The hot spot now sits below Yellowstone Park and is responsible for extraordinary volcanic eruptions that occur about once every million years.

23.3 Neogene Climate

The Neogene was generally a time of cooling, marked by expansion of the Antarctic ice sheet and onset of northern hemisphere glaciation. However, the late Oligocene and early Miocene saw a shrinking of Antarctic ice sheets, and this warm spell peaked in the middle Miocene, prior to rapid cooling and expansion of ice sheets from the late Miocene to the present.

23.3.1 Mid-Miocene Climatic Optimum

The Miocene Epoch is also regarded as a generally warm period in recent Earth history, at least compared with more recent climate. Optimum temperatures occurred around 16 Ma and were followed by about 12 million years of generally cooling climate. The Mid-Miocene climatic optimum occurred at the same time that the Columbia River flood basalts were extruded in eastern Canada and Oregon. However, although this warm interval is apparent in oxygen isotope and paleontological records, it is not obvious in paleo-CO_2 records.

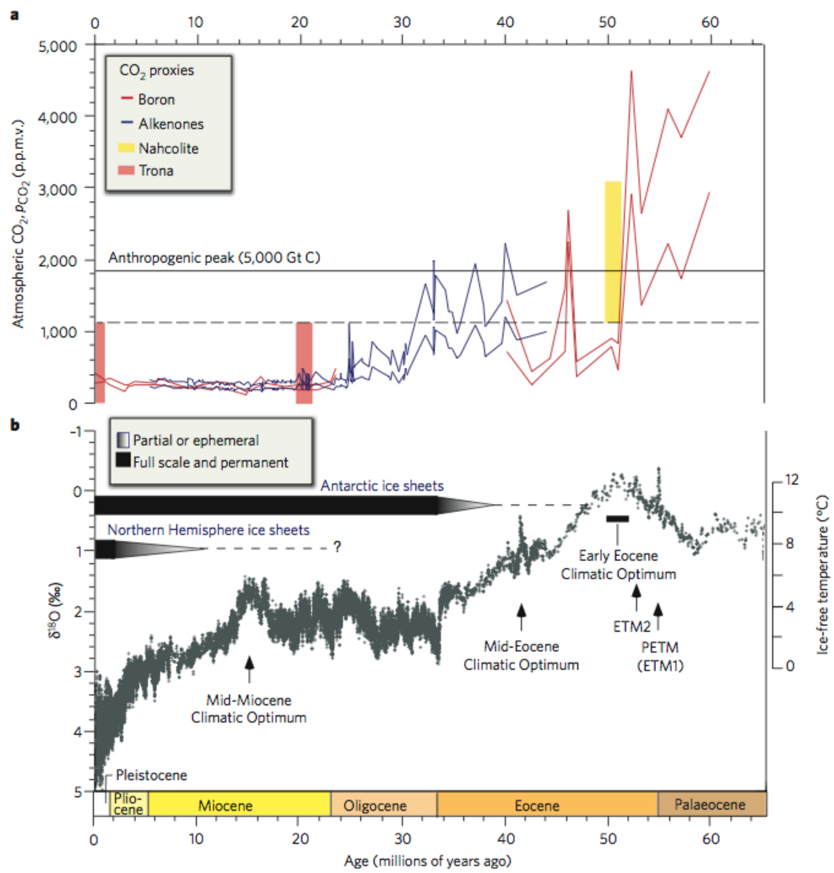

Figure 23.2: Reconstruction of benthic (deep water) $\delta^{18}O$ and pCO_2 during the Cenozoic (Zachos et al., 2008).

23.3.2 Uplift of Tibetan plateau

Although the collision between Asia and India started as long as 50 Ma, the Tibetan plateau did not go up until much more recently – perhaps about 10–8 million years ago. The uplift of the plateau likely led to the onset of the Indian-Asian monsoons and the increase in aridity in central Asia (e.g. the Gobi desert).

23.3.3 Onset of northern hemisphere glaciation

The northern hemisphere remained unglaciated until sometime in the Miocene, and major northern hemisphere ice sheets did not appear until about 3.2 million years ago. Whereas there remains much debate about the cause of the expansion of northern hemisphere ice sheets, one popular hypothesis is that it is linked to closure of the Isthmus of Panama, which would have deflected warm, saline waters of the Caribean northwards, setting up the North Atlantic Thermal Haline Circulation (THC) cell and delivering moisture to the northern high latitudes.

23.4 Life in the Neogene

To a large extent, life in the early Neogene was already very similar to that which occurred today. Invertebrate animals, in particular, have hardly changed since. However, a few groups of land mammals and plants continued to diverge rapidly during this interval. Specifically, both grasses and plants which include weeds and herbs diversified. At the same time, rats and mice (Muridae), songbirds, and snakes diversified impressively. Of course, from our anthropocentric vantage, the most important evolution was that of the primates, which we will discuss in more detail in the next chapter.

23.4.1 C4 Plants

A major Cenozoic innovation in plants was the development of a new style of photosynthesis. The *C4 plants* that use this novel form of photosynthesis are much more efficient in their utilization of carbon dioxide. Although the evolution of C4 photosynthesis probably occurred earlier in the Cenozoic, it only became widespread in the Miocene Epoch, almost certainly in response to low CO_2 levels. C4 photosynthesis is restricted to the angiosperms and includes many modern grasses and several important crop plants, such as maize, sorghum, and millet.

Because C4 plants use more of the CO_2 that they incorporate via diffusion through their stomata, they require fewer stomata than comparable C3 plants. One important side effect of this adaptation is that they are also more tolerant of drier climates, because they do not lose as much water through their stomata. Another side effect is that C4 plants fractionate carbon isotopes to a lesser extent than C3 plants (about half as much), since they make do with more of the CO_2 available, meaning they are less picky about which isotope they use. The expansion of C4 plants in the late Miocene is clearly seen in the carbon isotopic composition of pedogenic carbonates and in mammal teeth in Asia and North America, as well as in the carbon isotope record of marine carbonates.

Chapter 24

Pleistocene Glaciation and Earth System Evolution

Reading: Chapter 19 in Stanley; Chapter 17 in Wiccander and Monroe

The Pleistocene Epoch officially began at 1.8 Ma, when continental-scale northern hemisphere ice sheets stabilized. However, Northern hemisphere glaciation started earlier than this, with alpine glaciers and isolated ice caps dating back as far as the Eocene or Oligocene according to seem researchers. In any case, it appears that northern hemisphere ice sheets became important in global climate starting about 3.2 million years, roughly when the Isthmus of Panama formed and cut off tropical circulation between the Atlantic and Pacific oceans.

24.1 The Geological Record of Pleistocene Glaciation

Whereas early geologists regarded the unusual deposits and landforms left behind by ice sheets in Europe to be the product of the *Great Flood*, Louis Agassiz, building on the work of a handful of other scientists, began to sway geologists and the public that Europe and much of northern North America had been glaciated relatively recently. Evidence for this glaciation was in fact widespread and is now irrefutable. Is also clear that ice sheets have expanded and contracted cyclicly over the past several million years, at least.

- *Erratics*, drumlins, tills, etc.
- Evidence for shifting geographic distributions of biota
- Evidence of four distinct glacial epochs in till deposits in Europe and in North America
  - Nebraskan
  - Kansan
  - Illinoian
  - Wisconsonian
- These were just the largest of the glacial advances

24.2 The Oxygen-Isotopic Record of Glaciation

24.2.1 Marine carbonates

The way we view the Pleistocene glaciations changed dramatically with the advent of oxygen isotope analyses on marine carbonates (specifically on foraminifera). In the mid 1950's, Cesare Emiliani (of the famous Urey group at the University of Chicago) analysed the oxygen isotope composition of foraminifera in deep sea cores and discovered an amazingly cyclic record (Fig. 24.1). His pioneering work rapidly and thoroughly deconstructed the prevailing dogma (based on field geology of glacial deposits) on the number of Pleistocene glaciations and resurrected the long dormant theory of orbital forcing in climate (that is, Milankovitch theory). Emiliani originally interpreted the large fluctuations in δ^{18}O to be due mainly to *ice-volume* effects. That is, the sequestration of isotopically light water in large continental ice sheets during glacial maxima results in an increase in the δ^{18}O value of residual seawater. As it turns out, and not especially surprisingly, temperatures also play a role in the foram δ^{18}O record (Fig. 24.1). The precise role of temperature versus ice volume in the record is difficult to deconvolve, but fortuitously, both increases in ice volume and decreases in seawater temperatures drive the δ^{18}O values of marine carbonates to higher values. Thus, *over the Plio-Pleistocene glaciation, higher δ^{18}O values in marine carbonate records correspond to colder intervals, and lower δ^{18}O values to warmer intervals.* (This is an important point and should not be confused with the ice core δ^{18}O record, in which lower δ^{18}O values correspond to cooler temperatures.)

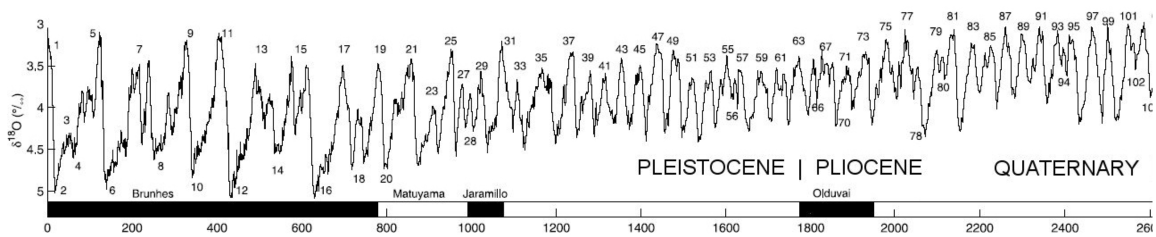

Figure 24.1: A compilation of marine benthic foraminifera δ^{18}O data over the past 2.5 m.y. (Lisiecki and Raymo, 2005). The bars on the bottom refer to magnetic polarity intervals and the numbers to *Marine Isotope Stages.*

A closer look at the marine δ^{18}O records reveals not only that there have been about fifty glacial cycles over the past 2.5 m.y., but that their frequency has changed through time: around 900 ka, the dominant frequency changed from ca. 40 k.y. to 100 k.y. Note that the δ^{18}O pattern, particularly over that last 900 ka, has a sawtooth pattern, similar to that in ice core records. This pattern suggests that global cooling occurs slowly, whereas global warming occurs quickly. The obvious advantage of the marine δ^{18}O record over the ice core records is that it extends back much further in time. Plus, these cores can be taken from many more regions, giving a more complete picture of spatial climatic variability in the past. To a large extent, the early work of Emiliani set in motion what is now the International Ocean Drilling Program (IODP).

The variation in oxygen isotopes in marine carbonates during glacial-interglacial cycles can be attributed to two synergistic causes:

- temperature dependent isotopic fractionation (between seawater and carbonate)

- the ice-volume effect

24.2.2 Ice cores

Like sediment cores, ice cores provide an exceptional record of glacial interglacial cycles, beautifully showing the saw-toothed pattern and interruptions to the warming in oxygen and hydrogen isotope ratios (and many other parameters) during the deglaciations (Fig. 2). However, ice core records only go back about a million years for Antarctica, and not even 100,000 for Greenland.

One important difference in the oxygen isotope record of the waxing and waning of ice sheets as seen in ice cores is the oxygen isotope ratios fluctuate much more than they do in marine carbonates. These fluctuations are controlled mainly by temperature, and in this case, the lowest δ^{18}O values correspond to the glacial maxima (so the inverse of marine carbonates, where it is the highest δ^{18}O that correspond to glacial maxima).

24.3 Milankovitch Cycles

As early as the early 19th century, scientists suspected that variations in Earth's climate were systematic and that they might be related to variations in Earth's orbit. In the early 20th century, Milutin Milankovitch both calculated past variations in solar forcing related to changes in Earth's three main orbital parameters and directly linked these variations to climate.

- Precession ($\sim$20 ky)
- Obliquity ($\sim$40 ky)
- Eccentricity ($\sim$100 and 400 ky)

The answer, it turned out, was in cool northern hemisphere summers (rather than in cold winters, which were also coupled with hot summers).

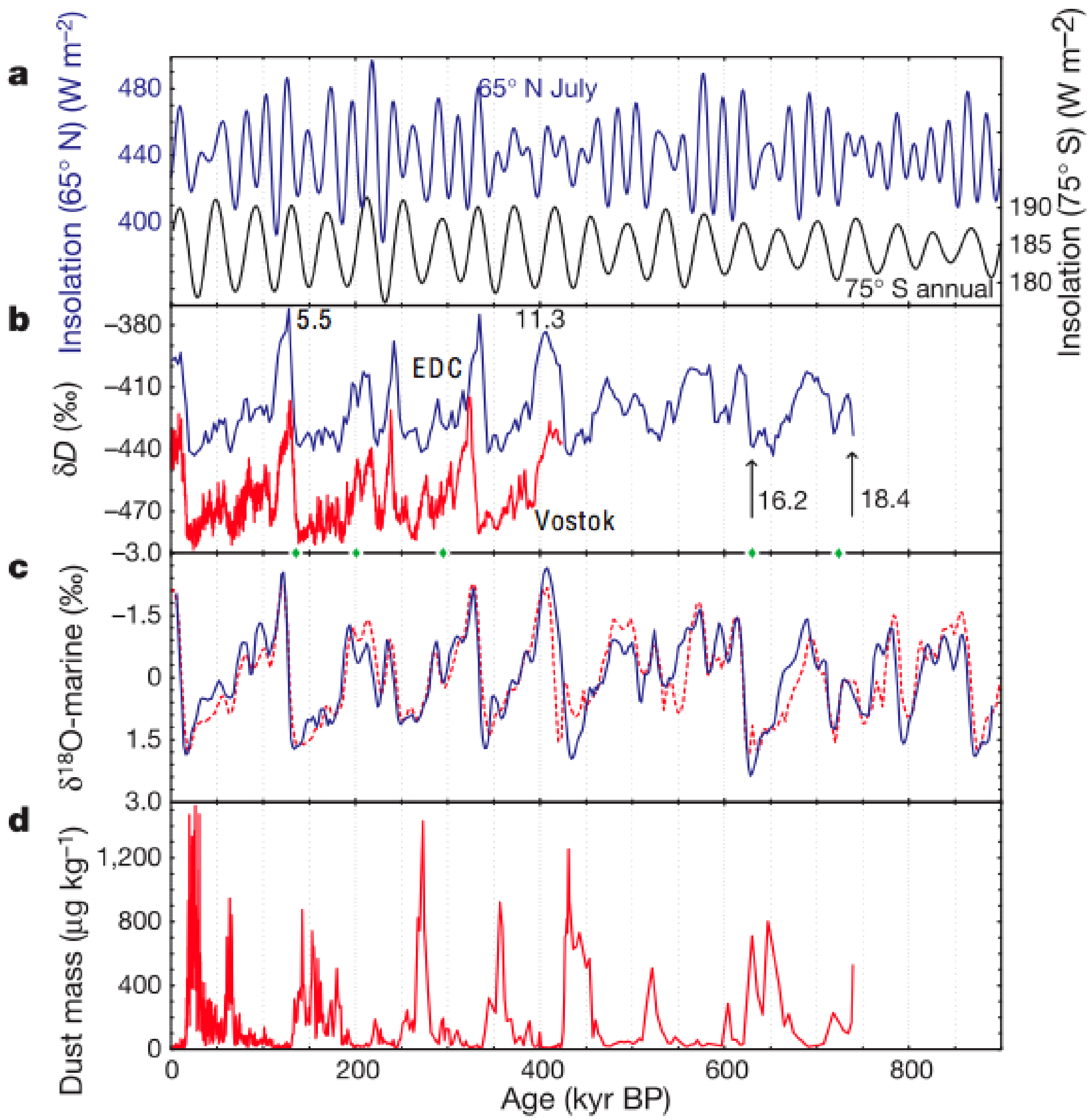

Figure 24.2: The δD records from the EPICA and Vostok ice cores in Antarctica, plotted with the δ^{18}O marine and dust records and mean insolation at 65°N and 70°S (community members, 2004).

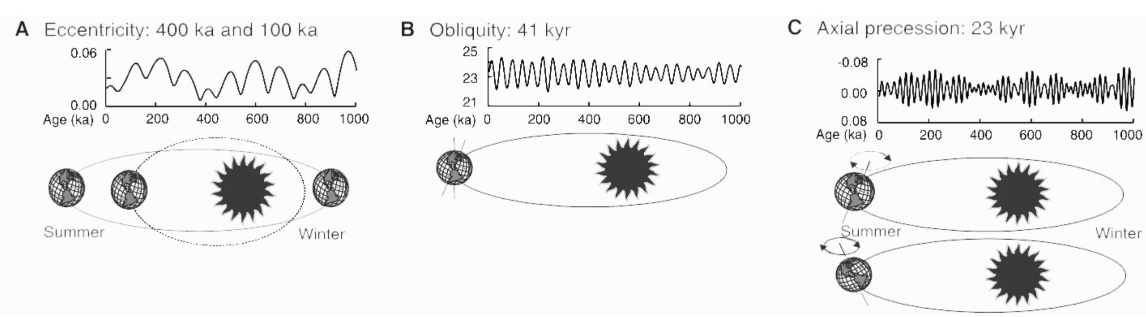

Figure 24.3: The three dominant orbital (Milankovitch) cycles (from Zachos et al., 2001).

Chapter 25

Origin and Evolution of Hominoids

25.1 Early Hominid Evolution

The earliest apes date to about 20 million years ago, and this now extinct group of apes radiated throughout Africa and Asia by about 15–16 Ma and included both tree-dwelling and land-dwelling species. These early apes gave rise to the Hominoidea, which includes both the family Pongidae (chimpanzees and gorillas) and hominids. The hominoids migrated through Africa and Asia by the middle Miocene and were diverse and widespread throughout the Miocene.

The hominoids declined dramatically in the late Miocene, and from this diminished group emerged the hominids (sometime between 8 and 5 million years ago), which include both humans and the the apes: chimpanzees, gorillas, and orangutans. *Sustralopithecines*, from whom modern humans are descended, emerged during this time. *Australopithecus*, was effectively half ape and half human and was shorten in stature (1–1.5 m). Its brain, in relation to body size, was not much larger than that of a chimpanzee, but its body was distinctly more human. It walked upright and two legs, although it could probably still manoeuvre adeptly in trees.

The human genus (*homo*) evolved from *Australiopithecus* by 2.4 billion years ago and it appears that by 2 million years ago, there were at least two species of *homo*. These early humans had brain capacities about twice the size of *Australopithecus*, smaller teeth, and a body that was not significantly different from that of modern humans. They began to use tools, known as *Oldowan* tools, and meat became a much more important part of their diet. The rather sudden appearance of *homo* coincided with a fairly dramatic shift in climate in Africa, during which it became cooler and drier as Northern Hemisphere glaciation dramatically expanded[1]

25.2 Hominin Evolution

The Hominins are commonly considered the *human clade*, which split last from other hominidae when the Chimpanzees (gebys *Panini*) branched off the Hominidae lineage some 5

[1]The author of the textbook (Steven Stanley) has written a separate and fascinating book on this connection between climate change and human evolution.

million years ago. *Homo erectus* was on of the early *Homo* species, and it closely resembled us. Two important and related difference were that its brain capacity was only about 1000 cm (compared to 1200 to 1500 in modern humans), and it had narrower hips. Presumably, these narrow hips prevented it from developing a larger brain. Amazingly, it appears that *Homo erectus* co-existed with other humans, including modern humans for a long time; fossils found on an isolated islands in Indonesia are dated at about 50,000–30,000 years old. Several other species also co-existed, including *Homo heidelbergensis* (700,000–200,000 years ago) and *Homo intercessor*, which more closely resembles *Homo sapiens* than *Homo heidelbergensis.*

The Neanderthals appeared in Europe by about 200,000 years ago. These creatures were once thought to be a early ancestors of modern humans, but appear instead to be a separate species that, based on molecular clock data, diverged from the ancestors of *Homo sapiens* about 500,000 years ago. They did, however, co-exist with humans and lived until about the peak of the last ice age maximum, around 28,000 years ago. *Homo sapiens* appeared about 150,000 ago in Africa and had migrated into Europe by about 40,000 years ago. This early *Cro-Magnon* culture incorporated diverse, specialized tools and even rather sophisticated art.

Incredible advances in the study of ancient DNA have revolutionized are understanding of hominin evolution and co-existence over the past 200 kyr. It is now clear that there were (at least) 3 separate lineages of humans that inhabited Europe and Asia as recently as the last Pleistocene: modern humans, Neanderthals, and the Denisovans (Prüfer and 44 others, 2013). The Denisovans have only been discovered recently, based on the genome sequence of a finger bone from a cave in the Altai Mountains of Siberia! (Reich et al., 2010). The Denisovans and Neanderthals appear to have branched from the modern human lineage about 600 ka and from one another about 400 ka. Their DNA also suggest that they interbred slightly with one another and with modern humans. Interestingly, they also show evidence of having interbred with a fourth, as yet unknown hominin (Fig. **??**), which must have persisted at least until 400 ka, since this mysterious DNA does not appear within the Neanderthals.

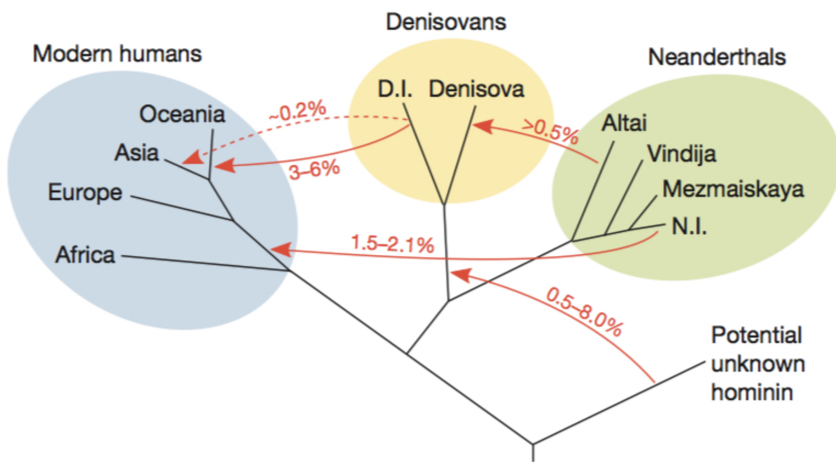

Figure 25.1: A newly developed evolutionary tree for humans and their close relatives (Prüfer and 44 others, 2013), based on recent DNA genome sequencing.

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